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Magnetic susceptibility of the rock matrix related to magnetic fabric studies

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Abstract—Non-ferromagnetic minerals constitute what is called the rock matrix, whose susceptibility (K_l) is directly accessible using only high magnetic fields. Measurements on minerals and a wide range of rock types show that K_l is mainly due to paramagnetism and hardly exceeds 10^{-3} SI, with an anisotropy degree (P) less than 1.35. Different methods to estimate the role of the matrix component in low-field susceptibility (K) and its anisotropy include petrological and chemical analysis, evolution of P vs K, low temperature studies and comparison of K with remanent magnetizations.

INTRODUCTION

MAGNETIC susceptibility is an intrinsic physical property of rock-forming minerals. A distinction is generally made between the ferromagnetic (s.l.) minerals, which can carry a remanent magnetization, and the non-ferromagnetic minerals, grouped under the term 'matrix', due to the fact that they generally correspond in volume to almost the entire rock. Since the work of Daly (1967) and Parry (1971) the importance of matrix susceptibility (K_l) has been evidenced by many authors, for instance Coward & Whalley (1979), Wagner *et al.* (1981), Henry & Daly (1983), Rochette & Vialon (1984), Schultz-Krutisch & Heller (1985) or Borradaile *et al.* (1985/1986).

However, this factor is sometimes underestimated by workers in this field who implicitly assume that in a low magnetic field, K_l is negligible compared to the ferromagnetic susceptibility K_f. The general two-fold origin of low-field susceptibility K, with $K = K_f + K_l$, is the only parameter usually measured, thus frequently is ignored, perhaps because of the greater number of paleomagnetic and related rock magnetic studies than studies of rock susceptibility, or also because of the lack of appropriate instruments. Experimental techniques favour the investigation of the ferromagnetic component which is easily studied by the remanent properties.

The only direct and general way of studying the matrix component is to use sufficiently high magnetic fields. In fact K_l is independent of the field while ferromagnetic minerals reach saturation, thus giving a zero differential susceptibility. In the following, K_l will be referred to as the high field linear susceptibility (at room temperature if not indicated otherwise).

A good instrument for measuring K_l has to combine a high field obtained with a superconducting or Bitter magnet, a high sensitivity (better than 10^{-6} SI) and a large enough sample volume necessary to study common rocks. Saturating fields of a few Teslas may be necessary, for example, in hematite-bearing rocks.

This paper, aimed in particular at giving an overview of K_l in rocks, will be based mainly on measurements obtained with a high-field cryogenic magnetometer (Rochette *et al.* 1983); this instrument measures the induced magnetization of 1 cc samples with a sensitivity of 2×10^{-4} A/m under a field range of 0–4 T, and a temperature range of 2–400 Kelvin (K).

THE DIFFERENT TYPES OF MATRIX SUSCEPTIBILITY

Any material has a negative, temperature independent susceptibility or diamagnetic susceptibility. Some compounds are exclusively diamagnetic like quartz, feldspars, calcite, water or organic compounds (Foex 1957, Carmichael 1982). Diamagnetic susceptibility does not vary to a great extent, having a mean value of about -14×10^{-6} SI. Calcite is the only well-known diamagnetic mineral with a significant degree of anisotropy ($P = K_{max}/K_{min}$) of 1.13 (Owens & Bamford 1976). Rather questionable data from Finke (1910) indicate a value of $P = 1.34$ in orthoclase, while quartz is almost isotropic (Hrouda 1986).

Magnetic ions of transition and rare-earth elements, when diluted in minerals, give rise to paramagnetic behaviour. In nature, these elements are practically all in negligible quantities compared to iron, except in specific rocks with a high content of manganese, chromium, nickel or uranium, for example. The paramagnetic susceptibility follows the Curie–Weiss law $K_l = Cd/(T - \theta)$, where C is the Curie constant, proportional to the ion amount, d the relative density, θ a constant specific to the mineral and T the absolute temperature. C values per g of metal are 744 and 984×10^{-3} K, respectively, for Fe^{2+} and Fe^{3+} ions. Such behaviour has been evidenced by low temperature studies of K_l in many rock types. The examples in Fig. 1 obtained in granites show that θ is near zero when an iron-poor mineral like muscovite is present, while it reaches 10–30 K for an iron-rich mineral like biotite.

