Metamorphic control of magnetic susceptibility and magnetic fabrics: a 3-D projection

NORIHIRO NAKAMURA¹ & GRAHAM J. BORRADAILE²

¹Department of Geo-Environmental Science, Tohoku University, Sendai 980-8578, Japan (e-mail: n-naka@mail.tains.tohoku.ac.jp)
²Geology & Physics Dept., Lakehead University, Thunder Bay, P7B 5E1, Canada (e-mail: gjborrad@lakeheadu.ca)

Abstract: Magnetic fabric changes due to progressive metamorphism are poorly understood. Bulk magnetic susceptibility (κ) is known to increase with metamorphic grade but anisotropy changes have been neglected. To combine information on anisotropy with bulk susceptibility, we introduce a projection with three axes: κ, ellipsoidal eccentricity (Pₚ, the so-called 'anisotropy degree', despite the fact that this is quantified) and ellipsoid symmetry (Tₛ) as independent variables. The projection reveals that metamorphic fabrics can be discriminated successfully in the 3-D projection, with distinct, significant regression surfaces for crustal metamorphic rocks metamorphosed successively in greenschist, amphibolite, and granulite facies. This emphasizes that bulk magnetic susceptibility (κ) and its anisotropies ('magnetic fabric') evolve in response to metamorphic process, not just strain. Moreover, post-tectonic granitic plutons, upper mantle harzburgites and serpentinitized mantle rocks also have characteristic regression surfaces relating κ, Pₚ and Tₛ in the new projection.

Metamorphic processes are solid-state transformations of minerals and textures of a pre-existing rock due to changes in temperature, pressure, fluid-rock interaction, strain and recrystallization (e.g., Miyashiro 1994). Generally, magnetic susceptibility (κ) may change with metamorphic grade from greenschist to granulite facies with depth (Hrouda 1982; Shive et al. 1988, Jackson & Tauxe 1991). Under some circumstances, 'magnetic isograds' may be identified, for example, in a zeolite to amphibolite facies progression recorded in lithologically monotonous black shale (Rochette 1987). The effect of metamorphic grade on magnetic susceptibility may reflect, in the broadest sense, a generalized metamorphic cross-section of the continental crust, as well as lithologically homogeneous deep crustal and upper mantle rocks. The bulk susceptibility (κ), anisotropy of low field susceptibility (AMS) and other forms of a magnetic anisotropy are commonly used as tools to investigate the petrofabric, mainly of tectonically deformed rocks (Hrouda 1982; King et al. 1982; Henry 1983; Jackson & Tauxe 1991; Rochette et al. 1992; Borradaile & Henry 1997). Whereas most AMS studies concern tectonically deformed rocks, the relative roles of strain and metamorphic recrystallization are not readily distinguishable. To discern multivariate dependent relationships among magnetic anisotropy and bulk susceptibility due to different metamorphic facies, we examine four distinct protoliths in the common facies series from greenschist through amphibolite to granulite facies. The protoliths considered include:

(1) Archaean greywacke, deposited in an accretionary prism (greenschist through granulite facies)
(2) Oceanic mantle harzburgites (Troodos ophiolite)
(3) Serpentinitized oceanic mantle (Troodos ophiolite)
(4) Post-tectonic I-type granite

AMS may be represented by a magnitude-ellipsoid whose anisotropy parameters are most effectively described by the eccentricity Pₚ and its shape Tₛ (Jelinek 1981). The Pₚ, Tₛ plot represents AMS more symmetrically than the Flinn-plot used to discriminate strain ellipsoids and fabric types in structural geology (Borradaile 2003a). Earlier studies tended to attribute the preferred orientations detected by AMS to finite strain. Principally, the shape parameter (Tₛ) and eccentricity (Pₚ) of the magnitude ellipsoid (Jelinek 1981), and the mean or bulk susceptibility (κ) can be used to infer the magnetic mineralogy, for example, the role of paramagnetic minerals versus remanence-bearing minerals, grain-size variations, and their relative contributions of their orientation distributions to the AMS. However, in most tectonically deformed rocks and many plutonic, nominally igneous rocks, both the paramagnetic silicates and the iron oxides have recrystallized (Housen et al. 1993). This may only affect the orientation-distribution but it may equally produce a

new mineral assemblage with a different partitioning of $\kappa$ among the magnetically anisotropic minerals. The prominent control on $\kappa$ is the concentration of accessory magnetite due to its enormous susceptibility, (e.g., common multidomain (MD) magnetite has $\kappa \approx 3.1$ SI; Heider et al. 1996) whereas the most susceptible rock-forming pure silicates have values $<0.002$ SI.

