\( K = \ln(a)/\ln(b) \) overcame that, with the advantage that strains combine as a product (Fig. 10b). Ramsay's \( K \) is no more advantageous than Flinn's but it relates more simply to a modern magnetic fabric parameter (below). Ramsay's convention \( X \geq Y \geq Z \) for the principal stretches (e.g., \( X = L_{\text{NEW}}/L_{\text{OLD}} \)) is now universally adopted whereas Flinn's pioneering studies (1962, 1965a,b) used the optical mineralogy convention \( (Z \geq Y \geq X) \). Early strain analysis described the shape of finite and incremental strain ellipsoids with Lode's parameter, simplified here by substituting recognizable Flinn ratios \((b,a)\).

\[
\nu = \frac{\ln(b) - \ln(a)}{\ln(b) + \ln(a)} \tag{38}
\]

(Lode 1926, Nadai 1963). A few other shape descriptors still find occasional use (Khan 1962):

\( L \) (lineation) = \( \kappa_{\text{MAX}}/\kappa_{\text{INT}} \) analogous to Flinn's \( \alpha \); not to be confused with 'L' in his qualitative field-fabric L-S scheme used by structural geologists (Fig. 12a).

\( F \) (foliation) = \( \kappa_{\text{INT}}/\kappa_{\text{MIN}} \) analogous to Flinn's \( b \).

These are quantifiable ratios; confusingly, in structural geology 'lineation' and 'foliation' are fabric elements, or their orientations. Jelinek (1981) introduced a shape parameter \( T \), defined identically to Lode's parameter \( (\nu) \) of continuum mechanics but for our purposes easily remembered in terms of \( F \) and \( L \):

\[
T = \frac{\ln(F) - \ln(L)}{\ln(F) + \ln(L)} \tag{39}
\]

Like Lode's parameter, and in contrast to Flinn's \( k \), it ranges symmetrically over the spectrum of ellipsoid shapes: \( T = +1 \) for oblate, \( T = -1 \) for prolate and \( +1 > T > -1 \) for the general ellipsoid (Fig. 10c). Jelinek's \( T \) and Ramsay's shape parameter \( K \) may be simply related since Ramsay's plot uses logarithmic space:

\[
T = 1 - \frac{1}{K} \tag{40}
\]

where \( K = \ln(a)/\ln(b) \) or \( K = \ln(L)/\ln(F) \).

The distance of the point from the origin \((b = 1; a = 1)\) represents some aspect of total strain/​anisotropy, e.g. using Ramsay's logarithmic plot:

\[
\sqrt{(\ln(a - 1))^2 + (\ln(b - 1))^2} \quad a \neq 1 \quad b \neq 1 \tag{41}
\]

Such concepts are application-dependent; e.g. in mechanics different finite strain states may have been achieved by equal work and the logarithmic shear strain parameter \((\bar{\varepsilon_s}, \text{eq. (44) below})\) may preserve that equivalence. In rock magnetism this quantity is mysteriously termed the anisotropy degree although it is readily quantifiable. Nagata's (1961) simple anisotropy degree is:

\[
P = \frac{\kappa_{\text{MAX}}}{\kappa_{\text{MIN}}} \tag{42}
\]

A 'corrected' (sic) anisotropy degree \( P' \) or \( P_j \) (Jelinek 1981) is now widely accepted, not least because it includes a reference to \( \kappa_{\text{INT}} \); \( P_j \) normalizes the principal susceptibilities to their mean. Jelinek used the arithmetic mean. However the geometric mean may be wiser, especially for large anisotropies, e.g. finite strain or AARM:

\[
\ln(P_j) = \sqrt{2} \left( \ln \left( \frac{\kappa_{\text{MAX}}}{\kappa} \right) + \ln \left( \frac{\kappa_{\text{INT}}}{\kappa} \right) \right)^2 \tag{43}
\]

It is similarly designed to the logarithmic shear strain parameter, \( \bar{\varepsilon_s} \), of continuum mechanics.

\[
\bar{\varepsilon}_s = \frac{1}{\sqrt{3}} \left[ \ln \left( \frac{X}{Y} \right)^2 + \ln \left( \frac{Y}{Z} \right)^2 + \ln \left( \frac{Z}{X} \right)^2 \right]^{1/2} \tag{44}
\]

(Nadai 1963), which usefully considers the three possible principal axial ratios, although its principal values are inconsistently normalized \( P_j \) has numerous advantages for us (and for structural geology). Moreover, it may be retrieved simply from Nagata's older \( P \)-values with

\[
\ln(P_j) = \ln(P) \sqrt{1 + \frac{T^2}{3}} \tag{45}
\]

Jelinek's \((P_j, T)\) plot assigns eccentricity and shape uniquely and conveniently to different Cartesian axes. However, for near isotropic eccentricities \((P_j \rightarrow 1)\), slight shape differences \( (T) \) appear just as significant as very large shape-differences at higher \( P_j \) (Fig. 10c). In geological strain analysis a superior but under-used annular plot shows shape \( (\nu) \) along 60° arcs versus eccentricity \( (\bar{\varepsilon}_s) \) as radii (Hossack 1967; Ramsay & Huber 1983). We introduce a \( \pi/4 \)-segment polar-plot, on which \( P_j \) is radius and T is arc-length; this representation has further advantages for strain and AMS. For example, shapes \((T) \) are not dispersed for low \( P_j \), and near-isotropic states plot unambiguously close to the origin. For AMS, the radial structure is easily reflected to plot diamagnetic anisotropies. The negative field also facilitates the re-plotting of AMS for counterintuitive inverse petrofabrics due to intrinsic inverse AMS in minerals such as.
calcite, SD magnetite and tourmaline (Rochette 1988; Rochette et al. 1992, 1999).

Our polar plot is just as beneficial with ODs and samples, as with individual specimen ellipsoids. Weak ODs cluster unfavourably near the origin of the Flinn plot (Woodcock 1977) or disperse unnaturally along the T axis of the Jelinek plot (Fig. 11a); near isotropic ODs cluster close to the origin on the polar plot (Fig. 11b). This is illustrated with a sample-suite of AMS tensors, for Archaean greywackes (data of Werner & Borradaile 1996) whose misleading dispersion in Cartesian coordinates (Fig. 11c) becomes more meaningful in polar projection (Fig. 11d).

Rock composition, bulk susceptibility (κ) and AMS ellipsoid-shape

It cannot be stressed too strongly that AMS ellipsoid shapes rarely correlate with finite strain magnitudes (Borradaile 1991) or with a causal
quantity. In contrast, the principal AMS axial orientations usually show some causal relationship with a geological process or event. Some elementary observations serve to illustrate this point; consider a suite of Archaean upper greenschist facies greywackes with mean low field susceptibility ($\kappa$) mostly controlled by biotite and pyrrhotite, with a somewhat bimodal frequency distribution, Figure 12a (Borradaile et al. 1988, 1993; Borradaile & Sarvas 1990; Borradaile & Spark 1991; Werner & Borradaile 1996). Simple comparison of the eccentricity

![Graph](image)

**Fig. 12.** AMS magnitude ellipsoid shape is dictated primarily by rock composition, expressed partly in terms of $\kappa$. (a) Bimodal frequency distribution of $\kappa$ for Archaean greenschists, in which biotite and pyrrhotite mainly control $\kappa$ (Borradaile et al. 1988, 1993; Borradaile & Sarvas 1990; Borradaile & Spark, 1991; Werner & Borradaile 1996). (b) For the same specimens, $P_j$ (eccentricity of AMS ellipsoid) is controlled by $\kappa$ (i.e. rock composition) for both high and low-$\kappa$ subgroups. Both correlations are significant at the 95% level, due to favourable sample-sizes. The high-$\kappa$ sub-sample represents pyrrhotite-rich specimens. (c) The symmetry or shape ($T$) of AMS ellipsoids as well as $P_j$, may be more influenced by rock-type and mineralogy than by metamorphism, strain or other secondary processes (some of data from Jackson & Borradaile 1991).
(\(P_j\)) of the AMS ellipsoid with \(\kappa\) shows a general dependence of \(P_j\) on rock composition, since \(\kappa\) proxies for rock-composition (Fig. 12b). All specimens considered globally give a deceptively strong correlation with rock-composition (~\(\kappa\)), \(\kappa\) appearing to account for 72% of the variance in \(P_j\) (\(R^2 = 0.72\)). Analysing the specimens in two, more homogeneous subgroups (\(\kappa < 300 \mu\)SI; \(\kappa > 300 \mu\)SI) that are dominated by pyrrhotite and biotite respectively reveal slightly weaker associations between rock composition (proxied by \(\kappa\)) and \(P_j\), however both correlate significantly at the 95% confidence level, in view of the sub-sample's sizes (Borradaile 2003). We cannot reject the hypothesis that rock composition controls anisotropy degree (\(P_j\)). Similar observations have been recorded elsewhere (Tarling & Hrouda 1993; Rochette 1988; Rochette et al. 1992).

The next logical question concerns the shape or symmetry of the AMS ellipsoid; is \(T\) controlled by factors other than strain or other secondary alignment processes? Compare the data for non-deformed Oligocene sandstone, specimens of which have been studied for other purposes (Borradaile & Mothersill 1991) with specimens of Cambrian Welsh Slate (Fig. 12c). The latter has an extremely penetrative, continuous fine cleavage and strains reported in excess of 60% shortening perpendicular to cleavage, whether it is purple or reduced to show green spots or completely made green (Wood et al. 1976). AMS and \(\kappa\) merge contributions from matrix chlorite (the green pigment) and hematite (purple-red pigment). Magnetite is important in green spots and in bedded-controlled green reduction patches (Jackson & Borradaile 1991). Despite its extreme metamorphic fabric and strain, the magnetite-bearing slate has almost complete overlap in \(P_j - T\) space with a non-deformed typical sedimentary rock (Fig. 12c).

