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Paper

Comments on Magnetic Petrophysics

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Abstract

Measurements of the magnetic properties of rocks constrain magnetic interpretation and resolve much of the ambiguity which afflicts the magnetic method. Magnetic petrophysical studies invariably indicate the importance of remanence as a source of magnetic anomalies. The susceptibility, remanent intensity and Koenigsberger ratio exhibited by a rock containing magnetic mineral grains is not only a function of the volume fraction of magnetic material with a given composition, but is also sensitively dependent on the domain state of the magnetic grains. Superparamagnetic, stable single domain and multidomain size ranges are given for magnetite, titanomagnetite, maghaemite, haematite and monoclinic pyrrhotite. The susceptibilities, remanent intensities and Koenigsberger ratios of superparamagnetic, single domain and multidomain grains of these minerals are plotted. Charts of typically observed magnetic parameters for various rock types are presented. Rocks often bear a multicomponent remanent magnetisation. The various components are often carried by grains with different coercivity or blocking temperature spectra and can be resolved using palaeomagnetic cleaning techniques. Surface samples typically contain palaeomagnetic noise which must be identified and removed if representative remanence values for the rock unit are to be determined. Under favourable conditions probable remanence directions of rock units can be inferred from a knowledge of the age of the rock and its locality and of apparent polar wander with respect to the crustal block within which the rock unit is located. Formulae are given for inversion of palaeomagnetic data to obtain magnetisation directions, and a table of primary remanence directions throughout the Phanerozoic is given for Australia. Apart from providing input to magnetic interpretation, many other applications of magnetic petrophysics are apparent. Palaeomagnetic dating of mineralisation, magnetic fabric studies for structural interpretation, magnetostratigraphy, detection of redox chemical remanent magnetisation effects and magnetic techniques of mineral identification and quantitative analysis, are some of the promising applications in the mineral exploration and processing industries.

Introduction

Magnetic surveying is one of the most frequently applied geophysical methods, yet interpretation of magnetic data in terms of geology lags far behind other geophysical techniques, notably seismic methods. The major problem in obtaining useful geological information from magnetic surveys is lack of knowledge about the magnetic properties of rocks. Like other geophysical methods, magnetics is afflicted by ambiguity in that models which account for observed anomalies are always non-unique. Determination of the magnetic properties of the rock types represented in the area under consideration serves to constrain interpretation by restricting the range of feasible models. Particularly when integrated with other information (geological control, drilling data, other geophysical methods, etc.) knowledge of magnetic properties can often lead to an interpretation which is highly probable or even effectively unique in geological terms.

Emerson (1979) proposed that as well as assisting magnetic interpretation in particular areas, compilation of a magnetic petrophysics data bank is essential for a deeper understanding of the relationship between geology, rock magnetism and magnetic signatures. The complexity of this relationship is evidenced by the dependence of the magnetic response of a rock unit on *inter alia* lithology, structure and geological history (palaeoenvironment, thermal history, alteration). A review of the magnetic mineralogy of sediments and metasediments with the emphasis on magnetite has been published by McIntyre (1980) and the magnetic mineralogy of igneous rocks has been discussed by Haggerty (1979). Additional information on the mineralogy of the magnetic oxides in igneous and metamorphic rocks can be found in Haggerty (1976) and Rumble (1976), respectively.

Some interesting conclusions from the publications cited above include:

(1) Magnetic surveying frequently delineates premetamorphic sedimentary environments in metasediments, rather than lithological boundaries. Magnetite formation during metamorphism depends mainly on the iron oxidation state ratio inherited from the sediment. Formation of magnetite is favoured by: low silica, low titanium, dehydration, excess aluminum, equilibration to lower temperatures in titaniferous rocks, and absence of carbon (McIntyre 1980).

(2) Magnetic mineralogy of igneous rocks is correlated both with bulk chemistry and mode of emplacement.

Acidic suites (granites and rhyolites) are characterised by typically 1–3% of oxidised phases (magnetite- and haematite-rich components) whereas basic suites (gabbros and basalts) contain commonly 5–10% ulvospinel- and ilmenite-rich components as primary precipitated oxides. Plutonic rocks are characterised by deep-seated equilibration, exsolution and low-intensity oxidation states. Extrusive suites on the other hand are typified by high temperature oxidation. High susceptibilities are favoured by partial oxidation of ulvospinel-rich titanomagnetites to form magnetite-ilmenite intergrowths (common in basic suites), whereas advanced oxidation in subaerial basalts tends to obliterate magnetic response (Haggarty, 1979).

Reconnaissance sampling of rocks from a particular area provides information on which lithologies are likely sources of magnetic anomalies and enables better correlation of geology with observed magnetic signatures. In addition, knowledge of likely magnetisation directions and magnitudes can improve modelling of buried magnetic sources.

Input from magnetic measurements is also valuable when a drill-hole has been targeted on a modelled magnetic source. It is important to ascertain whether or not the intersected material accounts for the anomaly. The results may show, for instance, that disseminated magnetite in barren rock explains the anomaly and that further drilling is not justified. Alternatively, the results may suggest the target has been missed altogether, or the intersected body only partly accounts for the anomaly. This would encourage further drilling to locate a discrete nearby source which could be an ore body.

Maximum value can only be obtained from magnetic property measurements on drill-core samples if the core is oriented. Core orientation is possible with minimal disruption to normal drilling procedure and yields useful structural information (e.g. bedding and schistosity planes) as well as enabling determination of remanence directions and susceptibility ellipsoid axes. The ABEM Craelius Core Orientator is a commercially available device for accurate orientation. Adequate accuracy can be achieved in inclined drill-holes with a simple method requiring only a metal cone attached to a cable. The cone is lowered down the hole, whenever drilling is interrupted, until the tip comes into contact with the lowermost point of the hole. The cone is then raised a short distance and allowed to drop, making an indentation in the rock. Drilling then proceeds and when the next piece of core is extracted the lowermost part of the core is indicated by the indentation. This, together with the survey of the hole, determines the orientation uniquely and in practice an accuracy of 5° is often obtainable. In many cases drill core can be roughly oriented on the basis of the intersection of known bedding or schistosity planes with the core.

Domain structure and magnetic properties

The magnetic properties of a ferro- or ferrimagnetic mineral are fundamentally controlled by the domain structure of the mineral grains. The atomic magnetic moments within fine particles are aligned and the whole particle is magnetised to saturation in zero applied field along an easy axis of magnetisation. The particle then possesses single domain (SD) structure. Larger grains find it is energetically favourable to subdivide into a number of magnetic domains with non-parallel spontaneous magnetisations in order to reduce magnetostatic energy associated with free poles at the grain boundary. The magnetic domains are separated by domain walls within which the atomic magnetic moments are not aligned and therefore incur increased exchange energy.

Formation of domain walls to produce multidomain (MD) structure therefore involves a trade-off between reduced magneto-static energy and increased exchange energy. Because domain wall energy is a surface energy, whereas magnetostatic energy is proportional to volume, MD structure is favoured by large grain size. The threshold size for equidimensional grains at which MD structure becomes energetically favourable is called the critical SD size. In strongly magnetic materials grain shape also affects domain structure. The critical grain length for the SD-MD transition in elongated magnetite particles, for example, is much larger than the critical SD size for equidimensional grains. Equidimensional magnetite particles smaller than $\simeq 0.06\,\mu m$ have SD structure. Domain walls form in larger grains but SD-like properties are exhibited to some extent by grains up to $\simeq 20\,\mu m$.

At all temperatures above absolute zero, the magnetic properties of grains are affected by thermal agitation. An assemblage of SD particles will attain thermal equilibrium with a characteristic time constant (time for exponentially decaying remanence to attain $1/e$ of its original value) which is primarily influenced by grain volume. The remanence of the assemblage will decay in zero field, whereas an initially demagnetised assemblage in an applied field will build up a viscous remanence. The time constant for these processes is proportional to $\exp(E/kt)$ where E is the energy barrier between stable magnetisation states of the grains, k is Boltzmann's constant, and T is the absolute temperature. E is proportional to particle volume, so at a given temperature sufficiently small particles will have very short relaxation times whereas larger particles may have large relaxation times. Because of the exponential dependence on volume, the transition from unstable (relaxation time $\ll$ laboratory measurement time) to stable (relaxation time $\gg$ laboratory measurement time) is very sharp and at a given temperature we may define a critical grain volume, known as the blocking volume, below which the grains are unstable and above which they are stable. Unstable grains are known as superparamagnetic (SPM) because their behaviour in an applied field is analogous to that of paramagnetic atoms, with the difference that the magnetic moment of a SPM particle is thousands of times larger than that of a single paramagnetic atom (Bean & Livingston 1959).