Although magnetite’s grain-shape-controlled anisotropy degree ($P_j$) may be low, its contribution to the rock-fabric may be more significant than the much more anisotropic, but less susceptible, rock-forming silicates (Borradaile 1987). Although the paramagnetic matrix minerals of interest (chlorites, serpentine, micas, biotite, amphibole and pyroxene) show relatively low intrinsic bulk susceptibility, they may be contaminated by single-domain (SD) or pseudo-single-domain (PSD) magnetite as inclusions or exsolution grains. For pure paramagnetic silicates, $(\kappa)$ ranges from 50 $\mu$SI to a theoretical maximum of 2000 $\mu$SI (Syono 1960; Droop 1987) but their anisotropy is much higher than magnetite, and controlled by crystal symmetry.

![Diagram](image)

**Fig. 1.** Traditional 2-D Jelinek plot and $P_j$-$\kappa$ plot of metamorphic rocks from greenschist-facies through amphibolite-facies to granulite-facies may not always show effective visual discriminations due to complete overlap of the data.
not by grain-shape. Moreover, magnetite inclusions in paramagnetic minerals may increase bulk susceptibility to 10000 µSI (Borradaile 1994; Borradaile & Werner 1994; Lagroix & Borradaile 2000). However, in those cases, the orientation distribution of the inclusions is usually controlled by the host silicate lattice, thus reinforcing the matrix-fabric. This gives some insight, into the complications that metamorphism may cause, shifting the control on anisotropy between iron-oxide accessories and matrix silicates, according to mineral reactions during progressive metamorphism.

On the traditional 2-D Jelinek plot, the initial example of metamorphic rocks shows almost complete overlap of greenschist-facies, amphibolite-facies and granulite-facies data (Fig. 1a). This traditional $P_T - T_j$ plot also lacks information about bulk susceptibility that can discriminate the fabric-shape patterns with respect to $k$. Although the influences of metamorphic facies on $k$ may be evident (Fig. 1b), AMS is also influenced by facies. Metamorphic control of magnetic fabrics is presented on a 3-D projection for the common progressive metamorphic facies series (greenschist-amphibolite-granulite), in terms of $k$, $P$, and $T_j$. This is really a simplified approach to linear discriminant analysis; we choose a suitable viewing axis in coordinate-space such that clusters of data are distinguishable on the basis of some attribute, here metamorphic facies. Furthermore, most of our data groups (~ metamorphic facies) may be characterized by multiple regression surfaces in the same program used for three-dimensional viewing of data (in our case SigmaPlot 8.0 © software).

**Magnetic petrology**

Our AMS data is from published sources for Archaean metasedimentary rocks from sub-garnet greenschist-facies slates ($n = 224$) through amphibolite-facies schists ($n = 193$) to granulite-facies lower crustal gneisses ($n = 258$) and late Archaean granitoids ($n = 129$). All of these rocks consist of a homogeneously metamorphosed similar protolith, representative of mean-continental crustal chemistry: greywacke. Also mantle harzburgites and serpentinite samples ($n = 457$) from the Troodos ophiolite, Cyprus are used to represent metamorphosed oceanic upper mantle. These examples outline facies control on $k$ and AMS for typical continental crust and oceanic upper mantle.

The greenschist-facies slates consist of Archaean meta-greywacke subjected to one penetrative deformation accompanied by a low-grade metamorphism (Borradaile & Sarvas 1990). Magnetic susceptibility is dominated by ferromagnetic pyrrhotite and paramagnetic chlorite (thuringite), and biotite. A steep metamorphic gradient gives a smooth transition through to amphibolite facies rocks with an extensive growth of pyrrhotite (Werner & Borradaile 1996).