Only the non-reduced, completely purple Welsh slate is distinct in \(P_j - T\) space from the sedimentary rock. The logical conclusion is that metamorphism and strain may be far less significant than mineralogy in determining AMS ellipsoid shape (\(T\)) and eccentricity (\(P_j\)). It will be clear subsequently that AMS axial orientations are far more useful and more readily interpreted in terms of secondary processes.

Interpreting AMS in terms of mineral ODs

Determining AMS and \(\kappa\) for specific minerals

The roles of individual minerals may be approached along several different routes; some are restricted to estimates of \(\kappa\), while others permit an estimate of AMS or another magnetic anisotropy. Some permit the exclusion or isolation of contributions from ordered phases. Some are mechanically destructive and most preclude subsequent palaeomagnetic work on the same specimens. Exposures to high fields (Potter & Stephenson 1990; Stephenson & Potter 1996) and to low or high temperatures may change \(\kappa\) and AMS (Dunlop & Özdemir 1997). Although mineral-specimens may be measured just as standard rock-specimens, it is sometimes forgotten that no natural mineral is pure. The lower its intrinsic \(\kappa\), the more difficult it is to determine its AMS. Inclusions and exsolutions of magnetite and other magnetically ordered minerals are most troublesome; a few SD grains, undetectable by any microscopy may swamp the measurement of AMS in weakly paramagnetic or diamagnetic minerals (e.g. Fig. 2b, Table 2).

Calculation: Ideally, and for a pure mineral phase, \(\kappa\) may be calculated. Syono's (1960) formula (eq. 6) approximates this for a general paramagnet. This indicates upper limits near ~2000 \(\mu\)SI for stoichiometric, hypothetically pure, paramagnetic Fe–Mg silicates (e.g. chlorite, amphibole, biotite, pyroxene, epidote).

Single crystal measurements: Individual crystals, where suitably shaped, may be mounted in universal-orientation holders for AMS and for AARM measurement. Shape-effects are a minor consideration for most \(\kappa\)-values. Hysteresis parameters including \(\kappa_{hf}\) may be measured in selected orientations, for microscopic specimens (<100 mg), using an alternating-gradient magnetometer (Princeton Measurements MicroMag) (Borradaile & Werner 1994; Lagroix & Borradaile 2000a). Torque measurements sensitively detect crystalline anisotropy (Banerjee & Stacey 1967; Martín-Hernández & Hirt 2001, 2003).

Mineral separations: crushed or picked mineral aggregates may be density separated or magnetically separated to yield quite pure concentrations of individual minerals. Individual grains may be assembled and aligned mechanically or magnetically (Borradaile et al. 1985, 1987, 1990; Johns & Jackson 1991; Johns et al. 1992).

Leaching: Oxides and carbonates may be preferentially removed; especially where the solid specimens have been suitably perforated with holes or saw cuts; the same technique is used for 'chemical demagnetization' of palaeomagnetic specimens (Henry 1979). This may preferentially remove certain subfabrics according to grain-size or composition (Borradaile et al. 1990; Jackson & Borradaile 1991).
High-field techniques: High-field magnetization measurements show field-dependent anisotropy for ordered phases, with a maximum in fields near the coercivity, declining to zero as the phases saturate (Rochette & Filion 1988). AARM and AIRM exclusively characterize the ferromagnetic mineral fabric (Fuller & Kobayashi 1964; Daly 1967; Daly & Zinner 1973). High-field torque measurements in different fields allow separate characterization of the deviatoric anisotropy of the linear, paramagnetic fabric and the nonlinear (ferri- and antiferromagnetic) phases (Jelinek 1985; Martín-Hernández & Hirt 2001).

Low-temperature and high-temperature techniques: Remanence-bearing phases may be identified non-destructively by low-temperature thermomagnetic measurements, which show discrete changes in magnetization associated with structural phase transitions, isotropic points, and spin-flop transitions, e.g. magnetite (--titanomagnetite) (~120 K Verwey transition, ~130 K isotropic point), hematite (~260 K Morin transition), monoclinic pyrrhotite (Fe₇S₈; ~34 K) (O'Reilly 1984; Thompson & Oldfield 1986; Rochette et al. 1990; Richter & van der Pluijm 1994; Dunlop & Özdemir 1997; Moskowitz et al. 1998). Ti-rich titanomagnetites and titanohematites, including the ilmenite and ulvöspinel end-members are magnetically disordered at room temperature. Curie or Néel temperatures (2000–7000 °C) characterize most common remanence-bearing minerals, although heating may alter the mineralogy. Some minerals, especially SD magnetite, may show enhanced susceptibility just prior to complete thermal demagnetization (‘Hopkinson peak’). At high and low temperatures, \( \kappa \) is relatively easily measured but AMS determination requires sophisticated laboratory facilities (Rochette & Filion 1988). Temperature equilibration during measurements along different specimen axes is difficult, even in a relatively stable liquid nitrogen environment (77 K) but some data are available (Ihmé et al. 1989; Lüneburg et al. 1999, Parès & van der Pluijm 2002). Cryogenic AMS measurements may offer the most promise to partition anisotropic contributions between the paramagnetic, diamagnetic, and ‘ferro’-magnetic responses although the mineral sources might still be ambiguous.

Anomalous orientations of AMS in certain minerals

Although MD titanомagnetite-magnetite, or pyrrhotite is usually present in traces (~0.2 wt%), depending on the other minerals, their high \( \kappa \) may outweigh their concentration and they may dominate the rock’s \( \kappa \) and the rock’s AMS (Borradaile 1988). First, it is necessary to identify special orientation effects here, like Fuller’s (1961) ‘supergrain’ interaction effect (Fig. 2d). Recently, Hargraves et al. (1991) termed this ‘distribution anisotropy’, which may confuse students of petrofabrics for which the term location fabric less ambiguously, and with historical precedent, describes the fabric contribution from any lithological component discretely located in space, regardless of its internal isotropy or anisotropy (Turner & Weiss 1963). Magnetite causes a more common and serious special fabric effect in SD-form it shows an inverse fabric with its long axis //\( \kappa \) MIN (O’Reilly 1984; Stephenson et al. 1986; Potter & Stephenson 1988). Few rocks show a net inverse AMS (i.e. \( \kappa \) MAX // S) since unusually high concentrations of SD magnetite are required. However, lesser concentrations cause blended fabrics more commonly, in which an SD subfabric interferes with a ‘normal’ matrix (Rochette et al. 1992; Borradaile & Gauthier 2001, 2003b).

Matrix-forming calcite shows intrinsic inverse fabrics where there is insufficient competition from paramagnets and ferromagnets, as in many limestones. Calcite is diamagnetic, with the most negative susceptibility, \( \kappa_e//c \) (Rochette 1988; Ihmlé et al. 1989). \( \kappa \) is not precisely reported (~13 to ~14 \( \mu \)S, Nye 1957) but precise values for anisotropy (\( \kappa_s - \kappa_e // S \approx 1.172 ± 0.028 \mu \)S were obtained by torque magnetometry (Owens & Rutter 1978). Common metamorphic-deformation mechanisms align c axes for calcite (and also for quartz) steeply with respect to the XY-plane or to the plane of maximum shear (later, Fig. 22). In nature, calcite (and quartz) grains are flat-shaped in the basal plane (\( \perp \) c). Thus, an inverse fabric results with the most negative susceptibility \( \kappa_e//c \) and \( \perp S \). The elongate habits of calcite and of quartz (Fig. 3c, e) are not expressed in the matrix. No other major rock-forming minerals are yet reported with intrinsic inverse AMS; although in some leucocratic granites there is sufficient accessory tourmaline to yield inverse or blended fabrics (Rochette et al. 1994).

The Orientation Distribution (OD) of specimen AMS principal axes

Two principal axes (usually \( \kappa_{MAX} \) and \( \kappa_{MIN} \)) suffice to define the orientation of a few similarly oriented specimens since the orthogonality of the third principal axis follows naturally from
the second-rank tensor structure of AMS. However, there are distinct visual and interpretative advantages to viewing all three axes for each specimen on the stereoplot. The orientation of $\kappa_{\text{INT}}$ with respect to other structures may be a clue to blended subfabrics (Fig. 17a later); in general, permutations of all subfabric axes must be considered in combination (Borradaile & Sarvas 1990b; Rochette et al. 1992; Ferré 2002). For similarly oriented specimen-tensors, density contours are interpretable and the sample’s OD is revealed by the cluster-girdle patterns associated with the L-S fabric scheme (Flinn 1965b; Woodcock 1977) (Fig. 12a) though their distribution in $P_j - T$ space is possibly more meaningful in polar coordinates (cf. Fig. 12d, c).