The properties of greatest interest to magnetic interpretation (susceptibility, intensity and stability of remanence, Koenigsberger ratio) are strongly dependent on the predominant magnetic state (SPM, stable SD, MD) of the magnetic mineral grains in the rock. SPM grains, for example, possess very high susceptibility and zero remanence, and a small fraction can disproportionately affect the magnetic properties of the rock. SD grains of a magnetic mineral, on the other hand, have much lower susceptibility than corresponding SPM particles, but can carry a relatively intense remanence. SD grains respond to an applied field by rotation of grain magnetic moments. The susceptibility of MD grains is due to domain wall translation as well as domain moment rotation. The intrinsic susceptibility of MD grains is generally greater than that of SD grains, but the observed emu susceptibility of MD grains of highly magnetic minerals is limited by self-demagnetisation to $\leqslant 0.25$.

The SPM threshold size and critical SD size for equidimensional particles are given in Table 1 for various magnetic minerals. The properties of SD and large MD grains are reasonably well understood, but the behaviour of small MD grains which contain only a few domain walls is still somewhat enigmatic. No sharp transition in magnetic properties is observed experimentally at the SD-MD threshold size. The properties of small MD grains, commonly called pseudo-single domain (PSD) grains, are intermediate between those of SD and large (true) MD grains and vary systematically with grain size. The most important property of PSD grains is the capacity to retain relatively intense, hard and stable remanence analogous to that of SD grains. The relevance of PSD grains to magnetite-bearing rocks is that a substantial fraction of the grain size distribution commonly falls into the PSD range ($\simeq$ 0.1–15 μm) and PSD grains therefore dominate the remanent properties of these rocks.

TABLE 1

Domain structure transition sizes (slightly modified from Dunlop 1981)

Mineral	SPM threshold size (μm)	Critical SD size (μm)
Iron	<0.008	0.023
Magnetite	0.03	0.06
Maghaematite	0.02	0.06
Titanomagnetite (x = 0.6)*	0.08	0.40
Titanomaghaematite (x = 0.6, z = 0.4)*	0.05	0.75
Titanomaghaematite (x = 0.6, z = 0.7)*	0.09	2.40
Haematite	0.03	15.0
Pyrrhotite	0.018**	1.60

Equidimensional particles assumed. Critical sizes cited are at 20°C. *Titanomagnetite $Fe_{3-x}Ti_xO_4$ = x Usp + (1-x)Mt solid solution; titanomaghaematite $Fe_{(3-x)R}\,Ti_{xR}\,V_{3-R}O_4$, $R = 8/[8 + z(1 + x)]$; V = cation lattice site vacancy, z = Fe^{2+} oxidised/original Fe^{2+}. **Calculated from relaxation time equation assuming magnetocrystalline anisotropy constant $K_1 = 3 \times 10^5$ erg/cm^3 for pyrrhotite.

Diagnosis of the domain state of magnetic minerals has been discussed in a state-of-the-art review of rock magnetism by Dunlop (1981).

Contrary to popular belief, remanence is nearly always an important contributor to magnetic anomalies so a brief discussion of remanence acquisition is warranted. Thermoremanent magnetisation (TRM) is acquired when a rock cools from a high temperature in an applied field. As the magnetic grains cool through their blocking temperatures (the temperatures at which the relaxation times become comparable to laboratory measurement times) the induced magnetisation of the grains is 'frozen in' and becomes a permanent magnetisation. Above the blocking temperature the grains are SPM with high susceptibility. Thus induced magnetisation is high and produces an intense remanence, parallel to the applied field, when it becomes blocked. On further cooling the remanence intensity increases further, purely as a result of the increase in spontaneous magnetisation with decreasing temperature (this applies to most, but not all, magnetic materials). Therefore acquisition of TRM is an efficient mechanism for producing intense remanence in a relatively weak applied field. The remanence direction is a memory of the direction of the applied field and may bear no relation to the present field direction (e.g. in approximately 50% of cases the remanence is >90° from the present field).

Acquisition of chemical remanent magnetisation (CRM) occurs at low temperatures when initially SPM mineral grains grow through their blocking volume. CRM is similar in its properties to TRM and is carried by magnetic minerals formed at low temperatures. The natural remanent magnetisation (NRM) of haematite is usually CRM, and may be quite intense.

Isothermal remanent magnetisation (IRM) is produced by application of a large magnetic field which remagnetises all magnetic grains with coercivity less than the applied field. The NRM of outcrop samples is commonly contaminated by IRM associated with lightning strikes. If the applied field is sufficiently strong the rock acquires a saturation IRM (SIRM).

The range in susceptibility, k, NRM intensity (for NRM = TRM, CRM, SIRM) and observed Koenigsberger ratio for an ambient field of 0.5 Oersted (Q_n = NRM/0.5k) are plotted for various magnetic minerals as a function of magnetic state (SPM, SD, MD) in Fig. 1. Susceptibility and NRM intensity values refer to unit volume of the magnetic minerals. In order to estimate values applicable to rocks containing dispersed magnetic grains the data must be multiplied by the volume fraction of the magnetic mineral. The data are applicable to non-interacting grains (e.g. for MD grains the susceptibilities are corrected for self-demagnetisation) and therefore are most appropriate for the typical case of rocks containing small volume fractions of dispersed magnetic grains. Rocks containing more than 10 volume % of strongly magnetic material, tightly clustered magnetic grains or large grains subdivided by lamellae, will have their properties modified somewhat. Strongly interacting SPM particles behave as SD grains. Interactions increase the stable SD size range, increase the susceptibility, decrease the coercive force and reduce the remanence for SD and PSD grains (Dunlop 1981). For MD grains interactions increase the susceptibility and the remanence (Stacey & Banerjee 1974).

The data in Fig. 1 are based on theoretical relationships given in Stacey & Banerjee (1974) and on experimental results obtained by Uyeda (1958), Hargraves (1959), Carmichael (1961), Syono *et al.* (1962), Bin & Pauthenet (1963), Strangway *et al.* (1968), Kropacek & Krs (1968, 1971), Kropacek (1971) and Day *et al.* (1977).

Magnetic minerals

Antiferromagnetic, paramagnetic and diamagnetic minerals make negligible contributions to magnetic anomalies, so

