Magnetic susceptibility, petrofabrics and strain

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Abstract


Magnetic susceptibility is a non-destructive technique for quantifying the average fabric of a small sample of rock. The interpretation of the magnetic fabric is not always straightforward. However, the principal directions of the magnitude ellipsoid of susceptibility commonly show orientations consistent with the kinematic interpretations of folds, shear zones and other structural features. The directions may correspond with the orientations of strained objects or with the plane-linear mineral orientations. There will usually be multiple mineralogical sources of susceptibility, often involving silicates. If the sources are known, or if the susceptibility can be attributed to a single mineral species, it may be possible to establish a correlation between the strain ellipsoid and the susceptibility ellipsoid. This correlation will be of principal directions in many instances and occasionally there may be a weak correlation of strain magnitudes as well. In other circumstances it may be possible to establish a correlation between changes in susceptibility and the strain. Nevertheless magnetic fabric studies are not routine substitutes for strain analysis. Even where information on strain is not provided, the magnetic fabrics (and subfabrics) yield a measure of the preferred crystallographic orientation or preferred dimensional orientation of the minerals that may be ingressed profitably with other petrofabric data. Experimental deformation of certain synthetic aggregates indicates that directions of magnetic susceptibility spin rapidly with advancing strain, especially where the matrix grains undergo crystal-plastic deformation. In certain experiments, simple shear appears to change the intensity of magnetic fabric more effectively than pure shear. Experiments indicate also that the initial anisotropy of a rock-like material is not easily overprinted by deformation whereas field studies are equivocal.

Introduction

This review draws the attention of structural geologists to the value of magnetic fabrics but indicates current limitations in their interpretation. Note is also made of some relevant complexities of rock deformation that are not always emphasised in geophysical studies.

Shape changes and some aspects of kinematic histories of rocks, especially tectonically deformed rocks, may be inferred from grain and mineral fabrics. These procedures are time-consuming, difficult and require the presence of special structures or textures. In contrast, the anisotropy of magnetic susceptibility may be determined rapidly from a wide range of rocks in suitably shaped specimens (usually pieces of drill-core with a length = 0.82 x diameter). Its most attractive feature is that it sums the magnetic contributions of all components of the rock and presents an average, bulk fabric description. Its relationship with fabric of the rock arises because the most magnetically susceptible minerals can have distributions of shape orientations or lattice orientations influenced by the kinematic history of the fabric. If these influences apply similarly to the rest of the rock's fabric, then the magnitude ellipsoid of susceptibility may be a faithful representation of the total fabric.

There is a wealth of literature on the magnetic susceptibility of rocks (Hrouda, 1982; Urrutia-Fucugauchi and Odabashian, 1982; Zavoyskiy,
Textbooks discuss applications (Tarling, 1983) and instrumentation (Collinson, 1983); detailed studies date from 1942 (Ising). The earliest uses concern sediments and sedimentary rocks. Rees (1967) and Hamilton et al. (1968) showed experimentally that the movement of sediment produces characteristic susceptibility patterns. Numerous field studies either foresaw or followed this work with inferences of paleocurrent trends and stratification planes from the principal directions of magnetic susceptibility (e.g., Hamilton, 1963; Graham, 1966; Crimes and Oldershaw, 1967; Van den Ende, 1975; Hamilton and Rees, 1970, 1971; Hrouda and Janak, 1971).

A similar theme has been followed by workers on volcanic or other igneous rocks in which cryptic flow fabrics of the magma have been inferred from the orientation of the magnetic susceptibility ellipsoid (Khan, 1962; Hrouda and Chlupacova, 1980). Mimetic igneous fabrics have also been discerned (Hrouda et al., 1971). Magmatic flow fabrics and their magnetic anisotropy have been simulated experimentally (Wing-Fatt and Stacey, 1966). The alteration of magmatic rocks (Hrouda et al., 1972; Lapointe et al., 1986), the fabrics of ore deposits (Schwarz, 1974; Hrouda et al., 1985) and the deformational history of a chondritic meteorite (Hamano and Yomida, 1982) have all been studied through their magnetic fabrics. Most studies have concentrated on deformed rocks since their petrofabrics are noticeably anisotropic (Graham, 1954). For example, Singh et al. (1975), Hrouda (1976), Kneen (1976), Wood et al., 1976; Kägi et al. (1977) and Stott and Schwerdtner (1981) all were able to compare the principal directions of the susceptibility ellipsoid with the principal fabric elements in slates and low-grade schists.

Magnetic susceptibility anisotropy and mineralogy

Susceptibility is a measure of the response of the material to an applied magnetic field. This can be considered under three main classes (e.g., Uyeda et al., 1963; Strangway, 1970; Owens and Bamford, 1976) for minerals which are responding to weak applied fields, of the same order of magnitude as the Earth’s field. The majority of rock-forming minerals are either paramagnetic or diamagnetic. In the paramagnetic case the magnetisation ($M$) induced in the mineral is considered to support an applied weak field, or magnetic intensity ($H$) weakly, in a linear fashion:

$$M = k \cdot H$$

For the diamagnetic case, $k$ is negative. For rock-forming minerals $k$ is typically small, about $1 \times 10^{-4}$ SI units/unit vol. for some paramagnetic minerals and typically $-1 \times 10^{-5}$ SI units/vol. for diamagnetic minerals. Susceptibility is also sometimes given in cgs units which are $1/(4\pi)$ of an SI unit.) In geology it is also sometimes expressed on a unit mass basis as the specific susceptibility ($\chi$) rather than a unit volume basis ($k$). It is useful to report the densities so that comparison can be made between results reported in the two schemes. For diamagnetic and paramagnetic minerals the anisotropy of susceptibility is controlled by the crystallographic orientations of the mineral grains and is not appreciably influenced by grain shape. The anisotropies of paramagnetic matrix-forming minerals are quite high (see Table 1) by comparison with diamagnetic minerals (see Table 2).