![Diagram](image)

**Fig. 13.** (a) Flinn’s (1965) qualitative L-S fabric scheme describes ODs of preferred crystallographic or preferred dimensional orientations. It is now also used to describe ODs of magnitude-ellipsoids of tensors, for example finite strain or AMS. (b) ODs of axes of specimen AMS-tensors; above perfect L-fabric, below perfect S-fabric. (c) For a sample of AMS specimen-tensors, Jelinek’s confidence regions about the mean axes have a shape and symmetry that characterizes the sample’s OD in the L-S scheme. Here the OD is for an S > L fabric since the confidence regions about mean $\kappa_{\text{MAX}}$ and mean $\kappa_{\text{MIN}}$ are elongate in the plane perpendicular to $\kappa_{\text{MIN}}$. (d) Werner’s (1996) comparison of confidence regions for the sample’s mean axes, determined by different calculation methods for the same sample of AARM measurements.
When using fabric parameters and fabric plots, it is important to recall that \( P_{OD} \) values for the sample-OD may be unrelated to any or even the average specimen. For example, an L-tectonic fabric (point-concentration of \( k_{\text{MAX}} \)) with \( T_{OD} = -1.0 \) may be comprised of individual specimen-ellipsoids for any shape, even \( T = +1.0 \), provided their \( k_{\text{INT}} \) and \( k_{\text{MIN}} \) axes are dispersed in a self-cancelling arrangement in the great-circle girdle \( \perp \) to \( k_{\text{MAX}} \). (We discussed previously that in general \( P_{OD} < P_{\text{SPECIMEN}} \)).

The L-S character of an OD is also represented in the shape of the confidence regions for the axes of Jelinek’s mean-tensor (Fig. 13). ODs for L- or S-tectonites produce characteristic features and great-circle girdles of \( k_{\text{MAX}} \) and \( k_{\text{MIN}} \), respectively (Fig. 13b). Using Jelinek statistics, the confidence region for each mean-tensor axis (maximum, intermediate and minimum) is determined by retaining the essential mean-axes’ orthogonality. The confidence regions thus retain a symmetry which characterizes the OD’s prolateness versus oblateness in Flinn’s L-S scheme (Fig. 13c). Whereas confidence regions possess axial symmetry for the end-member L \( (T_{OD} = -1) \) and S \( (T_{OD} = +1) \) cases, a hypothetical general orthorhombic case has elliptical confidence regions which show the OD has \( S > L \) in, e.g. \( T_{OD} \sim +0.5 \) (Fig. 13c).

The mean tensor and its confidence regions faithfully characterize a sample of tensors. However, the use of traditional density contours on stereograms is still useful for reconnoitring ODs, e.g. identifying sample-homogeneity. Density contours assume no distribution-model so that each group of axes \( (k_{\text{INT}}, \text{etc.}) \) is treated as though they were independent of the other principal axes (Fig. 14a). That example shows that individual specimens have poorly concentrated principal axes, neither forming elliptical clusters, nor elliptical girdles; this strongly suggests that the sample is heterogeneous and that it should be analyzed in more homogeneous sub-samples. Any raw sample, however homogeneous, will generally show non-orthogonality of peak concentrations for \( k_{\text{MAX}} \) for \( k_{\text{INT}} \) and for \( k_{\text{MIN}} \); even maximum eigenvectors will generally be non-orthogonal (cf. Fig. 14b). In this sample, the AMS peak densities’ non-orthogonality and their inclination to the AMS mean-tensor are due to two competing subfabrics in the AMS; one of which is due to a magnetite subfabric here isolated by AARM. The differently oriented subfabrics shown by the AMS/AARM disagreement cause the peak-densities of AMS to be so unreliable (Fig. 14e).

Confidence regions for a mean AMS axis (e.g. \( k_{\text{MIN}} \)) have also been determined using Fisher or Bingham statistics, borrowed from palaeomagnetism where they are applied to vectors or directions. They assume circular or elliptical clusters on the sphere for directions that act as independent variables. In contrast, the distribution of \( k_{\text{MIN}} \) axes is constrained by the distribution of \( k_{\text{MAX}} \) and \( k_{\text{INT}} \) and their confidence regions may only be reliably determined by Jelinek statistics. Werner (1977) compared confidence regions for a homogeneous sample of AARM specimen-tensors and showed that Jelinek-statistics were superior and possessed the expected orthorhombic symmetry (Fig. 13d).

One approach evaluated by Werner used the boot-strap method to determine confidence regions for sample-mean. This may be used for confidence regions of directions, axes or indeed any variable (Fisher et al. 1987). Bootstrap randomly re-samples the original sample of \( n \) measurements, usually at least 200 times. Each of the 200 re-samples contains \( n \) orientations that duplicate some of the original measurements and omit others. The mean orientation of each re-sample is then plotted; the 200 pseudo-mean orientations produce a beguilingly simple elliptical concentration whose density distribution may be contoured to define the confidence region for a certain mean axis. This technique finds proponents and critics in AMS and palaeomagnetic applications (e.g. Werner 1997; Tauxe 1998, Borradaile 2003). It is subject again to the criticism that all the \( k_{\text{MAX}} \) orientations (and then all \( k_{\text{INT}}, k_{\text{MIN}} \)) must be treated independently, as if each \( k_{\text{MAX}} \) orientation was unconstrained by the other principal axes of the specimen tensor. It is more worrisome from a general statistical viewpoint that all bootstrapping requires that the sample be strictly representative of the population. If that is the case, one might just as well use a traditional, less deceptive approach, to calculate the confidence regions, e.g. by assuming a bivariate-Normal distribution on the sphere (Henry & Le Goff 1995) or a Fisher-Bingham distribution (Fisher et al. 1987). Sub-orthorhombic symmetry of bootstrapped confidence regions need not have the same interpretative significance as those derived by Jelinek statistics (below).

Jelinek (1978) provides a rigorous alternative procedure to determine the mean axial orientations and their confidence regions for a sample of tensors. Small samples, or unusually large confidence regions may limit their validity but sample-size rules may not be simply formulated as in classical frequency-distribution statistics. Orientation-statistics are harassed by the closure-problem, and the disadvantages of small sample-sizes may be offset by strong preferred
Fig. 14. (a) The peak-density for a certain set of axes, say $K_{\text{MAX}}$, for a sample of AMS specimen-tensors are usually non-orthogonal. Although inspection of density contours is useful to verify sample homogeneity, they do not characterize ODs for tensors faithfully [sample of Archaean gneisschists]. (b) In contrast, Jelinek's (1978) mean tensor retains the required orthogonality of the sample's mean principal axes [same data]. (c) Confidence regions for mean-tensor axes need not show orthorhombic symmetry [q.v. Fig. 132c]. (d) Multiple subfabrics or (e) specimens with anomalous $K$. When the sample is
orientation. Apart from providing the only valid determination of mean axes for a homogenous sample of specimen-tensors, Jelinek statistics permit the determination of confidence regions about the mean axes that recognizes the mutual orthogonality of principal axes for each specimen. For this reason, the symmetry of Jelinek’s confidence-cones for the mean-axial orientations defines the sample-OD in the L-S scheme and is ideally orthorhombic, for an homogeneous sample due to a single-event coaxial fabric-forming process (Fig. 13). However, for heterogeneous samples the confidence regions may sub-orthorhombic, two or more of them inclined to the symmetry planes (Borradaile 2001, 2003) (Fig. 14c). This is due to competing contributions from differently oriented subfabrics, or the distracting effects of some specimen-outliers (Fig. 14d, e). Jelinek statistics permit these competing contributions to be partially suppressed where the subfabrics, or the outliers, are significantly different in $\kappa$ from the remainder of the rock.

Distortion of the sample OD by the specimens of anomalous $\kappa$ may be reduced and orthorhombicity may be restored by equal-weighting of all specimens. Standardization divides $\kappa_{\text{MAX}}$, $\kappa_{\text{INT}}$ and $\kappa_{\text{MIN}}$ for each specimen by its $\kappa$; thus, all specimens are reduced to unit-susceptibility. In this way the orientation of a minor subfabric, or of a few outlying specimens, is much less significant in the OD and the mean-tensor’s orientation and the shapes of its confidence regions may be changed. For example, a few high-$\kappa$ specimens are responsible for the AMS bedding fabric in an arkose; standardization reveals a cryptic cleavage-fabric due to the lower susceptibility matrix (Fig. 14f).

Even without standardization, data-processing reconnaissance may also partially isolate subfabrics if ODs are compared for sub-samples of different $\kappa$. The AMS OD for high $\kappa$ specimens may be as diagnostic as measuring AARM in the same rocks. The OD of low-$\kappa$ specimens better defines the silicate matrix petrofabric than the OD for high $\kappa$ specimens (Borradaile & Gauthier 2001), (Fig. 14g).

Standardization may suppress an anomalous subfabric, re-establishing the orthorhombic symmetry expected for a single-generation L-S fabric. However, it may also suppress a tectonic component, e.g. revealing the bedding fabric in a slaty matrix (Fig. 15a,b). In rocks with multiple subfabrics, standardization may enhance tectonic fabric symmetry (Fig. 15c) or reveal a different-symmetry subfabric, for example a cryptic L-S fabric where the raw AMS OD suggested an L-fabric (Fig. 15d).

**Orientation-Distribution (OD) of AMS specimen tensors and magnetic fabric**

The L-S fabric concept combines the contributions of shape and orientation of three-dimensional objects or tensor magnitude-ellipsoids to define an OD. A single-event, homogenous L-S fabric must show orthorhombic symmetry for its OD (Fig. 13b, c). For simplicity, consider a sample of specimens, each with the same anisotropy. With a perfect, saturation alignment, the sample has the same anisotropy ($T_P$) as individual specimens. Any imperfection in the OD requires $P_{\text{SAMPLE}} < P_{\text{SPECIMEN}}$; it may sometimes also subdue ellipsoid-shape ($|T_{\text{SAMPLE}}| < |T_{\text{SPECIMEN}}|$).