Iron-bearing phyllosilicates have anisotropy (P) at room temperature from 1.1 to 1.37 (biotite) with minimum susceptibility perpendicular to the sheet plane

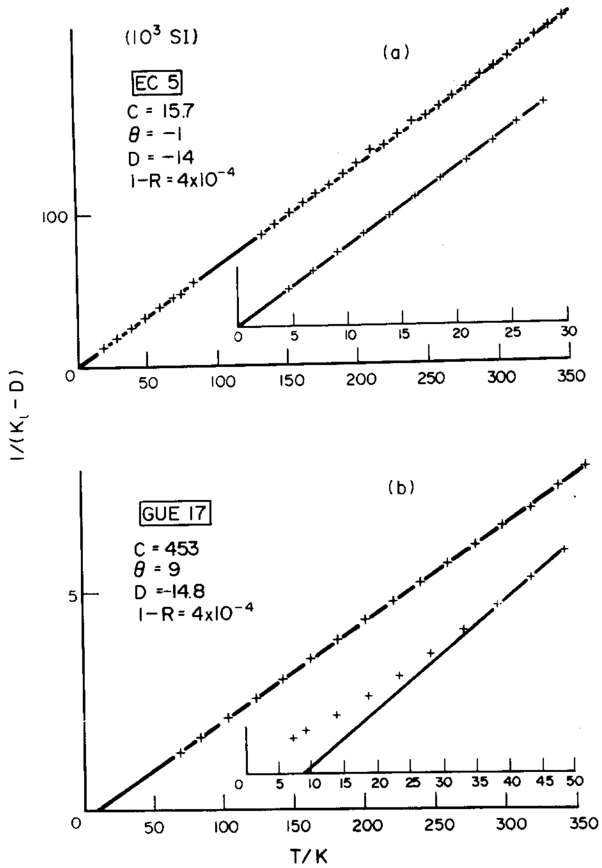

Fig. 1. Inverse of the matrix susceptibility K_l corrected for the diamagnetic term D (in 10^{-6} SI), for two samples of Hercynian granites from the French Massif Central: (a) muscovite-bearing (EC 5), (b) biotite-bearing (GUE 17). D is determined by obtaining the best regression coefficient R; the Curie constant C is indicated in 10^{-4} K and the paramagnetic Curie temperature θ in K (θ is given by the intersection with the temperature axis). Detailed low temperature curves are shown in the inset. In (a) the lack of points corresponds to the zone where K_l changes its sign, thus giving a poorly defined signal.

(Foex 1957, Ballet *et al.* 1983). Paramagnetic anisotropy in orthopyroxenes is well established (Wiedenmann *et al.* 1986), P being equal to 1.21 for $Fe_{0.87}\,Mg_{0.13}\,SiO_3$. Anisotropies between 1.08 and 1.4 have been reported for clinopyroxenes and amphiboles (Hrouda 1982) as well as in ilmenite (Foex 1957); however, the quantitative separation of actual paramagnetic anisotropy from anisotropy due to possible minute oriented inclusions of iron oxides is poorly documented. Finally, Fe^{2+} ions in carbonates also produce a paramagnetic anisotropy, the maximum susceptibility being along the c-axis; this is the case in siderite (Jacobs 1963) with P up to 4.2, as well as in calcite with paramagnetic impurities (work in preparation) with P around 1.1.

A third type of matrix behaviour, antiferromagnetism exemplified by goethite and hematite, leads to a linear susceptibility (K_{af}) smaller than the paramagnetic susceptibility. According to Hedley (1971) goethite has $P = 1.28$. At this point it is worth mentioning a difficulty arising from the present definition of matrix susceptibility: hematite contributes to both the matrix component, with an isotropic antiferromagnetic susceptibility, and the ferromagnetic component which is alone responsible for the anisotropy (Néel & Pauthenet 1952).

Apart from diamagnetism, paramagnetism and antiferromagnetism, another curious type of behaviour is illustrated by pyrite with a weakly positive isotropic susceptibility, independent of temperature (Serres 1953). Its susceptibility, $K_l = +10\text{–}30 \times 10^{-6}$ SI, is generally negligible in rocks, even compared to the diamagnetic component.

