Journal of Structural Geology, Vol. 14, No. 1, pp. 65 to 71, 1992
Printed in Great Britain

0191-8141/92 $05.00+0.00

Separation of a component of tectonic deformation from a complex magnetic fabric

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(*Received* 9 *April* 1990; *accepted in revised form* 31 *May* 1991)

Abstract—A method has been developed for separation of the deformational component from a complex rock fabric. For the separation, data on the final (measured) fabric and the primary (pre-deformational) fabric are necessary as input. The primary fabric data are obtained either from measurements in terrains where gradual transitions from primary to deformational fabrics occur or, in the case of magnetic anisotropy, this may be obtained from data in the literature on individual rock types. Examples of the use of the method are presented for rocks from the Nízký Jeseník Mountains of the NE Bohemian massif.

INTRODUCTION

IN THE determination of rock strain based on the measurement of strain markers, the pre-deformational strain marker shape is usually considered spherical. Even though this approach is realistic in some cases (e.g. post-depositional reduction spots in sedimentary rocks, 'bird's eyes' in lapilli tuffs, some kinds of ooids), there are also cases in which the pre-deformational shapes of the strain markers are non-spherical (e.g. pebbles in conglomerates, mafic enclaves in granitic rocks, the fabric ellipsoid constructed from the preferred orientation of minerals). For the correct determination of strain in the latter case, the pre-deformational non-spherical shape of the strain markers must be allowed for.

In the past two decades attempts have been made to use the anisotropy of magnetic susceptibility (AMS) as a strain indicator. The relationship between the AMS and strain has been investigated theoretically (Owens 1974, Hrouda 1987, Rochette 1988, Henry & Hrouda 1989), empirically (for review see Hrouda 1982, Borradaile 1988, Rathore 1988) and experimentally (Owens & Rutter 1978, Borradaile & Alford 1987, 1988). Even though the results of these studies are not absolutely unambiguous, it is obvious that in many rocks with deformational magnetic fabrics the relationship of the AMS to strain is very close, and the AMS, after calibrating this relationship for a particular region on pilot specimens, can be used for rapid strain determination.

As shown by the investigation of undeformed sedimentary and volcanic rocks (for summary see Hrouda 1982), the primary magnetic fabric is never isotropic and, consequently, the magnetic fabric of a deformed rock is always complex, i.e. it is represented by a superposition of the deformational magnetic fabric on the primary one, unless the deformational overprinting is very strong. On the other hand, the variability in the primary magnetic fabric in both sedimentary and volcanic rocks is relatively limited (see Hrouda 1982) and this can be used advantageously in the estimation of strain from AMS measurements.

The first attempt to separate the deformational component of the AMS from a complex AMS was probably made by Hrouda (1979a), who developed the substitutional ellipsoid method for the case of coaxial or orthogonal superpositions of magnetic fabrics. Later this method was generalized by Goldstein (1980), Hirt *et al.* (1988) and Park *et al.* (1988) for obliquely oriented magnetic fabrics that are due to irrotational strains.

The purpose of the present paper is to develop a method for the determination of rock strain using pre-deformational non-spherical strain markers for the separation of the deformational component from a complex AMS, even for the more general case in which strains are rotational.

THEORY

Let us consider a rock whose fabric originated by a superposition of tectonic deformation on a primary (either tectonic or non-tectonic) fabric. If the generation of the primary fabric is formally regarded as a kind of straining, the relationship between the final, tectonic and primary deformations is as follows (see Elliott 1970)

$$\mathbf{D}_f = \mathbf{D}_t\mathbf{D}_p \quad \text{or} \quad \mathbf{D}_t = \mathbf{D}_f\mathbf{D}_p^{-1}, \tag{1}$$

where $\mathbf{D}_f$, $\mathbf{D}_t$, $\mathbf{D}_p$ and $\mathbf{D}_p^{-1}$ are the final, tectonic, primary and inverse primary deformation matrices, respectively. (Henceforth, the inverse matrix will be denoted by the superscript -1 and the transposed matrix will be denoted by a prime.)

The deformation matrix is in general asymmetric and very non-illustrative for physical and geological interpretation. In order to obtain a more convenient representation, it is advantageous to factorize this matrix into other matrices more simple to interpret. Among many possible factorizations, polar decomposition, providing easy mathematical manipulations and simple physical interpretation, is used most often (Truesdell & Toupin 1960, Elliot 1970). In this decomposition, the deformation matrix is equal to the product of a symmetric matrix (representing the distortion) and an orthogonal matrix (representing the rigid body rotation). Because of the non-commutative nature of matrix multiplication, there is the left polar decomposition (in which the distortion follows the rotation) and the right polar decomposition (in which the rotation follows the distortion) (see Elliott 1970)

$$\mathbf{D} = \mathbf{TR} \quad \text{and} \quad \mathbf{D} = \mathbf{RU}, \tag{2}$$

where $\mathbf{R}$ is the rotation matrix, $\mathbf{T}$ is the left stretch matrix and $\mathbf{U}$ is the right stretch matrix (notice that $\mathbf{T} \neq \mathbf{U}$). Using these factorizations and the fact that the inverse of an orthogonal rotation matrix equals its transpose (Flinn 1979), equation (1) can be written as

$$\mathbf{D}_t = \mathbf{T}_t\mathbf{R}_t = \mathbf{T}_f\mathbf{R}_f\mathbf{R}_p'\mathbf{T}_p^{-1} \tag{3a}$$

$$\mathbf{D}_t = \mathbf{R}_t\mathbf{U}_t = \mathbf{R}_f\mathbf{U}_f\mathbf{U}_p^{-1}\mathbf{R}_p'. \tag{3b}$$