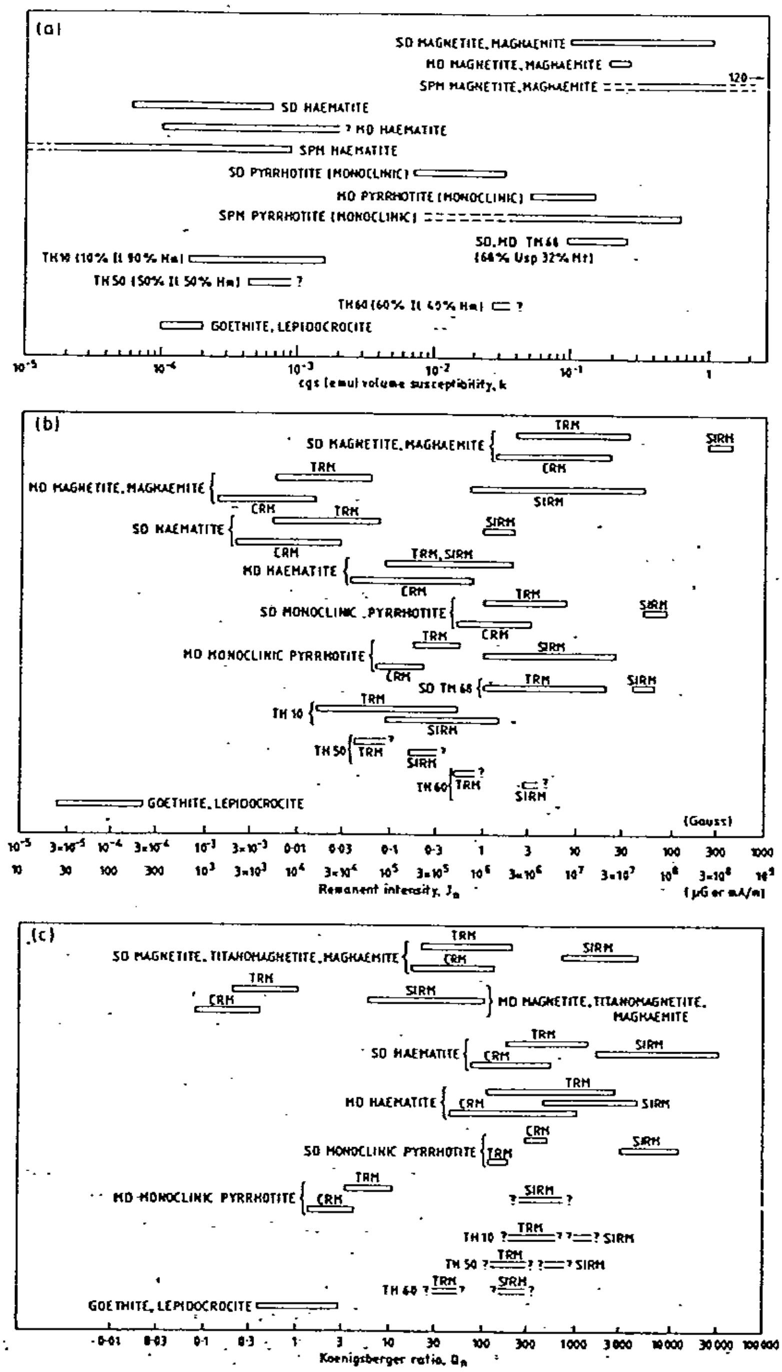


FIGURE 1

(a) Cgs (emu) susceptibility of magnetic minerals, referred to unit volume of magnetic material. (b) Intensity of NRM, J_n, referred to unit volume of magnetic material. (c) Koenigsberger ratios, Q_n, exhibited by various magnetic minerals as a function of domain state and remanence acquisition mechanism. SD = single domain, MD = multidomain, SPM = superparamagnetic, TRM = thermoremanent magnetisation, CRM = chemical remanent magnetisation, SIRM = saturation isothermal remanent magnetisation. Koenigsberger ratios are calculated assuming an ambient field intensity of 0.5 Oersted ($Q_n = J_n/0.5\ k$).

for the purposes of magnetic interpretation they may be regarded as non-magnetic. The most important magnetic minerals in rocks are titanomagnetites (stoichiometric and cation-deficient), titanohaematites and monoclinic pyrrhotite, but there are many other minerals which may occasionally contribute to magnetic anomalies. The properties of these minerals are discussed below.

Titanomagnetites

The general formula for titanomagnetites is $Fe_{3-x}Ti_xO_4$ $(0 \leqslant x \leqslant 1)$ representing 100 x % ulvospinel in solid solution with magnetite, and is often denoted by TM100x. The observed emu susceptibility of MD magnetite grains is about 0.25 whereas the susceptibility of SD grains ranges from 0.1—1.1. These values are relatively insensitive to titanium content up to ≃TM75 (see e.g. Day *et al.* 1977). Compositions which are more ulvospinel-rich than TM75 are paramagnetic at room temperature and therefore have very low susceptibility. The susceptibility is a sharp function of temperature very near the Curie-point and their exists a narrow range of compositions approximating TM80 which are strongly magnetic in frigid climates but non-magnetic in the tropics. TM68, which has a Curie temperature of ≃80°C, has a susceptibility of around 0.1 (SD) to

0.25 (MD) — very similar to that of magnetite. Cation-deficient titanomagnetites (titanomaghaematites) have susceptibilities in the same range as stoichiometric titanomagnetites.

A small quantity of SPM titanomagnetite grains can drastically affect the observed susceptibility. For instance, spherical magnetite grains of volume $10^{-17}cm^3$ are SPM at room temperature with an emu susceptibility of $\simeq 19$ (Stacey & Banerjee 1974). Therefore, as little as 0.01% of these grains by volume will contribute 1900 x 10^{-6} to the susceptibility of the rock.

Titanohaematites

Haematite (α-Fe_2O_3) is antiferromagnetic with a weak superimposed ferromagnetism due to slight canting of the spins, plus a contribution due to crystal defects. Fine-grained (< 15 μm) haematite has a low susceptibility ($\simeq 60 \times 10^{-6}$) but is capable of carrying a substantial remanence, up to 80 000 μG (8 000 gamma) for TRM acquired in 0.5 Oersted. Large (> 100 μm) crystals of haematite, which usually occur only in certain massive ores, have much larger susceptibilities (up to $\simeq 2\,000 \times 10^{-6}$) as well as strong remanence.

Pure ilmenite ($FeTiO_3$) is paramagnetic at room temperature. Intermediate composition titanohaematites ($Fe_{2-x}Ti_xO_3$) with x in the range $\approx$0.5 to $\approx$0.8 are ferrimagnetic at room temperature and consequently have high susceptibility and remanence.

Sulphide minerals

Pyrrhotite ($Fe_{1-x}S$) is often an important contributor to magnetic anomalies, particularly in mineralised terrains. Only monoclinic pyrrhotite with 4C superstructure is ferrimagnetic at room temperature. Monoclinic pyrrhotite is generally restricted to the composition $\simeq Fe_7S_8$, but intermediate pyrrhotites (e.g. Fe_9S_{10}) may acquire ferrimagnetism if quenched from $\simeq$200°C.

The rare minerals greigite (Fe_3S_4) and smythite ($[Fe,Ni]_9S_{11}$) are also strongly magnetic. Cubanite ($CuFe_2S_3$) is also reported to have a weak spontaneous magnetisation, similar in magnitude to that of haematite, so that where present it may contribute to the magnetisation of an ore body.

The magnetic properties of sulphide minerals are reviewed by Vaughn & Craig (1978).

Iron-nickel-cobalt alloys

Partial serpentinisation of ultramafic plutonic rocks produces metallic iron and alloys of Fe-Ni-Co-Cu which are highly magnetic and have high Curie temperatures. These minerals may account for deep crustal anomalies whose sources appear to lie below the Curie-point isotherm of magnetite (Haggerty 1979).

Other ferrimagnetic minerals

Some marine ferromanganese minerals, e.g. todorokite, may contribute to the magnetic signature of marine sediments (Henshaw & Merrill 1980).

A number of spinel minerals exist which are either ferrimagnetic in their own right, or form ferrimagnetic solid solutions with magnetite. These minerals are natural analogues of synthetic spinel ferrites with important technical applications. The magnetic properties of some spinel minerals are listed in Table 2.

TABLE 2

Magnetic properties of spinel minerals (data from Craik 1975; Kropacek 1971; Kropacek & Krs 1968; and Nagata 1961)

Mineral name	Chemical formula	Magnetic properties
Magnesioferrite	$MgFe_2O_4$	Ferrimagnetic, J_s = 119G, T_c = 440°C
Jacobsite	$Fe_{3-x}Mn_xO_4$	Ferrimagnetic for $0 \leqslant x \leqslant 2.5$, T_c = 50 – 580°C
Jacobsite	Fe_2MnO_4	T_c = 300°C, J_s = 398G
Franklinite	$ZnFe_2O_4$	Paramagnetic at room temperature, in ss with magnetite can be strongly magnetic
Chromite	$FeCr_2O_4$	Paramagnetic at room temperature
Chromite-magnetite ss	$Fe_{3-x}Cr_xO_4$	Ferrimagnetic for $0 \leqslant x \leqslant 1.2$, T_c = 20 – 580°C
Chromite-magnetite ss	Fe_2CrO_4	J_s = 250G, T_c = 200°C
Hercinite	$FeAl_2O_4$	Paramagnetic at room temperature, in ss with magnetite can be strongly magnetic

J_s = spontaneous magnetisation at 20°C (480G for magnetite); T_c = Curie temperature (580°C for magnetite).

The rare hexagonal ferrite mineral magnetoplumbite ($PbO.6Fe_2O_3$) is highly magnetic with a saturation magnetisation of 320 G and a Curie temperature of 450°C.

Iron oxyhydroxides

Goethite (α-FeOOH) and lepidocrocite (γ-FeOOH) are weakly and variably magnetic (Strangway *et al.* 1968; Vlasov & Gornushkina 1973). Typically the susceptibility of these minerals ranges from 120–200 x 10^{-6}. TRM intensities range from 20–230 μG with Koenigsberger ratios 0.4–3.

Anomalously magnetic minerals

A number of silicate and ore minerals which are intrinsically non-magnetic are nevertheless frequently found to exhibit relatively high susceptibility and remanence. The magnetisation of olivines, feldspars and pyroxenes is due to ultrafine titanomagnetite grains contained within them. In these cases the Koenigsberger ratio is high due to the SD structure of the majority of the TM particles, and the remanence may be quite strong. Magnetite inclusions also appear to explain the magnetisation of zircons, cassiterite and other ore minerals.