The arguments applied to a sample of many specimens apply equally to the AMS of one specimen and the AMS of its constituent aligned minerals. It is also a conceptual objection to the correlation of finite-strain magnitudes and AMS magnitudes for a specimen (Borradaile 1988, 1991). Even a well-aligned tectonite specimen usually has a modest $P_P$, much less than that of the dominant anisotropic mineral, due to self-cancelling effects of unfavourably oriented examples of that mineral, and also due to other minerals. In nature, the relationships of mineral-AMS to specimen-AMS are obfuscated in several ways:

(i) Mineral AMS axes may only be coaxial with crystal axes for high symmetry; orthorhombic (Fig. 1d), trigonal (Fig. 3c,e) as well as cubic and hexagonal classes. Most minerals are monoclinic; they may have only one AMS axis parallel to a crystal axis. For triclinic minerals (e.g. plagioclase), all AMS axes are inclined to crystal axes.

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standardized by dividing each specimen’s principal susceptibilities by $\kappa$, the contribution of the anomalous subfabric or outliers to the OD is suppressed. (f) Standardizing specimen-tensors [dividing their principal susceptibilities by their $\kappa$] may neutralize the distracting contribution of a subfabric or outliers of anomalous $\kappa$. Thus a cryptic S-fabric is revealed when the AMS of weakly cleaved arkose is standardized; the masking contribution of a high-$\kappa$ bedding subfabric is suppressed. (g) Without resort to tensor-standardization or other techniques like AARM, inspecting sub-samples of AMS tensors according to their $\kappa$-ranges may also reveal different subfabrics (ophiolite dykes, Cyprus; Borradaile & Gauthier 2001).
(a) Borrowdale Volcanic Slate
\[ \kappa = 381\, \mu\text{SI} \]
\[ n = 74 \]
\[ L = X \]

(b) Mean tensors: before & after standardization
\[ L = X \]

(c) Trout Lake Granite: before & after standardization
\[ S \gg L \quad (n = 75) \]

(d) Kapuskasing Gneiss: L becomes L=S after standardization
\[ (n = 60) \]

Mean tensor axes for sample of \( n \) specimens
- max
- int
- min

Specimen tensors normalized

95% confidence limits on mean tensor's axes

Fig. 15. (a) AMS of Borrowdale volcanic slate defines a good \( S \gg L \) fabric. (b) Mean-tensors of raw data emphasize field observations of \( S \gg L \) fabric but standardization reduces the symmetry and strength of the S-component; this may be revealing a matrix-bedding fabric. (c) Standardization enhances fabric symmetry and slightly changes interpreted mean-orientations. (d) Gneisses show an overall S-fabric for non-standardized AMS, mainly due to a low high-\( \kappa \) specimens. Standardization shows that the silicate matrix fabric possesses an L-fabric. However, the L-directions of both fabrics are similar.

(ii) For low-symmetry minerals with poorly corresponding susceptibility axes and crystallographic axes, an OD of crystals may fail to define consistent AMS orientations.

(iii) A one-to-one mapping of magnitudes with petrofabrically convenient crystal axes is not mandatory (Fig. 3c, e). This is due to intrinsic-crystallographic inverse AMS, apparently shown also by goethite, tourmaline and possibly by cordierite (Rochette et al. 1992).

(iv) For SD magnetite, \( \kappa_{\text{MIN}} \parallel \) the long dimension due to the inverse AMS shape-effect (Fig. 3h). Other high susceptibility minerals that may show an inverse AMS include maghemite (Borradaile & Puurunen 1989), suitable titanomagnetite compositions and greigite (Aubourg & Robion 2002).
(v) Physical alignment mechanisms (crystal-plastic mechanisms etc.) are related to crystallography in complex ways, they do not automatically align the long habit of the crystal. For example, quartz and calcite c axes align steeply to S, sometimes in cones. Moreover, $\kappa_c / c$ so that inverse fabrics ensue.

(vi) Weakly aligned crystal axes and partially cancelling contributions from a diamagnetic matrix and low abundance minerals with $\kappa > 0$ compound the above problems, especially for $P_j \rightarrow 1.0$ (Fig. 3). Any specimen with $\kappa < 100 \mu S I$ may be suspect.

Thus, AMS orientations may at best only approximate the minerals’ OD. To follow from this monomineralic reductio ad absurdum, consider polymineralic rocks, with incompletely developed or even multiple sub-fabrics, with additional AMS contributions attributable to shape-controlled magnetite accessories or inclusions, and with potentially self-cancelling diamagnetic and paramagnetic components in low-susceptibility rocks. These confounding complications lead to the undeniable conclusions:

(a) AMS fabrics that characterize the overall OD in a tectonically meaningful manner are probably attributable to a dominant subfabric of a single mineral phase of large $\kappa$; with orthorhombic symmetry of its sample-mean tensor (e.g. Fig. 13c). These need not be due to magnetite; mafic silicates such as chlorite, biotite, serpentine or amphibole may dominate $\kappa$ and usually possess stronger ODs than magnetite.

(b) A sub-orthorhombic OD is probably due to a subfabric or a few extreme-$\kappa$ specimens in the sample (Fig. 14d, e). Subfabrics may be partially isolated by comparing standardized to non-standardized mean tensors (Figs. 14f, 15c, d), by comparing sub-samples of different $\kappa$ (Fig. 14g). More definitive answers may be obtained by separate rock magnetic experiments that determine AARM or pAARM, since these completely isolate the OD due to remanence-bearing minerals (Fig. 14b).

(c) Mathematical associations between ODs and AMS are applicable only to idealized scenarios (e.g. Hrouda & Schulmann 1990; Henry 1992; Hrouda 1993; Benn 1994). They usefully constrain thought-experiments and sampling strategies but few examples are directly applicable in nature (Housen et al. 1993a).

(d) AMS studies would have been discouraged if these complications had been realized at the time of Graham (1954). Fortunately, AMS axes commonly proxy for petrofabric orientations, in tectonic, sedimentary or igneous rocks, which is attributed mainly to (a) and (b) above (Hrouda 1982; Jackson & Tauxe 1991; Tarling & Hrouda 1993; Borraidaile & Henry 1997).

Mineral abundances: AMS and $\kappa$ for an ‘aligned’ rock

Consider AMS for a monomineralic rock; if its mineral-alignment is less than perfect $P_{\text{SPECIMEN}} < P_{\text{MINERAL}}$ and $T_{\text{SPECIMEN}}$ will differ from $T_{\text{MINERAL}}$. Rock specimen’s AMS is a subdued version of the constituent mineral’s AMS. Now, consider the addition of a second mineral, similarly aligned, for example dispersed accessory MD magnetite. $\kappa_{\text{SPECIMEN}}$ is approximated from the volume proportions of the two minerals. With 1% magnetite, $\kappa_{\text{SPECIMEN}} \approx 0.99 \kappa_{\text{BIOTITE}} + 0.01 \kappa_{\text{MAGNETITE}}$. Conservatively, assume the maximum bulk susceptibility for biotite (2000 μSI) and a value of 2 500 000 μSI for magnetite: $\kappa_{\text{SPECIMEN}} \approx 1980 + 25 000 = 26980 \mu S I$. [The diamagnetic contributions of minerals has been overlooked here since it usually only affects AMS orientations and magnitudes in weakly aligned and low-$\kappa$ rocks (Rochette 1987a, 1994a)].

Having recognized the affect of the dilution principle on $\kappa$, we must now consider its effects on AMS (i.e. $P_j$ and $T$). Combinations of coaxial AMS ellipsoids in this model show that $\text{AMS} \text{SPECIMEN}$ migrates from biotite towards that of magnetite with increasing magnetite abundance, measured in ppm (Fig. 16a). A similar model is shown for a calcite-magnetite aggregate which is a realistic model of magnetite concentrations (1–10 ppm) in some limestone (e.g. Hamilton et al., this volume); here one notes how the AMS moves from the diamagnetic field to the field of positive-$\kappa$ which involves a change of AMS symmetry ($T$ changes sign) and requires an inversion of principal axial orientations.

Fuller (1963) first appreciated the essence of polymineralic problems, although their far-reaching consequences for petrofabric and AMS interpretation could not be foreseen at that time. For example, on their own, such arguments throw doubt on any generally valid causal relationship between AMS and the magnitudes of alignment processes (Borraidaile 1988, 1991; Borraidaile & Henry 1997). Indeed, in nature, AMS and finite strain usually correspond
poorly (Borradaile & Mothersill 1984; Borradaile 1991) although there are a few well documented examples to the contrary where lithologies are uniform, strain gradients are small and metamorphic effects are minimal (Siddans et al. 1984; Hirt et al. 1988, 1993). More optimistically and actually more usefully, AMS orientations and finite strain axes do correspond well in nature (Hrouda 1982) and also in the laboratory (Borradaile & Alford 1987, 1988; Jackson et al. 1993).

Finally, where the different minerals have different alignments, mineral abundances also influence the orientations of the gross AMS axes for the specimen. For example, increasing concentrations of magnetite move $\kappa_{\text{MAX}}$ \text{SPECIMEN} progressively from the orientation of $\kappa_{\text{MAX}}$ PHLOGOPITE toward the orientation of $\kappa_{\text{MAX}}$ MAGNETITE \text{(Fig. 16c–e)}. With 1000 ppm magnetite, the differently oriented phlogopite matrix is expected to show negligible influence on the sample's AMS orientations \text{(Fig. 16e)}.

