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Magnetic, seismic, and other anisotropic properties of rock fabrics

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Magnetic fabric, as a resultant property, summed over individual grains in a rock, stands apart from other, bulk property measurements (e.g. seismic, thermal, dielectric anisotropy), which treat the rock as a continuum. Thus, magnetic anisotropy can be more directly related to preferred orientation of grains in a rock than can bulk property measurements. The latter, however, may permit remote, geophysical determinations of large scale 'fabric' features.

These contrasting aspects of fabric determination are discussed, drawing examples of magnetic fabric analysis from studies on naturally and experimentally deformed rocks, and seismic anisotropy from refraction studies of the crust and upper mantle.

1. Introduction

Anisotropy in any physical property of a material implies that the response of the material to an applied field depends on the direction in which the field is applied. In single crystals anisotropy may be an intrinsic property (Nye 1964), but in rocks it is usually determined by the integrated effects of many fabric elements, which may interact in complex ways. The specification of an anisotropic property necessarily implies that on the scale of interest the body may be treated as homogeneous.

Anisotropic properties are important in geology in two distinct, though related, ways. Firstly, a knowledge of anisotropy is important for a full understanding of geological processes – the influence of rheological anisotropy in deformation is an obvious example. Secondly, measurement of anisotropy may be used to provide information about the fabric of rocks. In comparison with other methods of studying fabric, the relative usefulness of any anisotropic property is determined by a balance between the ease and speed with which it can be measured, the extent to which the contribution of different fabric elements can be resolved and the amount of information (necessarily limited by symmetry) which is carried by the anisotropic property.

This paper is principally concerned with the usefulness of magnetic (§2) and seismic (§3) anisotropy in fabric studies; other anisotropic properties will be considered briefly in §4. In many ways magnetic and seismic anisotropy represent opposite extremes. The former is limited to measurements on the scale of a hand sample and for such specimens measuring procedures are well established. Since magnetic anisotropy is one of the least complex of the anisotropic properties of rocks, in terms of the multiplicity of fabric elements involved, its analysis and its consequent applicability and limitations are relatively well understood. By contrast seismic anisotropy is, at the outset, a more complicated phenomenon. It too may be measured on the laboratory scale, but the possibility that excites greater interest is that of *in situ* determination of anisotropy on a much larger scale, on regions of the crust and upper mantle which are inaccessible to direct observation, but of great interest tectonically. The physical constraints

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associated with such measurements make complete specification of the anisotropy difficult; present interest centres therefore on the geological significance of the measurements that have been obtained to date and on developing methods which will allow a more complete specification of the anisotropy in the future.

2. Magnetic anisotropy

Common rock forming minerals can be classed as dia-, para- and ferrimagnetic. In contrast to ferrimagnetics, dia- and paramagnetic minerals display only induced magnetization. It is convenient, therefore, to consider their response to a magnetic field first and then to consider the rather more complex behaviour of ferrimagnetic minerals.

(a) *Dia- and paramagnetic minerals*

In dia- and paramagnetic minerals the induced magnetization is linearly related to the applied magnetic field:

$$\boldsymbol{J} = \mu_0 \chi_{\mathrm{m}} \boldsymbol{H}. \tag{1}$$

If the mineral is anisotropic, χ_{m}, the magnetic susceptibility, is a second rank tensor quantity (Nye 1964, p. 54), the mean susceptibility being negative for diamagnetic and positive for paramagnetic minerals. If dia- and paramagnetic minerals occur together in a rock, their contributions to the resultant anisotropy cannot be resolved. In general, however, the stronger magnetization of the paramagnetic fraction will swamp that of the diamagnetic minerals. Quartz (which is nearly magnetically isotropic) and calcite are common diamagnetic minerals, whereas ferromagnesian minerals are paramagnetic.

One might expect equation (1) to involve not only the applied external field, but to include a contribution from the field resulting from the specimen magnetization. For dia- and paramagnetic minerals, however, the susceptibility is so low ($\approx 10^{-4}$) that such effects may be neglected (Nye 1964, p. 57). This is an important simplification, for it means, firstly, that crystal shape is not important and secondly, that grain interaction effects (which would involve the spatial relationships between grains, i.e. a textural fabric element) may be ignored, so that, for a distribution of grains in a rock, the resultant susceptibility anisotropy is simply the sum of the grain susceptibility tensors, referred to a suitable reference frame:

$$\boldsymbol{X}_{\mathrm{m}} = \Sigma v_i \chi_{\mathrm{m}i}. \tag{2}$$

The susceptibility tensor can be measured by determining specific vector components of the magnetization induced when a field is applied in a number of given orientations to a sample. This principle underlies the application of spinner magnetometers and of inductance devices to anisotropy measurement (see Collinson *et al.* 1967). In practice it is often advantageous to rotate the sample and thus to measure only susceptibility differences (which yield the deviatoric part of the susceptibility tensor), the mean susceptibility being determined separately. Alternatively, torque meters may be used to determine the susceptibility differences. Since torque meter measurement is central to the subsequent discussion, it will now be considered in more detail.

