ON THE MAGNETIC SUSCEPTIBILITY ANISOTROPY AND ITS MEASUREMENT

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RESUMEN

En este trabajo se discuten algunos aspectos de la medición e interpretación de resultados de anisotropía de suscept'bilidad magnética. Se muestra que el tamaño y forma de la muestra a analizar, son parámetros muy importantes en la medición de anisotropía y, que en algunos casos pueden crear serias dificultades. Por ejemplo, en el caso de sedimentos rojos, se encontró que la fábrica magnética cambia de una fábri⌣a secundaria típica a una fábrica primar'a (deposicional) típica con la simple alteración del cociente longitud-diámetro (muestras cilíndricas). Cambios apreciables fueron también detectados en muestras de mayor susceptibilidad (granitos). La medición de estas muestras se realizó con un magnetómetro de rotación PARM-SM2. Comparación de mediciones usando este instrumento y otro de baja rotación (Digico) sugiere que los resultados de direcciones son comparables, no así las magnitudes que muestran d'ferencias hasta de un 10-18%, donde los resultados del PAR-SM2 parecen ser consistentemente menores. Cabe mencionar, que algunas muestras mostraron mucho mayores diferencias. Comparación de los resultados del Digico con los de un magnetómetro de torsión de bajo y alto campo magnético indica, que los resultados son distintos y que estas diferencias se deben a la dependencia de la anisotropía magnética, con el campo magnético usado en las mediciones.

ABSTRACT

Some aspects of measurement and interpretation of magnetic susceptibility anisotropy are discussed. It is shown that the finite specimen size and shape could create severe problems for measurements. For instance, it is found that for 2.5 cm diameter specimens of red bed units, the magnetic fabrics change from typical secondary fabrics to typical primary depositional fabries by simply varying the length- diameter ratio. Large changes in magnitudes and directions are also observed for higher bulk susceptibility samples (granites). These measurements were carried out using a PAR-

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SM2 spinner meter. Comparative determinations of magnetic anisotropy obtained using a PAR and a Digico spinner meter yield good overall directional agreement, but some 10-18% differences in magnitude with lower values corresponding to the PAR. Individual specimens showed in some cases even greater differences in magnitudes and directions. Comparison of results obtained with a Digico and low and high field torque meter suggests that magnetic fabric is not directly comparable. The discrepancies may be explained in terms of the different magnetic fields used in booth systems and the saturation hysteresis characteristics of the samples. Field dependence of magnetic anisotropy is briefly discused.

INTRODUCTION

Although the earliest works on the anisotropy of magnetic susceptibility (AMS) of rocks were carried out some four decades ago (Ising, 1943), its potentialities on geological problems have not yet been fully developed. This is partially due to difficulties in measuring the magnetic anisotropy and in interpreting these anisotropy measurements. Since the late 1960s, a variety of instruments have been developed and become more common in palaeomagnetic laboratories; this particularly since the advent of computers and commercial meters. At present, there are several different techniques for anisotropy and many more instruments, and it is of obvious importance to know the relative instrumental capability of the meter system one uses and how it compares with those of other workers. This paper attempts to discuss certain aspects of the measurement and interpretation of AMS data.

Most of the discussion concerning with the precision and limitations of in instruments is focused on spinner meters, namely Princeton Applied Research (PAR) spinner meters and Digico spinner meters, which are some the fastest instruments available. Results are compared with those obtained by low-and high-field torque meters, which are more time consuming. Since large sample collections and high precision are usually required, it is important to establish if precision is not lost in fast spinner systems and if slow torque systems are precise enough to compensate for the time consumed. Also, it is important to know how results are affected by factors such as shape and size of specimens, specimen inhomogenities, strong remanent magnetizations, orientation of specimens during collection and measurement, magnitude of bulk susceptibility, type of magnetic anisotropy (magnetic mineralogy), and magnitude of magnetic anisotropy.

Applications of AMS studies in magnetic fabric (petrofabric) and in other geological and geophysical problems are reviewed in a companion paper (in preparation). There, the analysis of magnitude and directional AMS data is of critical importance. To the variety of techniques, which partially reflects the variety and difficulties in instrumentation and experiment design, one should add the lack of adequate statistical methods for processing data, particularly those for the directions.

The majority of workers concerned with magnetic susceptibility anisotropy are using electromagnetic units and give the susceptibility as e.m.u. cm^{-3} Oe^{-1}, Gauss/Oe or CGS units, whereas some others who follow the resolution of the International Association of Geomagnetism and Aeronomy (IAGA) are using the SI system and give the susceptibility as SI units or just dimensionless. To help the reader follow the references and discussion a table of conversion factors is given in the appendix.

2. ISOTROPIC AND ANISOTROPIC MAGNETIC SUSCEPTIBILITY

The magnetization induced Ji in a magnetically isotropic medium by a relatively weak magnetic field $\bar{H}$ may be described by the relationship

$$\bar{J}_i = \frac{k\,\bar{H}}{1 + k\,N_A} \tag{1}$$

where k is the magnetic susceptibility of the medium and N_A is the demagnetizing factor of the medium. This expression is usually expressed in terms of the effective magnetic fiel H_{eff} which is defined by

$$\bar{H}_{eff} = \bar{H} - N_A\,\bar{J}_i \tag{2}$$

hence

$$\bar{J}_i = k\,\bar{H}_{eff} \tag{3}$$

In most practical cases $\bar{H} \simeq \bar{H}_{eff}$, so that an approximate relationship is usually quoted as

$$\bar{J}_i = k\,\bar{H} \tag{4}$$

The magnetic field $\bar{H}$ is related to the density of magnetic flux B (or 'true' magnetic field) by the relationship

$$\bar{B} = \mu\,\bar{H} \tag{5}$$

where μ is the magnetic permeability which is defined as

$$\mu = \mu_0 \ (1 + k) \tag{6}$$

with μ_0 as the magnetic permeability of free space.

When the medium is anisotropic, k and μ represent symmetrical second rank tensors, called the susceptibility and permeability tensors, respectively.

The susceptibility tensor may be represented by

$$K_{ij} = \begin{bmatrix} K_{11} & K_{12} & K_{13} \\ K_{21} & K_{22} & K_{23} \\ K_{31} & K_{23} & K_{33} \end{bmatrix} \tag{7}$$

This tensor represents a quadratic or second degree surface (ellipsoid or hyperboloid) of the form

$$K_{11} X^2 + K_{22} Y^2 + K_{33} Z^2 + (K_{12} + K_{21}) Xy + C K_{13} + K_{31})$$

$$XZ + (K_{23} + K_{32}) YZ = 1 \tag{8}$$

which reduces to

$$K_1 X^2 + K_2 Y^2 + K_3 Z^2 = 1$$

$$K_1 X^2 + K_2 Y^2 + K_3 Z^2 = 1 \tag{9}$$

For an ellipsoid with semiaxes equal to the reciprocal of the square roots of the susceptibility values k. From that ellipsoid, the susceptibility in any direction can be obtained as the reciprocal of the square root of the radius vector in that direction. When a magnetic field $\bar{H}$ is applied along a given direction of a rock, the apparent susceptibility is given by

$$K_H = K \left(\frac{\lambda^2}{1 + N_1 k} + \frac{\mu^2}{1 + N_2 k} + \frac{\nu^2}{1 + N_3 k} \right) \tag{10}$$

Where $N_1 + N_2 + N_3 = 4\pi$ and λ; μ, ν represent the direction cosines of $\bar{H}$. Two cases are of interest, for spherical bodies and for planar bodies, where $N_1 = N_2 = N_3 = 4/3\ \pi$ and $N_1 = N_2 = 0$, $N_3 = 4\ \pi$, respectively. If the magnetic grains of the rock are oriented at random and constitute a volume V, then the bulk susceptibility is given by (Nagata, 1967):

$$K_a = \frac{Vk}{3}\left(\frac{1}{1+N_1k} + \frac{1}{1+N_2k} + \frac{1}{1+N_3k}\right) \qquad (11)$$

That is, the more numerous the magnetic grains, the higher the bulk susceptibility.

3. TYPES OF MAGNETIC ANISOTROPY

Six main types of magnetic anisotropy in rocks have been distinguished so far (Bhathal, 1971), of which four result from some kind of alignment. These are: (1) shape anisotropy (alignment of grains); (2) magnetocrystalline anisotropy (alignment of crystallographic axes); (3) domain anisotropy (alignment of magnetic domains in multidomains particles); (4) textural anisotropy (stringing together of magnetic grains in lines or planes); (5) exchange anisotropy; and (6) stress-induced anisotropy.

3.1 *Shape anisotropy*

Shape anisotropy results from inequalities of the demagnetizing factors along different directions (see expressions 1 and 10). In rocks the magnetic grains are usually of irregular shapes, which presents problems in mathematically defining the demagnetizing factors and relating them to a coordinate system. In practice this is solved by approximating these irregularly shaped grains with some simple geometries such as ellipsoids and, in most situations, this appears to describe the observed anisotropies fairly accurately (Stacey, 1960b; Uyeda *et al.*, 1963). Stoner (1945) has shown that if a given body (grain) is uniformly magnetized, then its demagnetizing field is constant throughout it. He described methods of calculating the demagnetizing factors for ellipsoids and in particular gave tables for prolate and oblate spheroids. Measurements on any cross section of the spheroids will give demagnetizing factors along the axes of an ellipse ($a > b$), i.e. N_a and N_b.

From expressions (1) to (3) we have that

$$\overline{J_f} = k_o \dot{H} \qquad (12)$$

where k_o stands for the observed or apparent susceptibility and may be expressed by

$$K_o = \frac{K_f}{1 + N K_f} \qquad (13)$$

where k_f represents the intrinsic (bulk) susceptibility.

For ellipsoids of revolution the maximum difference in susceptibility is

$$\Delta K_{max} = \frac{K_i^2 (N_a - N_b)}{(1 + N_a K_i)\ (1 + N_b K_i)} \qquad (14)$$

Hence, the anisctropy due to shape is very small when the intrinsic susceptibility is small. It is thus very important in the members of the magnetite-ulvospinel series but not in the ilmenohaematite series (Stacey, 1963).