Although they usually occur as accessory minerals, a third important class are the ferromagnetic minerals. Magnetite and hematite are most important, although pyrrhotite can make a significant contribution (Fuller, 1976; Rochette, 1987). The susceptibilities of natural examples of these minerals are sensitive to composition but are positive and are large by comparison with the paramagnetic and diamagnetic rock-forming minerals. (Typical approximate values are 5 SI units/vol. for magnetite and $6 \times 10^{-3}$ SI units/vol. for hematite).

The response of magnetite depends non-linearly on the strength of the applied field but for the low fields at which routine susceptibility determinations are made the susceptibility anisotropy of magnetite is controlled by the shape of the mineral grain as long as the grains are well-spaced. (There may be differences in the determination of the magnitude of susceptibility because of its slight dependence of the field strength (Smith and Banerjee, 1987), however, the effects
TABLE 1  
Susceptibility data for common metamorphic minerals *

<table>
<thead>
<tr>
<th>Mineral</th>
<th>( k_{\text{max}} )</th>
<th>( k_{\text{med}} )</th>
<th>( k_{\text{min}} )</th>
<th>Susceptibility (SI units/vol. x 10^-6)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Actinolite</td>
<td>1.076</td>
<td>0.942</td>
<td>0.947</td>
<td>3560</td>
</tr>
<tr>
<td>Actinolite</td>
<td>1.083</td>
<td>0.927</td>
<td>0.899</td>
<td>6500</td>
</tr>
<tr>
<td>Hornblende</td>
<td>1.347</td>
<td>0.817</td>
<td>0.809</td>
<td>8920</td>
</tr>
<tr>
<td>Chlorite</td>
<td>1.052</td>
<td>0.992</td>
<td>0.958</td>
<td>333</td>
</tr>
<tr>
<td>Glaucophane</td>
<td>1.094</td>
<td>1.000</td>
<td>0.908</td>
<td>787</td>
</tr>
<tr>
<td>Chlorite</td>
<td>1.093</td>
<td>1.000</td>
<td>0.866</td>
<td>358</td>
</tr>
<tr>
<td>Chlorite</td>
<td>1.287</td>
<td>1.059</td>
<td>0.734</td>
<td>70</td>
</tr>
<tr>
<td>Chlorite</td>
<td>1.128</td>
<td>1.023</td>
<td>0.866</td>
<td>1550</td>
</tr>
<tr>
<td>Chlorite</td>
<td>1.103</td>
<td>1.020</td>
<td>0.921</td>
<td>370</td>
</tr>
<tr>
<td>Biotite</td>
<td>1.114</td>
<td>1.066</td>
<td>0.812</td>
<td>1230</td>
</tr>
<tr>
<td>Biotite</td>
<td>1.098</td>
<td>1.095</td>
<td>0.832</td>
<td>1180</td>
</tr>
<tr>
<td>Biotite</td>
<td>1.107</td>
<td>1.096</td>
<td>0.824</td>
<td>998</td>
</tr>
<tr>
<td>Biotite</td>
<td>1.108</td>
<td>1.107</td>
<td>0.814</td>
<td>1290</td>
</tr>
<tr>
<td>Phlogopite</td>
<td>1.106</td>
<td>1.091</td>
<td>0.838</td>
<td>1140</td>
</tr>
<tr>
<td>Muscovite</td>
<td>1.129</td>
<td>1.072</td>
<td>0.820</td>
<td>165</td>
</tr>
</tbody>
</table>

* From Borradaile et al. (1987) where further details may be found.  
\* Pure specimens, by personal communication through Karel Zapletai from Hrouda et al. (1985a). These specimens also show good agreement between the mean susceptibility reported here and the susceptibility predicted theoretically from the precisely determined compositions.

on anisotropy are expected to be small.) Magnetite is the only mineral of importance in which the grain shape fabric directly influences the magnetic fabric (Uyeda et al., 1963). Thus the susceptibility anisotropy of magnetite will in practice depend on the particular habit of the grains in question and which is nearly equidimensional for detrital examples. For example, detrital magnetite has been reported with an anisotropy of 1.108: 0.964: 0.936 by the author, and Rees (1965) reports another example with susceptibilities in the ratio 1.06: 0.97: 0.97. Hydrothermal magnetites (Heider et al., 1987) can have much higher dimensional anisotropies (e.g. 2:1), as do some magmatic examples. The problems arising from the shape effects of magnetite are compounded by the grouping of grains to form elongate masses. There can be tectonic causes for this, for example where pre-existing magnetites have been strung together to form elongate domains by pressure solution or other forms of metamorphic differentiation. In sediments trains of magnetic grains may be formed by colonial bacteria (Karlin et al., 1987; Chang et al., 1987).

TABLE 2  
A. Weakly susceptible rock-forming minerals (units are SI/vol. x 10^-6)

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Mean ( \kappa )</th>
<th>St. dev.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Plagioclase (oligoclase)</td>
<td>-2.78</td>
<td>0.34</td>
</tr>
<tr>
<td>Quartz</td>
<td>-13.4</td>
<td>0.80</td>
</tr>
<tr>
<td>Calcite</td>
<td>-13.8</td>
<td>0.34</td>
</tr>
<tr>
<td>Talc</td>
<td>+5.4</td>
<td>0.82</td>
</tr>
</tbody>
</table>

B. Anisotropy of weakly susceptible minerals

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Anisotropy ratio, ( k_{\text{max}}/k_{\text{min}} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Quartz</td>
<td>1.00</td>
</tr>
<tr>
<td>Calcite</td>
<td>1.11</td>
</tr>
</tbody>
</table>

Mineralogical sources of magnetic susceptibility in rocks

The magnetic fabric relates primarily to the petrofabric of the significantly susceptible mineral...
TABLE 3
Susceptibility and anisotropy of iron minerals (units are SIF/vol.)