If an anisotropic sample is suspended in a uniform magnetic field, it experiences a couple which arises from the interaction of the induced magnetization and the applied field, since the two will, in general not be parallel. If reference axes x_1, x_2 are defined in the horizontal plane relative to the specimen, and if the applied field makes an angle θ with the x_1 direction, the torque about a vertical axis on a specimen of volume v *in vacuo* is:

$$T = \tfrac{1}{2}\mu_0 v H^2\{(X_{m11} - X_{m22}) \sin 2\theta - 2X_{m12} \cos 2\theta\}. \tag{3}$$

The torque is periodic in 2θ, with amplitude proportional to the square of the field. From measurements in three orthogonal planes the deviatoric part of the susceptibility may be calculated.

(b) *Ferrimagnetic minerals*

Although magnetic anisotropy associated with pyrrhotite is known (Fuller 1964), magnetite and haematite are the most important ferrimagnetic minerals in rocks. Since they behave rather differently, they will be considered separately (see also Bhathal 1971).

(i) *Magnetite*

Magnetite is strongly magnetic. In contrast to dia- and para-magnetic minerals, the effects of the magnetic field resulting from the magnetization induced in a grain cannot, therefore, be ignored. At low magnetic fields, where its behaviour is reversible, magnetite in bulk possesses an isotropic susceptibility. The magnetization of a grain depends, however, on both the applied field and the demagnetizing field, arising from the induced magnetization, and the latter is a function of grain shape. Thus, for an ellipsoidal grain

$$\boldsymbol{J} = \mu_0 \chi_m(\boldsymbol{H} + \boldsymbol{H}_d) \tag{4}$$

where

$$\boldsymbol{H}_d = -\boldsymbol{N}\boldsymbol{J}/\mu_0 \tag{5}$$

and $\boldsymbol{N}$ is the shape factor, a tensor quantity (Osborn 1945). However, the magnetization of a grain can be related directly to the applied field if, by recasting equation (4), we introduce an anisotropic apparent susceptibility tensor

$$\psi_m = \chi_m(1 + \chi_m \boldsymbol{N})^{-1}. \tag{6}$$

The torque on a magnetite grain will then be given by equation (3) and torque meters can be used to measure the anisotropy (King & Rees 1962).

At very high magnetic fields the magnetization saturates and equation (3), for the torque on a specimen, is replaced by

$$T = -vJ_s^2\{(N_{11} - N_{22}) \sin 2\theta - 2N_{12} \cos 2\theta\}/2\mu_0, \tag{7}$$

where J_s is the saturation magnetization. The torque is, as before, periodic in 2θ, but its amplitude is now independent of field strength. In the saturating case higher harmonics in the torque curve can arise from magnetocrystalline anisotropy of a grain, but if, in a rock, grain shapes are uncorrelated with the crystal axes, preferred orientation by grain shape will randomize this contribution (Stacey 1960).

At intermediate field strengths the magnetic behaviour of magnetite is complicated by discontinuous changes in magnetization. These show up in torque meter measurements as rotational hysteresis (Day *et al.* 1970), which is defined as the work done when a sample is rotated through 360°:

$$W_r = \int_0^{2\pi} T\,\mathrm{d}\theta. \tag{8}$$

Thus it may be measured by the first term of a Fourier analysis of the torque curve. For magnetite and titanomagnetites rotational hysteresis characteristically rises sharply to a peak at

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about 2×10^5 A m^{-1} and decreases smoothly with increasing field strength. Maghemite, which in other respects behaves similarly to magnetite, shows a sharp peak in rotational hysteresis at 1×10^5 A m^{-1}.

The high induced magnetization in magnetite means that magnetic interactions between grains can only be ignored if the magnetite content is low and therefore inter-grain distances are large. This is fortunately the case in many rocks and the simple relation of equation (2) can be applied. Where, however, the grains are not homogeneously dispersed, the possibility of textural anisotropy must be considered. Additionally, in rocks of high magnetite content, interaction effects will require that the overall specimen shape is taken into account.

(ii) *Haematite*

Haematite has a weak magnetization that lies in the basal plane. For perfect single crystals, measured torque curves agree well with calculations made on the assumption that the magnetization can rotate freely within this plane (Porath & Chamalaun 1968). The calculated torque is:

$$T = J_s H \sin\theta \sin^2\phi / \sqrt{(1 + \tan^2\theta \cos^2\phi)}, \tag{9}$$

where ϕ is the angle between the basal plane and the plane of measurement and θ is the angle between the intersection of these two planes and the applied field. The form of the torque curve is no longer simply periodic in 2θ and depends on the angle ϕ. It is noteworthy that this equation predicts a linear dependence of torque on field strength.