**Blended sub-fabrics and the rock's gross AMS orientations**

The fortunate recognition of 'anomalous' principal orientations associated with discrete sub-fabrics in pressure-solution cleavage lead to the
recognition of blended AMS subfabrics (Fig. 17a). In the case of the pressure-solution anomaly \( \kappa_{\text{MAX}} \) is parallel to the intersection of cleavage with bedding, the \( \beta \)-lineation, rather than \( X \), due to the summation of orthogonal subfabrics from the bedding plane and from the stylolite-plane (Borradaile & Tarling 1981). Penetrative schistosity may also produce \( \kappa_{\text{MAX}}/\beta \), which is disconcerting since it is not predictable from petrographic or field observations (Borradaile & Sarvas 1990a;b). Figure 17b. The discrete location subfabrics may include crenulation-cleavage surfaces versus their microcrysts, many types of metamorphic differentiation, mylonitic lamination and S-C fabrics and pervasive subfabrics are also common but may require microscopic identification (e.g. slaty cleavage cutting massive beds). Provided the specimen samples the mineral-ODs in each subfabric representative, the specimen-AMS will blend the subfabric's differently oriented AMS. Of course the manner in which the subfabrics blend depends on their angular relationships as well as their anisotropy (\( P_j \) and \( \kappa \)); for convenient simplicity we consider subfabrics with nearly orthogonal AMS ellipsoids. Depending on the subfabrics' AMS-ellipsoid-shapes (\( P_j \) and \( \kappa \) ), their relative contributions in terms of volume and \( \kappa \), numerous permutations exist, of which but one example is shown in Figure 17c. Rochette et al. (1992, 1999) evaluated possible specimen AMS axes from varied volumetric proportions of coaxial normal and inverse subfabrics (Fig. 17d). The orientations of the specimen's principal AMS axes switch dramatically in orientation as the volume-fraction of the inverse subfabric increases, and the shape of the specimen's AMS ellipsoid changes dramatically and non-progressively, as shown on the polar \( P_j - T \) plot (Fig. 17d). Ferré (2002) considers more complex situations. A matrix algebra approach to this problem is elegant (Hrouda 1992) but its sensitivity to commutativity may be overlooked in nature and it may be best reserved for cases where the sequence of truly discrete subfabrics is known. Although considerations of coaxial/orthogonal subfabrics provide useful interpretative principles the AMS subfabrics will generally be inclined to one another in nature.

Occasionally, one or more subfabrics may contribute counter intuitively to the gross AMS producing \( \kappa_{\text{MAX}} \) at a high angle to S. These deserve consideration and three distinct categories reported:

(a) Minerals with intrinsic crystalline-inverse AMS, e.g. tourmaline and goethite have \( \kappa_{\text{MIN}}/c \), which is their long axis. Their ODs thus have \( c//X \) lineation characterized by \( \kappa_{\text{MIN}} \), not \( \kappa_{\text{MAX}} \).

(b) Diamagnetic minerals like quartz and calcite. Although \( \kappa_{c}/c \) and \( \kappa_{c} \) is the most negative ('maximum', sic) principal susceptibility, common crystal-plastic deformation mechanisms (e.g. c-basal glide) align \( \kappa_{c} \perp \) to S.

(c) SD magnetite shows an inverse-effect AMS, its measured \( \kappa_{\text{MIN}} \) // grain long axes.

Occasionally, ambiguous subfabrics occur in counter-intuitive orientations, e.g. \( \kappa_{\text{MAX}} \) steeply inclined to bedding in weakly deformed sediment (Borradaile et al. 1999a; Aubourg & Robion 2002) or \( \perp \) to XY in experiments (Borradaile & Puumala 1989). The present state of knowledge does not exclude unknown alignment processes or unverified inverse-AMS mineral-responses (e.g. greigite, maghemite).

In a perfect inverse-situation, an S-petrofabric will have a perpendicular L-fabric AMS. Conversely an L-petrofabric mineral alignment will possess an orthogonal S-symmetry AMS. In other words, \( T_{\text{PETROFABRIC}} \) and \( T_{\text{AMS}} \) are of different sign and the OD and AMS ellipsoids have perpendicular major axes. The importance and potential of AMS for the common matrix-forming diamagnetic minerals, calcite and quartz was appreciated early (Owens & Bamford 1976; Owens & Rutter 1978), though not their inverse-fabric complication. Inverse AMS has been recognized for some limestone (inter alia Rochette 1988; Ilmlé et al. 1989; de Wall et al. 2000; Hamilton et al., this volume) and is predicted for quartzite (Hrouda 1986) and is reported rarely for quartz-rich rocks like tonalite-gneiss [author, unpublished data]. These rocks' AMS may be diamagnetic or, in the presence of impurities, weakly paramagnetic. It is very important to recognize that AMS axial orientations may be difficult to interpret due to self-cancelling sources of positive and negative susceptibility, especially for specimens in the range \(-15 < \kappa < 50 \mu \text{SI}\) (Fig. 3a,b) and the presence of any ferromagnetic contamination may hinder interpretation (Borradaile & Stupavsky 1995).

The inverse-effect fabrics of SD magnetite are due to shape anisometry, the measured \( \kappa_{\text{MAX}} \) and \( \kappa_{\text{MIN}} \) correspond to short and long grain axes' orientations respectively. SD grains may occur as lattice-controlled inclusions/exsolutions in silicates, and therefore at best indirectly related to tectonic fabric. In ophiolite, copious SD production by ocean-floor metamorphism provides examples of inverse-subfabric contributions (Rochette et al. 1992; Borradaile
Fig. 17. (a) Combined AMS subfabrics due to bedding (S₀) and cleavage (S₁ spaced and stylolitic due to pressure solution). The combination yields $\kappa_{\text{MAX}}//\beta$ ($\beta$ = intersection lineation of bedding and cleavage) not // X (Rheinisches Schiefergebirge; Borradaile & Tarling 1978). (b) A penetrative schistosity (S) subfabric pervades and blends with a bedding (S₀) subfabric also to give $\kappa_{\text{MAX}}//\beta$-lineation & fold-hinge; the two subfabrics are penetrative (Archaean schists; Borradaile et al. 1988). (c) One of many possible permutations of coaxial subfabrics; however, subfabrics may also combine non-coaxially in nature. (d) The coaxial blending of a normal and an inverse AMS; the inverse fabric may be due to a counter-oriented subfabric or due to an
& Gauthier 2001, 2003b). However, perfect inverse AMS responses due to the SD effect are rare since the AMS of \( \sim 10^6 \) SD grains may be cancelled by the AMS of a single \( 10 \mu \text{m} \) MD magnetite grain. More commonly, an inverse-SD subfabric partly cancels a normal subfabric to produce a blended fabric.

Interpreting AMS\(_{\text{ROCK}}\) is still more challenging where it comprises non-coaxial subfabric permutations. Certain subfabrics may be isolated, enhanced or suppressed by experimental or statistical procedures. These may partly resolve ambiguous interpretations, as follows.

Experimentally, AARM (or even AIRM), isolates the subfabric OD of all ‘ferromagnetic’ grains, or part of the ferromagnetic population, if it is selected using a coercivity range in the pAARM technique. Fortunately remanence-anisotropy for all magnetite, including SD, shows a normal-fabric with \( A_{\text{MAX/long axis}} \) and \( A_{\text{MIN/short axis}} \) (Jackson & Tauxe 1991), compatible with the associated AMS (e.g. Fig. 3g, h; Fig. 23 e, f later). Hematite and pyrrhotite also have compatible, ‘normal’ AARM and AMS. Goethite is reported to have an inverse-crystalline AMS (Özdemir & Dunlop 1996) but it is rarely abundant and its alignment mechanisms are unknown.

**Mineral-abundances: the rock’s \( \kappa \) and principal susceptibility magnitudes**

The rock’s magnitudes \( \kappa_{\text{MAX}} \geq \kappa_{\text{INT}} \geq \kappa_{\text{MIN}} \) are influenced by mineral-proportions and their ODs (e.g. eq. 18; Owens 1974). For initial simplicity, assume two minerals are coaxially aligned. Henry (1983) and Henry & Daly (1983) showed simple relationships for susceptibility-magnitude contributions with slightly varying mineral concentrations. The model predicts that for a suite of specimens principal magnitudes (\( \kappa_{\text{MAX}}, \kappa_{\text{INT}} \)) are linearly dependent on \( \kappa \), and this is verified in nature (Fig. 18a) if the premises are reasonably satisfied. ‘False positives’ may also be observed, in which linear correlations are observed, but the model assumptions do not hold, and different interpretations are required (Johns & Jackson 1992; Rochette et al. 1992).

A similar concept underlies comparisons of \( \kappa \) and \( A \) for ARM in granulometry studies, Figure 18b (King et al. 1982; Jackson et al. 1988). Extended to anisotropy (AMS versus AARM), different relationships occur with different concentrations of the constituent subfabrics (Fig. 18c) (Borraďaile & Lagroix 2001). Matrix \( \kappa \) and accessory-subfabric \( A \) magnitudes may be isolated and one may infer relative contributions of matrix versus ferromagnetics to AMS from the graphs. Non-linearity may imply subequal contributions from more than two phases (e.g. Fig. 18c).

Henry’s (1989) triangular tensor plot, adapted from one used in strain analysis (Hsü 1966), investigates the combined effects of \( \kappa \)-variation and orientation. Closure restraints of the triangular plot and the usual restrictive premises, e.g. high-symmetry crystals, hamper this approach (Fig. 18d). Triangular graph axes represent each of the three standardized principal magnitudes, e.g. \( 0 \leq (\kappa_{\text{MAX}}/\kappa) \leq 1 \), with advantages and disadvantages. Ranges of permissible standardized \( \kappa_{\text{MAX}}, \kappa_{\text{INT}} \) and \( \kappa_{\text{MIN}} \) values for the tectonite lie along lines in the triangular diagram.