A. Anisotropy controlled by shape (Borradaile et al., 1987)

<table>
<thead>
<tr>
<th>Mineral</th>
<th>$k_{max}$</th>
<th>$k_{int}$</th>
<th>$k_{min}$</th>
<th>Mean value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Magnetite (crushed metamorphic)</td>
<td>1.063</td>
<td>0.989</td>
<td>0.951</td>
<td>5.8</td>
</tr>
<tr>
<td>Magnetite (deritative)</td>
<td>1.108</td>
<td>0.964</td>
<td>0.936</td>
<td></td>
</tr>
</tbody>
</table>

B. Crystallographic anisotropy

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Anisotropy ratio</th>
<th>Maximum susceptibility value</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hematite</td>
<td>&gt; 100:1</td>
<td>$1 \times 10^{-4}$</td>
<td>/basal planes</td>
</tr>
<tr>
<td>Pyrrhotite</td>
<td>c. 10,000:1</td>
<td>1.0</td>
<td>/basal planes</td>
</tr>
<tr>
<td>Pyrrhotite</td>
<td>10.24:3.68:0.027</td>
<td>9.10:3.91:0.028</td>
<td></td>
</tr>
<tr>
<td>Ilmenite-hematite</td>
<td>15:1</td>
<td>0.4</td>
<td>/basal planes</td>
</tr>
</tbody>
</table>

C. Mean values of other iron minerals (Carmichael, 1982)

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Mean susceptibility (approximate)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Magnetite</td>
<td>6.2</td>
</tr>
<tr>
<td>Ilmenite</td>
<td>1.8</td>
</tr>
<tr>
<td>Pyrrhotite</td>
<td>1.5</td>
</tr>
<tr>
<td>Titanomagnetite</td>
<td>0.05</td>
</tr>
<tr>
<td>Hematite</td>
<td>0.007</td>
</tr>
<tr>
<td>Siderite</td>
<td>0.005</td>
</tr>
<tr>
<td>Limonite</td>
<td>0.003</td>
</tr>
<tr>
<td>Pyrite</td>
<td>0.001</td>
</tr>
</tbody>
</table>

Species. If all the minerals in a rock have similar magnitudes of susceptibility, then it should be expected that the magnetic susceptibility ellipsoid bears some close relationship to the gross petrofabric. However, where the susceptibility is chiefly carried by an accessory mineral, the susceptibility ellipsoid will only be representative of the rock's fabric to the degree that the accessory minerals imitate and coeval with the host fabric. Hitherto, most papers on deformed rocks concentrated on rocks in which the accessory iron oxides, particularly magnetite and hematite, were the dominant source of susceptibility. However, the incoming of pyrrhotite at the expense of pyrite during progressive metamorphism may also contribute to the susceptibility of certain greenschist facies metamorphic rocks (Rochette, 1987; Borradaile et al., in press).

Bulk susceptibility values are readily available (International Critical Tables, 1929; Weast and Astle, 1978; Carmichael, 1982), whereas anisotropy data are rarer, beginning with the report of Voight and Kinoshita (1907). The earliest determinations used sufficiently large fragments of single crystals and the same approach continues.
Fig. 2. Susceptibility anisotropies of metamorphic minerals separated from metamorphic rocks (from Borradaile et al., 1987), and the field (stippled) of common metamorphic rocks (from Hrouda et al., 1978). $P'$ and $T$ are explained in the Appendix. $A_l$, $A_2$—actinolite; $B_l$, etc.—biotite; $C_l$, etc.—chlorite; $C_R$—crocidolite; $G$—glaucophane; $H$—hornblende; $M$—muscovite; $P$—phlogopite. Triangle—metamorphic magnetite, partly crushed; triangle in circle—detrital magnetite.

for ore minerals (Uyeda et al., 1963, Schwarz, 1974, 1975; Rochette, 1988), hornblende (Parry, reported in Wagner et al., 1981), augite and calcite (Owens and Bamford, 1976) and quartz (Hrouda, 1986). However, for many rock-forming metamorphic minerals large single crystals are unavailable or have unrepresentative compositions. To overcome this problem samples large enough for determination can be synthesised by orienting small (c. 1 mm) grains into an aggregate (Borradaile et al., 1986). This technique can be applied to any markedly anisometric single-crystal grains that can be physically manipulated or magnetically arranged into a strong preferred crystallographic orientation (Fig. 1). The technique yields a minimum estimate of the anisotropy because it is not possible to produce a perfect alignment of crystallographic axes in this manner. Nevertheless, it is the only practical method at the present time for isolating the contribution of specific mineral grains with fine-grained habits to the susceptibility and anisotropy of susceptibility of a given rock.

Anisotropy data for some important rock-forming minerals mainly from metamorphic rocks, and some bulk anisotropy data for other important minerals are given in Table 1 and Fig. 2.

Contributions to whole-rock susceptibility and its anisotropy

Initially most work on metamorphic or deformed rocks concentrated on examples in which the susceptibility was due to iron oxides. However, several workers have addressed the more general situation in which there may be paramagnetic contributions also (Coward and Whalley, 1979; Henry, 1983; Rochette and Vialon, 1984; Borradaile et al., 1986; Lamarche and Rochette, 1987) or even diamagnetic contributions (Hrouda, 1986). Such effects become important where magnetite is present in small amounts, usually less than 1 wt.\% in a rock that has a matrix of moderately susceptible silicates. Detailed magnetic studies have been undertaken involving magnetic susceptibility determinations at different field strengths (Coward and Whalley, 1979; Hrouda and Jelinek, in press), at different temperatures (Rochette and Vialon, 1984) and the effects of temperature on remanent magnetisation (Kligfield et al., 1982). Such studies have identified the mineralogical source of the susceptibility. However, those techniques are not routine and it may be beneficial to physically separate the minerals in a few typical samples, so that susceptibility measurements may be carried out on these individually.

Henry (1983) and Henry and Daly (1983) inferred multiple sources of susceptibility from the measurements directly, by using multiple specimens of similar lithologies but different intensities of fabric. The method uses a plot of principal susceptibilities of individual specimens against their mean or bulk susceptibility. The data for each of the principal susceptibilities lie on straight lines that intersect at a point that has equal values on both graph axes. This can be considered as a gradual transition in anisotropies from the most anisotropic case at the right of the graph, through lesser anisotropies to a theoretically predictable equivalent isotropic case in which all principal susceptibilities equal the mean susceptibility (Fig. 1).
Fig. 3. Diagram from Henry (1983), showing the variation in anisotropy of adjacent specimens or parts of a specimen. The axes are individual principal susceptibilities (vertical) and the mean or bulk susceptibility (horizontal).