In fine grained haematite the movement of the magnetization in the basal plane is constrained by a marked anisotropy, dominantly uniaxial, arising from internal stresses within the grain (Stacey & Banerjee 1974). This anisotropy strongly influences the behaviour of fine grained haematite at both low and high fields.

At low fields the magnetization is deviated only slightly from the easy direction. If we can assume that the easy directions are not preferentially aligned, it follows, from Stacey & Banerjee (1974, p. 87) that the basal plane susceptibility arising from this cause will be about 12.5×10^{-4}, or higher, if the anisotropy, which is grain size dependent, is less pronounced. This is superimposed on an isotropic susceptibility of 2.5×10^{-4}. Thus, in low fields, haematite has a strongly oblate anisotropic susceptibility and will give rise to torque curves of 2θ form described by equation (3).

At high fields the anisotropy causes a large rotational hysteresis which can persist to very high values of the field (over 2×10^6 A m^{-1}; Vlasov *et al.* 1967). From this evidence of incomplete saturation, we cannot, in general, expect the behaviour of fine grained haematite to approach closely, in field strengths readily available, the behaviour described by equation (9).

(*c*) *Comparison of the anisotropic behaviour of dia-, para- and ferrimagnetic minerals*

From the preceding discussion we have established that in low magnetic fields, where ferromagnetic behaviour is reversible, the behaviour of all magnetic minerals can be described by specifying an anisotropic susceptibility, whether of crystalline, shape or magnetocrystalline origin. Where the anisotropy can be ascribed to the presence of a single mineral, low field measurements may provide the most convenient method of defining the anisotropy. Where more than one component is involved, measurements over a range of higher magnetic fields allow the presence of the components to be recognized and, in favourable cases, to be resolved.

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Rotational hysteresis is diagnostic of the presence of the different classes of minerals, for dia- and paramagnetic minerals do not show the effect, while the rotational hysteresis of magnetite peaks at relatively low fields and that of haematite only at much higher fields. The contributions of the minerals to the anisotropy can be resolved, in principle, by separating torque curves, measured at fields above those required for saturation of the ferrimagnetic minerals, into components which show a linear (haematite, equation (9)), and quadratic (dia- and para-magnetic, equation (3)) dependence on field and a component independent of field strength (magnetite, equation (7)). In practice, however, this scheme is difficult to apply where haematite components are involved since the fields required for saturation are very high.

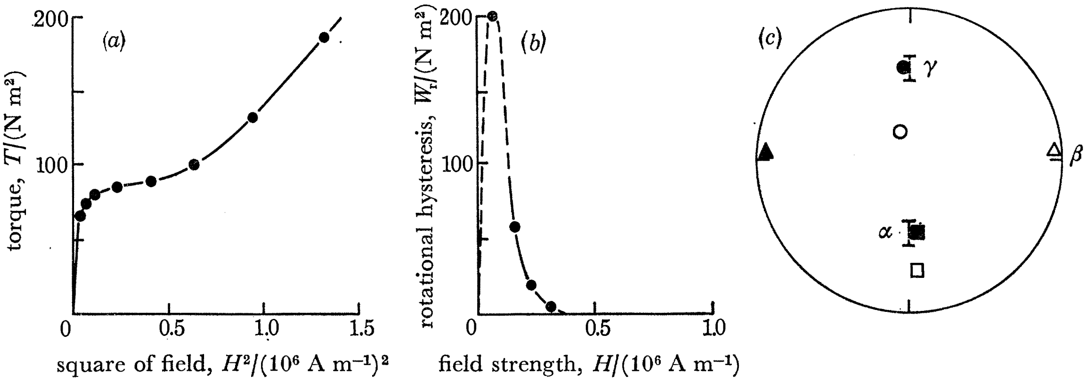

FIGURE 1. Torque meter measurements on an augite crystal. (*a*) Torque amplitude plotted against square of field strength indicates the presence of ferrimagnetic and paramagnetic components. (*b*) Rotational hysteresis against field strength suggests that the ferrimagnetic component is maghemite (cf. Day *et al.* 1970). (*c*) Equal area stereogram of the principal susceptibility (solid symbols) and demagnetizing factor (open symbols) directions (maximum, □; intermediate, △; minimum, ○); α, β and γ are the optic axes of augite.

An example of the separation of saturating and paramagnetic components is given in figure 1 (data of G. R. Parry, private communication). Measurements on an augite crystal indicate a ferrimagnetic component related to inclusions along cleavage planes and a paramagnetic fabric related to the optic axes of the crystal.