**Isolating separate subfabrics experimentally**

Physical discrimination of mineralogical sources is less ambiguous than statistical procedures, especially where one subfabric is ‘ferromagnetic, with low or moderate coercivity. Suitable laboratory approaches include:

(a) Anisotropy of complex electromagnetic susceptibility, measured in high-frequency a.c. fields may characterize the subfabric of high-conductivity minerals, e.g. sulphides and graphite (Vincenz 1965; Clark et al. 1988; Borraďaile et al. 1992; Ellwood et al. 1993; Worm et al. 1993). Eddy currents induced by the time-varying applied field generate a positive quadrature and negative in-phase a.c. response (the latter of which can be mistaken for diamagnetism in a single-frequency, single-temperature measurement).

(b) Measuring \( \kappa \) at various low and high temperatures may identify the predictably inverse-T paramagnetic response, or the temperature-independence of diamagnetics (Rochette & Fillion 1988; Richter & van
Fig. 18. (a) Simple AMS-magnitude relations may exist for two dominant minerals (Henry 1983) although such relationships do not guarantee a simple concentration-control (Johns & Jackson 1991). (b) ARM intensities versus $\kappa$ indicate the relative importance of remanence-bearing versus paramagnetic minerals (King et al. 1982). (c) AMS magnitudes versus AARM magnitudes in the same specimens; homogeneous coaxial subfabrics with a non-linear relationship. Mafic silicates, MD magnetite, ilmenite and chromite all contribute significantly to AMS whereas only magnetite contributes to AARM (data of Borradaile & Lagroix 2001). (d) The effect of mineral abundances on AMS magnitudes and axial orientations has been tentatively investigated for simple systems with saturation alignment and high-symmetry crystals (Henry 1989).

der Pluijm 1994; Parés & van der Pluijm 2002), Figure 4. For 'ferro'-magnets, temperature-dependence is sensitive to crystal structure and composition. (e.g. Dekkers 1989a, b; Dekkers et al. 1989; Moskowitz et al. 1998).

(c) The practice of laboratory heating to enhance and characterize a subfabric
(Perarnau & Tarling 1985) is not generally endorsed and is discussed conveniently under metamorphism, below, although that is a natural process.

(d) Isolating or suppressing magnetic responses from ferromagnetic subfabrics. Magnetization measured in the presence of high fields (>300 mT) ignores the response of minerals that are saturated at that level, such as magnetite. Consequently, the 'matrix' AMS is isolated (Rochette & Fillion 1988; Kelso et al. 2002; Ferré et al. 2004). High-field torque measurements in different fields permit separate characterization of the saturated ferromagnetic anisotropy (field-independent above saturation) and the anisotropy of the linear high-field susceptibility of dia-, para- and antiferromagnets (Jelinek 1985; Martín-Hernández & Hirt 2001). Alternatively, and more easily, the anisotropy of the ferromagnetic subfabric is measured independently, if its coercivity is low enough to permit successive remagnetizations and measurements along different specimen-directions. The latter techniques include AIRM (Daly & Zinser 1973; Stephenson et al. 1986), AARM (McCabe et al. 1985; Jackson & Tauxe 1991; Trindade et al. 2001) and GRM (Stephenson 1981a; Stephenson & Potter 1987). Hematite, goethite and some pyrrhotite are usually too coercive to respond to the latter two (AF-based) techniques with the fields available.

(e) Where the tensor for one subfabric is isolated, it is rarely possible to isolate the remaining OD by subtraction from the whole-rock OD. For example, Hrouda et al. (2000) show that in general the non-ferromagnetic contribution to AMS is not isolated when an AARM-defined subfabric is subtracted from the 'parent' AMS fabric.

The most reliable techniques are (d); particularly AIRM and AARM (see also Potter, this volume). AIRM requires the measurement of IRMs applied in different directions, usually with a pulse-magnetizer. Advantages are high signal/noise ratio and speed (rapid magnetization process; multi-component measurements allow tensor determination from three orthogonal magnetization steps). The procedure is more time-consuming if fields exceed the non-linearity threshold since multiple measurements are needed to determine each directional Rayleigh coefficient. Experimentation with pilot specimens at various applied fields may show it is possible to determine sensible AIRM in the linear regime for ≤50 mT. However, Stephenson et al. (1986) recommended <5 mT. Despite its weaknesses, the AIRM method may produce some sensible comparisons with AMS fabrics and field microstructures (Borradaile & Dehls 1993). AARMs offer more reliable fabric interpretations, which offset the demands in time and technology (Jackson 1991). AARM was first used by McCabe et al. (1985) to isolate a magnetite-OD from the whole-rock AMS, thus identifying a feeble tectonic magnetite subfabric overprint on bedding. AARM also permits the isolation of subfabrics within the magnetite subfabric; this is partial AARM (pAARM). pAARM is possible since most magnetite has coercivities below the peak field in routine AF demagnetizers; each pAARM may then be determined within a selected coercivity range within the decaying AF from its peak value, in an AF demagnetizer. Thus, one may measure pAARM for magnetite grain-subfabrics with different coercivity (Jackson et al. 1988, 1989a, b; Jackson 1991; Trindade et al. 2001; Aubourg & Robin 2002; Aubourg et al. 2000; Nakamura & Borradaile 2001a, b, 2003[100]) (see Fig. 23 later). The pAARMs define subfabric ODS characterized mostly by different grain-sizes or different degrees of internal stress. This technique may be useful with other ferromagnetic minerals if they have coercivities within the range of routine laboratory AF demagnetization.

Finally, AARM is especially useful in assessing palaeomagnetic recording fidelity (Kodama & Sun 1992; Kodama 1997; Gattacceca & Rochette 2002), since it is often tensorial even in AFs much larger than the d.c.-field nonlinearity threshold.

Interpreting AMS in terms of tectonic processes

The community has recognized broad interpretative categories for single-generation AMS fabrics, including:

1. Finite strain
2. Strain history (kinematics)
3. Deformation mechanisms (grain-scale lattice realignment)
4. Metamorphism
5. Stress (incremental strain)

Finite strain

Field studies and rock-mechanical laboratory experiments have revealed some useful
relationships between AMS and the orientation of the finite strain ellipsoid (axes $X \geq Y \geq Z$) both in coaxial strain histories and those involving a notable shear-component. In the simplest, coaxial case, there may be a one-to-one mapping of $X, Y, Z$ orientations with those of $\kappa_{\text{MAX}}, \kappa_{\text{INT}}$ and $\kappa_{\text{MIN}}$. (Fig. 15a, b, c, d; see Fig. 24b, d later). Such simple angular relationships may be expected where the tectonic fabric successfully overprints or may be clearly distinguished from an earlier fabric (inter alia Hruda 1982; Rochette & Vialon 1984; Siddans et al. 1984; Benn & Allard 1988; Hirt et al. 1988, 1993) or where, as in experiments, the initial fabric may be near-isotropic by design (Borradaile & Alford 1987, 1988; Borradaile & Puumala 1989).

However, a coaxial strain history requires finite strain axes remaining constant with respect to the material as strain accumulates. In detail, this cannot be true and the non-material $X, Y, Z$ axes may spin in response to changing strain increments caused by inconsistent stress axial orientations: all strain histories are non-coaxial to some extent (Flinn 1962; Ramsay 1967; Ramsberg 1975). Fortunately, the degree of non-coaxiality is beneath the detection-level in many cases; obvious exceptions are shear zones and mylonite zones. Most examples assume an accumulation of strain in which successive strain ellipsoids become progressively more anisotropic. Unfortunately, for extreme vorticities, later incremental strain ellipsoids may de-strain the previous ellipsoid so that the strain history is non-progressive, leading even to pulsating strain histories (Ramsberg 1975; Ramsay & Huber 1983). The latter are probably restricted to high shear-strain zones, e.g. mylonites.

Increasing field evidence, experimental evidence and theoretical considerations cast doubt on any meaningful cause-effect correlation of strain magnitudes on AMS $(P, T)$. Occasional apparent graphical correlations may give the illusion of a process-caused relationship but the real, lurking variable (Borradaile 2003), may be mineral-abundance and orientation-distributions (Borradaile 1988; Johns & Jackson 1991; Johns et al. 1992, Rochette et al. 1992).

Strain history (kinematics)

Few rocks possess strain markers that reveal $XYZ$ orientations and shear sense. However, more often AMS axes are compatible with tectonic structures that have well-known relationships to finite-strain orientations (Aubourg et al. 1997, 1999, 2000; Frizon de Lamotte et al. 2002; Saint-Bezar et al. 2002). Thus, magnetic petrofabrics usually identify principal strain axes $(XYZ)$ and sometimes independently confirm shear-sense from multiple subfabrics generated during a non-coaxial strain history. The sequence of three-dimensional fabric ellipsoids will necessarily be limited to the small number of identifiable magnetic subfabrics, e.g. paramagnetic matrix, MD-magnetite and PSD-magnetite, possibly supplemented by a field-fabric, e.g. quartz-feldspar schistosity. Since these subfabrics develop sequentially, their relative angular relationships reveal the shear-sense (e.g. Fig. 19a). Studies along the dextrally transgressed Archaean terrane boundaries in northern Ontario reveal successive subfabric axes compatible with dextral shear accompanied by upwards extension to the E-NE (Fig. 19b) (Borradaile & Dehls 1993; Borradaile et al. 1993; Werner & Borradaile 1996). Higher metamorphic grade on the northern sides of the boundaries and some traditional field evidence for dextral shear support these conclusions. The emplacement of plutonic bodies (or igneous ones, for that matter) may also be investigated using asynchronous subfabrics. Origins suggested for Archaean gneiss domes include emplacement by diapiric inflation or secondary doming. Clearly, the latter would disturb AMS axes similarly at each site. In contrast, syn-metamorphic inflation would produce different angular relationships between $\kappa_{\text{MAX}} - \kappa_{\text{INT}}$ and $XY$ (schistosity) at different sites according to the amount of radial inflation. For the Ash Bay dome of northern Ontario, the AMS foliation-dome, due largely to a late magnetite subfabric, is more subdued than the gneissic-foliation-dome, favouring an inflation origin (Fig. 19c; Borradaile & Gauthier 2003a).