THE BULK MATRIX SUSCEPTIBILITY IN ROCKS

When neglecting θ compared to room temperature (T = 295 K) in the expression of paramagnetic susceptibility, a general formula for K_l is obtained:

$K_l \simeq -14 + d(25.2t + 33.4t') + K_{af}\,10^{-6}$ SI, where d is the relative density of the rock, t and t' the weight % of metal in the rock for Fe^{2+} and Fe^{3+}, respectively.

Figure 2 shows the range of K_l values for given values of iron content, assuming a fixed relative density of 2.65. It appears clearly that the main part of K_l is usually paramagnetic and that K_l can rarely exceed 10^{-3} in common rocks.

Typical values of K_l obtained in different rock types with the cryogenic magnetometer (Table 1) confirm this view, together with the wide variability of K_l. The comparative study of K_l and K_f also shows that the matrix is often dominant in the total low-field susceptibility of rocks whose K value is less than 10^{-3} like many granites or metamorphic and sedimentary rocks. Thus, when interpreting K data such as that obtained from susceptibility well logging or magnetic anomaly surveys, the role of the matrix has to be considered in these rocks.

K_l will vary according to lithology, being for example proportional to the phyllosilicate amount—at a given composition of these minerals—in granites and shales; this last case is demonstrated by Fig. 3 obtained in Jurassic and Eocene Alpine black shales. Higher values

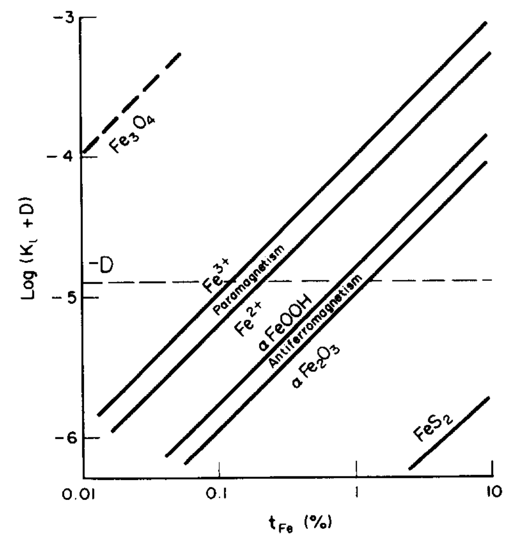

Fig. 2. Values of K_l in rocks as a function of iron content in metal weight % for different mineralogy with a fixed relative density of 2.65; a mean value for the susceptibility of magnetite (Fe_3O_4) is indicated by the dashed line.

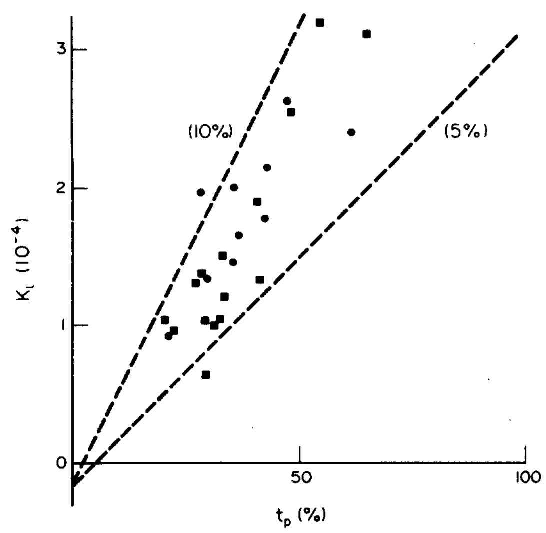

Fig. 3. Correlation between K_l and amount of phyllosilicates t_p for Jurassic (squares) and Eocene (circles) Alpine shales after Rochette & Lamarche (1986) and Lamarche & Rochette (1987); calculated curves correspond to 5 or 10% of Fe^{2+} ions in the phyllosilicates.

will be observed in mineralized levels, usually enriched with paramagnetic ions, present for example in siderite and iron-rich chlorite (Table 1). On the contrary oxidizing alteration (including weathering) will decrease K_l because paramagnetic iron becomes antiferromagnetic (Rochette & Vialon 1984).