3). This pattern was predicted by Henry's model of additive combinations of ferrimagnetic contributions from traces of magnetite with contributions from paramagnetic silicates. In a natural example, the range in anisotropies on his special plot could be an indication of the degree to which mineralogical constitution affects the magnetic anisotropy of a rock. Other influences cannot be ruled out, however. The variation in mineral composition between different specimens, or the variation in the degree to which the ferrimagnetic grains have similar preferred orientations to the other mineral grains, could also play a role.

Studies of susceptibility at low temperatures and of remanent and saturation magnetisation have confirmed that there are paramagnetic contributions from silicates and that these can swamp the ferrimagnetic contributions (Coward and Whalley, 1979; Henry and Daly, 1983; Rochette and Vialon, 1984). These contributions cannot yet be attributed to a specific mineral without physically separating the grains (Hrouda and Jelinek, in press).

The reverse procedure, predicting whole-rock anisotropy from multiple mineralogical sources is easier. Consider a rock with a perfect preferred crystallographic fabric of mica grains with their basal planes perfectly parallel and their long axes perfectly aligned in that plane. In the simple case of a monomineralic rock with perfectly aligned grains, the mineral anisotropy is equal to the anisotropy of the rock. If susceptibility contributions from different mineral sources can be summed as Henry (1983) and Hrouda (1986) suggest, the effect of adding a small proportion of magnetite to a paramagnetic matrix can be predicted. For simplicity, assume that the magnetite grains are dispersed and present as a small trace (<1%) with a perfect shape alignment coaxial with the preferred crystallographic fabric of the matrix. Although magnetite has an enormous susceptibility magnitude, its anisotropy (directly related to its shape) is usually small, especially if it is of detrital origin. On the other hand, metamorphic silicates have moderate susceptibility magnitudes but very high anisotropies (Table 1). (Whether the susceptibility anisotropy is intrinsic or due to inclusions is immaterial here.)

The magnetic anisotropies of metamorphic silicate grains are more extreme than those of most metamorphic rocks. Thus a trace of magnetite combined with much mica in a schist produces an anisotropy intermediate between that of the minerals. In Fig. 4 this thought experiment is carried out using “model rocks” with 60% plagioclase (a diamagnetic matrix with almost negligible susceptibility), traces of magnetite in the proportions stated on the figure, and about 40% of the named metamorphic silicate. From Fig. 4 it can be seen that the anisotropy of the rock is virtually equal to that of magnetite, if the latter is present as >1%. This is important for those interpretations of magnetic fabrics in terms of “March-type analysis” (see below) in which disseminated markers are virtually solely responsible for the magnetic fabric. But, where the magnetite is present in smaller proportions, the anisotropy of the rock may be influenced more by the composition of the rock than by its petrofabric or strain. Thus, compositional variation between specimens could produce differences in anisotropy even if the specimens had identical fabrics or strains.

Recently, a mathematical model has been investigated in which paramagnetic silicates are the only source of susceptibility (Hrouda, 1987). With increasing preferred crystallographic orientation the anisotropy is shown to increase until it equals that of the crystal lattice. Hrouda notes that this...
Fig. 4. Flinn-type diagram of susceptibility anisotropies for hypothetical rocks indicating the effects of variable proportions of magnetite and a paramagnetic mineral. Each “rock” is represented by a tie-line, and is composed of approximately 40% of the named mineral plus magnetite (in the small fraction indicated by the numbers on the tie-line) and 60% feldspar. As the proportion of magnetite in the rock is increased (from $10^{-6}$ at the distal end of the tie-line in each case), the magnetic anisotropy of the rock decreases towards the anisotropy of magnetite. The anisotropy of the particular magnetite chosen in these models is located near the convergence of the tie-lines (From Borradaile, 1987b.)

The relationship can be disturbed by the smallest traces of ferrimagnetic material and is difficult to calibrate in terms of strain. Moreover, a general difficulty of this type of mathematical model (both in magnetic fabrics and petrofabrics) is that it treats the grains as dispersed crystallites of no finite size which are free to develop their ideal crystal shape. In nature, grains are in contact and in the case of paramagnetic silicates in metamorphic tectonites their preferred orientations pose special problems. For example, grains with crystallographic orientations at high angles to the preferred orientation (schistosity) are restricted in growth. Thus their reduced volumes will decrease their contribution to the rock anisotropy (see Fig. 13b).

Magnetic anisotropy and finite strain

Principal susceptibility directions

In rocks with strong tectonic fabrics the principal magnetic susceptibility directions are often parallel to prominent structural directions (Kligfield et al., 1977) or related to fold geometry (Owens and Bamford, 1976). For example, the minimum susceptibility is usually perpendicular to a primary cleavage. The maximum susceptibility may by parallel to the maximum extension in the rock (e.g., Fig. 6b) contained in the cleavage and defined by stretched objects or mineral lineations (e.g., Singh et al., 1975; Kligfield et al., 1977; Coward and Whalley, 1979; Rathore, 1979; Goldstein, 1980). In contrast the maximum susceptibility may be parallel to the intersection lineation of bedding and cleavage (e.g., Hrouda 1976, also Fig. 5).

These two different patterns have been interpreted in a number of ways. One interpretation is that the maximum susceptibility is parallel always to the maximum extension. It implies that the intersection lineation of bedding and cleavage is the maximum extension direction. This appears to be unusual because an intersection lineation is the fortuitous interference of a tectonic fabric with the disturbed bedding surfaces. In some areas the magnetic fabrics may be reasonably consistent in orientation while the (single-generation) fold hinges and intersection lineations may vary considerably in orientation (Fig. 6). There is no good reason for supposing that the intersection lineation must necessarily have some genetic relation to the strain ellipsoid.