Deformation mechanisms (grain-scale processes)

Since AMS is mostly due to crystallographic alignment (shape alignment in the special case of a high-$\kappa$ mineral like magnetite) it is important to consider deformation mechanisms, which realign crystal lattices. Structural geologists have long been aware that passive continuum mechanical models inadequately model the actual process by which crystalline aggregates develop preferred crystallographic orientations (PCO), (e.g. Nicolas 1987; Nicolas & Poirier 1976). For example, a modern petrofabrics and material science recognizes that alignment processes are due to environmental conditions such as mean pressure, deviatoric stress, incremental-strain (~stress) history, vorticity, temperature...
and fluid pressure (Nicolas & Poirier 1976; Means et al. 1981; Poirier 1985; Nicolas 1987; Blenkinsop 2000). They recognize also that many intrinsic material properties influence alignment: crystal symmetry and process-activation energies and dynamic quantities such as dislocation density, grain-boundary shape and mobility, grain-size and disaggregation.

**Single-crystal processes**

**Rigid body rotation**

Dating from the earliest days of structural geology (Wettstein 1886), this notion considers an object aligning its long axis by freely spinning along a locus towards X. The trace of its projection, for example on the XY plane, is given by the change from initial angle with X (θ) to final angle (θ') via $X'/Y = \tan(\theta)/\tan(\theta')$. Its locus in three-dimensions is fixed by combining the line’s projections on the principal planes (Fig. 20a–c; Flinn 1962). Rotation is not necessarily by the shortest route, depending on the Y-extension; normals to planar elements spin along inverse paths toward Z. March's (1932) method of strain analysis assumes this alignment model; thus, in any given direction the density $\rho$ of normals to 'mica flakes' is predicted by the stretch; $\rho_2 = Z^{-3}$. The most unrealistic assumption is that the initial OD is uniform; its unwise application in AMS and field strain studies may be attributed to the over-interpretation of some excellent but specialized rock-mechanics laboratory experiments (Nicolas & Poirier 1976; Tullis 1976). Although this is a common supporting process at low strain, in nature, spinning objects impinge and are impeded by the matrix.

The association of $\kappa_{\text{MAX}}$ and $\kappa_{\text{MIN}}$ orientations with X and Z respectively in nature is not sufficient evidence to invoke 'Marchian rotation' or any other purely geometrical continuum mechanical model. A state of finite strain may be achieved by an infinite number of strain-histories and rheology (and rheological history) is poorly constrained. For example, simple experiments with analogue materials (Fig. 20d)
Coaxial strain history movement paths for normals to flakes, e.g. mica.
[Reverse paths and concentration patterns for long-axes to acicular minerals]

(d) Progressive flattening $X > Y > Z$
SD maghemite + plasticine
(average AMS) $Z \approx 0.5$

(e & f) Progressive flattening $X = Y > Z$  [$Z \approx 0.75$]
rock mechanics experiments initially "isotropic"

(g) Rock mechanics shear experiment
initially "isotropic"

(h) Progressive simple shear
MD magnetite & plasticine

$k_{MAX}$ $k_{MIN}$ $k_{INT}$

later $\bigcirc$ $\bigotimes$ $\bigoplus$
initial $\blacklozenge$ $\bullet$ $\blacklozenge$

Fig. 20. (a–c) March (1932) envisaged the re-distribution of orientations of linear elements in space during a coaxial strain history. This used strict continuum mechanics assumptions (uniform initial OD, non-impingement of elements, non-material properties of elements, etc.). The model provides a limiting case to constrain models and thought-experiments but is a poor approximation to any Natural alignment process. (d–e) AMS experiments with model materials and rock deformation experiments show superficial similarities with the movement paths predict the by the March model but are very sensitive to initial weak anisotropies; shape changes (e.g. $T_j$) may be non-progressive (Borradaile & Puumala 1989). (g–h) Noncoaxial strain history experiments show complex movement paths for AMS axes (Borradaile & Alford 1987, 1988). The fact that certain principal axes approach compatible strain axes is insufficient to justify March model rotations as the sole mechanism.
or rock mechanical experiments with rock-analogues (Fig. 20e–h) reveal complex movement paths of AMS axes and non-progressive changes in AMS ellipsoid shape (Borradaile & Alford 1987, 1988; Borradaile & Puumala 1989). Those experiments showed a strong sensitivity to the slightest initial anisotropy that even involved the rapid switching of orientations for AMS axes.

The following processes usually accommodate more strain effectively and are common in nature.

Preferred nucleation
New minerals may grow at the expense of others and their orientations may be controlled in several ways. First, overgrowths will be favoured on suitably aligned crystals with long habits (e.g. chlorite, mica, and actinolite), enhancing any initial alignment (Oertel 1983; Fig. 21a). Alternatively, neomineralization may create new lattices with their compliances appropriately controlled by the prevailing stress system; this could produce perfect saturation alignments of nuclei in high-grade rocks. Unfortunately, stress-nucleation metamorphic alignments may only record an ephemeral state of stress during nucleation. Subsequent stress or even finite strain increments may fail to leave a record in the fabric.

Differentiation processes
Many metamorphic differentiation processes produce heterogeneous and location fabrics. Pressure solution commonly produces textures readily sampled within an AMS specimen. A low strain-rate process, it is most effective at low temperature (<3000 °C), requiring a preferably mobile fluid-phase. Solutes such as quartz and calcite diffuse through the fluid to some local or distant reservoir from sites of high impingement stress. In some cases 60% of a rock has been transferred by pressure solution. The insoluble residue of phyllosilicates, aligned along stylolitic cleavage, commonly contributes

Fig. 21. Metamorphic effects, sensu stricto, have largely been neglected and under-appreciated in the analysis of AMS fabrics. (a) PCO may develop due to the suppression of overgrowths on unfavourable oriented nuclei (Oertel 1983). (b) Stress-controlled nucleation crystallization may develop a saturation alignment that is aligned incompatibly with subsequent stress/strain history. (c) Non-coaxial strain histories may produce differently oriented subfabrics. During progressive metamorphism, minerals such as chlorite and biotite may form at several times in differently oriented subfabrics. (d) A rare study of progressive metamorphism, AMS and AARM revealed and addressed many natural complexities that should be considered more often (Housen & van der Pluijm 1991, 1993). Clearly, chlorite OD, 'cleavage', AMS and AARM subfabrics develop at different times and in different orientations.
Fig. 22. A deformation mechanism map (rear wall of diagram) represents dominant crystal-deformation and grain-alignment mechanisms in terms of environmental variables (e.g., temperature and differential stress), usually for a fixed grain-size (e.g., Nicolas 1987). The mechanisms may operate in subdued form outside their designated field. In nature, most of these mechanisms may not proceed without particulate flow of the grains (Borroma1e 1981), a phenomenon that may be enhanced by fluid pressure. In principle, particulate flow may be entirely dependent on the intracrystalline processes or independent of them.

a significant AMS subfabric. A bedding-cleavage blended fabric may result with $\kappa_{\text{MAX}}/\beta$ not $X$, since the cleavage-stylolites are rarely penetrative, (Fig. 17a). Like other deformation-mechanisms pressure solution conditions may be identified on a deformation mechanism map (Nicolas 1987), extended to include also the intergranular processes, collectively particulate flow (Borroma1e & Tarling 1981, 1984), (Fig. 22).

**Diffusion**

Particularly at high homologous temperature, say $>0.5T_M$ ($T_M$ is the melting temperature in K), cation-diffusion pervades the lattices (Nabarro-Herring Creep) whereas diffusion follows grain boundaries at lower temperatures (Coble Creep). Change of crystal shape may be significant and it is not limited by grain interactions as in the March model. Effectively, all grain boundaries are mobile with the available thermal energy. This permits alignments of minerals of even moderately weak anisotropy, giving rise to well-defined AMS fabrics, in rocks with seemingly poor tectonic fabrics such as deep continental granulites or mantle harzburgite (Benn & Allard 1988; Borroma1e et al. 1999b; Borroma1e & Lagroix 2001). Diffusive processes are particularly sensitive to grain-size and to grain-shape which affect the diffusion-route.

**Crystal plasticity**

Early petrofabric work recognized strongly aligned crystal optic axes for matrix minerals such as quartz, calcite, olivine and pyroxene despite their weakly anisometric shapes (Sander 1930). Until metallurgical concepts of dislocation
motion were available in the 1950s, alignment mechanisms were poorly understood. However, in geology, causal correlations with structural/ petrofabric elements (schistosity, mineral lineation, L-S fabric) were acknowledged (Turner & Weiss 1963). For example, many mechanisms align quartz or calcite c axes steeply to S-planes, otherwise expressed by mica-alignment. Crystal-dislocations rearrange and multiply during any strain history, assisting the validation of Von Mises' criterion, that five independent dislocation slip systems suffice to strain (~align) any crystal outline arbitrarily. Fewer systems suffice in practice since one system accommodates most slip, accompanied by intercrystalline motion (particulate flow). Characteristic, simple c axis distributions (small/great circle dispersions, point-clusters, etc.) are computer-modelled readily for the ideal polycrystal but in nature grain-interactions produce less neat petrofabric stereograms that may relate weakly to strain (and AMS).