The granulite-facies gneiss is an Archaean lower crustal granoblastic rock in the Kapuskasing Structural Zone formed at pressures between 0.7 and 1.0 GPa in the temperature range 650–750 °C (Percival 1983; Bursnall et al. 1994). The gneiss possesses poorly defined ‘visible’ petrofabric but well-defined magnetic fabrics (Borradaile et al. 1999). Major sources of susceptibility are free-MD magnetite, as accessory grains, and PSD magnetite inclusions in pyroxene host grains. The magnetite inclusions in mafic silicates occur as polycrystalline submillimetric veinslets. Paramagnetic mafic silicates, pyroxene, biotite and hornblende provide a similar combination to bulk susceptibility but show much higher anisotropy ($P_j$) than magnetite.

Plutonic I-type granites are represented by a sample suite of late Archaean post-tectonic granitic plutons (Trout Lake and Burnum Lake plutons) in north-western Ontario (Borradaile & Kehlenbeck 1996). Susceptibility ($k$) is controlled dominantly by MD accessory magnetite. These plutons have a bimodal frequency distribution of bulk susceptibility. The less common lower susceptibilities might be due to more extensive oxidation of magnetite or a reduced magnetite content so we have removed them from consideration. In these plutonic granitoids, AMS is dominated by the stress-induced preferred arrangement of significant magnetic domains within equidimensional multidomain magnetite, since there is no visible evidence of their preferred grain-shape alignment.

The data sets of mantle rocks are serpentinized harzburgites from the Troodos ophiolite, Cyprus (Borradaile & Lagroix 2001). Although serpentine is one of the most susceptible rocks, this is largely due to the role of abundant (‘accessory’) magnetite, a by-product of serpentinization (Rochette 1994; Dunlop & Özdemir 1997; Kido et al. 2001). Their magnetite inclusions within mafic silicates and the free-MD accessory magnetite produced by serpentinization dominate $k$. More than 50% of the bulk susceptibility is attributed to PSD magnetite inclusions in mafic silicates (Borradaile & Lagroix 2001). This highly susceptible PSD magnetite fabric camouflage the more anisotropic, but less susceptible mafic silicate fabric.
Fig. 2. Regression surfaces for magnetic properties of selected important protoliths in 3-D space, viewed along a suitable axis. Bulk susceptibility ($\kappa$); eccentricity $1 < P_j < \infty$ (sphere to ellipse) and symmetry, $-1 < T_j < -1$ (oblate to prolate). Note logarithmic scale for $\kappa$.

Multiple regression magnetic fabric parameters

A 3-D projection represents the metamorphic control of magnetic fabrics under progressive metamorphism with multiple linear regression surfaces: $\kappa = aP_j + bT_j + c$ (Fig. 2). The logarithmic axis facilitates the plotting of $\kappa$ due to its large range. We choose whatever orientation of axes of 3-D plot best reveals the relationship between $\kappa$, $P_j$, $T_j$ and metamorphic grade.

Regression permits statistical comparison of fabric-shapes by using Jelinek’s parameters of a symmetrical shape description: $T_j$ (oblate to prolate: $+1$ to $-1$) and an ellipsoidal eccentricity: $P_j$ (sphere to ellipsoid: $1$ to $\infty$) which includes reference to the intermediate value. The multiple regression surfaces are distinct and statistically significant at the 95% confidence level (Table 1). The test-statistics were used to determine whether the correlation coefficients ($R$) were significantly non-zero at the 95% level (see