Another interpretation is that maximum susceptibilities occur parallel to intersection lineations because of heterogeneous combinations of subfabrics. If some portions of the rock retain a...
bedding parallel oblate magnetic fabric due to sedimentation then this may “interfere” with the fabric due to tectonic causes. Under the right circumstances, which are not too special, this combination may produce maximum susceptibilities parallel to cleavage and bedding or perpendicular to the stretching direction (Fig. 7). For this explanation to be valid, indeed for this arrangement of susceptibility axis to be detectable, the interfering subfabrics must be present on scales smaller than the sample size used in susceptibility determinations (usually about 10 cm³). Recently, bedding-parallel fabrics have been detected in metasediments with isoclinal sheath folds (Borradaile et al., in press).

A third explanation, similar to the previous one, has been given for rocks that have been very weakly strained with pressure solution as the deformation mechanism. This mechanism often produces a regular, domoidal tectonic fabric of closely spaced laminae which concentrate insoluble material (including iron oxides and paramagnetic silicates) in thin sheets between rock that retains a sedimentary fabric. Where laminae are more closely spaced than the width of the core used to determine susceptibility, the net susceptibility is a combination of sedimentary and tectonic contributions. In this case, the maximum susceptibility is often parallel to the intersection lineation of the domains (“stylolitic cleavage”) and bedding (Fig. 8). Rochette and Vialon (1984) note that heterogeneous fabrics combining magnetic sub-fabrics of different ages prohibit any kind of quantitative correlation of strain and susceptibility.

In mylonite zones (Goldstein, 1980) and in major shear zones (Rathore, 1985) magnetic fabrics have orientations that indicate maximum susceptibilities compatible with maximum extension directions. Similarly, minimum susceptibilities are reported perpendicular to schistosities or foliations. The relationship to boudinage directions and mineral/stretching lineations has been cited to support the contention that there is a one-to-one correspondence of strain axes and principal susceptibility directions, in some cases. Hrouda (1979, reported 1977) seems to have been the first person...
to have considered the combining effects of magnetic fabrics of different ages. He described early susceptibility fabrics by a tensor $P$, superimposed tectonic fabrics by a tensor $D$ and suggested that they combine to form a resultant susceptibility tensor $R$ by:

$$D \cdot P = R \quad \text{or} \quad D = R \cdot P^{-1}$$

Goldstein (1980) applied the same approach to mylonitic fabrics. Some authors have simply assumed that it is permissible to use the orientations of magnetic fabrics within shear zones, rather like a schistosity, to determine the amount and sense of shear. Similar studies have been carried out subsequently on shear zones up to a regional scale in extent (e.g. Rathore, 1985). However, in experimental shear zones the author has found that earlier magnetic fabrics are difficult to overprint and magnetic fabric maybe an unreliable kinematic indicator.

Magnitudes of susceptibility and magnitudes of strain

The relationship between the shape of the magnitude ellipsoid of susceptibility and the intensity of the tectonic fabric is more complex than the relationship between directions, as discussed in the previous section. Usually, comparisons have been attempted between the shapes of the susceptibility and finite strain ellipsoids. The latter is usually derived from some macroscopic strain marker, such as pebbles, concretions, lappilli or rotated sand-dikes (e.g., Rathore, 1979; Borradaile and Tarling, 1981, 1984; Rathore and Henry, 1982; Clendenen et al., 1988; Hirt et al., 1988). Reduction spots have been used as strain markers from time to time; however, their pre-metamorphic origin is not always certain and in the Welsh examples some outcrops have susceptibilities that are too low to give reproducible magnetic fabrics. The susceptibility ellipsoid normally plots much nearer the origin of the Flinn-diagram than does the strain ellipsoid (Fig. 9). In some cases, logarithmic plots (Fig. 10) help in their discrimination. The relationships between strain and magnetic fabric are not straightforward. The most strained samples in an outcrop do not necessarily have the most extreme magnetic anisotropies. This is shown by the crossing tie-lines in Fig. 9. Moreover, there is always some question as to the validity of the comparison since the strain is commonly heterogeneous on the scale of a thin section whereas the homogeneously strained markers are much larger.

Susceptibility is usually measured from a sample about 10 cm$^3$, so that the susceptibility tensor will be defined on a smaller scale than the strain tensor. Another approach, although less common in the literature, is to compare the susceptibility
One of the earliest attempts to relate the strain and susceptibility tensors was by Kneen (1976). She plotted ratios of principal strains against ratios of principal susceptibilities and inferred correlations. Similar studies that compared strain ratios with susceptibility anisotropy used special logarithmic parameters to normalise values (Wood et al., 1976; Rathore, 1979, 1980; Rathore and Henry, 1982; Kligfield et al., 1982). These are termed $N_i$ (for strain) and $M_i$ (for susceptibility), where $i = 1, 2$ or $3$ for the appropriate principal value. The actual algebraic definition of $N$ and $M$ varies in the literature and care should be taken in comparing data from one paper to another. Unfortunately, in most of these cases the authors plotted the data for all three principal directions on the same axes and generally established correlation coefficients for the maximum, intermediate and minimum values simultaneously. Strictly speaking this is invalid (Fig. 11) and in one instance the correlation of anisotropies attempted in this manner has been disproved (Borradaile and Mothersill, 1984).

This may give a discouraging view of the correlations of strain and susceptibility but the data already published do give some indication that correlations may exist, although weaker than originally claimed. For example, comparing data for just minimum (normalised) strain and minimum (normalised) susceptibilities, it is possible to recognise modest correlations between susceptibility and strain in the case of the Welsh slates (e.g. Wood et al., 1976), here in Fig. 12. Clendenen et al. (1988) have recently recognised that this kind of weaker correlation between the strain of concretions and susceptibility can be statistically significant.

Thus although weak correlations of the kind:

$$k_2/k_1 = (X/Y)^n$$

may exist, these must only be of local significance...
and it seems unduly optimistic to consider them as universal relationships that may be extrapolated from one region to another. Within a single region, with a limited range of deformation mechanisms and lithological uniformity, it might be possible to extrapolate some correlations of magnetic fabric with strain.