3.2 *Magnetocrystalline Anisotropy*

This anisotropy is associated with the crystal structure, i.e. minerals that are intrinsically anisotropic. The effect of crystalline anisotropy is to preferentially align the magnetization along certain crystallographic axes, known as easy planes or directions. Thus, the work required to align a magnetization along a given direction other than an easy one is called the magnetocrystalline energy. This anisotropy is important in minerals of low intrinsic susceptibility and low crystallographic symmetry, such as haematite, pyrhotite and members of the ilemenite-haematite series (Uyeda *et al.*, 1963). For instance, in haematite at room temperature, the susceptibility may be about 100 times higher in the basal plane than in directions normal to it (Uyeda *et al.*, 1963). An interesting property is that its anisotropy differences decrease with increasing magnetic field, thus low fields may be best suited for measurement of this anisotropy type. Porath & Chamalaun (1966) suggested that low fields are preferable for studying haematite-bearing rocks since the Fourier spectrum of high-field measurements give significantly greater higher harmonics. In general, the higher harmonics should be small so that the magnetic anisotropy may still be approximated by and equivalent ellipsoid (Stacey, 1960b). At high fields the magnetocrystalline effects of cubic minerals also start to be noticeable and so crystalline alignments of magnetite crystals then become important.

3.3 *Domain Anisotropy*

This anisotropy results from alignments of magnetic domains in multidomain particles (Stacey, 1963). For a given grain, the susceptibility is greatest parallel to the domain directions K_{11}, and minimal normal to them, K_1. Stacey (1963) showed that K_{11} is more dependent on domain configuration and coercive force characteristics than K_1, and suggested that if the domains are aligned the anisotropy would be small. Domain alignment may be produced in the laboratory by applying alternating fields, thus this treatment may modify the anisotropy alternating fields, thus this treatment may modify the anisotropy characteristics. Bhathal & Stacey (1969) presented experimental results of this effect by using basaltic and synthetic magnetite samples. They applied the AF fields to stationary samples, first along easy directions and then at 90° directions. Theoretical studies on these effects are given by Stacey (1963) and Bhathal *et al.* (1969).

3.4 *Textural Anisotropy*

This anisotropy is considered rare among the normal range of rocks investigated in palaeomagnetism (Bhatal, 1971). It was first studied by Grabovsky & Broadskaya (1958) and it results from the stringing together of magnetic grains in lines or planes. The effect is greater in the case of high susceptibility grains such as magnetite grains (Stacey, 1960_a).

3.5 *Exchange Anisotropy*

This anisotropy is due to magnetic interactions between different types of magnetic minerals, such as between antiferromagnetic and ferromagnetic minerals or between ferrimagnetic and ferromagnetic minerals (Bhathal, 1971). Considerable attention has been devoted to exchange anisotropy in regard to the mechanisms of self-reversal. Stacey (1963) indicated that exchange interactions may give rise to self reversal phenomena and that exchange anisotropy is one of the most stable mechanisms. Self-reversals of remanent magnetization were considered capable to explain most of the reversed magnetization of rocks (e.g. Neel, 1955; Blackett, 1962), however, they are now regarded as relatively rare phenomena in nature (Stacey, 1969; Stacey & Banejee, 1974). Nevertheless, the phenomenon of self-reversal is important in certain cases (e.g. Ozima & Ozima, 1967; Creer *et al.*, 1970; Creer, 1971).

3.6 *Stress-Induced Anisotropy*

Graham (1956) suggested that reversible elastic stresses applied to rocks would cause irreversible deflections of the remanent magnetization. Earlier studies by Kalashnikov & Kapitsa (1952) (Kalashnikov, 1954; Kapitsa, 1955) on the effects of stress in the remanent magnetization indicated that tectonic stresses produce magnetic effects. They even suggested the possibility of using the effects for forecasting earthquakes (Kalashnikov, 1954), and for other potential applications (Kapitsa, 1955). Stott & Stacey (1960) showed that in isotropic rocks (dolerite, porphyry and basalt) reversible uniaxial stresses acting during remanence acquisition produce equally reversible effects on the magnetization. Later, numerous studies have shown that stresses are able to impose both reversible and permanent effects on the remanent magnetizations (Nagata & Kinoshita, 1965; Kinoshita & Nagata, 1967; Stacey, 1962, 1967; Nagata, 1966).

4. MEASUREMENT OF MAGNETIC ANISOTROPY

4.1 *Low-field magnetic anisotropy*

Most earlier investigations of the magnetic anisotropy of rocks were carried out with suspended sample (torque) magnetometers (Ising, 1943; Granar, 1958). The specimen is suspend from a torsion head and a uniform AF field is applied. According to its anisotropy characteristics the specimen tends to rotate so that the direction of its maximum susceptibility in the plane of measurement coincides with the applied field direction. The torque imposed is a measure of the anisotropy (Ising, 1943). This method is probably one of the most sensitive ways of measuring anisotropy and some more recent versions of this method are described in King & Rees (1962), Stone (1963, 1967) and Stacey (1960b).

Other methods used for measuring anisotropy are: (1) astatic magnetometers, (2) alternating current (AC) bridge methods, (3) spinner magnetometers, and (4) cryogenic magnetometers. Astatic magnetometers have long been used for measuring susceptibility (e.g., Blackett, 1952). Two variations are used one which uses Helmholtz coils to provide a uniform magnetic field in the measurement space where the sample is held stationary in a predetermined position, and the rsults are given in terms of equation (4). In the other method, the sample is rotated inter the influence of the Earth's magnetic field (no Helmholtz coils required) and the deflecting field produced at the sensor elements of the magnetometer are the combination of susceptibility and remanence effects. There the deflection (dt) is of sinusoidal nature and the amplitude (dr) of the sine curve gives the remanence effect, whereas the displacement (di) gives the induced magnetization (susceptibility) effects. The relationship may be expressed by

$$d_T = d_i + d_r \sin \theta \qquad (15)$$

where θ represent the different azimuths at which the sample is rotated. A disadvantage in this method arises when samples have higher Q-values, there the deflections (di) may be comparable to the uncertainties in measuring (dr) and there will be large errors in the computed susceptibility. As (1967) has described the measurement of anisotropy with the astatic magnetometer.

In the AC bridge methods, the susceptibility is measured directly in different directions (Graham, 1954). Among these methods we may mention those described by: Collinson *et al.* (1963) in which rock samples change the reluctance of an airgap of a pair of balanced transformer rings, thus giving a measure of the susceptibility in a particular direction (see also Stone, 1967). Fuller (1967) discussed a similar instrument and some of the potential difficulties: Graham (1967) presented details of an AC bridge based on a Maxwell inductance bridge; Daly (1967) described a translation inductometer, capable of measuring samples up to 300 gr ($\leq$ 125 cm^3) with considerable precision up to values of 2.10^{-9}, the instrument also measures the remanent magnetization and performs AF demagnetization up to 20 mT (200 Oe) peak-fields; and finally Jelinek (1973) has also described an AC bridge; capable of measuring samples up to 8.10^3 cm^3 and with bulk susceptibilities up to about 10^{-5}.

The use of spinner magnetometers for anisotropy measurements has been described by Noltimier (1967). The samples are rotated at a given frequency about an axis normal to the sensor-coils under the influence of a known magnetic field. If the sample is anisotropic a time-varying signal of frequency twice the spinning frequency is induced; this signal is filtered to remove the effects of the remanence. Then, the amplitude and phase of the filtered signal give the magnitude and directions of maximum susceptibility in a plane normal to the rotation axis. This system was used earlier by Howell *et al.* (1958).

Finally, we may mention the use of cryogenic magnetometers recently proposed by Scriba & Heller (1978). Basically, the instrument carries out the measurements as in ballistic magnetometers (these have been described by e. g. Nagata, 1967) and has been used for measurement of remanent magnetization with considerable success (see Goree & Fuller, 1976). The instrument uses three pairs of Helmholtz pickup coils which are kept in a superconducting state and an applied low field (e.g. 0.01 mT was used by Scriba & Heller, 1978), this gives a signal in each coil pair composed by the induced and remanent magnetizations along three ortcgonal axes.

4.2 *High-field magnetic anisotropy*

Stacey (1960b) introduced the use of the high-field torque magnetometer for measuring magnetic anisotropy. The method uses the energy diferences resulting from the anisotropy of samples, and at high (saturating) fields the samples tend to align the axis of easiest magnetization with the field direction. Acccunts of this method using different instrumental techniques have been given by Banerjee & Stacey (1967) and by Cox & Doell (1967). Another instrument used for anisotropy measurements was introduced by Boetzkes & Gough (1975), it is based on a spinner magnetometer and the aplied field to the samples is of about 91.6 mT (916 Oe).

4.3 *Par Spinner Magnetometer*

In the PAR spinner magnetometer, an anisotropic sample at ½ f Hz in a DC field applied normal to the spinning axis results in a sinusoidal signal from the pick up coils (or flux gate sensor) of f Hz. This is similar to the procedure used for remanence measurement, but here the corresponding signal is of ½ f Hz, which is fittered out. By spinning the sample about a vertical 'z' axis, the signal depends on K_{11} - K_{22} and K_{12} (see equation 7); then, by reorienting the sample in a convenient way, the differences of terms of the principal diagonal and values of the other terms can be obtained. The maximum fields available are 0.19 mT and 0.083 mT (1.9 Oe and 9.83 Oe) for horizontal and vertical fields, respectively. The procedure is described by Noltimier (1971). To have the principal susceptibility values, an independent measurement of axial susceptibility is required. This requirement is common to spinner and torsion balance meters and should be considered in discussing the precision of AMS measurements.