Two alternative techniques to separate the matrix susceptibility, using only low-field measurements, have been proposed.

(i) The analysis of the thermal variation of K, from liquid nitrogen to room temperature, has been successfully used by Schultz-Krutisch & Heller (1985) to estimate the paramagnetic part of K_l, assuming a temperature independent ferromagnetic susceptibility K_f. However, this is a rough approximation in the general case because K_f varies with temperature according to magnetic transitions, variation of the anisotropy constant, and blocking of domain walls or of superparamagnetic grains.

(ii) Passing a large volume of crushed rock through a vibrating magnetic separator with an adequate field gradient can result in the physical separation of highly magnetic and weakly magnetic fractions, whose low-field susceptibilities correspond to K_f and K_l (Henry & Daly 1983, Borradaile *et al.* 1985/1986). This method assumes complete mechanical separation between the matrix and the ferromagnetic grains, a condition that is hard to achieve when the latter are very fine or present as minute inclusions within silicate grains, for example. A high enough susceptibility contrast is also needed, which is not the case in rocks bearing hematite and biotite or ilmenite, for example. Contamination between the two fractions has thus to be carefully tested using, for example, microscopic investigations or saturation remanence measurements. Another drawback of this method arises from the possible change of K_f intensity during crushing, due to a change in shape and size of grains, and also to magnetic interactions, enhanced in the magnetically enriched fraction.

It thus appears that these techniques are not always efficient and, in general, less precise than high-field measurements.

Table 1. Matrix susceptibility K_l measured in high-field with the cryogenic magnetometer for different rock types; m is the percentage of matrix in the low-field susceptibility. This is a synthesis of measurements, largely unpublished, on 160 samples

Rock type	Origin and references	K_l (10^{-6} SI)	m (%)
Platform and continental margin sediments:	French, Swiss and Austrian Alps, Bassin de Paris, South Massif Central, Greece		
—shales and mudstones	Rochette & Lamarche (1986)	25–310	23–86
—schists	Lamarche & Rochette (1987)	35–440	77–100
—pyrrhotite-bearing schists	Rochette & Fillion (1987)	47–492	2–97
—limestones	Rochette (1987)	−12–90	1–82
—sandstones (including red ones)	+ unpublished data	10–450	80–92
—mineralized beds		691–993	84–96
—gypsum (Triassic evaporite)		−11	
Mn, Fe-rich oceanic sediments	Oman, Thomas *et al.* (1987)	585–810	97
Hercynian granites:	French Massif Central, Algeria Jover & Bouchez (1986), Bernier *et al.* (1987)		
—biotite-bearing	Guéret, Millevaches	63–530	95–100
—muscovite-bearing	Millevaches, Echassières	−9–33	58–100
—pyrrhotite bearing	Guéret, Beni Toufous	174–414	6–70
Caledonian magnetite-bearing granite	Crifel (Scotland), Bouchez (pers. comm. 1987)	90–493	1–2
—olivine basalt	Cameroun	651	6
—gabbros (ophiolitic, continental)	Oman, Cameroun	245, 700	29, 89
—peridotite (fresh iherzolite)	Lanzo (Italian Alps)	528	67
—amphibolite	Belledonne (French Alps)	665	100
—metavolcanite	Borrowdale volcanics, English Lake District	314	59
Ophiolitic diabase	Oman	336	90
pure monocrystals:			
—quartz	synthetic	−14.7	
—calcite	Alpine vein	−12.5	
—pyrite	Alpine schist	+10	

THE ANISOTROPY OF THE MATRIX RELATED TO MAGNETIC FABRIC STUDIES

The direct study of matrix magnetic anisotropy can be performed in high fields using a torquemeter (Hrouda & Jelinek 1987) or the cryogenic magnetometer in a rotating sample configuration (Rochette & Fillion 1987). In the following, only indirect access to the matrix anisotropy from low-field data will be discussed.