(*d*) *Interpretation of magnetic fabrics*

From the discussion above it is clear that in many rocks the resultant magnetic anisotropy can be ascribed to grain fabric alone (equation (2)), an important simplification of the general case discussed in the Introduction. However, the information content of a measurement is strictly limited, as well as the directions of the principal axes, only three magnitudes can be derived. These must reflect the magnetic mineralogy, possibly complex, responsible for the anisotropy, and also its preferred orientation. Even if the measured anisotropy arises from a single mineral species, it is clear that any detailed analysis of the data can only be attempted in the light of an externally supplied model.

The possibilities and limitations of magnetic fabric analysis are illustrated by the examples discussed below. The first refers to a well controlled laboratory situation in which the number of variables is limited, the second to a more complex natural situation in which control must be inferred from the geology. In both examples the anisotropy is demonstrably of dia- or para-magnetic origin; thus they contrast with other published examples of anistropy measurements, in which the fabric has been shown, or has simply been assumed to be of ferrimagnetic origin.

In describing measurements of anisotropic susceptibility, the magnitude parameters, which might most obviously be taken as the principal susceptibilities, have been regrouped to give the following quantities:

$$\bar{X}_m = \frac{X_{m,\max}+X_{m,\mathrm{int}}+X_{m,\min}}{3} \quad m = \frac{X_{m,\max}-X_{m,\mathrm{int}}}{X_{m,\mathrm{int}}-X_{m,\min}} \quad H = \frac{X_{m,\max}-X_{m,\min}}{\bar{X}_m},$$

where $|X_{m,\max}| \geqslant |X_{m,\mathrm{int}}| \geqslant |X_{m,\min}|$ are the principal susceptibilities. This combination follows naturally from the measurement procedure since the directional data and the shape parameter m can be derived directly from the torque meter measurements, whereas the mean susceptibility $\bar{X}_m$ and the strength of the anisotropy H require in addition a measure of one axial susceptibility. $\bar{X}_m$ reflect the volume and nature of the magnetic mineralogy and will not be considered further.

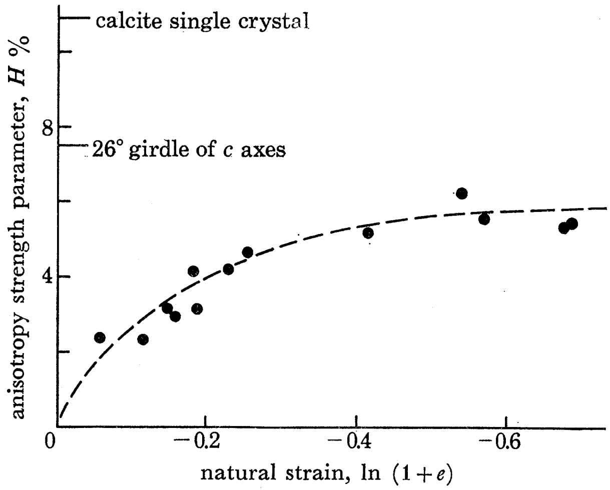

FIGURE 2. Development of diamagnetic anisotropy in samples of Carrara marble experimentally deformed under uniaxial compression at temperatures between 200 and 500 °C. The magnetic fabric is prolate about the compression axis.

The first example refers to measurements on samples of Carrara marble, experimentally deformed by uniaxial compression at temperatures between 200 and 500 °C (E. Rutter, private communication). Calcite has a uniaxial diamagnetic susceptibility anisotropy, prolate about the c axis ($X_{m,\max} = -1.38\times10^{-5}$, $X_{m,\mathrm{int}} = X_{m,\min} = -1.24\times10^{-5}$), Since the observed magnetic fabric is prolate about the compression axis, the results may be interpreted as the development, with increasing strain, of a preferred orientation of c axes about the compression axis. From the magnetic measurements alone nothing more can be inferred. In figure 2 the results are compared with the anisotropy to be expected for a perfect alignment of c axes about the compression direction and with that calculated for a 26° girdle, an orientation favoured by glide on $\boldsymbol{r}$ $\{10\bar{1}1\}$.

Figure 3 summarizes the results of intensive sampling of a small fold in a greywacke unit from the Bannisdale syncline, Shap Fell, Westmoreland (Moseley 1968). That the anisotropy is of paramagnetic origin is demonstrated by a linear plot of torque against square of field. X-ray diffraction shows that chlorite is the only ferromagnesian mineral present and therefore it would be expected to dominate the susceptibility. Although scattered, the results show a consistent variation around the structure, suggesting a systematic modification of grain fabric.

Some insight into the nature of this modification may be gained by examining a simple model (Owens 1974).

The model assumes that the magnetic fabric results from a preferred orientation of plate-like particles of given anisotropy (here taken arbitrarily as 20 %). Parameters of a function adopted to describe the preferred orientation are chosen so that the magnetic anisotropy parameters correspond to those of an 'initial model': here a Bingham distribution is taken, which gives $m = 0.65$ and $H = 4.8\%$, with the minimum susceptibility axis perpendicular to the bedding.