4.4 *Digico Spinner Magnetometer*

In the Digico meter, the flux gate sensor is replaced by two orthogonal pairs of inductance coils with their axes horizontal and within which the sample is spinned. The field coils are driven up to 10 K. H which induces a r.m.s. field of up to 0.7 mT (70e) along the coils axes; a sample spinning induces a signal of twice the spin frequency in the other coil pair. In the absence of a sample the pick up coil system is unaffected by the field. For each of three spins about orthogonal axes in the sample, signal information in the form of 120 readings per cycle of the induced e.m.f in the pick up system is stored and fourier analysed by a small computer. The number of cycles (N) can be fixed according to the signal-to-noise ratio in a form similar to that for remanence measurements (Molyneux, 1971). The noise is inversely proportional to $\sqrt{N}$, and it is estimated by taking readings without any sample spinning. Singh (1975) has quotted a noise level of less than $0.01 \cdot 10^{-6}$ G/Oe ⟶ at $N=2^8$; however it should be mentioned that this noise level can vary, so that noise level checks should be constantly taken during routine measurements. In practice, the minimum detectable susceptibility difference can be less than those given in the literature (eg. 10^{-8} to 10^{-9} G/Oe at $N = 10^7$; Collinson, 1976). Trying to increase the signal-noise ratio by increasing N can give unsuccessful results, and an alternative way for measuring weakly anisotropic samples may be by taking several readings.

5. EFFECTS OF SPECIMEN SHAPE AND SIZE

A critical factor in AMS measurement methods is to ensure that the size and shape of samples do not contribute an apparent anisotropy. If the size is too large relative to the space in which the inducing field is uniform, then susceptibility inhomogeneities may give rise to apparent anisotropy. This can be easily tested with coil detectors (Fuller, 1967) or ferrite cores (Christie & Symons 1969; Stone, 1967), and meters for large samples ($\leqslant$ 125 cm³; Daly, 1967 and $\leqslant$ 8000 cm³; Jelinek, 1973) have been developed. With respect to the sample shape, the problem has been investigated theoretically (e.g. Noltimier, 1971) and experimentally (e.g. Scriba & Heller, 1978) and a variety of ideal shapes has been suggested. The relative importance of these effects on AMS has not been fully realized, and most workers consider that for the majority of rocks, samples in the shape of equidimensional cylinders do not contribute a significant apparent AMS (e.g. Collinson, 1976). Experiments carried out using several distinct rock types indicate that significant effects may be present, and that what is an 'equidimensional' cylinder should be further investigated. The effects can be so large that different AMS fabrics are obtained by relatively small variations in sample shape (e.g. some 2.4% in the lenght diameter ratios for cylinders).

From equation (11) and discussion in 2 we can observe that an ideal shape for measurements is given by a sphere as this has equal demagnetizing factors in all directions. Furthermore, it permits to make idealizations concerning the distance to the sensor elements, as the sample simulated by a dipole located at the center of the sphere. However, in practice, spherical specimens would create problems with handling and orientation so that cylindrical or cubic specimens are commonly used. In the case of the Digico and PAR-SM2, the specimens are usually cylinders of about 2.54 cm-diameter, and there are significant effects due to the shape and the finite size.

For cylindrical specimens, the shape-induced anisotropy is given by (Rees, 1965):

$$\Delta K = \frac{K^2 (N_a - N_b)}{(1 + N_a K)\ (1 + N_b K)} \qquad (16)$$

where N_a and N_b are the demagnetizing factors along the long axis and in a cross-section plane, respectively. From here ΔK es zero if $N_a = N_b$, in the extreme case when $N_a = 2\pi$ and $N_b = 0$, i.e. a needle-shaped specimen (Nagata, 1961) we have

$$\Delta K_{max} = \frac{K^2\, 2\pi}{1 + 2\pi K} \qquad (17)$$

Thus, the effect depends on the magnitude of the bulk susceptibility (k). Scriba & Heller (1978) have plotted the demagnetizing factors for cylinders (Sharma, 1966) versus the ratio length diameter (1/d) of the cylinder and found that $N_a = N_b$ when 1/d is about 0.9185 (Figure 1a). These authors have also given a relationship between the ratio 1/d and the sensor-specimen distance (this as a/d from which small distances 'a' are preferable (Figure 1b).

Kent & Lowrie (1975) have provided an analysis for cylindrical specimens in terms of

$$\frac{\text{anomalous anisotropy}}{\text{intrinsic (true) anisotropy}} \left(= 15 \right) \frac{s}{a}^{2} \frac{J_1}{K_a - K_b} = C \frac{(1-1_o)K}{(K_a - K_b)\ 1_o} \qquad (18)$$

where these parameters are explained in Figure 2. There, k_a and k_b refer to the maximum and minimum susceptibilities in the measurement plane. The relative importance of the 'anomalous' anisotropy depends on the instrumen-

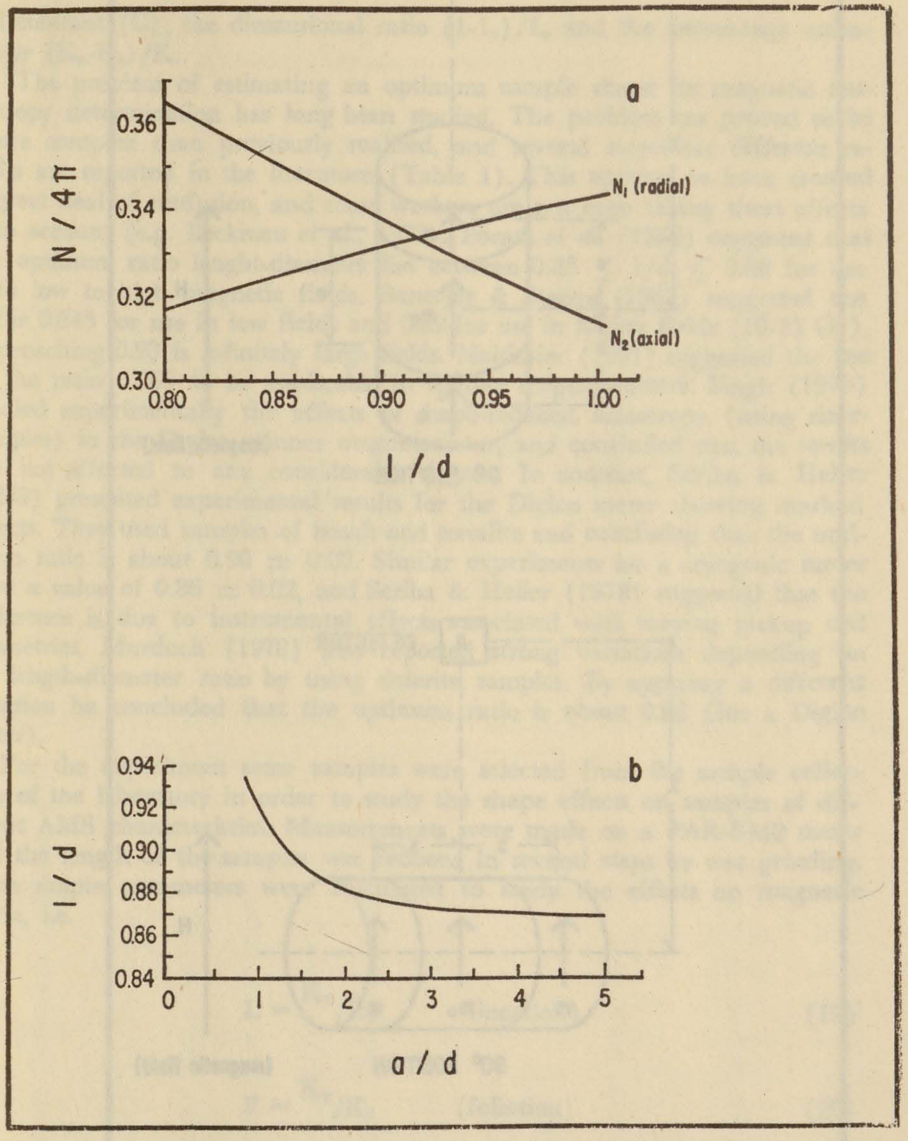

Fig. 1. (a) Variation of axial (N_z) and radial (N_r) demagnetizing factors with different length-diameter (l/d) ratio with changing sensor-distance to diameter ratio (a/d). After Scriba and Heller (1978).

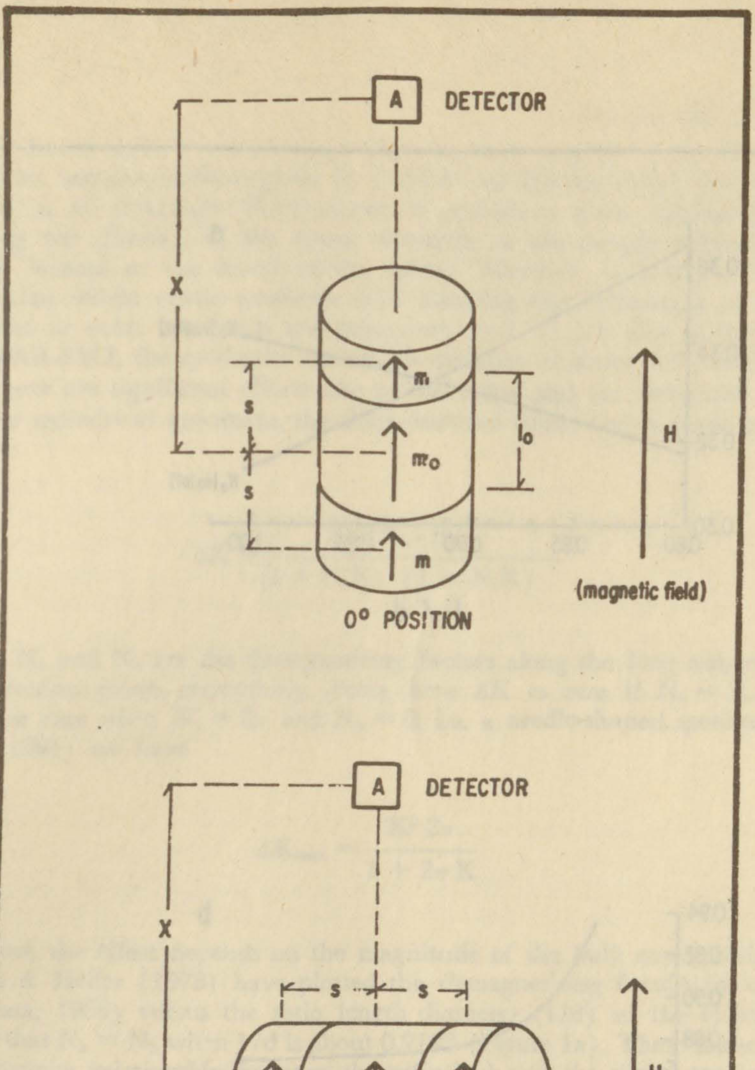

FIG. 2. Explanation of parameters resulting in anomalous anisotropy. Cylindrical specimens could be subdivided into an optimum specimen (optimum 1/d ratio) and two additional portions. A magnetic field (H) induces a magnetic moment (mo) in the central specimen and magnetic moments (m) in the external portions. The distances x from the sample center to the detector are different in the 0° position and the 90° position. Modified from Kent and Lowrie (1975).

tal constant (C), the dimensional ratio $(l-l_0)/l_0$ and the percentage anisotropy $(K_a-K_b)/K$.