<table>
<thead>
<tr>
<th>Rock type</th>
<th>$n$</th>
<th>$R$</th>
<th>Test-statistic ($R\sqrt{n}$)</th>
<th>$a$</th>
<th>$b$</th>
<th>$c$</th>
<th>$&gt;t_{0.05/2}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Greenschist-facies schist</td>
<td>193</td>
<td>0.39</td>
<td>5.42</td>
<td>954.5</td>
<td>-394.2</td>
<td>-644.5</td>
<td>OK</td>
</tr>
<tr>
<td>Amphibolite-facies schist</td>
<td>258</td>
<td>0.57</td>
<td>9.16</td>
<td>31396.5</td>
<td>-2442.3</td>
<td>-32898.2</td>
<td>OK</td>
</tr>
<tr>
<td>Granite-facies gneiss</td>
<td>94</td>
<td>0.24</td>
<td>2.33</td>
<td>30217.0</td>
<td>-1321.2</td>
<td>-7123.7</td>
<td>OK</td>
</tr>
<tr>
<td>I-type granite</td>
<td>436</td>
<td>0.34</td>
<td>7.10</td>
<td>19469.0</td>
<td>-1103.8</td>
<td>-17585.3</td>
<td>OK</td>
</tr>
<tr>
<td>Harzburgite</td>
<td>150</td>
<td>0.28</td>
<td>3.43</td>
<td>-2496.3</td>
<td>-8434.8</td>
<td>24608.1</td>
<td>OK</td>
</tr>
</tbody>
</table>

The multiple regression is defined by a linear equation: $\hat{\kappa} = aP_j + bT_j - c(P_j \geq 1; -1 \leq T_j \leq 1)$. The significance of the correlation coefficient is determined by comparing the test-statistic, ($R\sqrt{n}$), with $t_{0.05/2}$ according to conventional distribution-theory statistics (Borradaile 2003a, b). OK = Significant correlation surfaces at the 95% confidence level are indicated by ‘OK’.

References:
1. Archean Quetico belt, Canada (Borradaile & Servas 1991).
2. Archean Quetico belt, Canada (Werner & Borradaile 1996).
3. Kapuskasing Structural Zone (Borradaile et al. 1999).
4. Rainy and Barnum Lake, Canada (Borradaile & Kehlenbeck 1996).
5. Troodos ophiolite, Cyprus (Borradaile & Lagroix 2001).
6. Troodos ophiolite, Cyprus (Borradaile & Lagroix 2001).
Table 1, Figs 3 & 4). Although the correlation coefficients may be relatively low (Table 1), the regression surfaces are significant at the 95% confidence level, because the test-statistics \( |R|/\sqrt{n} \) exceeds 1.96 and \( n \) is a large sample-size (Borradaile 2003a,b). Formally, we should not reject the hypothesis that the regression surfaces provide a faithful generalization of the magnetic parameters with metamorphic grade. Although some data sets show considerable dispersion, particularly with regard to \( T_J \) (ellipsoid shape), these are notable differences of the multiple regression surfaces with respect to the \( P_J - \log \kappa \) axes (see Figs 3 & 4). The graphs show regression lines and 95% confidence envelopes following classical, distribution-theory statistics. This emphasizes a significant discrimination from the 2-D Jelinek plot by employing bulk susceptibility, avoiding an overlap of \( P_J - T_J \) data.

Figure 3a shows that the AMS data \((n = 224)\) in greenschist facies slates lie predominantly in the oblate field \((T_J > 0)\) and the regression surface indicates the \( P_J - k \) correlation (Borradaile & Sarvas 1990; Rochette et al. 1992; Borradaile & Henry 1997). In amphibolite facies schists, \( P_J \) value is less anisotropic and AMS symmetry shows more neutral ellipsoid shapes \((T_J \to 0)\), which may be due to the growth of stubby biotite at the expense of well-aligned chlorite in greenschist facies grade (Fig. 3b). Moreover, bulk susceptibility, \( \kappa \), in amphibolite facies is

---

**Fig. 3.** Traditional \( P_J - \kappa \) plot with 95% confidence envelopes (dotted lines) for the regression line (solid lines), new 3-D projection and Jelinek plot: (a) greenschist-facies slate. (b) amphibolite-facies schist. (c) granulite-facies gneiss. Note that sample size is \((n)\) and regression coefficient is \((R)\). All 3-D regression surfaces imply correlations significantly different from zero at the 95% level.
Fig. 4. Traditional $P_T$-$\kappa$ plot, with 95% confidence envelopes (dotted lines) for the regression line, new 3-D projection and Jelinek plot: (a) granite and oxidized granite, (b) harzburgite and serpentine. Note that sample size is $n$ and regression coefficient is $(R)$. All 3-D regression surfaces imply correlations significantly different from zero at the 95% level.
relatively greater than in the greenschist facies. Each multiple regression surface discriminates the differences in bulk susceptibility and in AMS (see Fig. 2, Figs 3a & 3b).

