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Notes on Rock Magnetization Characteristics in Applied Geophysical Studies

D. A. Clark
CSIRO Division of Exploration
PO Box 136
North Ryde, NSW 2113

D. W. Emerson
Department of Geology and Geophysics
University of Sydney
NSW 2006

Introduction

The magnetic method is firmly established as an investigative procedure in applied geophysics, and it is increasingly recognised that an understanding of the elements of magnetic petrophysics assists in optimising field data interpretations. These notes expand on the rock magnetism data previously presented by Emerson (1979) and Clark (1983); the data presented are based on magnetic property studies in the CSIRO Division of Exploration Geoscience and the Department of Geology and Geophysics University of Sydney and on published studies and compilations. The systematic collection of petrophysical data by many workers over the last decade has greatly expanded the size and scope of available data. Works listed in the Bibliography should be consulted for further information.

A rock's total magnetization, a vector quantity, comprises two components — induced and remanent. Induced magnetization is dependent on the applied field and is thus a dynamic property. The induced magnetization is related to a small applied field through a constant of proportionality — the susceptibility k. Remanence is a permanent magnetization that can be regarded as a static property as it is independent of weak applied fields.

Iron bearing minerals exert a dominant influence in rock magnetism. Ferrimagnetic magnetite accounts for about 1.5 percent of crustal minerals and it is by far the most important magnetic mineral in applied studies. However, all iron bearing minerals impart paramagnetism to rocks and these can be important influences in areas lacking magnetite and other sources of strong magnetization.

These notes briefly review magnetic mineralogy and the methods of measuring susceptibility and remanence. Magnetic susceptibility values are then presented with a discussion of their significance and this is followed by an outline of remanence effects in common rock types. Demagnetization is not discussed.

Magnetization and susceptibility terms and units can be confusing. Magnetization is the magnetic moment per unit volume. Magnetic volume susceptibility (k) is the magnetic moment acquired per unit applied field per unit volume; it is the ratio of a magnetization to a field and as such is dimensionless. Volume susceptibilities are mostly used in these notes. Mass or specific susceptibility (χ) may be obtained by dividing the volume susceptibility by the density. Molar susceptibility is the mass susceptibility multiplied by the molecular weight.

In the cgs system susceptibilities can be labelled emu or gauss/oersted (G/Oe) to identify the quantities. S.I. susceptibilities are 4π times the cgs values. Magnetizations (remanent, induced) often are cited in microgauss (μG, = 10^{-6} emu/cm^3) or gammas (γ, 10^{-5} G = 10 μG) or milliamp/metre (mA/m S.I., numerically equal to μG in cgs).

Key words: susceptibility, magnetization, Koenigsberger ratio, remanence, magnetic mineralogy, magentic petrology

Magnetic Mineralogy

Magnetite and maghemite have high intrinsic susceptibilities and can carry large permanent magnetizations. Their influence dominates in most magnetic surveys. Monoclinic pyrrhotite is ferrimagnetic and is also an important magnetic mineral in hard rock exploration; Clark (1983) has documented the magnetic properties of this mineral.

The compositions of the Fe-Ti oxide minerals that play a major role in rock magnetism are shown in Figure 1a. The rhombohedral titanohematites are second in importance only to the titanomagnetites. Titanohematites containing between 50 mole% and 80 mole% ilmenite are strongly magnetic and are efficient carriers of remanence. Compositions closer to ilmenite are paramagnetic at room temperature and hematite-rich compositions are only weakly magnetic. Titanomagnetites with less than about 80 mole% ulvospinel are ferrimagnetic at room temperature. Although the spontaneous magnetization and Curie temperature of titanomagnetites decrease steadily with increasing titanium content (Figures 1b, c), the susceptibility and specific intensity of TRM are not strongly dependent on composition for the ferrimagnetic phases. Thus high titanium content in magnetite does not generally produce weaker magnetic properties, contrary to popular opinion. In fact, grain size is a more important factor influencing magnetization of titanomagnetite-bearing rocks (see Figure 1d, e).

Laboratory and Field Measurement of Rock Magnetisations

The principles governing measurement of susceptibility and remanence are shown schematically in Figure 2. The resultant magnetization of a rock is the vector sum of the induced and remanent magnetizations.