Susceptibility, remanence and Koenigsberger ratio of rock-types

A large collection of rock samples representing a wide range of lithologies has been built up by the CSIRO Rock Magnetism Group during the course of company collaborative projects as well as for palaeomagnetic studies. The

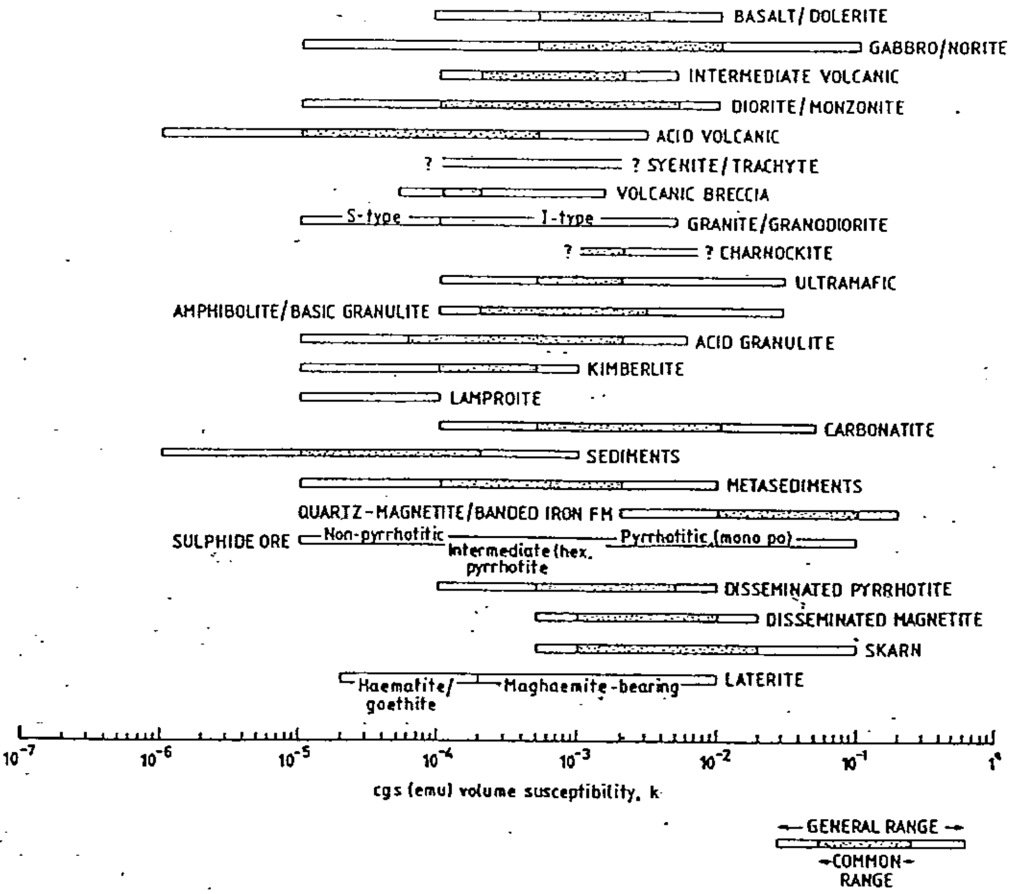


FIGURE 2

Range of cgs (emu) susceptibilities exhibited by common rock types (to convert to SI susceptibilities multiply values by 4π).

range of susceptibilities, NRM intensities and Koenigsberger ratios observed in different rock types are given in Figs 2, 3 and 4, respectively. Figures 2–4 are based on the CSIRO data bank, supplemented by data from published petrophysical studies, particularly Puranen *et al.* (1968), Henkel (1975, 1976, 1977), Ketola *et al.* (1976), Gupta & Burke (1977) and Krutikhovskaya *et al.* (1979).

The most significant feature of Figs 2–4 is that remanence is frequently an important contributor to the magnetisation and that Koenigsberger ratios greater than unity predominate in most rock types, particularly the more magnetic lithologies. Thus the common practice of ignoring remanence in interpretation cannot be justified.

A large range of susceptibilities and NRM intensities is found for all rock types. Although there is not a strong apparent correlation between magnetic properties and lithology, upon deeper examination a relationship between the two often emerges. For example I-type granitoids are more magnetic than S-type granitoids (Ellwood & Wenner 1981; P. W. Schmidt, pers. comm.). A similar relationship holds for ortho- and para-amphibolities. The magnetic properties of pyrrhotitic ores depend on the relative proportions of magnetic monoclinic pyrrhotite and non-magnetic intermediate pyrrhotite.

The categories 'disseminated pyrrhotite' and 'disseminated magnetite' do not correspond to a particular lithology or geological setting, but to an exploration concept. Sediments or volcanics containing disseminated magnetite or pyrrhotite, associated with subeconomic mineralisation and of considerable thickness, are often encountered during drilling targeted on a modelled discrete magnetic source. These categories refer then to the rocks responsible for the magnetic anomaly when no magnetic ore body is found.

The relatively low susceptibilities of kimberlites ($< 1000 \times 10^{-6}$) may seem surprising in view of their magnetite content, but titanomagnetites in kimberlites are generally ulvospinel-rich with appreciable magnesioferrite, spinel, hercynite and chromite in solid solution (Haggerty 1976). The Curie temperature is lowered by the cation substitutions and in many cases is below room temperature in which case the titanomagnetite is effectively non-magnetic. The magnetic signature of kimberlites may be due to ferrimagnetic picroilmenites (ilmenite-haematite-geikielite solid solutions) as well as small concentrations of magnetite-rich spinel phases.

Another magnetic parameter which may affect interpretation is anisotropy of susceptibility. Whilst the weak anisotropy exhibited by most rock types may be interpreted in terms of a magnetic fabric with potential applications to structural studies, the effect on magnetic anomalies is generally negligible. Exceptions include banded-iron formations and some ore bodies which are highly anisotropic. As a numerical example consider a banded-iron formation with $k_{max} = k_{int} = 2k_{min}$. Depending on the orientation of the banding with respect to the earth's field the effective susceptibility will be within −40% and +20% of the bulk susceptibility, the extremes corresponding to ambient field normal and parallel to the