The problem of estimating an optimum sample shape for magnetic anisotropy determination has long been studied. The problem has proved to be more complex than previously realized, and several somewhat different results are reported in the literature (Table 1). This appears to have created a great deal of confusion, and some workers are not even taking these effects into account (e.g. Beckman *et al.*, 1977). Porath *et al.* (1966) suggested that the optimum ratio lenght-diameter lies between $0.85 \leq l/d \leq 0.88$ for use with low to high magnetic fields. Banerjee & Stacey (1967) suggested the value 0.845 for use in low fields and 0.89 for use in higher fields (10-15 Oe), approaching 0.90 is infinitely large fields. Noltimier (1971) suggested the use of the ratio 0.865 to be applicable in spinner magnetometers. Singh (1975) studied experimentally the effects of shape-induced anisotropy (using slate-samples) in the Digico spinner magnetometer, and concluded that the results are not affected to any considerable degree. In contrast, Scriba & Heller (1978) presented experimental results for the Digico meter showing marked effects. They used samples of basalt and tonalite and concluded that the optimum ratio is about 0.90 ± 0.02. Similar experiments for a cryogenic meter gave a value of 0.86 ± 0.02, and Scriba & Heller (1978) suggested that the difference is due to instrumental effects associated with varying pickup coil geometries. Murdoch (1978) also reported strong variations depending on the length-diameter ratio by using dolerite samples. By applying a different criterion he concluded that the optimum ratio is about 0.86 (for a Digico meter).

For the experiment some samples were selected from the sample collection of the laboratory in order to study the shape effects on samples of different AMS characteristics. Measurements were made on a PAR-SM2 meter and the length of the samples was reduced in several steps by wet grinding. Some simple parameters were calculated to study the effects on magnetic fabric, i.e.

$$L = K_1/K_2 \qquad \text{(lineation)} \qquad (19)$$

$$F = K_2/K_3 \qquad \text{(foliation)} \qquad (20)$$

$$A = K_1/K_3 \qquad \text{(anisotropy degree)} \qquad (21)$$

where, high value of L, F and A indicate high degree of preferred linear-

TABLE 1. SAMPLE SHAPE EFFECTS.

Magnetic field applied (instruments)	optimum lenght-diameter ratio (1/d)	Reference
1.- Low to high fields	0.85 1/d 0.88	Porath *et al.*, 1966.
2.- Low fields	0.845	Banerjee and Stacey, 1967.
intermediate fields	0.89	
high fields	$\simeq$ 0.90	
3.- Low fields (spinner magnetometer)	0.865	Noltimier, 1971
4.- Low fields (Digico meter)	0.90 $\pm$ 0.02	Scriba and Heller, 1978
Low fields (cryogenic meter)	0.86 $\pm$ 0.02	
5.- Low fields (Digico meter)	0.86	Murdoch, 1978
6.- Low fields (Digico meter)	no appreciable effects on slate samples	Singh, 1975.
7.- Low fields (parastatic magnetometer)	0.88	Christie and Symons, 1969

parallel and planar-parallel orientation and of anisotropy degree, respectively. In addition the following parameters were calculated:

$$q = \frac{K_1 - K_2}{1/2\ (K_1 + K_2) - K_3} \quad (q - \text{factor}) \qquad (22)$$

$$h = \frac{K_1 - K_3}{K_2}\ (100\%) \quad (\text{anisotropy percentage}) \qquad (23)$$

where values of q between 0.06 and 0.67 suggest primary depositional fabrics, and values outside this range suggest secondary strain-controlled fabrics (Hamilton & Rees, 1970). The parameter h represents the degree of anisotropy (see equation 16).

Results from some red beds (Table 2) indicate that large changes occur by reducing the lenght-diameter ratios. An interesting result is given by the variation of the q factor, which suggests that the fabric could be interpreted as a secondary fabric or as a primary, fabric at different ratics (Rees, 1965; Hamilton and Rees, 1970). Directional changes of the principal susceptibility axes (Figure 3) support these interpretations, showing also large changes. The K_1 axes were parallel to the long axes of the samples and the K_3 axes were in the horizontal plane; this pattern was inverted as the length-diameter

TABLE 2. Typical examples of AMS variations as a function of lenght-diameter ratio. Measurements were using a PAR-SM2 spinner meter.

Sample	$X(10^{-6})$	l/d	K_1/K_2	K_2/K_3	K_1/K_3	q	h
RB-1	1.6	1.0	1.095	1.049	1.149	1.008	0.142
		0.9	1.047	1.037	1.086	0.794	0.083
		0.8	1.021	1.030	1.052	0.530	0.050
		0.7	1.036	1.039	1.076	0.648	0.073
RB-2	1.48	1.00	1.099	1.038	1.141	1.150	0.136
		0.95	1.087	1.046	1.137	0.995	0.131
		0.90	1.065	1.030	1.097	1.055	0.094
		0.85	1.042	1.020	1.063	1.034	0.062
		0.80	1.036	1.041	1.078	0.627	0.075
		0.75	1.039	1.037	1.077	0.707	0.074
		0.70	1.040	1.035	1.076	0.743	0.073

ratio was modified. Results from a granitic sample with higher bulk susceptibility and anisotropy degree display a similar pattern (Table 3 and Figure 4). By examining the anisotropy percentage values (and the anisotropy

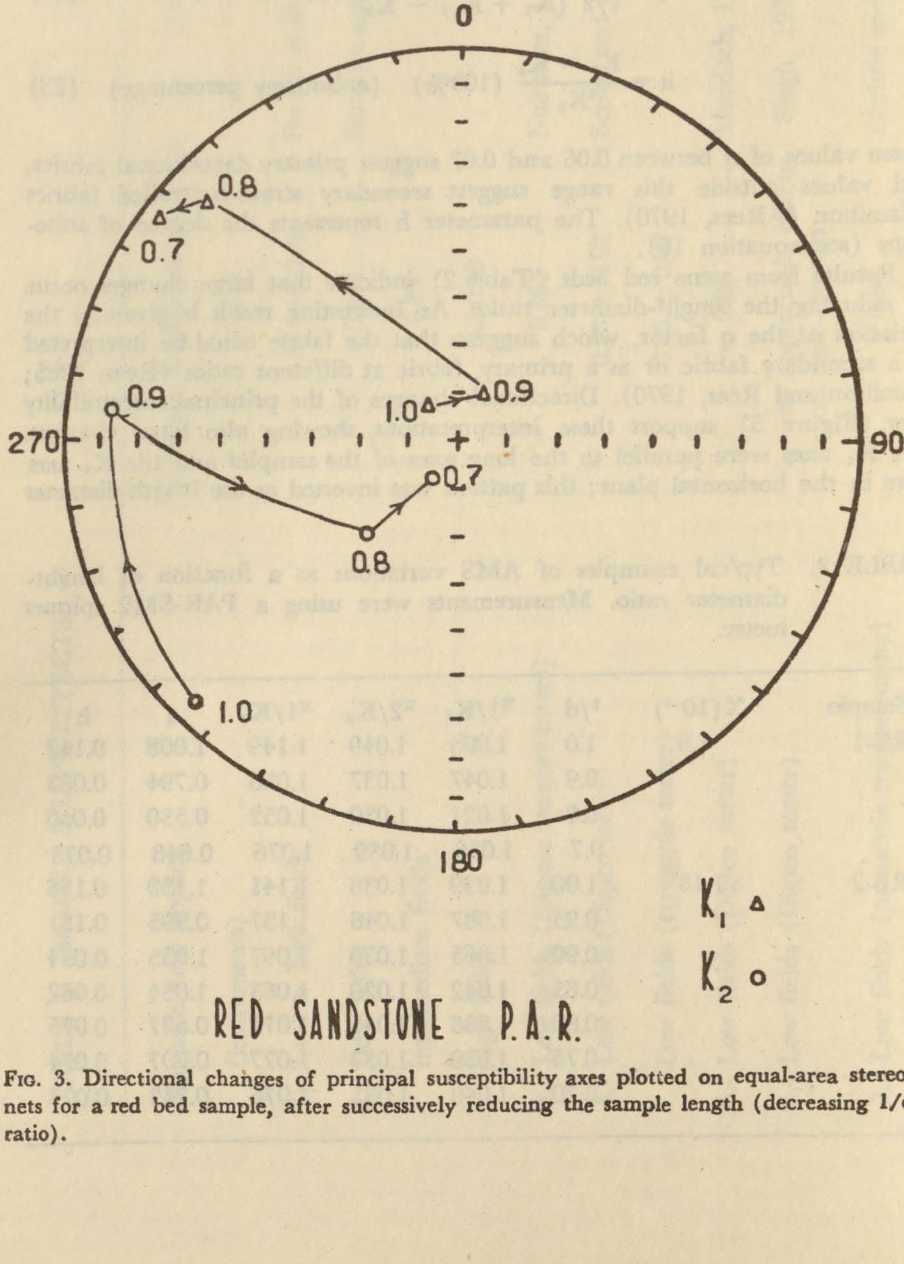

FIG. 3. Directional changes of principal susceptibility axes plotted on equal-area stereonets for a red bed sample, after successively reducing the sample length (decreasing l/d ratio).

degree A), it appears that the values first decrease to a minimum between 0.8 – 0.9 and then increase again. (Figure 5). The range agrees well with the optimum lenght-diameter ratios suggested (Table 1), but no definite conclusion can be given from the results.