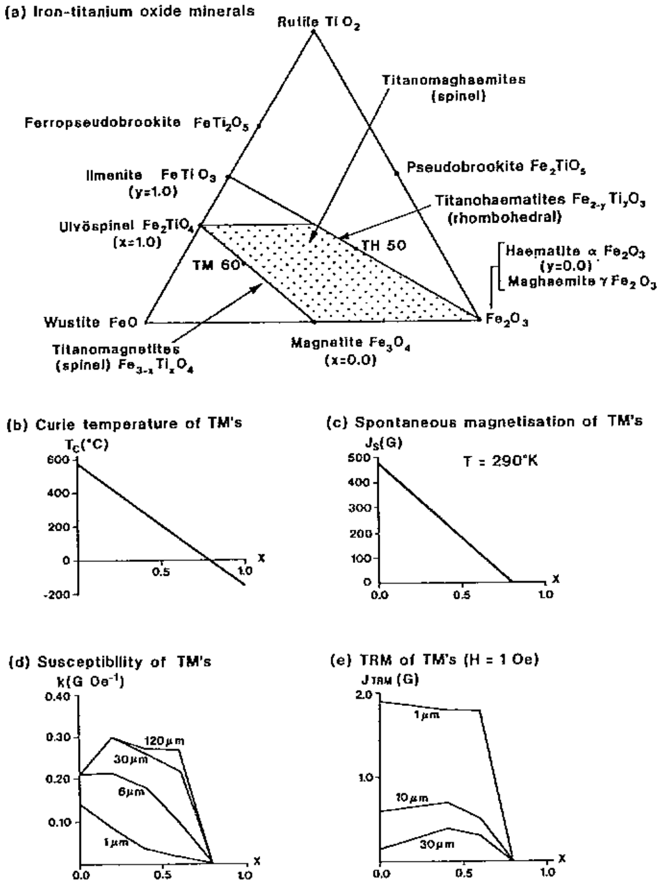

FIGURE 1

Both methods of susceptibility measurement rely on detecting changes in the reluctance of a magnetic circuit when a permeable specimen is introduced into the flux path; this is analogous to the change in resistance of an electrical circuit. The change in inductance of the transformer core when the specimen is placed in the gap produces a signal unbalance that is dependent on the specimen susceptibility. Placing the hand-held susceptibility meter against the surface of a magnetic sample changes the inductance of the core winding, and hence the resonant frequency of the LC oscillator. The change in frequency is dependent on the susceptibility.

Alternating fields used in these instrument are deliberately kept low in frequency (<5 kHz) to avoid exciting conductive responses from magnetite which itself is an excellent semiconductor with a resistivity of about 10^{-5} ohm m. Measurements are sometimes made at two frequencies a decade apart (e.g. 430 Hz and 4.3 kHz) to identify hyperfine magnetic grains where magnetic viscosity causes a lower susceptibility at the higher frequency.

Spinner fluxgate magnetometers are sufficiently sensitive to measure the remanence of small rock specimens for all but the very weakly magnetised rocks. The specimen rotates within a magnetic shield and produces a sinusoidal magnetic field at the fluxgate sensor. The amplitude and phase of the fluxgate signal are used to determine the intensity and direction of the remanence component in the plane orthogonal to the spin axis. By repeating the measurement with the specimen reoriented the complete remanence vector is determined. Cryogenic magnetometers are very sensitive instruments, capable of measuring remanences as low as 10^{-9} G (1 μA/m S.I.). The specimen is lowered into the measurement region within a superconducting shield and the change in flux passing through SQUID detectors is measured. If three detectors are used the full remanence vector can be determined almost instantaneously.

Susceptibility

A considerable amount of susceptibility information has been condensed for presentation in Figure 3 where susceptibility ranges for common rock types are presented. It is evident that each rock type exhibits a wide range of susceptibilities and that susceptibility 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*) as shown in Figure 4. Accessory minerals are generally ignored in petrological classifications. 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 compositions and microstructures 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_2 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.

The variations in magnetic properties for given lithologies are 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 3 is that the susceptibilities of a number of rock types have distinctly bimodal distributions. In the case of granitoids this reflects the existence of two distinct categories that have only recently been recognised: 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 Chappel and White (1974), whereas ilmenite-series granitoids are more reduced and are usually S-type. The new classifications, which have important petrogenetic 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.