From the preceding data on minerals it is clear that in natural rocks the actual anisotropy degree of the matrix component cannot exceed 1.35, except in *c*-axis oriented siderite-bearing rocks. However, higher apparent values can be observed in low-susceptibility rocks where an isotropic diamagnetic susceptibility, such as that of quartz, compensates for a positive anisotropic component; this situation, examined theoretically by Hrouda (1986), has since been evidenced in the muscovite-bearing hololeucocratic granite of Echassières in the French Massif Central by Bernier *et al.* (1987) (see also Table 1). Figure 4(a) from this work shows that compensation by the diamagnetism of quartz and feldspar artificially increases the real anisotropy due to muscovite near zero susceptibility, thus leading to a close dependence of P upon K_m; in fact one sample has a P value equal to 2.33 while it is not defined in another due to a different sign between K_{max} and K_{min}. After recalculating an anisotropy degree (P') based on K values corrected for the diamagnetic term determined as in Fig. 1, a more petrologically independent value is obtained, near 1.025 (Fig. 4b).

To avoid biasing P due to a variation in the positive component, it would thus be useful in many cases—except where a calcite fabric is expected—to use susceptibility values corrected for the diamagnetic term when it is not negligible, i.e. when $K < 10^{-3}$. Such a correlation between mean susceptibility and the anisotropy degree may prove useful in other situations where the low-field susceptibility is shared in variable proportion by an anisotropic matrix mineral and a ferromagnetic mineral with a stronger anisotropy. Ferromagnetic grains in fact usually have a high intrinsic anisotropy. P values less than 1.3 are only observed in subisometric magnetite grains; consequently very special conditions, such as late isotropic crystallization of the ferromagnetic phase alone are needed to observe a matrix anisotropy stronger than the ferromagnetic anisotropy in rocks. For example, the anisotropy of Swiss Jurassic schists is due partly to phyllosilicates and partly to pyrrhotite which appears progressively according to metamorphism (Rochette 1987). This variable amount explains the positive correlation observed between P and K_m in the pyrrhotite-bearing rocks (Fig. 5): for low and high values of K_m the anisotropy is mainly due to phyllosilicates and pyrrhotite, respectively.

Theoretical curves, assuming a constant paramagnetic component with $P_l = 1.3$ and a variable pyrrhotite component with $P_f = 1.5$–2, fit the data well considering that part of the P fluctuation is due to variable deformation and that K_l also varies but in a smaller range than K_f. Such curves are constructed with the following formula, derived for a neutral ellipsoid:

$$P = [K_m(P_l + 1)P_f - K_l(P_f - P_l)]/[K_m(P_l + 1) + K_l(P_f - P_l)]$$

(if $K_m = K_l$ then $P = P_l$ and if $K_m \gg K_l$ then $P = P_f$).

In fact simulations show that the curve is practically independent of the shape of the ellipsoid.

The same plot has been tested with the site-mean data of Rathore & Kafafy (1986) in various sedimentary and