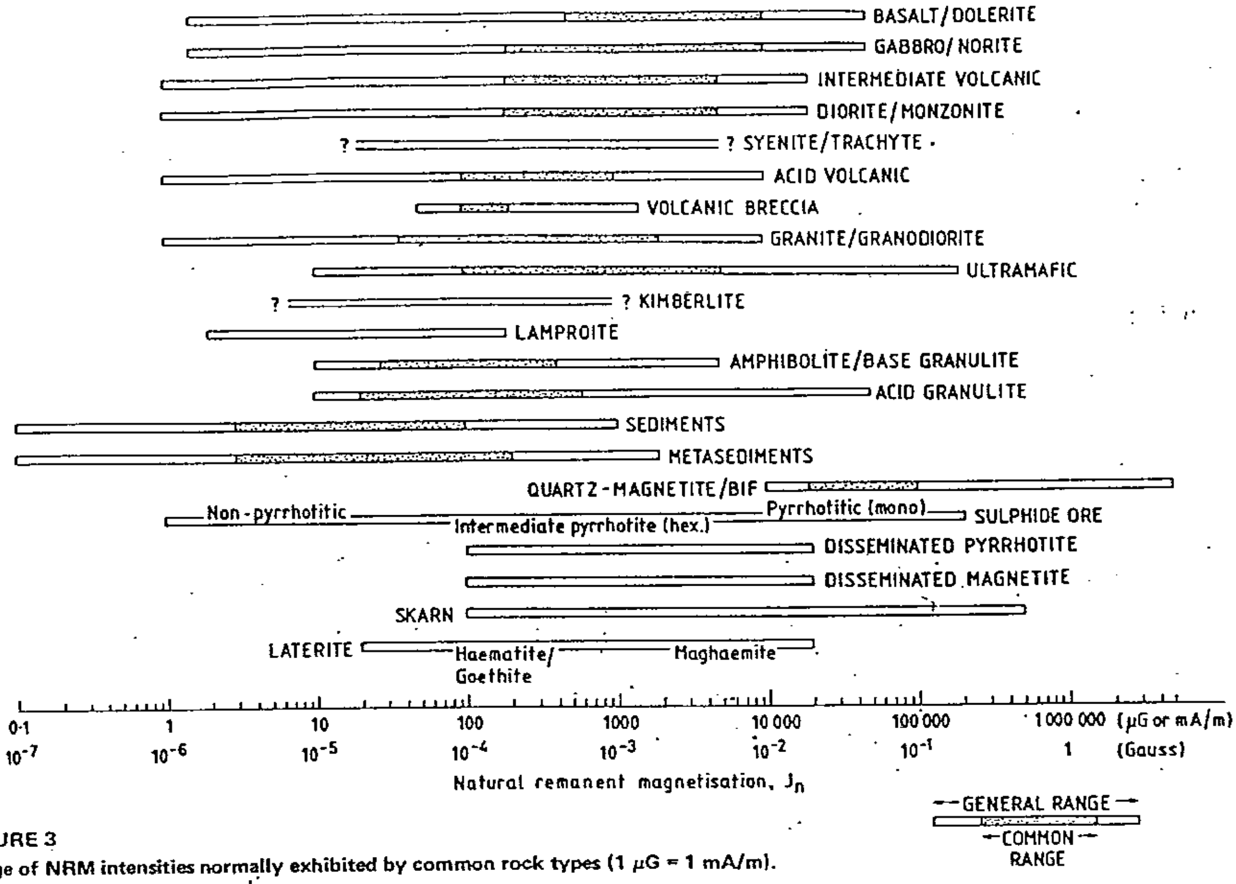


FIGURE 3

Range of NRM intensities normally exhibited by common rock types (1 μG = 1 mA/m).

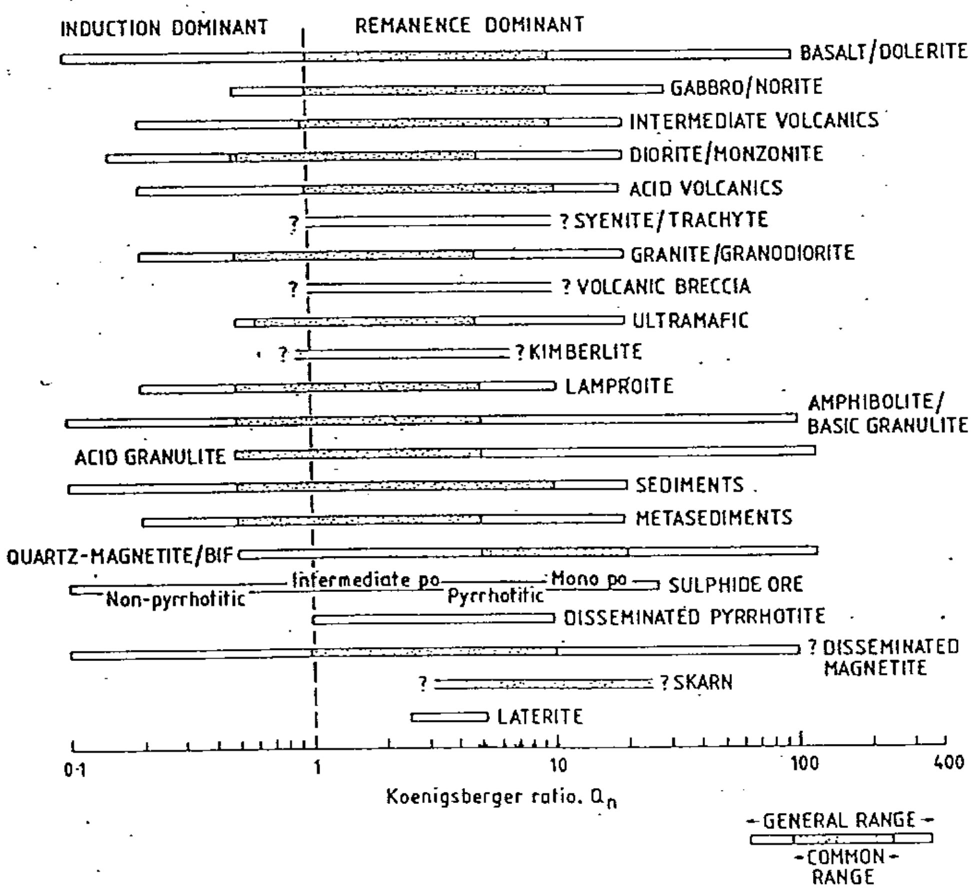


FIGURE 4

Range of Koenigsberger ratios normally exhibited by common rock types. The Koenigsberger ratio is the ratio of the remanent magnetisation intensity to the induced magnetisation intensity in the ambient Earth's field (typically ≃ 0.5 Oersted).

banding, respectively. In these cases there is no deflection of magnetisation direction. The maximum possible deflection of the magnetisation from the ambient field direction is 19°. The effective susceptibility is then 80% of the bulk susceptibility.

The commonly observed range of susceptibility anisotropy values in different rock types has been reviewed by Janak (1972, 1973).

Multicomponent magnetisations

The remanent magnetisation of rocks is often complex with several remanence components present. At the time of formation the rock acquires a primary remanence, or imprint. During its subsequent history the rock may acquire a thermal or chemical overprint, or even be remagnetised completely, in which case the secondary magnetisation is a reprint. Secondary components of magnetisation may also be viscous remanent magnetisation (VRM), acquired parallel to the earth's field by grains with relaxation times shorter than the latest geomagnetic polarity epoch, due to weathering or lighting strikes. In addition, rock samples may acquire spurious remanence components during or after collection as a result of exposure to magnetic fields.

In general the various components of magnetisation reside in different magnetic grains or zones of grains which have different stabilities to various palaeomagnetic cleaning techniques. These techniques include alternating field (AF), thermal, chemical and low temperature demagnetisation. Application of cleaning techniques, together with modern methods of palaeomagnetic data analysis such as vector diagrams, principal component analysis, remagnetisation circles, Hoffman-Day plots, etc., allows determination of the various remanence components present in the rock.

The relevance of these palaeomagnetic methods to petrophysical studies intended for input to magnetic interpretation is that NRM measurements on rock samples are often not representative of the bulk of the rock unit. Outcrop samples are often affected by weathering and lightning strikes. Samples collected in mines and quarries are often contaminated by exposure to magnetic fields. Drill core samples may, in special circumstances, acquire an axially-directed overprint during drilling using highly magnetic drill barrels.

To illustrate these points a number of specific examples are discussed.

Featherbed volcanics (rhyodacites) from North Queensland

A prominent negative anomaly is associated with an outcropping volcanic unit in an area considered possibly favourable for tin mineralisation. Of the five sites sampled one exhibited well-grouped reversed NRMs which were stable to both thermal and AF cleaning, with no evidence of other components. This stable NRM is believed to be primary Carboniferous TRM. The other sites were all lightning affected, with erratically varying NRM intensities and Q values, and scattered directions. Some samples had normal polarity NRM directions and these would have confused interpretation had they been considered representative. However, lightning-induced IRM is magnetically softer than the primary TRM. Therefore AF cleaning was able to remove preferentially the IRM, and the cleaned directions corresponded with directions from the unaffected site. The observed anomaly could then be accounted for using the reversed primary direction and intensity values from the lightning-unaffected site. There was therefore no need to explain the negative anomaly by postulating the presence of a reversely magnetised ore body similar to those elsewhere in the area.

Mogo Hill basaltic diatreme, Sydney Basin

The geology, petrophysics and geophysics of this intrusion have been described by Emerson & Wass (1980) and the palaeomagnetism is discussed by Schmidt & Embleton (1980). An 800 gamma negative anomaly is associated with the body, indicating the overall *in situ* magnetisation of the intrusion is reversed. However, the measured NRM directions are very scattered with normal polarities as common as reversed. The remanence is mostly carried by MD titanomagnetite grains which are magnetically soft and which have readily picked up random IRM components. The samples were collected in a quarry, so they are presumably not lightning affected. These randomly directed IRMs are readily removed by AF cleaning, sometimes in fields as low as 5–10 Oersted, and the cleaned directions are very well-grouped about a mean direction which is near vertical downwards. The AF cleaned direction is consistent with the mean magnetisation of the body inferred from magnetic modelling.