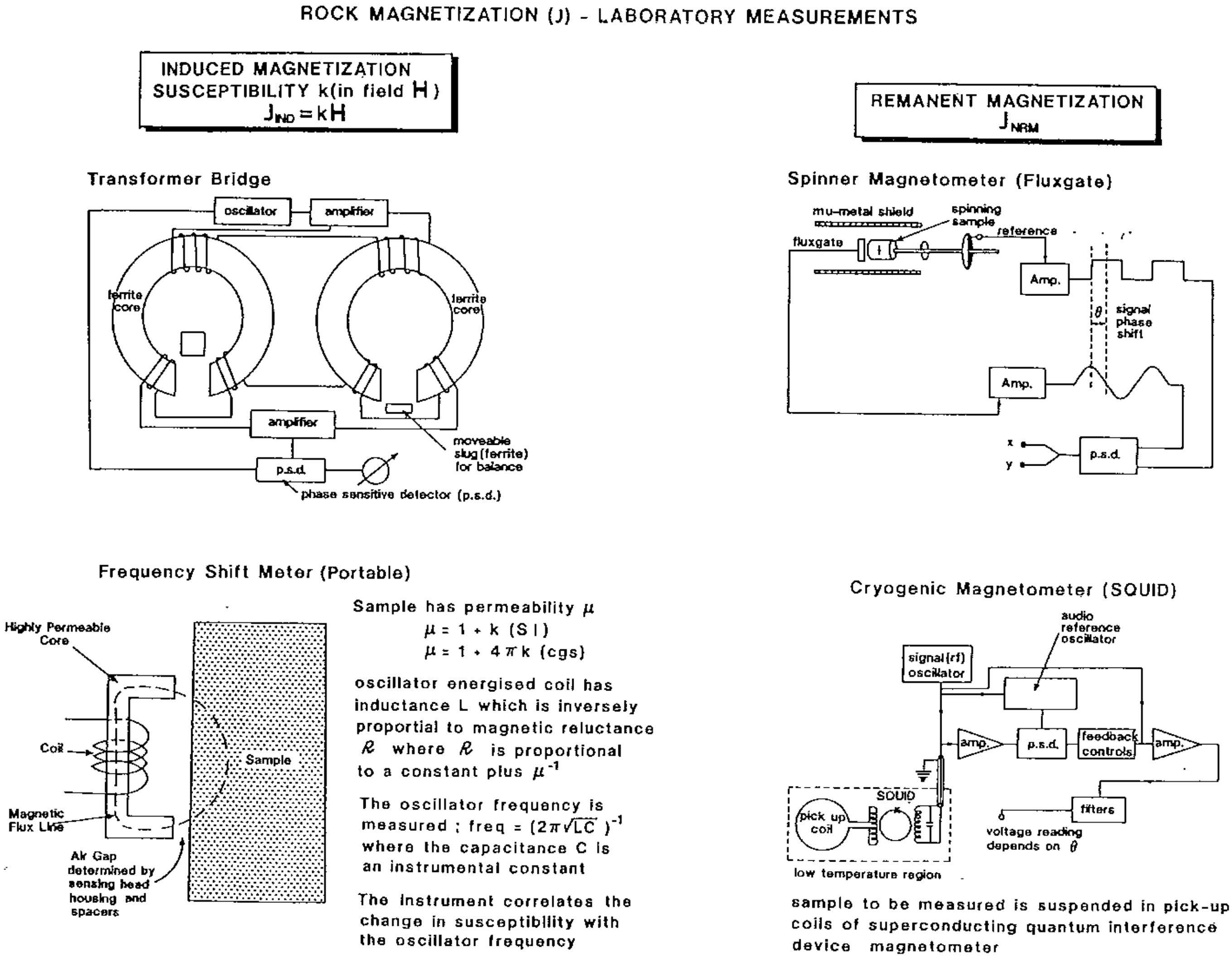

FIGURE 2

More generally, bimodal susceptibility distributions represent distinct subpopulations within each rock type for which ferrimagnetic minerals are absent and present respectively. Iron in the weakly magnetic subpopulation is incorporated into paramagnetic silicate minerals, predominantly as Fe^{2+}. Iron rich garnets, olivines, pyroxenes, amphiboles and micas are minerals that may exhibit susceptibility values up to several hundred μG/Oe. Similar rocks that are moderately to strongly magnetic contain significant Fe^{3+} which is incorporated into magnetite. Very highly oxidised rocks, however, tend to contain hematite rather than magnetite and are therefore also weakly magnetic. Within each magnetic subpopulation susceptibility tends to increase with basicity.

The greater abundance of paramagnetic mafic minerals in rocks with lower SiO_2 increases the paramagnetic contribution to the susceptibility. This produces a small, but systematic, difference in the susceptibilities of paramagnetic acid and basic rocks. The susceptibility of a paramagnetic rock can be calculated from its chemical composition. The susceptibility is the sum of contributions from paramagnetic ions, including ferrous iron, ferric iron, manganese, chromium etc. In practice, for most rocks only iron need be considered and the difference between ferrous and ferric species can be neglected, without greatly affecting the accuracy of the estimated susceptibility. Then the mass susceptibility is given by: χ_{SI} (10^{-8} m^3/kg) $\approx 2.86W_{Fe} = 2.22W_{FeO}$ or $\chi_{cgs} \approx 2.27W_{Fe} = 1.77W_{FeO}$ (μG cm^3/g Oe), where W_{Fe} is the weight per cent Fe and W_{FeO} is the weight per cent total iron as FeO (Puranen, 1989). As an example, consider a granite with 2 wt% FeO and a gabbro with 12 wt% FeO. The calculated mass susceptibilities, assuming the rocks are devoid of ferrimagnetic minerals, are 4.4×10^{-8} m^3/kg = 3.5 μG cm^3/g Oe and 2.7×10^{-4} m^3/kg = 21 μG cm^3/g Oe respectively. The corresponding volume susceptibilities, assuming densities of 2.65 g/cm^3 = 2650 kg/m^3 for the granite and 3.0 g/cm^3 = 3000 kg/m^3 for the gabbro, are 12×10^{-5} SI = 9 μG/Oe for the granite and 80×10^{-5} SI = 64 μG/Oe for the gabbro. 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