FIGURE 3. The magnetic fabric of a fold in greywacke (Moseley 1968, fold I). The principal susceptibility directions (maximum, □; intermediate, △; minimum, ○) are plotted as equal area stereograms projected on to the plane of section.

The response of this distribution to pure shear, in a plane perpendicular to the intermediate susceptibility axis, can be calculated for a range of strain magnitudes and orientations, if the simplifying assumption is made that the particles respond to strain as passive planes. In principle, then, by comparison of observations with the model predictions, the changes in m, H and the orientation of the principal susceptibility axes should provide sufficient information to solve for the relative strain. In practice, for the model adopted, the angular dependence of H follows that of m very closely (figure 4b), reflecting a weak dependence of these parameters on the assumed form of the distribution. Therefore an alternative strain dependent parameter must be found. To this end figure 4a records both the variation of the parameter m as a function of strain magnitude and orientation and also that of the angle between the deformed bedding and the maximum susceptibility axis. From this plot measurements of these two parameters can be used to solve for the irrotational component of strain; the rotational component then follows from the geometry. An obvious limitation, however, in the application of this method to a case like that of figure 3 is the requirement that the orientation of the bedding must be extrapolated from the attitude of the upper bounding surface.

Figure 4*c* summarizes the results of the application of this analysis to a two dimensional idealization of the field data. A quantitative interpretation of the results is clearly unreasonable, in the light both of the sweeping assumptions made in setting up the model and of the relatively poor comparison of observation and prediction in figure 4*b*, which may indicate that the parameters of the assumed model were not well chosen. Nevertheless, the *pattern* of variation in