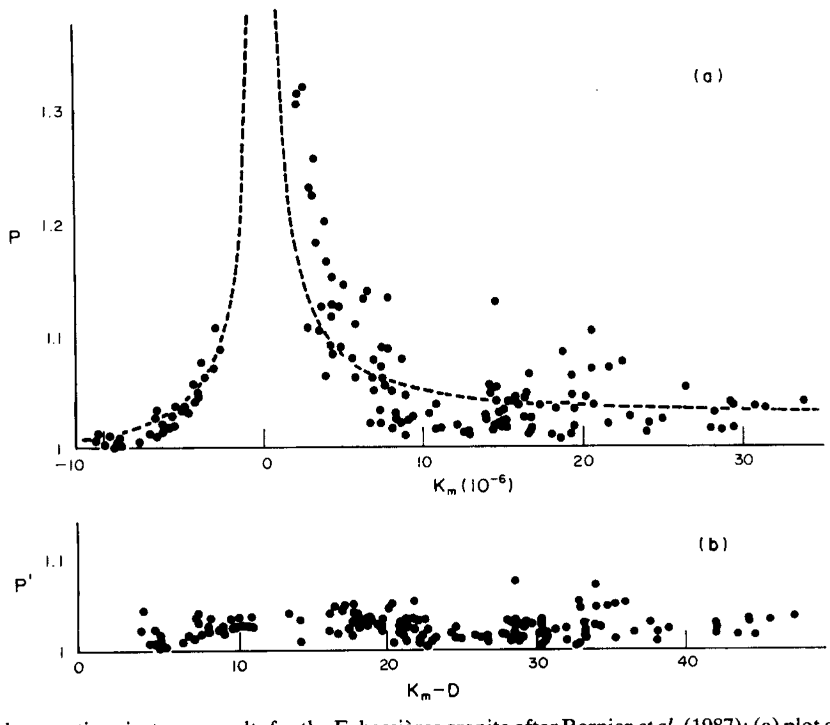

Fig. 4. Low-field magnetic anisotropy results for the Echassières granite after Bernier *et al.* (1987): (a) plot of the anisotropy degree, P, against mean susceptibility K_m, (b) corrected anisotropy degree, $P' = (K_{max} - D)/(K_{min} - D)$ against $(K_m - D)$; the theoretical curve assumes a paramagnetic anisotropy $P_l = 1.02$ and is shape independent.

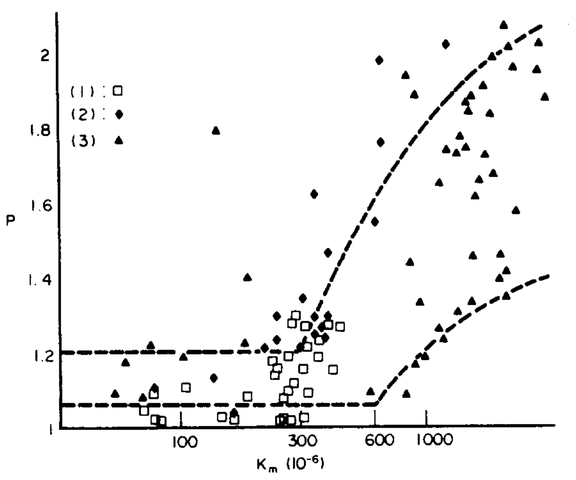

Fig. 5. Plot of the anisotropy degree P against mean low-field susceptibility K_m for all samples of the Swiss Jurassic schists after Rochette (1987): different symbols are used for sites outside the pyrrhotite zone (1), inside (3) or intermediate (2); theoretical curves are derived (see text) for $K_l = 300 \times 10^{-6}$, $P_l = 1.2$, $P_f = 2.2$ (upper curve) and $K_l = 600 \times 10^{-6}$, $P_l = 1.05$, $P_f = 1.5$ (lower curve). *Note added in proof:* M. D. Fuller produced in 1963 (*J. geophys. Res.* **68**, 293–309) highly similar data from pyrrhotite-bearing Welsh slates, and interpreted it as here.

volcanic rocks of the Lake District, where it was assumed that susceptibility was only due to magnetite. However Borradaile *et al.* (1985/1986) reached an opposite conclusion by studying rocks of the same formation, where they attributed the entire susceptibility to chlorite. High-field measurements made in Grenoble on one sample (collected by the author) effectively revealed a considerable matrix susceptibility, equal to 314×10^{-6}. The P vs K_m plot (Fig. 6) actually shows the same behaviour as in the Swiss schists, indicating that below approximately 300×10^{-6} the anisotropy is carried mainly by the matrix (probably phyllosilicates) because P is not correlated to K_m. Above this limit the contribution of magnetite to the anisotropy increases with the amount of magnetite and thus with K_m.

Such situations have been used by Henry & Daly (1983) to put forward a mathematical method to further isolate quantitatively the ferromagnetic and the matrix anisotropy, using linear regression between the terms K_{ij} of the susceptibility tensor and the mean susceptibility K_m. However, this method is only suitable when, at the sample scale, the matrix susceptibility is constant and the ferromagnetic susceptibility sufficiently variable. These conditions can be fulfilled in rocks where matrix and ferromagnetic grains originate from independent phenomena—like lithology and metamorphism in the case of the Swiss schists—but not at all when they are cogenetic. For example, in a granite with biotite (matrix term) and pyrrhotite (ferromagnetic term) work in preparation reveals that both contributions K_l and K_f to the susceptibility vary by the same factor of 1.5. Independent tests of the necessary requirements on K_l and K_f have thus to be performed before using the Henry method because the self-test of their computation, based on the quality of the linear regression between K_{ij} and K_m, is not at all decisive for this purpose. In fact perfect linear regression can be obtained if K_l and K_f are proportional and moreover if K_f is negligible.