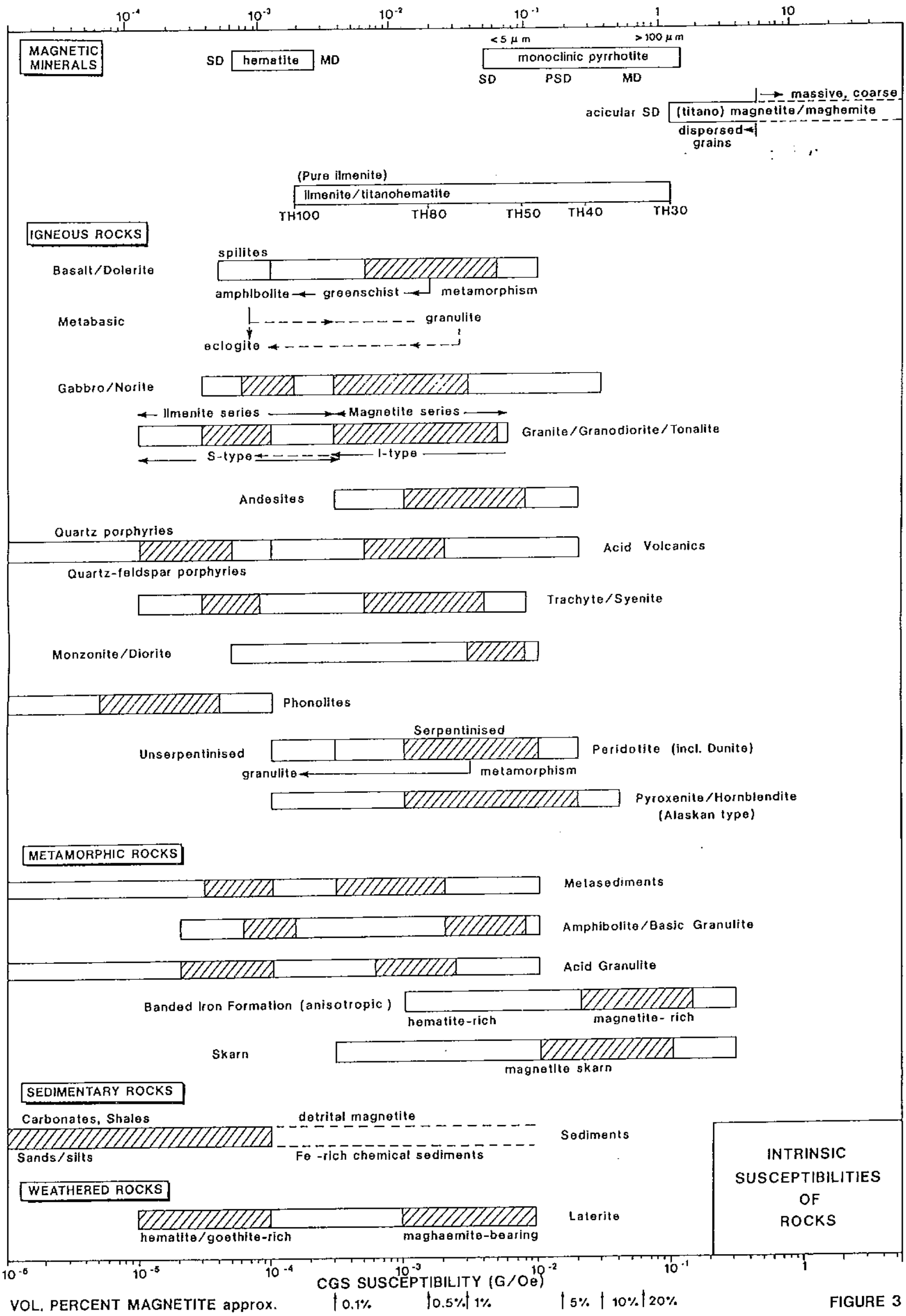
SI SUSCEPTIBILITY
10⁻⁴ 10⁻³ 10⁻² 10⁻¹ 1 10
MAGNETIC MINERALS
SD hematite MD
<5 μm >100 μm
monoclinic pyrrhotite
SD PSD MD
massive, coarse
acicular SD (titano) magnetite/maghemite
dispersed grains
(Pure ilmenite)
ilmenite/titanohematite
TH100 TH80 TH50 TH40 TH30
IGNEOUS ROCKS
spilites
Basalt/Dolerite
amphibolite greenschist metamorphism
Metabasic
granulite
eclogite
Gabbro/Norite
Ilmenite series
Magnetite series
Granite/Granodiorite/Tonalite
S-type
I-type
Andesites
Quartz porphyries
Acid Volcanics
Quartz-feldspar porphyries
Trachyte/Syenite
Monzonite/Diorite
Phonolites
Serpentinised
Unserpentinised
Peridotite (incl. Dunite)
granulite
metamorphism
Pyroxenite/Hornblendite (Alaskan type)
METAMORPHIC ROCKS
Metasediments
Amphibolite/Basic Granulite
Acid Granulite
Banded Iron Formation (anisotropic)
hematite-rich
magnetite-rich
Skarn
magnetite skarn
SEDIMENTARY ROCKS
Carbonates, Shales
detrital magnetite
Sediments
Sands/silts
Fe-rich chemical sediments
WEATHERED ROCKS
Laterite
hematite/goethite-rich
maghaemite-bearing
INTRINSIC SUSCEPTIBILITIES OF ROCKS
10⁻⁶ 10⁻⁵ 10⁻⁴ 10⁻³ 10⁻² 10⁻¹ 1
CGS SUSCEPTIBILITY (G/Oe)
VOL. PERCENT MAGNETITE approx.
0.1% 0.5% 1% 5% 10% 20%