The ultramafic intrusion at Mt Derriwong, NSW

This body is reported on by Emerson *et al.* (1979). Better agreement between the observed and calculated magnetic anomalies was obtained when the AF cleaned remanence direction was used rather than the mean NRM direction. The cleaned and raw remanence directions differ by 35°. Although it could be argued that departures from the chosen model geometry could account for the difference, it is considered unlikely because the interpretation is constrained by petrophysical and geological information and corroborated by gravity modelling. The cleaning data suggest the presence of two components in the NRM, the softer of which is unrepresentative of the bulk of the intrusion. The explanation is uncertain but the more easily removed component which approximates the present field direction may be associated with weathering. If this is so it is fortuitous that the secondary magnetic minerals produced by the weathering are predominantly of low coercivity. If the secondary magnetisation is viscous in origin it should be present throughout the body, contrary to the conclusions based on modelling.

Similar behaviour has been observed by Ellwood (1981). AF cleaning of weathered granite samples removed unstable secondary components revealing a hard stable component with direction identical to the NRM direction of fresh granite. The soft secondary component is attributed to maghaemitisation of primary magnetite.

Tennant Creek (NT) massive magnetite

Drill-core samples were strongly remanently magnetised such that measured NRMs, if representative, would have produced a much larger anomaly than that observed over the orebody. Block samples were found to have much lower NRMs, consistent with modelling. The drill-core NRMs were axial, suggesting the presence of a spurious, drilling-induced component. Laboratory tests showed that this very magnetically soft material was capable of acquiring remanence during drilling, probably due to vibration in an ambient field. AF cleaning successfully removed the drilling-induced overprint and allowed determination of the original *in situ* NRM. It is believed that this type of spurious overprint is relatively rare and is restricted to rocks whose magnetic mineralogy is dominated by very soft grains. In this case the *in situ* Q_n should be less than unity and the NRM should be dominated by a viscous component parallel to the induced magnetisation, allowing assumption of induction in modelling but with an effective susceptibility enhanced by the remanent component.

Pyrrhotitic sediments of the Cobar area, NSW

A number of linear magnetic anomalies in the Cobar area are associated with steeply-dipping pyrrhotite-bearing sedimentary horizons. Oriented drill-core samples from these units fall into two categories. Specimens from one type of sample are characterised by high NRM intensities, high Koenigsberger ratios and well-grouped normal polarity NRM directions. The other type of sample has relatively low NRM intensities and Q values, and individual specimen directions are streaked along a great circle — suggesting a two-component system. The problem arises of estimating representative values of the magnetic parameters for input to magnetic interpretation. The susceptibility of the sediments is inadequate to account for the anomalies and this raises the question of whether remanent magnetisation of the sediments is responsible or whether a discrete magnetic source has been missed.

Thermal and AF cleaning of the pyrrhotitic sediment samples elucidates the NRM measurements by revealing the nature of the multicomponent magnetisation present in these rocks. All samples bear two components of magnetisation: a ubiquitous soft normal component overprinting a hard component, which may be normal or reversed. The soft normal and hard normal components are similar in direction and where both are present in a sample the NRM intensity is augmented and the specimen NRM directions are tightly clustered. On the other hand, a hard reversed component, where present, partially cancels the soft normal component, reducing the NRM intensity and producing a streaking of directions due to variations in the proportions of soft and hard components present in individual specimens. Normal and reversed hard components appear to be of similar intensity and are equally common. Reversals occur on a scale of 10–20 m stratigraphic thickness. Therefore the hard components probably largely cancel out and the overall magnetic anomaly is attributable to the ubiquitous soft normal component, which is several times the induced magnetisation. Thus palaeomagnetic cleaning enables a much more confident interpretation of the magnetic property measurements, where at first sight the sampling seemed to be inadequate.

In conclusion, indications are that palaeomagnetic cleaning may be useful in determining representative remanence components. Analysis of multi-component magnetisations through various cleaning techniques and vector analysis shows promise as an improved method of assigning remanence parameters to sampled rock formations. Provided the sampling is carried out over a section which was originally representative, the magnetisation components applicable to the whole rock unit are (1) stable ancient magnetisations and (2) superimposed viscous components. The palaeomagnetic noise, which it is desirable to remove before using NRM measurements as input to interpretation, is represented by magnetisations carried by secondary magnetic minerals produced by weathering, IRMs (whether lightning-induced or post-sampling), and (rarely) components associated with vibration or shock, as in drilling- or blasting-induced magnetisations.

More than one cleaning technique may be required to investigate the remanence components as the efficacy of each technique is dependent on the nature of the components. For instance VRM is readily removed by thermal demagnetisation but not so readily by AF demagnetisation. The reverse applies to IRM.

Ancient geomagnetic field directions

The data presented in Fig. 4 demonstrate that remanence contributes substantially to the magnetisation of most rocks. Information on probable remanence directions for rocks of different ages is therefore of considerable interest. The configuration of the ancient geomagnetic field during the Phanerozoic is now quite well defined by palaeomagnetic studies and this information has been adapted for use in magnetic interpretation.

A number of assumptions are involved in application of this information to interpretation:

(1) The rock unit acquires a magnetisation parallel to the ambient field at the time of formation. This assumption, with few exceptions, is supported by a vast body of palaeomagnetic data.

(2) The ancient field direction is assumed to correspond to that of an axial geocentric dipole. This is a fundamental tenet of palaeomagnetism and is believed to hold for ancient geomagnetic field configurations averaged over 10^3–10^4 years. A magnetisation acquired more rapidly than this, for instance by a lava flow, will represent a spot reading of the ancient field direction, which may depart from the average dipole field direction typically by up to 15° ($\simeq$ 10° for palaeopolar regions and $\simeq$ 20° for palaeoequatorial regions). In general, secular variation of the geomagnetic field is averaged out for plutonic rocks, which acquire their magnetisation during slow cooling; for sediments, which acquire remanence during compaction and diagenesis; and for volcanic piles where extrusion has spanned thousands of years. In these cases the mean remanence direction corresponds to the dipole field direction.

(3) The rock unit has not been tectonically disturbed since formation. Correction of remanence directions for the effects of tectonic tilting are discussed later. It is also assumed that the rock unit has been an integral part of the Australian continent since formation, i.e. the locality is not a displaced terrain.

(4) The rock unit has not undergone metamorphism or alteration since formation. This restriction is the most serious limitation to applicability of this approach, because in many cases the primary remanence of rocks is partially or wholly overprinted during post-formation geological events. If the rock is wholly overprinted the analysis is still valid provided the age of remagnetisation is considered rather than the age of formation. For example, upper greenschist facies or higher grade regional metamorphism

of magnetite-bearing rocks will reset the magnetisation. The relevant age of magnetisation in this case is the age of post-metamorphic cooling. Similar considerations apply to total replacement of the magnetic mineral assemblage during hydrothermal alteration, late diagenetic alteration, lateritisation, etc.

The case of partial overprinting is more complicated. If a secondary component of magnetisation is superimposed on a primary remanence the NRM direction will lie in the plane containing the primary and secondary remanence directions. Although in many cases primary and secondary remanence directions of the same polarity may be sufficiently similar that a reasonable estimate of the NRM direction could be made, frequently the two components will be of opposite polarity and the resultant direction will be very uncertain. For instance, if the normal polarity directions from the age of formation and the age of partial remagnetisation differ by 20°, the NRM direction is determinable to within 10°. However, if the geomagnetic field was normal at the time of formation and reversed at the time of remagnetisation the NRM direction could lie anywhere along a 160° arc.

Complex geological histories which have produced complex magnetisations with three or more components preclude estimation of NRM directions from theoretical considerations alone. In these cases sampling is essential for estimation of NRM directions. Palaeomagnetic cleaning of the samples is also desirable as knowledge of the magnetisation components increases confidence in estimation of representative remanence parameters.

Palaeomagnetic poles and palaeofield directions

Palaeomagnetic pole positions with respect to Australia for the Phanerozoic are listed in Table 3 (taken from Embleton & Schmidt 1981). These poles are interpreted as representing the south geographic pole position with respect to Australia. Corresponding palaeofield directions have been calculated for a reference point in the centre of the continent (24°S, 134°E — near Alice Springs) assuming the palaeopole is a south magnetic pole, i.e. ignoring reversals of the geomagnetic field. Therefore, it should be borne in mind that for all the given field directions the reversed sense could apply. For example at −65 M.yr, the direction could be either (27°, −73°) or (207°, +73°). In practice an ambiguity of 180° is not an obstacle to interpretation as the polarity of the magnetisation will usually be obvious from the form of the anomaly.