The approaches reviewed so far in this section are essentially empirical and in most cases the susceptibility was attributed to disseminated magnetite. The studies of Coward and Whalley (1979) and of Henry and Daly (1983) were exceptions: they recognised multiple sources of susceptibility and Kligfield et al. (1982) also investigated the sources thoroughly. The latter authors demonstrated susceptibility anisotropy was due to iron oxides whose distributions and orientations were controlled tectonically by their position on the surfaces of the oolites used as strain markers. The other studies mentioned adopt the premise, or imply, that magnetite grains were distributed throughout the rocks and were tectonically aligned to produce a magnetite shape-fabric which, in turn, dictated the magnetic fabric. The underlying mechanistic model that appears to have influenced this train of thought is attributed to March (1932).

This model (e.g., Wood et al., 1976) outlines the situation in which platy elements, that do not impinge upon one another, and that are preferably randomly oriented in their initial configuration become systematically re-oriented by homogeneous strain and are not impeded by the matrix (Fig. 13a). Consequently, the orientation-distribution of the poles to these platelets after constant-volume strain is directly related to the strain. Thus for a given direction \( i \), the strain \( e \) is related to the normalised density \( d \) of platelet poles by:

\[
e_i = d^{1/n} - 1
\]

In the structural geology literature, some consideration has been given to the aspect ratios of the platelets as well as accompanying volume changes of the aggregate (Tullis, 1976; Tullis and Wood, 1975). However, these modifications still fall short of reality. The March model is difficult to adapt to three-dimensional grains, and even more difficult to reconcile with the premises of passive behaviour and non-impingement of the rotating grains on each other. A worse shortcoming is the inherent assumption that the grains are spatially separated whereas in a typical metamorphic rock the variation in orientation imposes restrictions on the degree to which differently oriented crystals grow (Fig. 13b). Nevertheless, it has found applications in magnetic studies (e.g., Cogné and Gapais, 1986) and studies of comparable mathematical models (Hrouda, 1980), and is frequently cited in petrofabric studies (Oertel, 1983).

Hrouda (1987) has extended these arguments to situations in which it is known that paramagnetic
sheet silicates, such as chlorite and biotite, dominate the anisotropy of susceptibility. An interesting development of this study is that in slates of the Carpathians, in which the anisotropy is due to paramagnetic phyllosilicates, the degree of anisotropy is proportional to the bulk or mean susceptibility. This has also been reported in Archean slates (Borradaile et al., in press).

Strain response models and multiple sources of susceptibility

Owens (1974) first drew attention to two important items that must be taken into account in any interpretation of susceptibility anisotropy in terms of strain. Firstly, the mineralogical source of the susceptibility should be defined. If there are multiple sources of susceptibility the direct application of the simple "March" model or any other model assuming a penetrative homogeneous magnetic response is invalid. Rochette and Vialon (1984) and Henry (1983) enforced this argument in different ways. Slight variations in the proportions of minerals with different magnetic anisotropies may, under certain circumstances, produce changes in the bulk anisotropy that override any changes in anisotropy due to strain (Borradaile, 1987b).

The second major point raised by Owens was that the actual strain response model has a major influence on the way in which the anisotropy accumulates from the pre-strain situation. For simplicity, he considered three models. The first was a "line/plane" model that we can compare to the March model already referred to above (Fig. 13a). The second was the "viscous" or Jeffery model in which the magnetic grain is likened to a rigid ellipsoid undergoing strain-induced motion in a viscous medium. The third "passive" model, likens the magnetic grains to passive strain markers. Structural geologists may compare this to the model proposed by Ramsay's school of structural geology to approximate the shape and consequent orientation changes of, for example, a tectonically deforming, passive, ellipsoidal marker. The shape-ratio of the strain ellipse is 2:1 as indicated. The "line/plane" or March model and where they do not appear to imply any similarity to the $R_r/c_p$ model (e.g. Cogne and Gapais, 1986).

The mechanical response is usually neglected in the correlation of strains and susceptibility anisotropies. Indeed, it is likely that many rocks and many types of deformation mechanisms are unsuitable for establishing a correlation using the premises of simple mathematical models involving a single mode of deformation. For example, if there are relict components of sedimentary magnetic fabrics mixed with the tectonic fabrics, or magnetic fabrics of different tectonic ages, a model assuming homogeneity will fail. Similarly, deformation structures that produce domainal fabrics that differently influence the magnetic fabric will cause complications if the magnetic fabric is sampled on a scale that does not take this into account.

In particulate flow (Fig. 15) the movement of grains may be poorly ordered. Some grains move more than their neighbours, rotate more or even exchange neighbours despite the fact that on a suitably large polygranular scale the strain may be approximately homogeneous. Movement in polygranular units is also possible (Fig. 16) as has been shown by experiments on analog materials (Means, 1977). The scale on which magnetic susceptibility is defined will determine whether or not this presents a problem.

The deformation mechanisms may be still more complicated. The deformation may cause the rock
Fig. 15. Particulate flow of a simulated (two-dimensional) aggregate. Deformation proceeds from (a) through (c). The motions of particulate is somewhat disordered whereas, on a large scale, the strain may appear to be approximately homogeneous as shown by the "strain ellipse" defined by the black particles.

Fig. 16. Experimentally produced particulate flow of groups of alumina grains involving intergranular displacement surfaces. (From Means, 1977.)