FIGURE 3

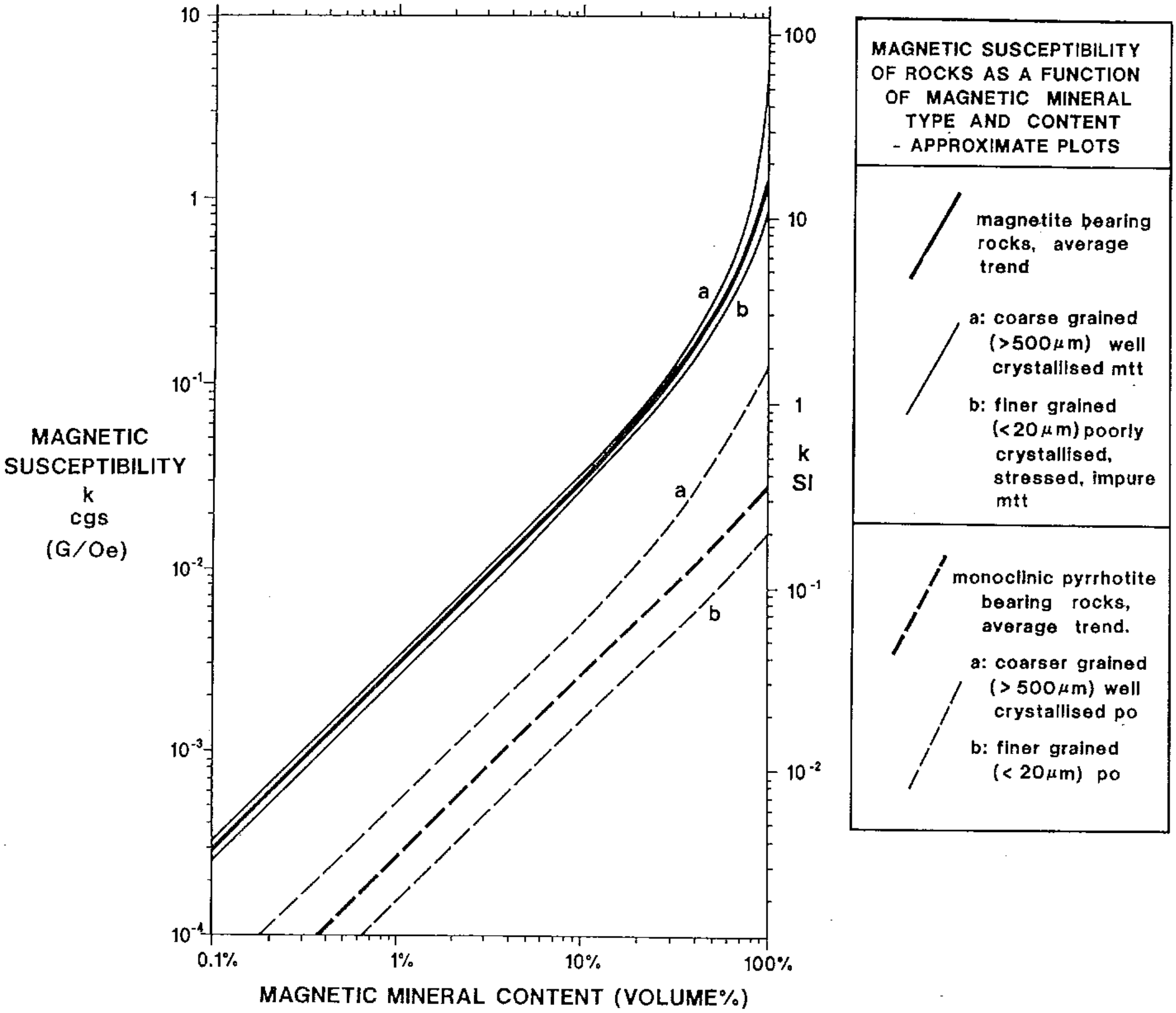

FIGURE 4

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 variations in the abundance of ferrimagnetic minerals. The overall tendency for magnetite content of ferrimagnetic rocks to increase with basicity is somewhat obscured when igneous rocks from different provinces are compared. For consanguinous rocks, in particular, there is a general correlation between susceptibility and basicity. Andesites generally have similar, or slightly lower, susceptibilities than related basalts.

Rhyolites have a distinctly bimodal susceptibility distribution. Ferrimagnetic rhyolites tend to be somewhat less magnetic than the more basic members of the series, and many rhyolites are paramagnetic. Trachyandesites and trachytes generally have moderate to high susceptibilities, comparable to or somewhat less than susceptibilities of related alkali basalts, but the corresponding phonolites are usually weakly magnetic. Within the ferrimagnetic subpopulation of each lithology, magnetic properties can also be related to geochemistry. For tholeiitic rocks in both oceanic and continental settings iron and titanium-rich variants have been found to have substantially higher susceptibilities, reflecting greater modal titanomagnetite, than similar rocks with lower Fe and Ti contents (e.g. Anderson *et al.*, 1975; de Boer and Snider, 1979).

"Clean" carbonates and clastic sediments have very low susceptibilities. Some immature sandstones are quite magnetic as they contain significant quantities of detrital magnetite which may provide an indication of provenance. Sediments deposited in the presence of metal-bearing solutions, associated with volcanic activity for example, may contain appreciable magnetite, or possibly pyrrhotite. Such sediments may be transitional to syngenetic massive mineralisation or transitional to banded iron formation.

Magnetite-rich banded iron formations are not only strongly magnetic but are characterised by high anisotropy of susceptibility. The susceptibility parallel to bedding is typically greater than the susceptibility normal to bedding by a factor of 2-4. The susceptibility values for banded iron formation shown in Figure 3 are bulk susceptibilities, i.e. the average of the susceptibilities along any three orthogonal directions. 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 influence significantly the form of magnetic anomalies. Exceptions include banded iron formations and some rocks and ores that contain pyrrhotite with a strong preferred orientation. Susceptibility anisotropy is a variable that may be included in the range of magnetic models presented by Emerson, Clark and Saul (1985) and Clark, Saul and Emerson (1986).