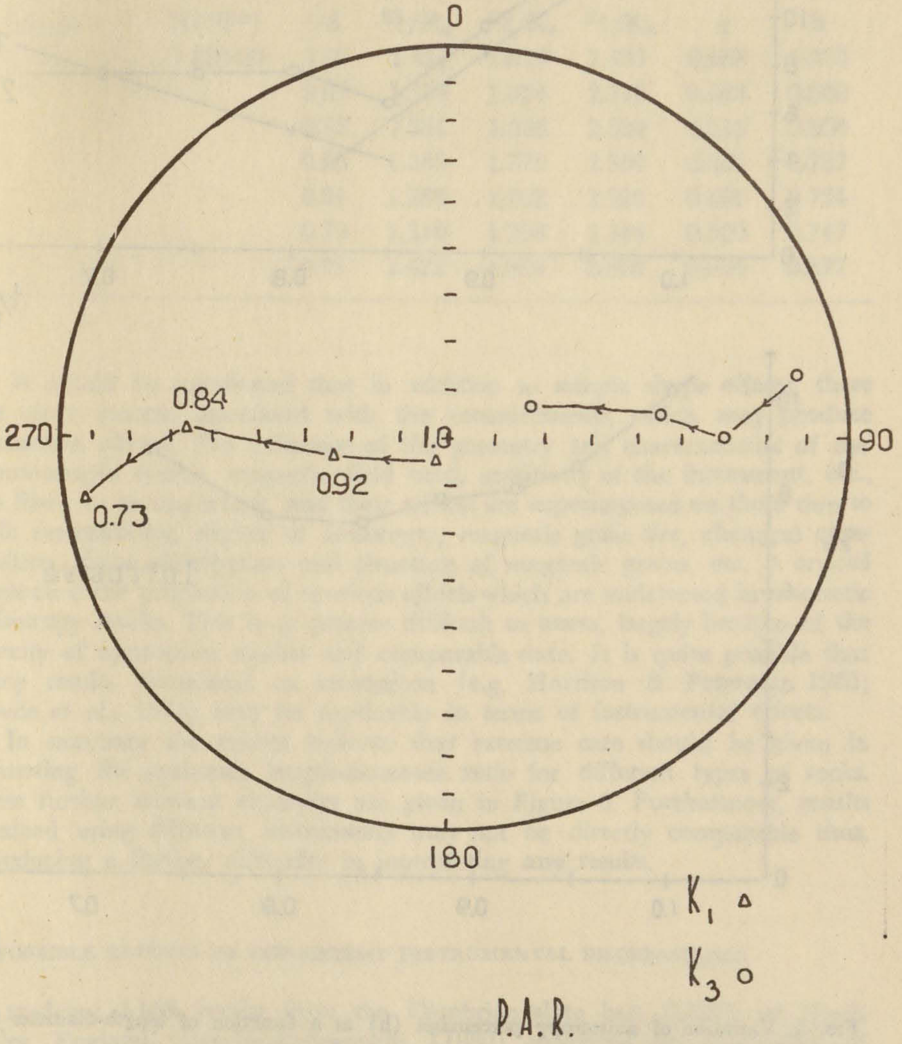

FIG. 4. Directional changes of principal susceptibility axes plotted on equal-area stereonet for a granite sample (see Table 3).

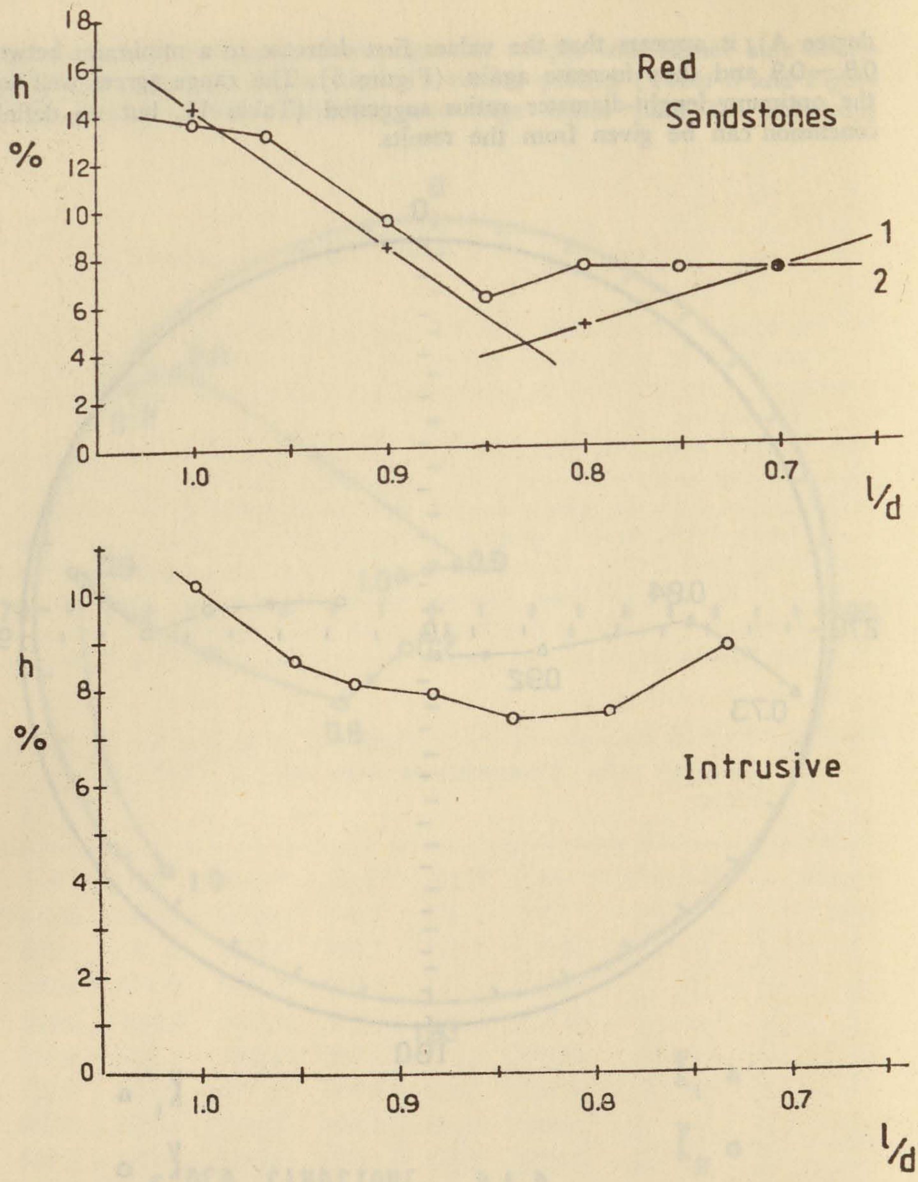

Fig. 5. Variation of anisotropy percentage (h) as a function of length-diameter ratio for two red bed samples.

TABLE 3. AMS variations as a function of lenght diameter ratio for a gran-
the optimum lenght-diameter ratios suggested (Table 1), but no definite
itic sample. Measurements were taken using a PAR-SM2 spinner meter.

Sample	$X(10^{-6})$	l/d	K_1/K_2	K_2/K_3	K_1/K_3	q	h
G-1	7 220.00	1.00	1.451	1.816	2.635	0.668	1.009
		0.95	1.376	1.824	2.510	0.588	0.828
		0.92	1.331	1.896	2.524	0.519	0.804
		0.88	1.365	1.729	2.360	0.604	0.787
		0.84	1.289	1.802	2.323	0.490	0.734
		0.79	1.310	1.798	2.344	0.520	0.747
		0.73	1.422	1.834	2.608	0.634	0.877

It sohuld be mentioned that in addition to sample shape effects, there are other sources associated with the measurements, which may produce anomalous effects. The influence of the geometry and characteristics of the sensor-sample system, magnetic field used, sensitivity of the instrument, etc., are likely to be important, and their effects are superimposed on those due to bulk susceptibility, degree of anisotropy, magnetic grain size, chemical composition, shape distribution and structure of magnetic grains, etc. A crucial quest on is the proportion of spurious effects which are undetected in magnetic anisotropy results. This is at present difficult to assess, largely because of the scarcity of appropiate studies and comparable data. It is quite possible that many results considered as anomalous (e.g. Harrison & Petersen, 1965; Løvlie *et al.*, 1971) may be explicable in terms of instrumental effects.

In summary the results indicate that extreme care should be given in estimating the optimum length-diameter ratio for different types of rocks. Some further relevant examples are given in Figure 6. Furthermore, results obtained using different instruments may not be directly comparable thus, introducing a further difficulty in interpreting any results.

6. POSSIBLE EFFECTS OF CONSISTENT INSTRUMENTAL DISCREPANCIES

By studying AMS results from the Cambrian slate belt (CSB) of North Wales, England, Urrutia-Fucugauchi (1980) suggested the possibility of instrumental discrepancies between a low-field torque magnetometer and a Digico spinner magnetometer.