The lower continental crustal section exposed at Kapuskasing, Ontario (Percival & West 1994) shows granulite-facies metamorphism. At this high metamorphic grade, bulk susceptibility is higher but also $P_t$ drastically increases more rapidly with increasing $k$ (Fig. 3c). The fabric symmetry $T_j$ evolves from oblate for low $k$ specimens to neutral or prolate at higher susceptibilities. This may be due to a nucleation of needle-like veinlet magnetite inclusions in pyroxene host minerals during retrograde metamorphism.

In northern Ontario, certain late Archaean granitic plutons intruded forcefully and post-tectonically, and their feldspar megacryst fabric is clearly magmatic or, at least related to late-magmatic inflation. The granites' bulk susceptibilities are high (Fig. 2, Fig. 4a), classifying them as magnetite-series granitoid (Ishihara 1979). The distribution of these granites in $P_t$, $T_j$, $k$ space is relatively insensitive to anisotropy. $P_t$ or $T_j$, although data still show distinct anisotropic fabrics. These anisotropies may be due to a stress-controlled alignment of intra-granular domain walls in multidomain magnetite since there is no obvious shape alignment of grains (Borradaile & Kehlenbeck 1996), despite relatively constant susceptibility (see also Fig. 4a). These high susceptible post-tectonic granitic plutons are responsible for high induced magnetization in Earth’s magnetic field, which is reflected in the aeromagnetic anomaly. The lower susceptible oxidized granite is less anisotropic but bulk susceptibility is sensitive to $P_t$.

The Troodos harzburgite exposes oceanic upper mantle rocks, and shows high-temperature, solid-state flow textures of the serpentinitized and retrogressed mafic silicates. The regression surface for harzburgite lies close to the $T_j$, $k$ plane because anisotropy degree ($P_t$) increases slowly with $k$ (Fig. 4b). The shape ($T_j$) is insensitive to bulk susceptibility. It varies from oblate due to the S-fabric and intrinsic crystallographic-symmetry of the mafic minerals, to prolate in specimens in which magnetite dominates (highest $k$). The eccentricity $P_t$ is lower, perhaps due to aligned PSD magnetite inclusions within host-silicates, which may be controlled by their host-crystal (Borradaile & Lagroix 2001). Serpentinitized harzburgites are less anisotropic but of significantly higher susceptibility (Fig. 4b), due to MD magnetite as a by-product of serpentinitization, although some lower susceptible serpentinized harzburgites data are plotted in similar region as harzburgite. Data indicates that $T_j$ dispenses broadly from +1 (oblate) to -1 (prolate) and is independent of $k$.

Summary

Some earlier studies imply that the magnetic susceptibility and magnetic anisotropy of rocks did not depend on metamorphism (Williams et al. 1985; Shive & Fountain 1988; Werner & Borradaile 1996). However, their sampling may not have involved uniform protoliths, nor a wide enough range of metamorphic conditions. We sampled lithologically distinct protoliths metamorphosed under very different metamorphic facies, from greenschist to granulite for continental crust and certain granites, and for oceanic upper mantle. Three low-field magnetic parameters, ($k$, $P_t$, $T_j$), provide a successful visual discrimination of greywackes metamorphosed in greenschist/amphibolite/granulite facies. They also characterize upper mantle rocks, their serpentinitized equivalent and I-type granite. The success of discrimination in this new three-dimensional projection testifies to the fact that metamorphic facies is at least as important as strain in controlling the magnetic fabric. This is corroborated and quantified by multiple regression statistics.

We thank M. Jackson and B. Housen for their helpful review. We are grateful to T. Werner for valuable review comments and for providing AMS data from an Archaean greenstone belt. The National Sciences and Engineering Research Council of Canada funded GB. N. N. is grateful to the JSPS postdoctoral fellowships for research abroad. This work was partly supported by the 21st century Center-Of-Excellence (COE) program (Earth Sciences) of Tohoku University.

References