Magnetic susceptibility anisotropy parameters are summarised in Figure 5. Most rocks are only weakly anisotropic. Susceptibility anisotropies of undeformed sediments are generally in the range 1.00-1.05, occasionally up to around 1.08. Anisotropy in these rocks mainly reflects a bedding parallel foliation, with a weak superimposed lineation that reflects the palaeocurrent direction during deposition. Weakly metamorphosed sediments tend to have somewhat higher anisotropies, typically 1.05-1.1, and the magnetic fabric is more complex, reflecting tectonic overprinting of the primary fabric. Extrusive rocks exhibit relatively weak magnetic anisotropy (A = 1.0-1.2). Viscous lavas with flow banding (e.g. rhyolites) tend to have higher anisotropies than basic lava flows. Magnetic lineations generally indicate the flow direction in lavas. Plutonic rocks generally have anisotropy magnitudes ranging from near 1.00 (i.e. almost isotropic) to relatively high values (A $\approx$ 1.5). Occasionally even higher values, up to about 1.8, are found. The highest anisotropies occur in strongly foliated granitoids and in some cumulates that have well-defined bands rich in magnetic minerals.

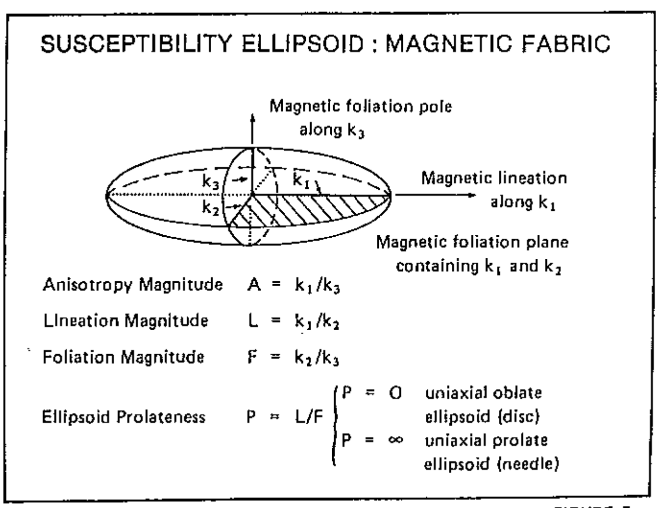

FIGURE 5

Strongly foliated metamorphic rocks, such as slates, schists and gneisses are more anisotropic than other common rock types. Anisotropy magnitudes of around 1.5 are common, and values up to 2 occur. In these rocks the magnetic fabric is usually dominantly oblate, with a pronounced magnetic foliation parallel to the petrofabric foliation, although strong magnetic lineations are also found. The most consistently anisotropic rocks are banded iron formations, which exhibit strong textural anisotropy due to the alternation of strongly magnetic and weakly magnetic bands. The anisotropy of BIFs is dominated by a bedding-parallel magnetic foliation. Thus the bedding corresponds to a plane of relatively high susceptibility. Within the bedding plane the susceptibility is usually fairly isotropic, unless the rocks are deformed on the mesoscopic scale. Anisotropy magnitudes for BIFs are typically 2 to 4, but higher values can occur. Other examples of high anisotropy include pyrrhotite-bearing rocks and ores (A = 1.1-2.0, average 1.4), and some massive hematite and hemo-ilmenite ores.

The iron content of the sediment (generally higher for pelites than for psammites) and the Fe^{3+}/Fe^{2+} ratio, which reflects the redox conditions during deposition and diagenesis, have a major bearing on the capacity of rock to develop secondary magnetite during metamorphism. 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, 1981). Pyrrhotite is the main magnetic mineral in many metasedimentary rocks, particularly in mineralised areas. Monoclinic (4C) pyrrhotite, which approximates Fe_7S_8 in composition, is the ferrimagnetic variety. More iron-rich varieties of pyrrhotite are very weakly magnetic.

Fresh basalts and dolerites have moderate to high susceptibilities. Hydrothermal alteration of these rocks usually reduces the susceptibility. Regional metamorphism to greenschist grade, and a *fortiori* to amphibolite grade, tends to demagnetise basic igneous rocks. Granulite facies metamorphism of these rocks, however, may produce secondary magnetite and produce highly magnetic units. Eclogite facies metamorphism destroys all magnetite and partitions iron into paramagnetic silicates. Gabbros exhibit a bimodal susceptibility distribution, reflecting paramagnetic and ferrimagnetic subpopulations. In most provinces, the ferrimagnetic group is more prominent. Very fine, generally acicular, (titano) magnetites within pyroxene, olivine and plagioclase grains are largely responsible for the relatively intense and stable remanence carried by many gabbros and other basic intrusives. These grains may be protected by their silicate hosts from metamorphic breakdown, so that gabbros may be somewhat less sensitive to low and medium grade metamorphism than their extrusive and hypabyssal equivalents.

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 intrusions are also moderately to highly magnetic. Unaltered komatiitic lavas (including spinifex textured peridotite, olivine orthocumulate and adcumulate — dunite — zones) are weakly magnetic. However, komatiites are almost invariably 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 ($Q_n < 1$). 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 demagnetizes serpentinites, which become paramagnetic at granulite grade. Subsequent retrograde serpentinisation, if it occurs, can produce a magnetic rock again.

Remanence

Natural remanent magnetization, NRM, is the raw permanent magnetisation of main interest in applied geophysics. The relative magnitude of remanence is often cited as the Koenigsberger ratio (Q_n) which is the ratio of the remanent to induced intensities. Ranges of Koenigsberger ratios for many rock types are shown Figure 6.