Wood *et al.* (1976) proposed a logarithmic relationship between the magnitudes of finite-strain and AMS of the form

$$\left(\frac{X_{fi}}{X_o} \right) = \left(\frac{l_{fi}}{l_o} \right)^a \qquad i = 1, 2, 3 \tag{24}$$

where X_f (l_f) and X_o (l_o) are the final and initial susceptibilities (axial dimensions) along a given principal susceptibility (principal strain) axis i. The relationship can also be expressed by

$$M_i = a\,N_i \tag{25}$$

Where a is the correlation exponent and M_i, N_i represent the principal natural

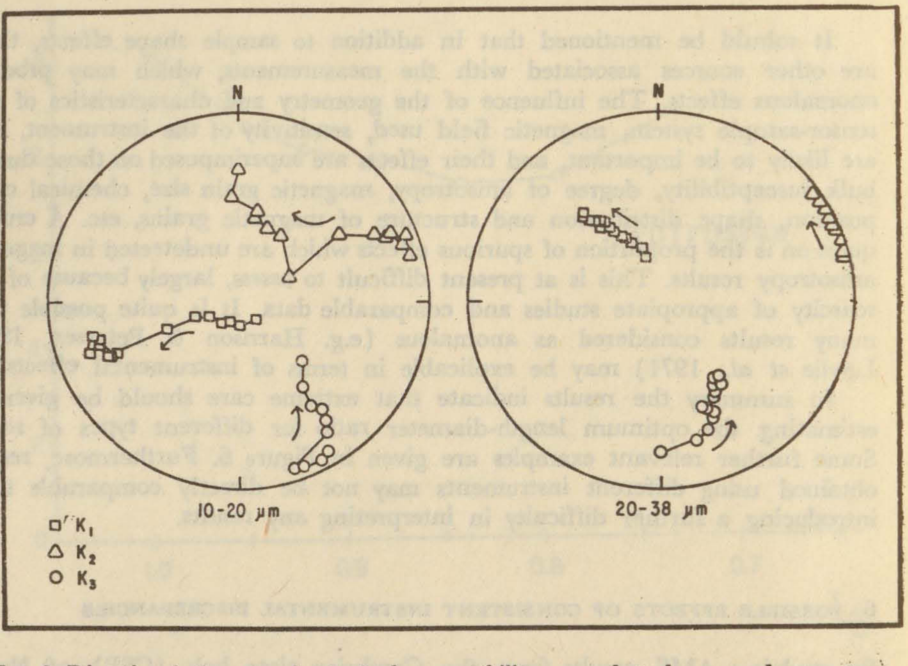

FIG. 6. Directional changes of principal susceptibility axes plotted on equal area stereonets for two samples composed of magnetite grains of different sizes, after successively reducing the sample length. Open symbols are negative inclinations and closed symbols are positive inclinations. After Ellwood (1979).

susceptibility deviators and principal natural strains, respectively (see Wood *et al.* (1976) and Rathcre (1979) for details).

Comparison of results obtained with a low-field torque and a Digico, in terms of the finite-strain AMS relationship, indicates that the torque M_i-values are consistently lower than those of the Digico (Figure 7).

To investigate further on the apparent discrepancy, the data were analysed using a logarithmic relationship (Kneen, 1976). This is in terms of the logarithms of axial ratios given by

$$\log P_1 = \log L \quad \text{and} \quad \log \frac{x}{y} = \log \frac{(\epsilon_1 + 1)}{(\epsilon_2 + 1)}$$

$$\log P_2 = \log A \quad \text{and} \quad \log \frac{x}{z} = \log \frac{(\epsilon_2 + 1)}{(\epsilon_3 + 1)} \qquad (26)$$

$$\log P_3 = \log F \quad \text{and} \quad \log \frac{y}{z} = \log \frac{(\epsilon^2 + 1)}{(\epsilon_3 + 1)}$$

where L, A and F are given in (19), (20) and (21) respectively, and $(\epsilon_1 + 1)$, $(\epsilon_2 + 1)$ $(\epsilon_3 + 1)$ are the principal strains.

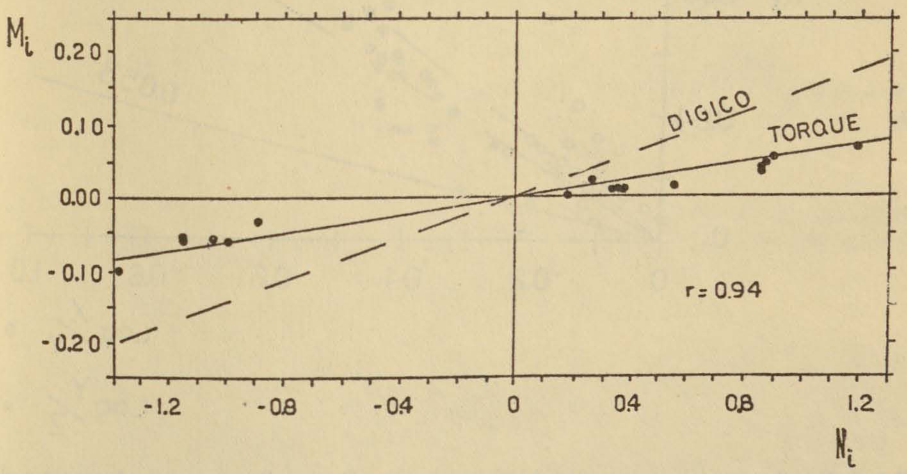

FIG. 7. Graph of principal natural susceptibility deviatiors (M_i) against the principal natural strains (N_i) for slates of the Cambrian belt of North Wales. Closed dots are data obtained with a low-field torque meter. The line is the best least-square fittings function to the data correlation coefficient r = 0.94). Compare with that for the Digico results (discontinuous line). After Urrutia-Fucugauchi (1980).

The AMS data obtained by the Digico meter are consistently higher than those given by the torque meter (Figure 8). The dispersion of results in this relationship is greater than that in the previous one, but there is again a consistent discrepancy.

Since the finite-strain results of both sets were obtained by the same techniques of X-ray texture goniometry and strain marker measurements (Wood, 1974; Wood *et al.*, 1976), the AMS results were examined in closer detail.

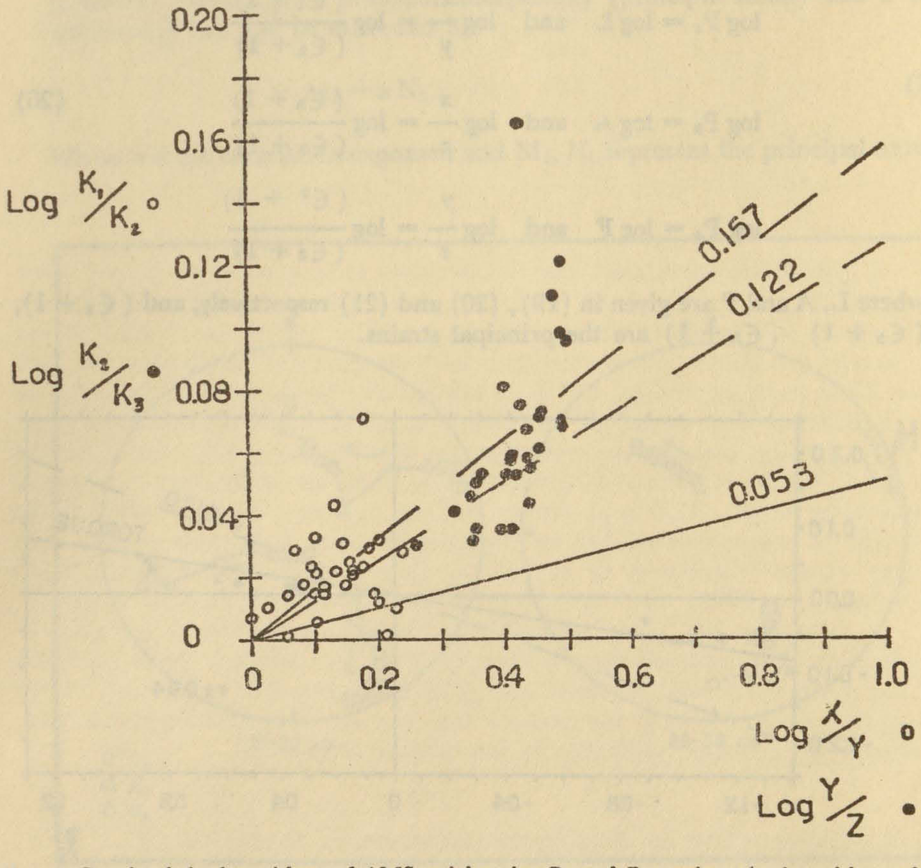

FIG. 8. Graph of the logarithms of AMS axial ratios P_1 and P_3 against the logarithms of strain axial ratios. Segmented lines correspond to best fitting linear regression relations where the numbers indicate the slopes. The correlation coefficients r are 0.77 (sope 0.157) and 0.86 (slope 0.122). Continuous curve is for data obtained with the low-field torque, whereas dots and segmented lines are for data obtained with the Digico spinner meter. After Urrutia Fucugauchi(1980).