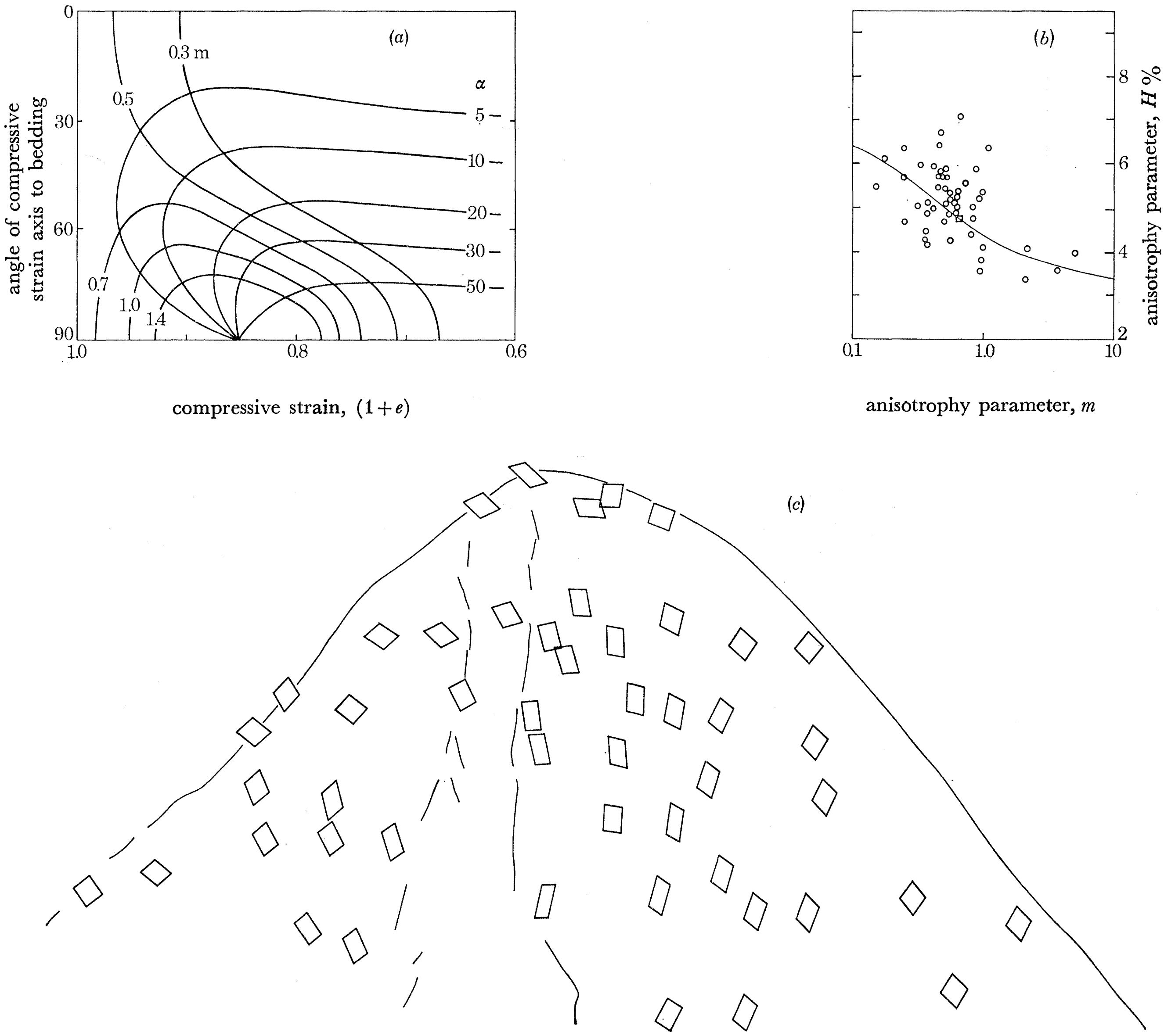

FIGURE 4. Interpretation of magnetic fabric results. (*a*) Model predictions of the magnetic anisotropy parameter, *m*, and the angle, *α*, between the magnetic foliation plane and the deformed bedding, as functions of the amount and orientation of pure shear. (*b*) The single-valued relation between the anisotropy parameters, *H* and *m*, predicted from the model, compared with the scatter of observed values. The □ corresponds to the initial values adopted for the model. (*c*) The pattern of relative strain variation around the fold of figure 3 inferred by application of the results of (*a*). The strain is indicated by a parallelogram which in the reference state was a square with sides parallel and perpendicular to the bedding.

strain intensity and orientation can be expected to be relatively insensitive to details of the model. That the overall pattern of variation in figure 4*c* is plausible may lend support to the hypothesis underlying the model, that the preferred orientation, in this case at least, has developed by a mechanism which involves a significant component of geometrical rotation, though further control is desirable to test this point.

Taken together, these examples illustrate a number of points. Firstly, they demonstrate that systematic changes in magnetic fabric may accompany deformation, though even in mineralogically simple cases, the interpretation of these changes may not be obvious or unambiguous, particularly if modification of a pre-existing fabric is involved. Secondly, the quantitative relation of magnetic fabric to strain is indirect and must involve a model for grain fabric modification. The limited information provided by magnetic measurements alone is of little help in formulating such a model and alternative, more detailed methods of grain fabric analysis (e.g. X-ray texture goniometry, universal stage microscopy) must be employed. The most obvious attributes of magnetic fabric measurements, to be weighed against their relatively limited information content, are their speed and sensitivity. That they permit higher density sampling than could be contemplated with alternative means of fabric analysis has been demonstrated (figure 3). Thus their role in a comprehensive scheme of grain fabric analysis is twofold – in the early stages as a sifting technique, for locating areas on which more powerful (and expensive) techniques may be brought to bear, and in the final stages as a convenient means of testing quantitative predictions made about the grain fabric.

3. Seismic anisotropy

Seismic anisotropy is observed over a wide range of scales, in the laboratory in single crystals and hand specimens and *in situ* on scales from a few metres to several hundred kilometres, and is conventionally understood to imply the dependence of measured compressional and shear wave velocities upon the direction of observation. The phenomenon is not restricted to any one type of rock – indeed Tilman & Bennett (1973) have suggested that 'anisotropy is probably the rule rather than the exception in describing the elastic behaviour of rocks of hand sample size'.

The theory of elastic wave propagation in anisotropic media is well understood; briefly, any generalized anisotropic medium may be completely described by specification of its density and the 21 elastic coefficients which form the cartesian components of a fourth order elastic tensor (Nye 1964); special combinations of these elastic coefficients determine the seismic anisotropy, that is, compressional and shear wave velocities as a function of direction. Broadly speaking, the wide variety of phenomena that can affect the elastic coefficients of rocks, and thus produce different types and magnitudes of seismic anisotropy, may be classified as related either to the preferred orientation of cracks, or to the preferred orientation of minerals; we consider these two classes separately.

(*a*) *Crack anisotropy*

It is well known from laboratory and *in situ* studies that crack porosity can significantly reduce seismic velocities even in low-porosity igneous rocks; if the cracks have a preferred orientation then natural seismic anisotropy will result. On the other hand, anisotropy can be induced *in situ* in rock containing non-preferentially oriented cracks if a non-hydrostatic stress field is applied;

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cracks normal to the field close and hence velocity tends to increase in this direction while cracks aligned parallel to the stress field are relatively unaffected.