The susceptibilities of most rocks primarily reflect their magnetite content, but remanence intensity, whilst also correlated with modal magnetite, is sensitive to other factors, particularly grain size and microstructure of magnetic minerals and geological history. The natural remanent magnetization of rocks is often multicomponent in character, i.e. it represents the vector sum of remanence components, each carried by a different subpopulation of magnetic grains, acquired at different times and therefore generally with differing directions. Variations in the relative proportions of remanence components throughout a rock unit produce scatter in the remanence direction, as well as variations 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 about 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_n values less than unity. Granitic rocks and metamorphic rocks with secondary magnetite usually contain relatively coarse-grained multidomain magnetite, accounting for the generally

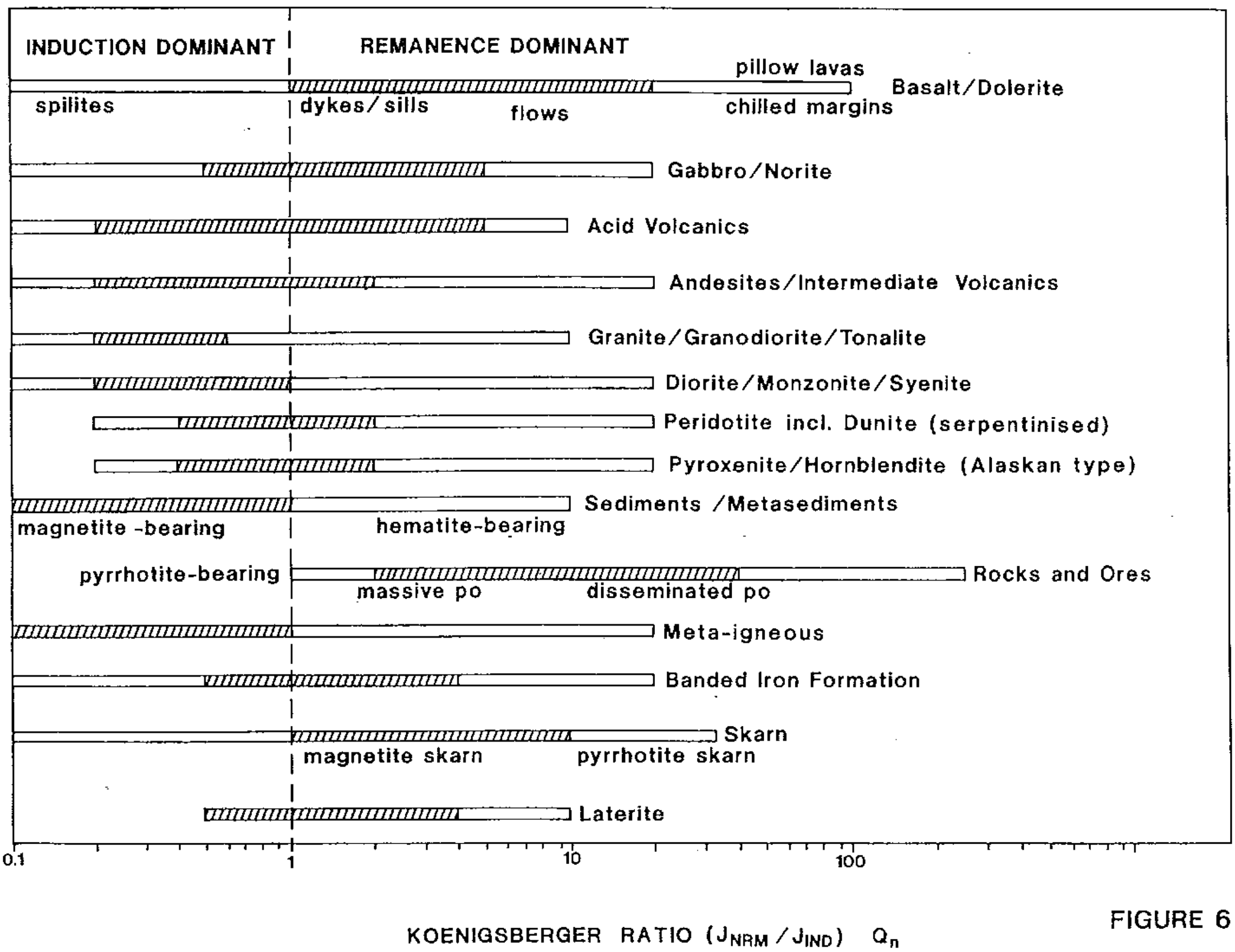

FIGURE 6

low Q_n values of these rocks. On the other had, young, rapidly chilled basaltic rocks (e.g. pillow lavas) exhibit very high Koenigsberger ratios, due to the fine grain size of the titanomagnetites. In basaltic rocks the Q_n 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_n 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, due to their coarse grain size, with the notable exception of some gabbroic intrusives for which the remanence is dominated by fine (titano) magnetite inclusions in silicate grains. Remanence carried by hematite and monoclinic pyrrhotite is characterised by high Q_n values, but hematite is only weakly magnetic and therefore hematite-rich rocks are rarely responsible for substantial anomalies, unless magnetite is also present. On the other hand, pyrrhotite-bearing rocks often carry a relatively intense remanence, which may be ancient and quite oblique to the present field. The remanence carried by magnetically soft multidomain magnetite, which is the dominant magnetic phase in many rocks, is dominated by viscous magnetization. This remanence is subparallel to the present field and therefore augments the induced magnetization, enhancing the effective susceptibility; this is common in serpentinites. Thus most anomalies can be interpreted in terms of magnetization by induction, even when typical Koenigsberger ratios are comparable to unity. However the anomaly amplitudes may be larger, for a given source geometry, than measured susceptibilities indicate, due to the viscous remanent magnetization. Neglect of remanence may therefore mislead quantitative interpretation, even though the anomaly form is consistent with magnetization parallel to the present field.