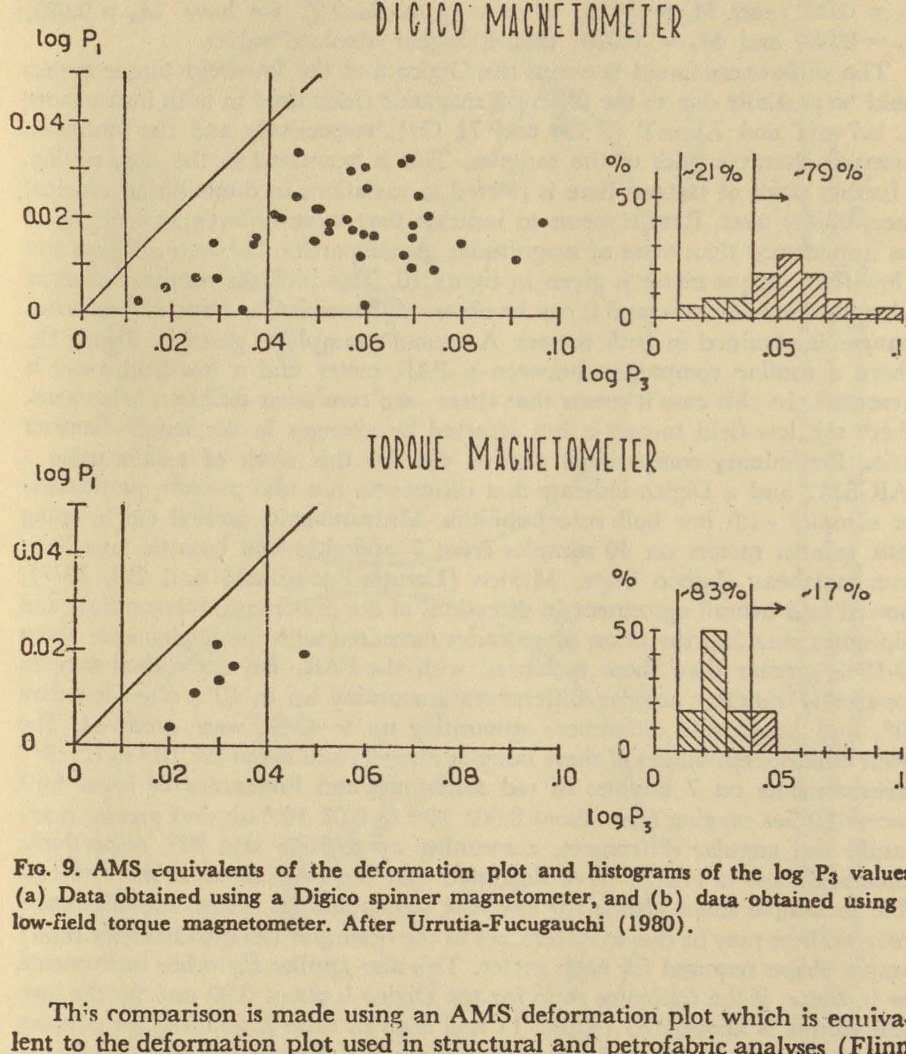

Fig. 9. AMS equivalents of the deformation plot and histograms of the log P_3 values. (a) Data obtained using a Digico spinner magnetometer, and (b) data obtained using a low-field torque magnetometer. After Urrutia-Fucugauchi (1980).

This comparison is made using an AMS deformation plot which is equivalent to the deformation plot used in structural and petrofabric analyses (Flinn, 1978). The Digico-determined P_3 values seem to have been systematically displaced from the range $0.98 < P_3 < 0.92$ to the region P_3 0.92 (i.e. from 0.01 - 0.04 to > 0.04 in Figure 9). To check if this displacement can explain the discrepancy in the finite-strain AMS relationship, the principal natural susceptibility deviators M_i were expressed in terms of the axial ratios P_1 and P_3. Numerical calculations show that the discrepancy can be due to that effect, e.g. at a constant $P_1 = 1.02$, the M_i values with P_3 0.98 are $M_1 = 0.021$,

M_2 = 0.001 and M_3 = 0.019, whereas with P_3 0.92 we have M_1 = 0.083, M_2 = 0.064 and M_3 = -0.020; that is higher absolute values.

The differences found between the Digico and the low-field torque meters could be partially due to the different magnetic fields used in both instruments, i.e. 0.7 mT and 7.1 mT (7 Oe and 71 Oe), respectively and the saturation hysteresis characteristics of the samples. This is examined in the next section. A further point of interest here is related to variations in direction of principal susceptibility axes. Results seems to indicate that these differences could be of less importance than those of magnitudes. A comparison between a Digico and a low-field torque meter is given in figure 10. This includes results of changes in length-diameter ratio and it can be observed that a similar trend of directional changes is obtained in both meters. A second example is given in Figure 11, where a similar comparison between a PAR meter and a low-field meter is presented. In this case it seems that there are consistent different behaviours, where the low-field torque is less affected by changes in the length-diameter ratio. Preliminary comparisons carried out for this work of results using a PAR-SM2 and a Digico indicate that differences are also present, particularly for samples with low bulk susceptibilities. Measurements carried out by using both spinner meters on 40 samples from 7 andesitic and basaltic lava flows from northeast Jalisco State, Mexico (Urrutia-Fucugauchi and Pal, 1977), showed and overall agreement in directions of the maximum, intermediate and minimum axes, but the mean magnitudes measured with the Digico were about 10-18% greater than those measured with the PAR. For individual samples apparently random angular differences amounting up to 40°, often less than 20°, and magnitude differences amounting up to 45%, were observed. The mean bulk susceptibilities of these samples ranged from about 0.1 10^{-3} to 1.5 10^{-3}. Measurements on 7 samples of red sandstones and limestones of lower bulk susceptibilities ranging from about 0.001 10^{-3} to 0.02 10^{-3} sdowed greater magnitude and angular differences, amounting up to 56% and 90°, respectively. Kent and Lowrie (1975) have suggested that the PAR meter is more susceptible to sample shape effects for low bulk susceptibilities. Thus, the differences observed here may be due to differences in the optimum (length-diameter ratio) sample shape required for each meter. This also applies for other instruments, for instance, if the optimum ratio for the Digico is about 0.90 and for the low-field torque is about 0.85 (Table 1) and a given ratio is adopted for samples, then measurements will indicate discrepancies (i.e. Figure 10) merely because it was not used the appropiate ratio for each instrument. M. Stupavsky (private communication cited by Ellwood, 1978) has indicated that the Schonsted spinner meter does not accurately define the AMS when remanence intensity is high relative to the principal susceptibility magnitudes and anisotropy degree. Finally, the precision and comparability of instruments used for bulk susceptibility measurement can also affect the results. In the case of spinner systems it was mentioned that an independent determination of the bulk susceptibility

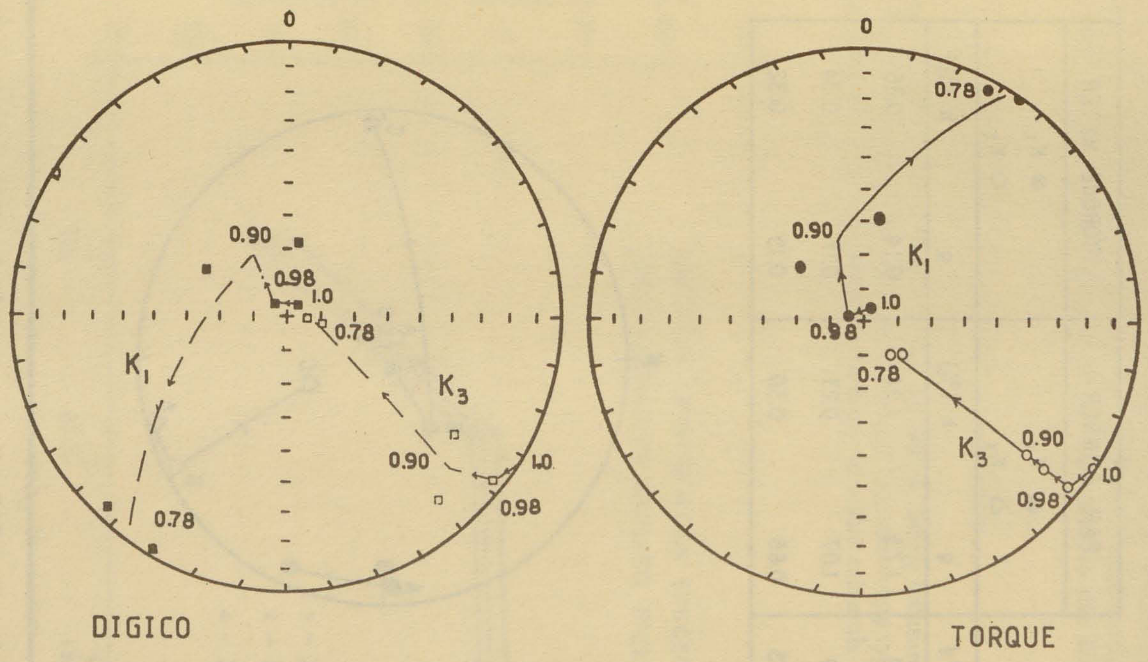

FIG. 10. Comparison between a Digico spinner meter and a low-field torque meter. Closed symbols represent K_1 directions and open symbols represent K_3 directions. Number indicate the length-diameter ratio. Figures constructed using data given by Ellwood (1978).

		PAR SPINNER		TORQUE METER	
		▲ K_1 △ K_3		● K_1 ○ K_3	
	l/d	q	h (%)	q	h (%)
A	1.4	1.73	1.21	0.14	0.35
B	1.0	1.07	0.23	0.17	0.34
C	0.75	0.65	0.30	0.19	0.32

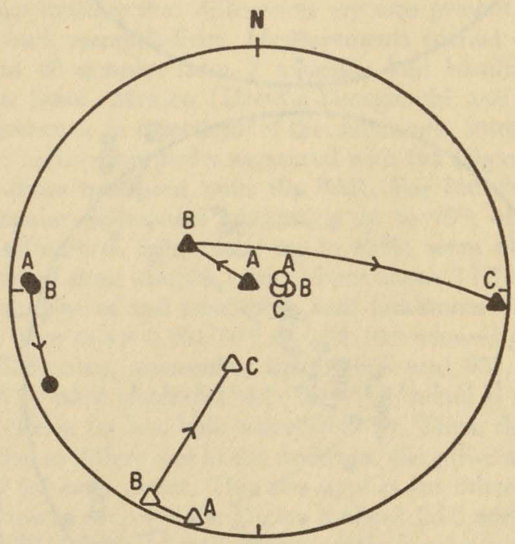

Fig. 11. Comparison between a PAR spinner meter and a low-field torque meter. Variation of anisotropy with length-diameter ratio for a deep sea sedimentary sample. After Kent and Lowrie (1975).

is needed; if different inducing magnetic fields are used, then this may introduce a source of error. Fcr instance, it was found that measurements using the susceptibility bridge (Collinson *et al.*, 1963) are consistently lower than those obtained using the susceptibility bridge attached to the Digico meter.