Fig. 17. Possible deformation processes in metamorphic rocks that could be significant in the evolution of magnetic fabrics. c—intergranular displacement surfaces (may be invisible); d—ductile (may be plastic) deformation of grains; e—stylolitic surfaces and grain-rotation; f—generation of grains; g—microfaults (cleavage slip); h—rotation of blocks or groups of grains between faults or other discontinuities; m—pressure-solution material transfer (may be out of system); n—new minerals crystallise.

to be a discontinuum on various scales. Lozenges of rock may rotate, mass removal may occur locally or even to outside the system, while grains plastically deform by intercrystalline processes and slide and roll past one another (Fig. 16). At the same time, new minerals may form with their crystallographic orientations controlled by the contemporary stress system. Any of these factors could complicate the strain-response model of the rock and the development of its magnetic fabric. While all of these processes are unlikely to occur at once, the roles of several of these (or others not discussed) have to be assessed before the magnetic fabric can be correlated with the bulk strain. Amongst these complications, probably the most common one is the growth of new minerals as for example in phyllites, schists and gneisses. In these cases, the explanation of schistosity may be bound up with the origin of the metamorphic texture more than with the strain of the rock (Fig. 17) (Borradaile et al., 1986).

Experimental studies

Although there has been some early interest in the effects of stress on magnetic susceptibility (e.g. Nagata, 1966, 1970; Kean et al., 1976), this was associated with strains within the elastic limit. Of interest here is the susceptibility changes that occur with large strains due to deformation that is macroscopically ductile, if not actually plastic at the crystalline level. The problem in the laboratory
environment is that even at the slowest strain rates that are practicable (\(10^{-6} \text{ to } 10^{-9} \text{ s}^{-1}\)) most rocks and minerals are not sufficiently ductile at room temperature to fail by the natural mechanisms of metamorphic rocks.

Owens and Rutter (1978) first tackled this subject with a study of marbles and of calcite. Although the diamagnetic susceptibility of calcite is low, the mineral is relatively easily deformed in a plastic manner at moderate confining pressures. It is also stable at modest experimental temperatures so that Owens and Rutter were able to enhance the ductility by using temperatures up to 500°C in combination with pressures of 150 to 300 MPa (1.5 to 3 kbar) at strain-rates of approximately \(10^{-4} \text{ to } 10^{-3} \text{ s}^{-1}\). They achieved strains up to 50% axial-symmetrical (coaxial) shortening of the cylindrical specimens.

The marbles they used had essentially isotropic magnetic and optical-crystallographic fabrics initially. The magnetic anisotropy rapidly reached a plateau intensity at about 45% shortening. The optical fabrics and the susceptibility fabric rapidly became axisymmetric to the shortening axis of the specimens. This study did much to confirm the value of magnetic fabrics as indicators of strain and kinematics of deformation but most natural studies are based on ferrimagnetic or paramagnetic materials.

Natural paramagnetic and ferrimagnetic materials are difficult to deform in a realistic manner without raising the temperature. Heating will not normally be possible without inducing mineralogical changes that might mask susceptibility changes due to deformation. Borradaile and Alford (1987) approached this problem by using a synthetic aggregate of sand enriched with magnetite that was cemented with commercial Portland cement. The latter has a negligible (diamagnetic) susceptibility in comparison with the aggregate and permitted the aggregate to shorten by 35% in a ductile manner by axial symmetrical shortening. The deformation did not involve cataclasis of the sand grains. Rather, the deformation was accommodated by the shearing of the cement gel. This permitted a grain-shape alignment to develop by rolling and sliding of the grains. The magnetite grains were of the same grain-size as the matrix sand so that these experiments essentially addressed the problem of the rotational behaviour of the clasts. (In most natural sandstones the iron oxide grains are much smaller than the sand clasts.) The experiments were at carefully controlled strain-rates of \(5 \times 10^{-5} \text{ s}^{-1}\) and confining pressures of 150 MPa (1.5 kbar). The maximum and intermediate principal directions of susceptibility of these specimens rotated very rapidly towards the plane of flattening, at rates exceeding the rate of rotation of a line undergoing homogeneous strain (cf. the “line/plane” model). It was inferred that the magnetites rotated rigidly, possibly with some plastic deformation. The rotation was even more rapid when the matrix was composed of calcite instead of silicate mineral grains. For initial susceptibility fabrics that were suitably oriented it was found that the magnetites rotated rigidly, possibly with some plastic deformation. The rotation was even more rapid when the matrix was composed of calcite instead of silicate mineral grains. For initial susceptibility fabrics that were suitably oriented it was found that the change in the degree of anisotropy was related to the strain by a power law (Fig. 18). This serves to emphasise the importance of the initial magnetic fabric. Relatively little can be deduced of the strain history from the final magnetic fabric alone. Just as the initial ellipsoidal shape and orientation is of paramount importance in the strain analysis of ellipsoidal objects so is a knowledge of similar information for the initial magnetic fabric ellipsoid. Although a power law relationship between the change in orientation of specially oriented magnetic fabrics exists in certain axial-symmetrical (“pure shear”) experiments.
Recent work on experimental shear zones of synthetic materials at 150 MPa extends these conclusions to the case of transpressive deformation — a non-coaxial strain history (Borradaile and Alford, in press). The susceptibility ellipsoids rotate rapidly to stable orientations in which the minimum susceptibility is at a high angle to the shear zone boundaries. The magnetic anisotropy changes very rapidly in the shear zones: they produce a stronger magnetic fabric than axial symmetrical shortening at the same strain. This is shown for two typical incremental tests on the same synthetic calcite-matrix material at 150 MPa in Fig. 19. A further complication recognised in the experiments is that the magnetic fabrics may be planar \((F > L)\) whereas the mineral fabrics are constricted \((a > b)\) or plane strain \((a = b)\). It is not known if these observations are transferable to natural shear zones.