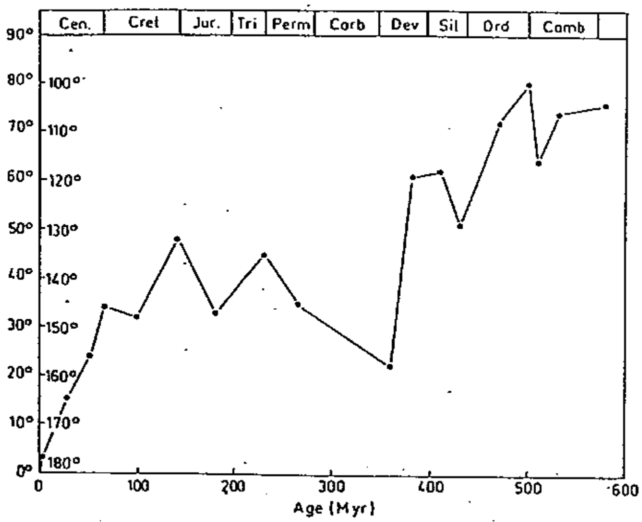
The angular deviation of the ancient geomagnetic field direction from the present dipole field direction is plotted in Fig. 5. It is clear that substantial departures of remanence directions from parallelism or antiparallelism with present field directions are to be expected.

TABLE 3

Geomagnetic field directions through the Phanerozoic

Geological period	Age (M.yr)	Pole position Lat.	Pole position Long.	Geomagnetic field direction at 24°S, 134°E Dec. (°)	Inc. (°)	α_{95} (°)
Quaternary-Late Tertiary	0-4.5	87°S	86°E	2	−44	1
Mid-Late Tertiary	20-34	75°S	99°E	11	−55	4
Early Tertiary	40-60	66°S	127°E	4	−66	6
Cretaceous-Tertiary boundary	65	51°S	112°E	27	−73	2
Middle Cretaceous	100	53°S	158°E	334	−71	4
Mid Jurassic-Late Jurassic	130-150	25°S	167°E	278	−74	9
Late Triassic-Early Mid Jurassic	160-200	48°S	180°E	317	−64	3
Early Triassic	230	30°S	147°E	300	−83	4
Late Carboniferous-Late Permian	240-290	48°S	137°E	354	−77	3
Middle-Late Devonian	350-370	63°S	77°E	28	−54	14
Early-Middle Devonian	370-390	47°S	267°E	330	+13	12
Middle-Late Silurian	400-420	48°S	356°E	26	+16	19
Early-Middle Silurian	420-440	39°S	34°E	51	−16	9
Ordovician	450-490	0°	22°E	83	+36	22
(Delamarian)	500	27°N	72°E	127	−22	11
Early Late Cambrian	510	15°N	30°E	98	+32	23
Middle Cambrian	520-540	34°S	18°E	50	+16	7
Latest Precambrian-Early Cambrian	600-560	49°S	345°E	20	+22	23

The successive pole positions represent the interpreted apparent movement of the south geographic pole with respect to Australia; declination = azimuth of field direction measured positive clockwise from true north; inclination = dip of field direction measured positive downwards from horizontal; α_{95} = radius of the error circle about the mean direction, at the 95% confidence level.



IGURE 5

Departure (θ) of the ancient geomagnetic field direction from the present dipole field direction at 24°S, 134°E. The dual scale (0°-°0° and 180°-90°) refers to opposite polarities of the geomagnetic ald.

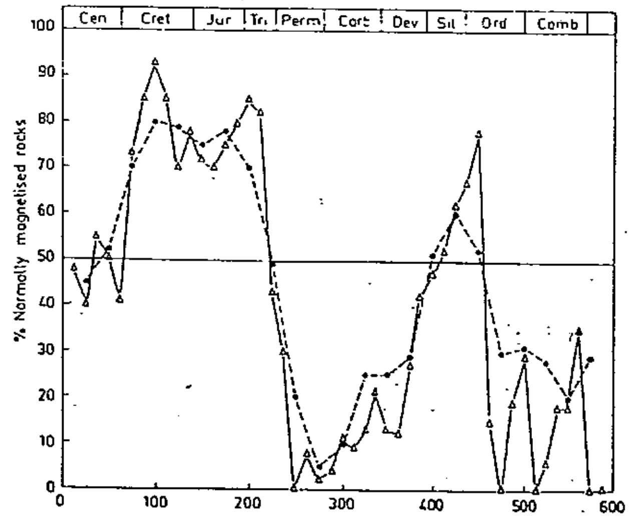


FIGURE 6

Geomagnetic polarity ratios (expressed as percentage of normally magnetised rocks) as a function of rock age for the Phanerozoic. Both 25 M.yr and 50 M.yr overlapping averages are shown. Some data used in the 50 M.yr averages are excluded from the 25 M.yr averages due to insufficient age precision: (Δ) 25 M.yr overlapping averages; (•) 50 M.yr overlapping averages.

Although geomagnetic field reversals have occurred fre-uently in the past, reversals are not evenly distributed iroughout geological time. Quiet intervals, which last typically for the order of ten million years (≃10 M.yr), are periods of infrequent reversals and strong polarity bias. uiet intervals are separated by disturbed intervals of milar duration, during which reversals are frequent but which have polarity bias correlated with adjacent quiet intervals. Polarity bias intervals (preferred normal or versed polarity) typically last ≃100 M.yr.

The polarity of the geomagnetic field throughout the Phanerozoic has been reviewed by Irving & Pulliah (1976). ata from their paper have been incorporated into Fig. 6, which plots percentage normal polarity for 25 M.yr and 50 M.yr windows. Figure 6 can be used in conjunction with able 3 to define the remanence direction corresponding to ı age of magnetisation by reducing ambiguity in polarity. For example a Permian magnetisation will almost certainly be reversed and the estimated direction from Table 3 will ierefore be (354°, −77°) reversed, i.e. (174°, +77°). It ın be seen from Fig. 6 that polarities are more or less equally mixed during the Cenozoic, are predominantly normal for the Mesozoic and predominantly reversed for e Palaeozoic, except for the Silurian and Upper Ordovician.

Because the Australian continent subtends about 35°, palaeofield directions at sites remote from central Australia ill differ somewhat from those at the reference point. As ı illustration the present field direction in northernmost Australia is about 20° shallower than the direction in central Australia, whereas the direction is about 20° steeper Tasmania. Therefore, it may be desirable in some cases to recalculate directions given in Table 3 for the locality of interest. The maximum possible error as a result of neglec-ıg the dependence of field direction on locality is about vice the angular separation of the locality and the reference point (24°S, 134°E). In most cases the error will be much less than this.

If the locality latitude and longitude are (λ', ϕ) and the palaeopole position given in Table 3 (or its antipole — whichever is the closer to the locality) is (λ, ϕ') then the palaeo*colatitude*, p, is given by (refer Fig. 7):

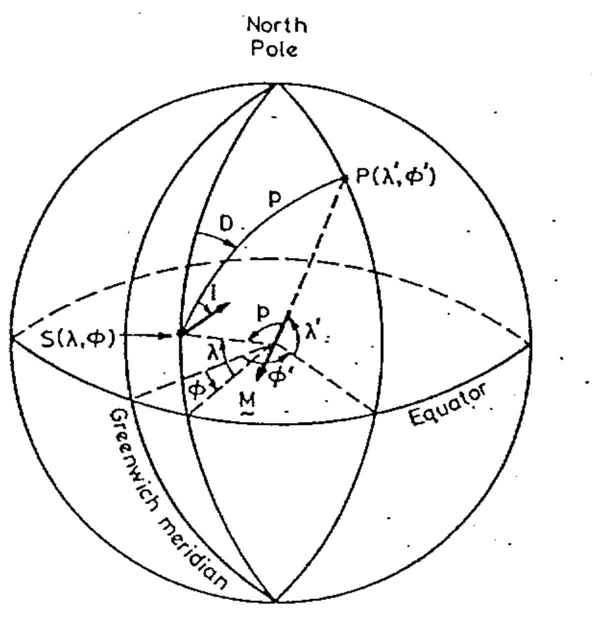


FIGURE 7
Definition of quantities for calculation of palaeolatitude (90-p) and inclination (I) given λ, ϕ the sampling latitude and longitude and λ', ϕ' the palaeopole position. M = ancient axial geocentric dipole. Tan I = 2 cot p.

$\cos p = \cos\phi \cos\lambda \cos\phi' \cos\lambda' + \sin\phi \cos\lambda \sin\phi' \cos\lambda' + \sin\lambda \sin\lambda'$

The palaeo*latitude* of the locality is $90° - p$.