The comparison between K and the intensity of a remanent magnetization (RM) has been used by Lowrie & Heller (1982) in stratigraphic profiles to decide if susceptibility and remanence were carried by the same mineral or not. This method appeared qualitatively successful in estimating the importance of matrix in K, although it should be borne in mind that an absence of correlation between K and RM could also mean that they are carried by different ferromagnetic minerals; this is the case in some Jurassic shales (Rochette & Lamarche 1986), where K is due to magnetite while saturation isothermal RM is mainly due to goethite or hematite. To check this, Lowrie & Heller used different types of remanence: low and high field IRM together with anhysteretic and natural RM.

Finally, from literature and the author's experience, it appears that an anisotropic matrix component more often corresponds to phyllosilicates. In this case the structural interpretation is very easy in terms of a planar markers model: the magnetic foliation corresponds to the plane of preferred orientation of the silicate sheets—i.e. flow plane, bedding or schistosity—while the magnetic lineation corresponds to the zone axis of the sheets.

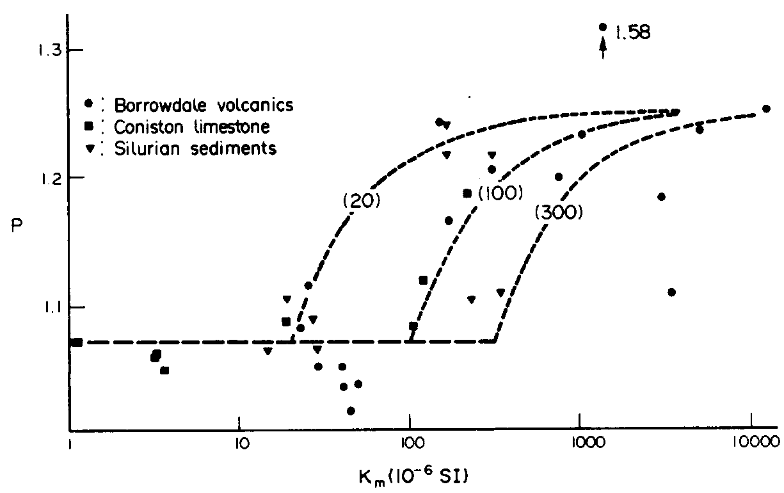

Fig. 6. Same plot as in Fig. 5 for site-mean data of the sedimentary and volcanic rocks studied by Rathore & Kafafy (1986) in the Lake District. Theoretical curves are drawn for $P_l = 1.07$, $P_f = 1.25$ and different K_l values: 20, 100 and 300×10^{-6}.

This zone axis can be assumed parallel to the flow direction in granites (Guillet *et al.* 1983, Jover & Bouchez 1986), to the current direction in sediments (Hounslow 1985) and to intersection or crenulation in schists (Rochette & Vialon 1984). When other minerals are responsible for the matrix anisotropy the interpretation has to be made in terms of preferred orientation of the corresponding crystallographic axes (see for example Owens & Rutter 1978 and Wagner *et al.* 1981).

CONCLUSIONS

The minerals of the rock matrix, especially the paramagnetic minerals, often play a major rôle in the magnetic susceptibility of rocks. The direct and precise study of the matrix part (K_l) of the low-field susceptibility (K) needs specific high-field techniques which are currently available but actually not in general use. However, some easy indirect ways of tackling this problem are proposed here.

(1) A good estimate of K_l and an upper limit of the corresponding anisotropy P_l can be obtained from the petrological analysis of the rock and the reference values listed in this paper; by comparison with the measured K and P values, the rôle of the matrix can be determined.

(2) The matrix will be negligible when $K \gg 10^{-3}$ SI or $P \gg 1.35$ except in rocks of near-zero susceptibility where a correction for diamagnetism is necessary. On the contrary a predominance of the matrix has to be suspected when $K < 300 \times 10^{-6}$ and $P < 1.35$, especially in phyllosilicate-bearing rocks.

(3) When K is variable in the formation, a correlation between K and P and between K and a remanent magnetization intensity (NRM or better ARM or IRM) may prove very useful.

(4) In favourable conditions, susceptibility measurements at low temperature as well as magnetic sorting can be used to determine K_l.

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