Conclusions

The interpretation of natural magnetic fabrics may be more complex than was initially recognised but progressive changes in magnetic anisotropy do occur sympathetically with regional strain gradients, despite accompanying regional metamorphism (e.g., Hrouda and Janak, 1976; Hrouda, 1976; Coward and Whalley, 1979). Indeed, Rochette (1987) has shown that the metamorphic production of pyrrhotite can enhance the susceptibility and produce a magnetically detectable isograd. Also, in certain rocks, there are at least weak correlations between ratios of principal strains and the ratios of principal susceptibilities. Numerous examples of the parallelism of principal susceptibility directions and structural directions (or even principal strain directions) support the contention that genetic relationships do exist between strain and susceptibility in some cases. Nevertheless, these correlations are not universal and it is not possible to predict which natural situations will yield susceptibility data that will faithfully represent principal strain directions, or principal strain magnitudes, or both. For example, magnetic fabrics do not always seem to be reliable kinematic indicators for shear zones and in the case of polyphase deformation later magnetic fabrics do not uniformly overprint earlier ones. Magnetic fabrics should not be used for routine methods of "strain analysis" without further study. A more profitable avenue is to determine how the susceptibility and strain tensors relate, taking account of the actual strain-response model (deformation mechanisms), and the actual mineralogical sources of susceptibility in that rock, including phases produced by metamorphism. A fruitful approach would be to consider the tensorial combination of coexisting magnetic subfabrics and successive episodes of magnetic fabric development along the lines initiated by Hrouda (1979).

Relationships between susceptibility and strain are in vogue. However, the earliest uses of magnetic studies simply to investigate "bulk" petrofabrics are not yet been fully developed or integrated with detailed petrofabric studies by optical or X-ray means. Furthermore, studies of remanence and of susceptibility at different field strengths can yield information on magnetic subfabrics of different grain-sizes or different mineralogy. This information is essential in the study of metamorphism of magnetic phases as well as their preferred orientation. The anisotropy of remanence is also a profitable area of study.
because only the ferrimagnetic traces contribute to that sub-fabric.

Finally, the nondestructive nature of magnetic fabric analysis provides a tool for investigating the progressive deformation or progressive metamorphism of a single specimen in laboratory experiments. Experiments on synthetic aggregates indicate that the initial anisotropy of materials is not easily overprinted by small strains at low temperature. They demonstrate also that the magnetic fabric ellipsoid spins very rapidly where the chief magnetic mineral is magnetite. This is more noticeable when the matrix grains undergo crystal-plastic deformation. Experimental shear zones produce greater changes in magnetic shape fabrics than axial-symmetrical shortening ("pure shear") at the same strain: they can also produce magnetic fabrics differing in symmetry from the grain fabric. Such results develop an awareness for the greater complexities of natural structures.

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Postscript

Since this paper was accepted for publication P. Rochette has drawn attention to unusual, "inverse" magnetic fabrics probably produced by single-domain magnetite. (Thesis, University of Grenoble, 1988, 211 pp.)

APPENDIX

Presentation of anisotropy of susceptibility

Magnetic susceptibility is a second rank tensor property, conveniently envisaged as a magnitude ellipsoid (see Hrouda, 1982, p. 38) with three principal axes: \( k_{max} \), \( k_{int} \) and \( k_{min} \) or \( k_{11} \), \( k_{22} \), \( k_{33} \). The average or bulk susceptibility is commonly defined as the arithmetic mean of the three principal values (Ellwood et al., 1986—Symposium recommendations), but it is not clear why the arithmetic mean is a more appropriate representation of bulk susceptibility than the geometric mean. Susceptibility anisotropies for rocks are so low (Figs. 9 and 10) that there is often little difference between these two expressions. For example, seventeen specimens of strongly strained fine schist of the English Borrowdale Volcanic Series have principal susceptibilities in the ratio 1.04:1.02:0.943.

Hrouda (1982) lists 25 different parameters that have been used at various times to represent anisotropy. The most useful are sometimes termed the "magnetic lineation" and the "magnetic foliation":

\[
L = \frac{k_{max}}{k_{int}} \quad \text{and} \quad F = \frac{k_{int}}{k_{min}}
\]

Note that some authors chose alternative definitions for \( L \) and \( F \) (e.g. Coward, 1976). The parameters \( L \) and \( F \) are plotted vertically and horizontally respectively, and compare directly with the ratios used to describe the strain ellipsoid on a Flinn-diagram:

\[
a = \frac{X}{Y} \quad \text{and} \quad b = \frac{Y}{Z}
\]

where \( X \geq Y \geq Z \) are the principal strains. In some instances it is clearer if the natural logarithms of \( L, F, a \) and \( b \) are plotted.

If the ratios of principal susceptibilities are listed individually, it is helpful to normalise them so that their product is unity. The normalised versions of the principal susceptibilities are related to their ratios by the following relations:

\[
k_{max} = L^{1/3} \cdot F^{1/3} \quad k_{int} = L^{-1/3} \cdot F^{1/3} \quad k_{min} = L^{-1/3} \cdot F^{-2/3}
\]

It is advisable also to present a mean (bulk) susceptibility value. This enables the reader to assess the significance of the results (Conference recommendations, Ellwood et al., 1986). Unfortunately this has often not been done in the past. Commonly one finds that anisotropies are poorly reproducible for rocks with susceptibility < 10⁻⁴ SI/vol. Providing cones of confidence about principal directions and standard errors for principal magnitudes is also useful in this regard.
Fig. A1. The relationship between Jelinek’s (1981) $P'$ and $T$ parameters and the conventional anisotropy parameters $L$ and $F$.

The ratios $L$ and $F$, like their counterparts in strain ($a$ and $b$) are very useful in describing the ellipsoid’s shape or sense of anisotropy. However, the distance of the point $(L, F)$ from the origin is also a measure of the degree of anisotropy, that is to say, its departure from sphericity. Jelinek’s (1981) parameters $p'$ and $T$ overcome this. $T$ represents sense of anisotropy and ranges from $-1$ (prolate) to $+1$ (oblate). $p’ (> 1)$ expresses degree of anisotropy. These are defined as:

$$T = \frac{2(\ln k_{max} - \ln k_{min})}{\ln k_{max} - \ln k_{int}} - 1$$

$$P’ = \exp \left[ 2 \left( a_1^2 + a_2^2 + a_3^2 \right) \right]$$

where $a_1 = \ln(k_{max}/k_{ref})$, etc. and $k_{ref} = (k_{max} \cdot k_{int} \cdot k_{min})^{(1/3)}$. The relationship of $P'$ and $T$ to $L$ and $F$ is shown in Fig. A1.

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