Laboratory experiments confirm theoretical predictions for the most part, though not as neatly and still with the simplification that "tri-axial" rig experiments usually impose macroscopic coaxial axial-symmetric strain (X = Y > Z), limited to ≤40% Z-shortening. Strain rates are normally high (~10^{-5} to 10^{-6} s^{-1}); >10^6 times faster than regional orogenic deformation, so that petrofabrics are replicated only for extreme processes (hot diffusion versus cool cataclastic flow). Stress-relaxation test permit processes to be inferred at much lower strain rates but the strains and textures may be undetectable. Crystal-plastic ODs are mostly temperature-dependent so that metamorphic grade may be useful in the interpretation of AMS.

Experimental petrofabric ODs are usually produced at higher strain-rates and temperatures than their natural counterparts, and experimental strain histories are usually coaxial (with X = Y ≠ Z). Whereas the stereogram usefully represents AMS and petrofabric ODs in a general natural tectonic or geographical reference frame, a crystal-axis coordinate system clarifies the relationships between ODs and strain for high-symmetry minerals with the experimental strain-symmetry (Owens & Rutter 1978). For high symmetry minerals the crystal axis coordinate system also corresponds to the AMS axes; important experimental evidence is available for trigonal quartz and calcite; orthorhombic olivine and orthopyroxene (e.g. Nicolas & Poirier 1976). Using crystal axes as a reference frame, the inverse pole figure illustrates the OD of the Z axes, recalling its limitation to \((X = Y)\) and high-symmetry minerals in a polycrystal with an isotropic initial OD (Fig. 23) (Nicolas 1987). The inverse pole-figure shows that in a plastically aligned polycrystal, petrofabric patterns are already quite complex and not always simply unimodal. For example, the distribution of some critical AMS axis (e.g. \(c/c\) axis, for calcite/quartz) may be dispersed along a small circle and difficult to interpret in terms of strain or kinematics.

Equating neat theoretical or experimental PCOs too closely with AMS is perhaps premature, especially since for single-crystals, low-symmetry usually weakens the relation between AMS and crystal-controlled deformation mechanisms (Fig. 1). Higher symmetry may permit more slip systems and the greater potential for crystal-plasticity. It may even be difficult to validate known strain histories from laboratory-produced petrofabrics for quartz, calcite, olivine and orthopyroxene (Nicolas & Poirier 1976). A further layer of symmetry constraints obfuscates the interpretation of strain-orientations from AMS. In nature \textit{vis-à-vis} most experiments, further complications ensue since the OD is influenced by metamorphic grade (~temp.), \(X ≠ Y\) and the strain history may be non-coaxial.

\section*{Aggregate processes}

\textbf{Dynamic recrystallization}

Deformation-rheology under metamorphic conditions is often governed by competition of work-hardening and recovery from that increased strain-energy state. Depending on temperature and strain-rate recovery may reduce grain-size to different degrees, in the first instance forming subgrains, rather than truly independent grains. Elongate ribbon grains, sub-grain and grain rotation may accompany these processes but the tendency is toward grain-size reduction. These processes are well known in crustal quartz and feldspar and mantle pyroxenes. However, AMS in such tectonites may be an indirect consequence of the alignment of other, higher-\(\kappa\) minerals.

\section*{Deformation mechanisms, metamorphism and AMS}

Structuralists conveniently collect thoughts on deformation mechanisms on a deformation mechanism map (Fig. 22). Differential stress, temperature, strain-rate and perhaps fluid pressure are important physical controls on deformation mechanisms. The first two controls are commonly chosen for axes of the graph, and
Fig. 23. (a) Geographic or specimen coordinate systems are the usual and logical reference frame to study AMS applications. (b, e) However, the inverse pole figure is superior to interpret crystalline-deformation mechanisms in triaxial experimental deformation experiments (Tullis 1973; Owens & Rutter 1978). Crystal axes provide the reference frame in which we plot the Z axes (shortening-directions) responsible for crystal alignment (Nicolas 1987). (d, e) The procedure only has advantage for macroscopically coaxial strain, with \( X = Y \), as in most experimental deformation. Its use is mostly limited to high symmetry minerals such as trigonal quartz & calcite; orthorhombic olivine and orthopyroxene (Nicolas & Poirier 1976). These are fortunately of great interest in tectonics and petrofabrics (Fig. 1). (f) Possible expected crystallographic orientations with respect to Z in lherzolite (Nicolas & Poirier 1976). (g) In non-coaxial strain histories, the stereographic projection must still be used. Here AMS and calcite c-axes are compared for specimens with S-C shear zone textures (Borradaile & McArthur 1990).
strain-rate contours are plotted on this plane (rear wall of Fig. 22). However, diffusive processes, including pressure solution are grain-size dependent so that is sometimes chosen as an axis. The maps may be standardized to compare isomechanical materials, e.g. shear stress may be normalized to the material's shear modulus and temperature may be expressed as a fraction of the melting temperature \( T_M \) in K, or even re-expressed as a composite P-T crustal-depth function (Ranalli 1987). Simplified maps, valid for one mineral and one grain-size, like Fig. 22, show ranges of conditions in which certain processes are most competitive. The mechanisms shown are not restricted to their labelled field. As a thought-guiding and problem-solving tool, the concept of a deformation mechanism map assists interpretations of magnetic fabrics. Simple continuum-mechanical 'strain-response' models bear little resemblance to the processes that aligning crystal lattices. AMS orientations are mostly referred to some structural/petrofabric feature, which in turn, may be defined more-or-less directly by finite strain axes \((X, Y, Z)\) axes from an L-S fabric) or by a kinematic pattern (non-coaxiality from some special structure, e.g. S-C fabric, shear zone). However, in terms of material processes, multiple deformation mechanisms may operate. For example, dislocation creep may align lattices. Although the AMS might be a simple function of the PCO for that reason (Fig. 23) other mechanisms may operate simultaneously; e.g. in low temperature environments pressure solution may also occur. Moreover, the deformation mechanisms on the rear-plane of the map are not independently capable of reshaping grains in a natural aggregate. Heterogeneous grain-strain demands intergranular motion, discussed next.

**Particulate flow**

In all rocks, grain-strain requires intergranular motion. Finite strain of a polycrystalline, polyminerallc assemblage is impossible without intergranular motion. Various described as grain-boundary sliding or neighbour switching in different sciences, it is broadly particulate flow. This may be entirely dependent on grain-deformation as implied in the metallurgical and high-grade metamorphic contexts; it is a geometrical necessity to accommodate the incompatibility of new contiguous grain shapes (e.g. Flinn 1965a; Nicolas & Poirier 1976; Poirier 1985). However, for low temperature and high-strain rate environments particulate flow may be independent of the crystal-alignment processes and this may affect AMS interpretation (Borradaile 1981; Borradaile & Tarling 1984).

As an extreme case, soft-sediment deformation or fault-rock cataclasis involve extensive particulate flow that are respectively without and independent of grain-deformation. Fluid pressure may enhance independent particulate flow, particularly at low metamorphic grade, and suppress the onset of grain-strain that would otherwise be evident. Commonly grain-strain may be less than that of the aggregate and in a general non-coaxial strain history, the strain-analysis technique may determine which of several possible sets of \(X, Y, Z\) axes are reported and which of them may be related to AMS. Of course, mineral fabrics always control AMS, prima facie. Thus, interpreting AMS orientations requires some consideration of deformations mechanisms (Fig. 22) and the choice of strain analysis technique, some of which exclude the contribution of intergranular motion (Ramsay 1967; Ramsay & Huber 1983). The choice of strain analysis technique depends on the requirements; many methods measure grain-strain (\(R^tF^t\) of grains), some emasure intergranular strain (centre-to-centre method) and some may measure both (strain of large objects) (Borradaile 1981).

**AMS and metamorphism**

AMS orientation or ellipsoid-shape may vary with metamorphic grade or facies (Robion et al. 1997; Rochette 1987[Q101]; Nakamura & Borradaile, this volume). However, AMS is rarely interpreted in terms of metamorphic processes, due to their complexities. A few studies draw attention to the role of recrystallization/neomineralization mostly in classic areas, or with classic lithologies (Rochette 1987a; Bina & Henry 1990; Henry 1990; Housen & van der Pluijm 1991; Jackson & Borradaile 1991; Housen et al. 1993b, 1995). The principal problem is the extent to which the pre-existing tectonic fabric is masked by metamorphic recrystallization. In high-grade metamorphic environments the new fabric may nucleate in a near-perfectly aligned, saturation OD that obscures evidence of earlier subfabrics (Fig. 21a,b). On the other hand, at low grade, metamorphic recrystallization may be localized and develop subfabrics affecting only certain minerals, or certain location sub-fabrics (e.g. pressuresolution seams, strain-shadows, S-C foliations) (Aubourg et al. 1995; Robion et al. 1995).

These studies combine all the sophistication recommended to reduce interpretation-ambiguity; mineralogical/crystallographic considerations, subfabric ODs, multiple mineralogical controls, and subfabric isolation by AARM or