The magnetic inclination, I, is given by:

$I = \tan^{-1} (2 \cot p) = \tan^{-1} (2 \cos p/[1-\cos^2 p]^{1/2})$

The magnetic declination, D, is given by:

$\cos D = [(\sin\phi \cos\lambda \sin\lambda' - \sin\phi' \cos\lambda' \sin\lambda) \sin\phi + (\cos\phi \cos\lambda \sin\lambda' - \cos\phi' \cos\lambda' \sin\lambda) \cos\phi] / (1-\cos^2 p)^{1/2}$

Note that $0° < D < 180°$ if the pole meridian lies east of the site meridian, $180° < D < 360°$ if the pole meridian lies west of the site meridian (the shortest distance between meridians being taken).

Once the palaeodirection (D, I) is derived it can be compared for polarity with Table 3 and reversed if necessary in order to produce an equivalent table for the locality.

It is often useful to calculate the angle, θ, between two directions (D, I) and (D′, I′). The equation relating these parameters is:

Effect of tectonic tilting

Tectonic movements after acquisition of remanence will rotate the remanence vector with the rock unit (Fig. 8). The anticipated remanence directions will therefore have to take this into account. It is sometimes possible to interpret detailed geological structure on the basis of bedding-remanence relationships around folds.

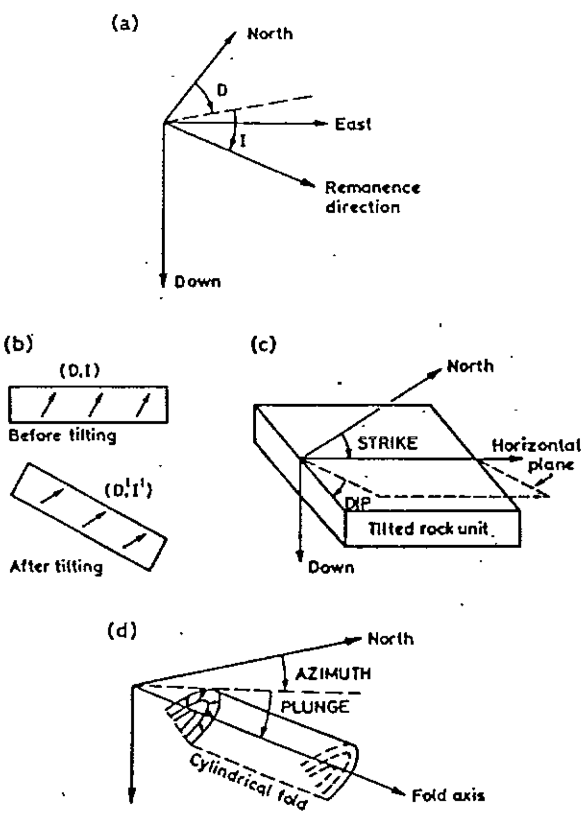


FIGURE 8
Effect of tectonic tilting on remanence directions.

A particularly simple correction applies to the case of tectonic rotation about a vertical axis. This occurrence is, however, relatively rare and difficult to recognise in practice. The magnetic inclination is unaffected by the rotation and the declination is augmented by the rotation angle, defined positive clockwise.

A much more common case is that of cylindrical folding about a horizontal fold axis. Define the bedding strike positive clockwise from north. The dip is then defined with the strike direction to the right as the observer faces the rock unit. The dip is $> 90°$ if the rock unit is overturned. Then if the pre-folding remanence direction was (D, I) the post-folding declination and inclination are given by:

$D' = \tan^{-1} [\tan(D\text{-STRIKE}) \cos DIP - \tan I \sec (D\text{-STRIKE}) \sin DIP] + STRIKE$

$I' = \sin^{-1} [\sin I \cos DIP + \cos I \sin (D\text{-STRIKE}) \sin DIP]$

If the angle between the initial declination, D, and the strike is greater than 90°, then 180° must be added to D′. Alternatively, the strike direction closest to the horizontal remanence component can be taken, defining DIP as positive if this strike direction goes from left to right and as negative if the strike goes from right to left as seen by the observer.

In structurally complex areas, fold axis plunges may have to be taken into account. In terms of fold axis azimuth and plunge and bedding strike and dip the rotated vector is given by:

$D' = \tan^{-1}[(\cos I \sin(D\text{-AZIMUTH}) \cos DIP^* - \sin I \sin DIP^*) / (\cos I \cos (D\text{-AZIMUTH}) \cos PLUNGE - (\sin I \cos DIP^* + \cos I \sin (D\text{-AZIMUTH}) \sin DIP^*) \sin PLUNGE)] + AZIMUTH$

$I' = \sin^{-1} [(\sin I \cos DIP^* + \cos I \sin (D\text{-AZIMUTH}) \sin DIP^*) \cos PLUNGE + \cos I \cos (D\text{-AZIMUTH}) \sin PLUNGE]$

DIP^* is the bedding dip when the effect of plunge is removed and is given by:

$DIP^* = \tan^{-1} [\tan DIP \cos PLUNGE \cos (AZIMUTH - STRIKE)]$

The 180° ambiguity in D′ due to the properties of the $\tan^{-1}$ function is easily resolved by inspection.

An alternative to these rather complicated formulae is a graphical method. Anyone familiar with the stereonets used by structural geologists and palaeomagnetists can readily carry out the necessary rotations.

Applications of rock magnetism

Apart from providing input for magnetic interpretation, a number of other potential applications of rock magnetism to the mining industry are apparent. These include the following.

Redox CRM technique (Bacon & Elliott 1981)

This is a newly proposed method for detection of sulphide ore bodies beneath volcanic cover. The concept is based on remobilisation of iron in younger volcanics covering a then-active redox potential cell associated with an ore body. Measurements of susceptibility and NRM intensity of samples collected along profiles appear to exhibit a characteristic signature above ore bodies which may be buried beneath a considerable thickness of volcanics.

Magnetic fabric

Magnetic susceptibility anisotropy can be interpreted in terms of magnetic fabric (foliation plane plus lineation) which is analogous to conventional petrofabric and which has similar application to structural interpretation. The advantages of the magnetic method are the ease and rapidity of fabric determination compared to microscopic techniques, and the great sensitivity enabling detection of mesoscopic fabrics which are too indistinct to map in the field. Applications include determination of the bedding plane in poorly-bedded sediments, estimation of finite strain ellipsoids in tectonites, and determination of the mode of emplacement of igneous rocks.

Down-hole susceptibility logging

As well as providing useful data for magnetic interpretation, susceptibility measurements on drill-core samples or using a down-hole probe (which is preferable) can sometimes be used for stratigraphic correlation between drill holes. In the amphibolites of the Broken Hill area, for instance, subtle lithological variations which are difficult to trace by conventional means, are reflected in substantial susceptibility variations, allowing matching of susceptibility signatures between holes.

Magnetic analytic techniques

A number of potential applications of magnetic techniques to mineral processing are apparent: Measurements of magnetic properties of the circulating medium (e.g. magnetite plus ferrosilicon) in heavy media separation plants should provide a rapid means of monitoring the quantity and composition of the magnetic material at various points of the plant. At present, time consuming wet chemical analysis is employed. A similar application is apparent for monitoring magnetite in coal washeries. Magnetic properties of different magnetites may also be used to ascertain their suitability for coal washing. Flotation of sulphide minerals is often affected by the presence of various forms of pyrrhotite. Another problem presented by some pyrrhotites is spontaneous combustibility. Thermomagnetic measurements could be used to analyse ore for the different pyrrhotites and therefore guide ore treatment.

Palaeomagnetic dating of mineralisation

Hitherto little application has been made of palaeomagnetism in studying ore genesis and thermal history. In certain cases palaeomagnetism could distinguish between syngenetic and epigenetic origin of the mineralisation. Palaeomagnetic dating is also applicable to weathered profiles such as laterites, bauxites and gossans.

Magnetostratigraphy as a tool for geological correlation

Investigation of magnetic viscosity of SPM grains in rocks and its effect on transient electromagnetic (TEM) methods

The widespread occurrence of SPM material in overburden accounts for anomalous tails at late delay times in the coincident loop mode (Buselli 1982). SPM behaviour is now recognised as being more common in rocks than formerly believed and this fact will affect procedures for TEM surveys.

Conclusions

It is hoped that these comments will illustrate the great potential of rock magnetic studies for application to geological and geophysical problems. In particular there is a pressing need for magnetic petrophysical studies to improve our understanding of the relationship between geology and magnetic anomalies.

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