Nur (1971) has discussed the theory of both naturally occurring and stress-induced anisotropy in rocks with cracks, and Anderson *et al.* (1974) have extended this treatment to the case in which the cracks are fluid-filled. For the simple case in which the cracks are parallel and 'penny-shaped' and sit in an otherwise homogeneous isotropic material, only five independent elastic coefficients are required and, in bulk, the material is transversely isotropic (Stoneley 1949): the predictions of such simple models compare favourably with laboratory crack distribution measurements. However, as Nur emphasizes, although velocities may be uniquely determined from the crack distribution, the reverse is not true. Thus, *in situ* seismic anisotropy measurements in cracked rock cannot, by themselves, yield information which can be explicitly related to the crack-forming mechanism, for example, to tectonic deformation; they *can*, however, be used to monitor stress changes and hence present-day tectonic deformation. Todd *et al.* (1973) – who also summarize several examples of naturally occurring and stress induced velocity anisotropy – use estimates of crustal stresses and laboratory velocity data to show that stress build-ups could induce *in situ* anisotropies of 13 % in compressional wave velocity and 8 % in shear wave velocity of cracked rocks. Anisotropy at such levels can be detected by conventional and straightforward seismic methods and are an order of magnitude greater than the anisotropy induced if a comparable homogeneous anisotropic stress is applied to a perfectly elastic homogeneous medium (Dahlen 1972).

Crack anisotropy ceases to be important at pressures greater than 1 or 2×10^8 Pa as all cracks will then be closed; thus, within the lower crust and upper mantle, seismic anisotropy can be solely ascribed to mineral associated anisotropy.

(*b*) *Mineral anisotropy*

Mineral anisotropy can be caused either by the alignment of individual grains or by the layering of one or more minerals.

Laboratory studies of elasticity and petrofabric studies provide clear evidence that many ultramafic rocks possess 'bulk' anisotropic properties; in particular a preferred orientation of olivine grains occurs in many ultrabasic rocks and, because of the large directional variations in the elastic properties of single olivine crystals (Verma 1960), the resultant bulk velocity anisotropy can often be greater than 10 % for both compressional and shear waves. A closer examination of individual samples often shows the olivine *b* crystallographic axes to be concentrated parallel to slow velocities while the *a* and *c* axes tend to lie in a girdle normal to the *b* axis – the material is then transversely isotropic (Christensen & Crosson 1967).

Layering in rocks can also produce seismic anisotropy; if the thickness of the layering is very much less than the seismic measurement wavelength, then, in bulk, the rock is transversely isotropic, the apparent elastic constants being averages over the layers (Backus 1962). This type of anisotropy is well known in sedimentary rocks and can occur in metamorphic and igneous rocks due to layering of minerals.

Thus, mineral-associated seismic anisotropy is related to and may be predicted from rock fabric (see, for example, Baker & Carter 1972); however, as with crack anisotropy, it is not possible to use seismic anisotropy measurements *alone* to determine rock fabric: they do, on the other hand, both complement and supplement conventional techniques of fabric analysis. Christensen (1971), for example, used *both* fabric and seismic anisotropy studies to examine the tectonic history of the Twin Sisters Dunite, Washington State. His results showed a complex but definite pattern of seismic anisotropy throughout the body in which the slow velocities tended to be horizontal and normal to the elongation of the body: mean velocities in a vertical direction were slightly higher than mean north–south velocities. This anisotropy is related to preferred olivine orientation, and Christensen interprets this fabric as having originated by recrystallization accompanying flow in the upper mantle with local modification by cataclasis and plastic flow; the dunite body thus appears to have been transported from the mantle as a solid body along a major thrust fault. Peselnick *et al.* (1974) have applied a similar combined approach to the Lanzo peridotite massif in the western Alps: this body is interpreted as a rooted intrusion of the upper mantle.

Mineral anisotropy is believed to be responsible for the best documented example of *in situ* seismic anisotropy, that is, the existence of upper mantle anisotropy in oceanic areas. The clear association of this anisotropy with directions of sea-floor spreading – maximum velocities are observed normal to magnetic lineations and parallel to transform faults – has led several authors (see Carter *et al.* 1972; Peselnick *et al.* 1974 for reviews) to suggest a direct connexion between the origin of this anisotropy and the mechanism of lithospheric generation and/or movement. Furthermore, the best (to date) example of *in situ* seismic anisotropy in a *continental* upper mantle (in southern Germany; Bamford 1973) has been related by Illies (1974) to the palaeo-stress field that governed Rhinegraben tectonism to the end of the Oligocene; in this case the maximum seismic velocity and the maximum stress direction are parallel to each other and to the graben axis.

(*c*) *Discussion*

We see, therefore, that, while observations of crack-associated anisotropy are probably most useful for the monitoring of present-day stress build-ups, observations of mineral associated seismic anisotropy within the upper mantle and lower crust may be fundamental to our understanding of plate and intra-plate tectonics; in particular it may be possible to discriminate between various tectonic models by comparison of model predictions with observed velocity data (Crosson 1972) – for example a transversely isotropic mantle would fit some models and not others. However, the realization of such an exciting possibility is dependent on the availability of sufficiently widespread, reliable and detailed measurements of seismic anisotropy; these are already available in the laboratory but *in situ*, although there is every reason to believe that the state of the art can be improved (Bamford 1975), measurements are few in number and tend to be either fairly qualitative, for example a positive indication that elastic anisotropy is required to explain some aspect of available observations, or quantitative only in a limited sense, for example, accurate measurements of the variation of compressional velocity in the horizontal plane only; a brief summary of the history of anisotropy measurements in oceanic areas illustrates these points very well.