Koenigsberger ratios for viscous remanence carried by multidomain magnetite have an upper limit of approximately unity, but are typically much lower, averaging about 0.2. Rocks containing predominantly somewhat harder multidomain magnetite grains may carry a stable ancient remanence, characterised by a larger Q_n value. The Koenigsberger ratio of thermoremanence carried by an unmetamorphosed igneous rock that contains predominantly such multidomain grains ranges typically from 0.5 to 5, with higher Q_n values favoured by smaller grain sizes. 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.

Estimation of the bulk remanent magnetization of a rock unit is not straightforward. The scatter of directions must be taken into account, as well as the distribution of intensities. Remanence makes a greater contribution to the anomaly associated with a unit that exhibits a consistent remanence direction and moderate Koenigsberger ratios than a unit that has highly scattered remanence directions on a mesoscopic scale, even though the samples may all have high Q_n 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_n 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 magnetisations identified in this way should be then analysed statistically as *vectors*, rather than analysing directions and scalar intensities separately.

Figure 7 illustrates the distinction between induced magnetization, viscous remanence and stable remanence. Figure 7a shows the changes in magnetization of an initially demagnetized sample in response to an applied field, which is switched on and off as shown in Figure 7b. Small applied fields, comparable in strength to the geomagnetic field, produce small, reversible changes in magnetization, i.e. the magnetization vanishes on removal of the field. This induced magnetization is approximately proportional to the strength of the applied field. Thus the susceptibility, which is defined as the induced magnetization divided by the applied field, is approximately independent of the field. If a larger field is applied and then removed an irreversible change in magnetization occurs — and isothermal remanent magnetization J_{IRM} has been imparted to the sample. This magnetization can be regarded as permanent on the experimental time scale. Figure 7c illustrates this behaviour

ISOTHERMAL AND VISCOUS REMANENCE

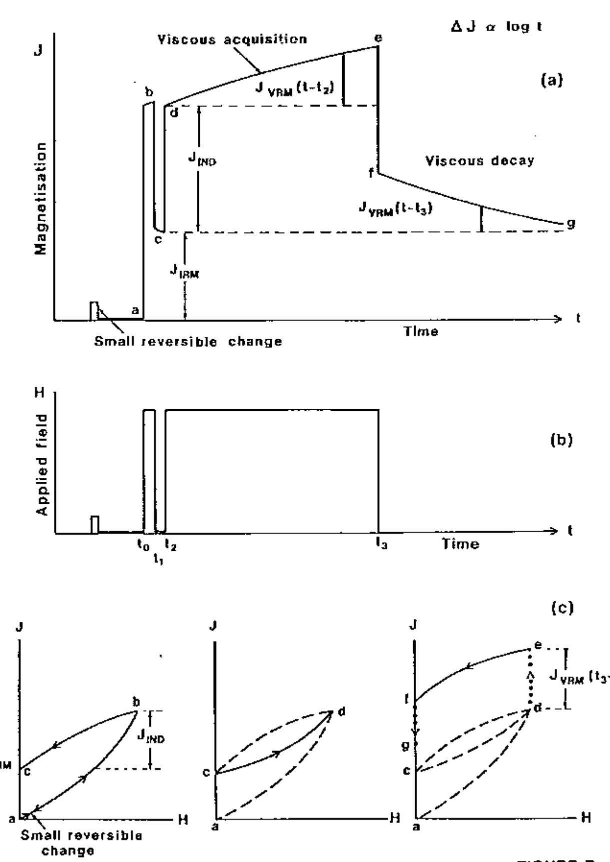

FIGURE 7

in terms of the hysteresis loop, plotting magnetization J versus applied field H. The initial portion of the J-H curve is approximately linear and is traversed almost 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 the point b, the total magnetization is the sum of the induced magnetization, which by definition is the component that vanishes when the field is removed, and the isothermal remanence, which is an essentially permanent magnetization on the experimental time scale.

If the field is again switched on, the magnetization follows the path cd and returns initially to its former value (J_{IRM} + J_{IND}). If the applied field remains on, however, the total magnetization increases gradually with time, from d to e. After an initial period that depends on the grain size distribution, the increase in magnetization is usually found to be approximately proportional to the logarithm of time, over several decades of t. When the field is removed this excess magnetization, which is known as viscous remanent magnetization (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 magnetization that is intermediate in character between induced magnetization and more stable forms of remanence. The separation of magnetization into induced, viscous and stable components is not exact, because the distinctions depend on the time scale under consideration. In general the sample may not be initially in a demagnetized state. If the sample carries a stable remanence component initially the magnetization plots in Figures 7a, b are simply shifted upwards.

Concluding Remarks

The complexities of magnetic petrophysics and the pitfalls of geologically naive magnetic interpretations have long been recognised. Magnetic petrology aims to elucidate the nexus between geology and magnetic patterns by integrating petrological and rock magnetic techniques; this field of study is now making moderate progress and has great promise for the future. The existence of large data bases of magnetic properties, coupled with a deeper understanding of the processes that create, destroy and alter magnetic minerals, should allow increasingly realistic *geologic* interpretations of magnetic surveys in the future.

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