7. FIELD DEPENDENCE OF MAGNETIC ANISOTROPY

Here it should be stressed that the anisotropy being measured depends strongly on the field used for the measurement and the anisotropy type of the specimen. For instance: (1) crystalline anisotropy of cubic minerals is negligible at low

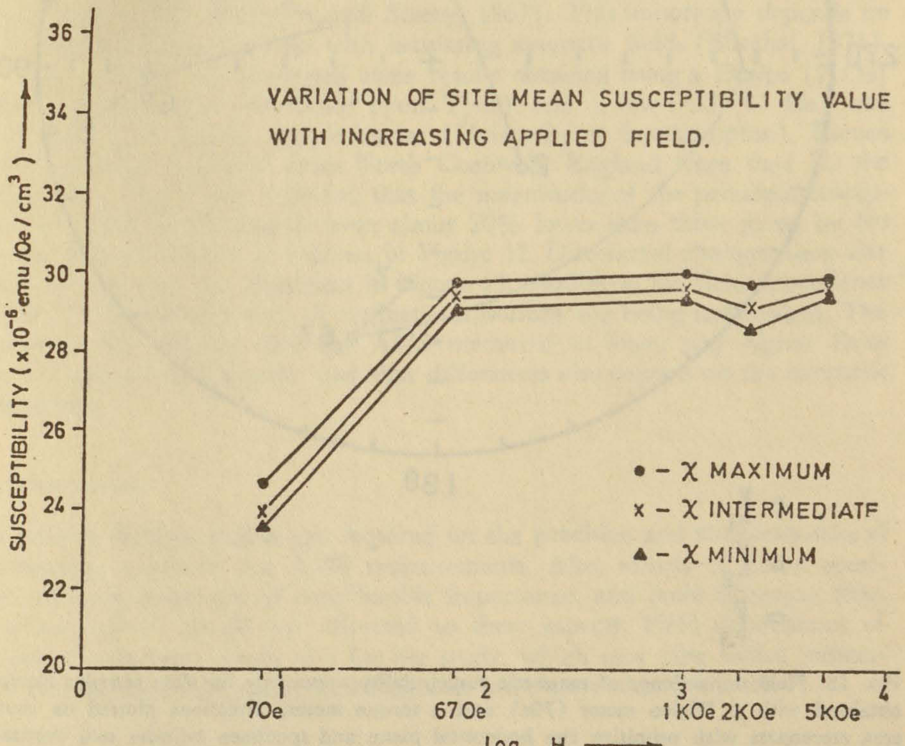

FIG. 12. Field dependence of magnetic susceptibility anisotropy for slate samples. Results obtained with a Digico meter (7Oe) and a torque meter. Magnitudes plotted represent site-means. After Singh (1975).

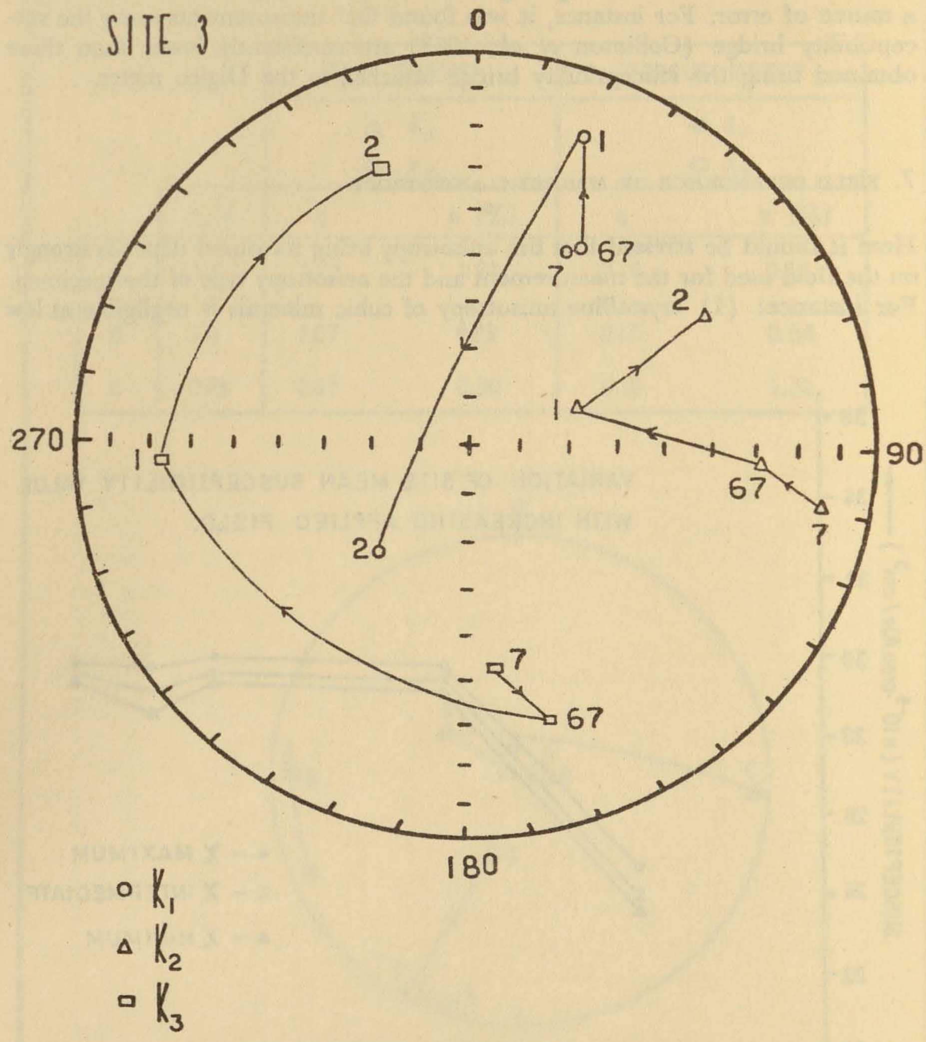

FIG. 13. Field dependence of magnetic susceptibility anisotropy for slate samples. Results obtained with a Digico meter (70e) and a torque meter, directions plotted on equal area stereonets with primitive the horizontal plane and specimen cylinder axis vertical. Symbols are o = K_1, Δ = K_2 and □ = K_3. Directions represent site-means. Numbers indicate magnetic fields used, 70e, 67 Oe, 1KOe, and 2KOe. Figure constructed using data given by Singh (1975).

fields and increases at high fields; (2) anisotropy of minerals of high crystallographic symmetry (such as members of the magnetite-ulvospinel series) at low fields is due to shape effects; (3) domain anisotropy is higher at low fields as high fields destroy the domain alignments; (4) exchange anisotropy is only observable in high fields; (5) textural anisotropy at low fields combines its effects with shape anisotropy, whereas at high fields becomes separated; (6) stress-induced anisotropy is usually observable at low and high fields; and (7) shape anisotropy and magnetocrystalline anisotropy are usually observable at both low and high fields, however, their effects depend on the magnetic mineralogy. Here, there is uncertainty about which fields is best to use. For example some workers use high fields (1.15k Oe) for magnetite bearing rocks (e.g. Gough *et al.,* 1977), whereas most workers use low-fields (less than 75 Oe); in the case of haematite bearing rocks Porath & Chamalaun (1966) have recommended the use of low fields, however other workers use high fields (e.g. Stacey, 1963; Banerjee and Stacey, 1967). This anisotropy depends on the field, showing a decrease with increasing magnetic fields (Bhathal, 1971).

Singh (1975) has presented some results obtained using a Digico (7 Oe) meter (University of Newcastle upon Tyne) and a low-field (67 Oe) and high field (1-5 k Oe) torque meter (University of Southampton). Eleven slate specimens collected from North Cornwall, England were used for the comparison, and it was reported that the magnitudes of the principal susceptibilities given by the Digico were about 20% lower than those given by the torque systems. Results are shown in Figure 12. Directional changes were also observed and they are illustrated in Figure 13. Studies on the field dependence (from 10e to several k Oe) of magnetic anisotropy are being undertaken. The results so far indicate that the AMS measured at lower and higher fields are not directly comparable, and that differences also depend on the magnetic mineralogy.

8. CONCLUSION

In general, further studies are required on the precision and characteristics of instruments available for AMS measurements. Also, effects of finite specimen size and shape are of considerable importance, and more attention than previously given should be directed to these aspects. Field dependence of magnetic anisotropy needs also fur her study, which may give useful indications on magnetic mineralogy and magnetic anisotropy characteristics.

ACKNOWLEDGMENTS

I would like to thank F. T. Addison, D. H. Tarling, J. S. Rathore, L. Molyneux and F. Heller of useful discussions.

APPENDIX

A NOTE ON UNITS

For years, workers on palaeomagnetism and rock-magnetism have been using the unrationalized c.g.s. electromagnetic sistem or e.m.u. system. Recently, the rationalized M.K.S. (metre-kilogram-second) system on S.I. system has become widely used, and several attempts to introduced it in these fields have been given. The International Association of Gecmagnetism and Aeronomy (IAGA has addopted the new system of units. Some of the common units and corresponding conversion factors are given in this appendix.

Symbol	Quantity	Dimensions	S.I. unit	equivalence in e.m.u. units
H	Magnetic field	$\mu_o^{-1}B$	A/m	4.10^{-3} Oe
B	Magnetic density	$\mu_o^{1/2}M^{1/2}L^{-1/2}T^{-1}$	Weber/m²	10^4 Gauss
J	Intensity of magnetization	$\mu_o^{-1}B$	A/m	10^3 Gauss
P	Magnetic moment	$\mu_o^{-1}BL^3$	A/m²	10^3 Gauss cm³
X	volume susceptibility	μ_o	Henry/m	$10^7/16\pi^2$ Gauss/Oe
X_r	relative susceptibility	1		$1/4\pi$
X	specific susceptibility	$\mu_o L^3M^{-1}$	Henry m²/Kg	$10^7/16\pi^2$Gauss cm³/Oe g
σ	Specific magnetization	$\mu_o BL^3M^{-1}$	Weber m/Kg	$10^7/4\pi$ Gauss cm³/g
μ	permeability	μ_o	Henry/m	$10^7/4\pi$ Gauss/Oe

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