The first report of velocity anisotropy in an oceanic area was by Hess (1964); he compiled the results of a large number of linear seismic refraction profiles in the Pacific – each profile giving a reasonably accurate measurement of velocity in one direction only – and suggested that his results were consistent with a variation of up to 8 % in upper mantle compressional wave velocity with direction, the higher velocities being sub-parallel to the fracture zones. This suggestion prompted, in both the Pacific and Atlantic Oceans, a series of special seismic refraction experiments designed specifically to measure upper mantle compressional wave velocity

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anisotropy in the horizontal plane; these were generally successful. Morris *et al.* (1969), for example, measured an overall velocity variation of 0.6 km s^{-1} (7 %) in the upper mantle off Hawaii and confirmed that the maximum velocity was parallel to the fracture zones. To date, probably half-a-dozen such experiments have been completed in various oceans; we note that they have only measured compressional wave velocity and this only in the horizontal plane: furthermore, the nature of the seismic refraction method is such that these measurements refer at most to the top one or two kilometres of upper mantle immediately below the Mohorovicic discontinuity.

There is no comparable information available at other depths in the oceanic lithosphere although Christensen (1972) has carried out a similar compilation to Hess's, this time for the lower oceanic crust, and reached similar conclusions; he suggests anisotropy of several per cent in compressional wave velocity with velocity maxima corresponding fairly sensibly with fracture directions.

Other types of seismic anisotropy measurement are almost nonexistent in oceanic areas although Forsyth (1973) reports that his surface wave data from the eastern Pacific requires a 2 % velocity anisotropy *somewhere* in the lithosphere. There have been no shear wave velocity measurements in the oceans and we have no idea whether the anisotropy we observe is general anisotropy or transverse isotropy.

In continental areas even less information is available (Bamford 1975), and thus in both oceanic and continental areas we are at present some distance away from complete descriptions of the *in situ* anisotropy that is undoubtedly present.

4. Electrical and thermal anisotropy of rocks

The electrical behaviour of rocks in their natural state is determined principally by their porosity. Thus, in rocks of high porosity, current flows by electrolytic conduction in the pore water. Anisotropy of resistivity in this case will be determined principally by the relative spatial arrangement and size distribution of grains; as might be expected, lithological control of resistivity is well known (Keller & Frischknecht 1966).

Rocks of low porosity and those that have been artificially dried are, in general, insulators. Since the theory of polarization of dielectrics is closely allied to that of induced magnetization, it might be thought that dielectric anisotropy might prove equally useful in providing information about the grain fabric. There are, however, two important differences between the magnetic and the electric cases.

Firstly, the relative strengths of magnetic and electric susceptibility require that, whereas, in the case of induced magnetization, (susceptibility $\approx 10^{-4}$) grain interaction effects could be ignored, in the case of electric polarization (susceptibility ≈ 10) this in no longer true (Nye 1964, p. 70). Thus, although rocks can show pronounced electrical anisotropy, interpretation of measurements of dielectric anisotropy will be more complex than that of magnetic anisotropy, since both grain fabric and textural effects will contribute. Further, the induced polarization will depend on the overall specimen shape; thus separate specimens (usually disks) will be required for each directional measurement needed to define the anisotropy.

The other difference relates to the source of the polarization. Hill (1972), in perhaps the most comprehensive study to date of the electrical anisotropy of rocks, has demonstrated that electrical properties, at least in the frequency range normally used in measurement, are strongly influenced by effects of interfacial polarization and thus are unlikely to be closely specifiable. This is in contrast to the magnetic properties, which are primarily determined by mineral composition.

On a larger scale, the possibility of using electrical measurements for *in situ* determinations of anisotropy, though attractive, must be discounted, for potential theory precludes the unique determination of anisotropic electrical properties. This does not mean that electrical anisotropy is unimportant in this sphere; rather that a knowledge of the probable anisotropy is required so that the range of possible solutions in the interpretation of electrical soundings may be established.

Measurements of thermal anisotropy of rocks have been made and are of relevance both to detailed modelling of thermal regimes associated with tectonic processes and to heat flow measurement. However, the practical difficulties associated with thermal measurements suggest that they are unlikely to be widely applied in studying rock fabric. Indeed the work of Johnson & Wenk (1974) was directed towards establishing a correlation between seismic anisotropy and thermal anisotropy, so that, from the relatively straightforward measurements of the former, the latter might be predicted.

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