It follows from equations (1) and (3) that the tectonic deformation matrix can be calculated if the complete final deformation matrix and the complete primary deformation matrix are known, this is both the distortional and rotational components of the primary and final deformations must be known. The final stretch matrix can be obtained from the marker shapes or AMS measurement of the rock investigated, and the primary stretch matrix can be estimated in terrains where transitions from primary to deformational fabrics occur or, in the case of AMS, also from published data on the primary magnetic fabrics of individual rock types (the best studied rocks from this point of view are sediments—see for example Rees & Woodall 1975, Rees 1983). (The relationship between the AMS and strain will be discussed later.) Unfortunately, the rotation matrix can be obtained neither from the AMS data nor from the marker shape measurements; it can be obtained only rarely, for example if the orientations of two lineations or foliations are known before deformation and after deformation, and mostly it cannot be determined at all.

In many cases, the process of formation of the primary fabric does not involve a rigid body rotation (for example, deposition and gravitational compaction of a sedimentary rock). In this case the primary rotation matrix equals the unit matrix and there is no distinction between primary left and right stretch matrices; therefore, only one primary stretch matrix will be used. Then, the equations (3) simplify to

$$\mathbf{D}_t = \mathbf{T}_t\mathbf{R}_t = \mathbf{T}_f\mathbf{R}_f\mathbf{S}_p^{-1} \tag{4a}$$

$$\mathbf{D}_t = \mathbf{R}_t\mathbf{U}_t = \mathbf{R}_f\mathbf{U}_f\mathbf{S}_p^{-1}, \tag{4b}$$

where $\mathbf{S}_p = \mathbf{U}_p = \mathbf{T}_p$.

From these equations it is obvious that for calculation of the tectonic deformation one needs to know both the final and primary stretch matrices and the final rotation matrix. However, obtaining the last matrix, as pointed out earlier, is difficult. Nevertheless, in many cases the rotation is represented by the rotation of an originally horizontal plane about a horizontal or inclined axis (for example, buckle folding of bedding). Then, the final rotation matrix is (see Flinn 1979)

$$\mathbf{R}_f = \mathbf{R}_1\mathbf{R}_2\mathbf{R}_3\mathbf{R}_2'\mathbf{R}_1', \tag{5}$$

where $\mathbf{R}_1$, $\mathbf{R}_2$ and $\mathbf{R}_3$ are, respectively, rotations about a vertical axis (by the angle of trend of the fold axis), about an E–W horizontal axis (by the angle of plunge of the fold axis), and about the fold axis (by the angle of rotation about this axis). (For definition of these matrices see Flinn 1979.) If there are indications that the primary fabric rotated before tectonic distortion (for example, if bedding underwent buckle folding before tectonic straining associated with slaty cleavage development), the left polar decomposition (equation 4a) should be used in the calculation of the tectonic deformation matrix.

Sometimes there are indications that the primary fabric did not rotate before tectonic distortion, and the magnetic fabric underwent rigid body rotation after tectonic distortion (for example, if a sedimentary rock was vertically shortened after deposition and subsequently folded). In this case the right polar decomposition should be used (equation 4b). The left and right polar decompositions can be made numerically or, if a method for this is not available, the decompositions can be made as follows. The eigenvalues and eigenvectors of the matrices $\mathbf{D}_t\mathbf{D}_t'$ and $\mathbf{D}_t'\mathbf{D}_t$ (for left and right polar decompositions, respectively) give the squares of the principal tectonic strains (Sanderson 1982, appendix) and their orientations. Using equation (10) the tectonic left stretch or right stretch matrices can be calculated. Then, the tectonic rotation matrices can be determined

$$\mathbf{R}_t = \mathbf{T}_t^{-1}\mathbf{D}_t, \qquad \mathbf{R}_t = \mathbf{D}_t\mathbf{U}_t^{-1}. \tag{6}$$

From the rotation matrix the orientation of the axis of rotation as well as the angle of rotation can be calculated (see Flinn 1979). It should be noted that the tectonic rotation matrix, $\mathbf{R}_t$, can be calculated only if the final rotation matrix is known (knowledge of this is necessary for the calculation of the $\mathbf{D}_t$ matrix, see equations 4). It should also be noted that the $\mathbf{R}_t$ matrix does not equal the $\mathbf{R}_f$ matrix, because the $\mathbf{R}_f$ matrix comprises not only the tectonic rotation, but also the rotation arising from the superposition of the tectonic distortion on the primary distortion.

From the measurement of marker shapes or AMS only the final left stretch matrix ($\mathbf{T}_f$) can be directly obtained, because its principal directions are in the same positions as they were during straining. On the other hand, the right stretch matrix ($\mathbf{U}_f$) cannot be directly obtained from marker shapes or AMS measurement, because the principal directions of this have been rotated by the $\mathbf{R}_f$ matrix (see equation 1) after distortion. Nevertheless, the $\mathbf{U}_f$ matrix can be obtained as follows. The measured tensor representing final distortion is factorized into its diagonal form matrix and the orientation matrix (equation 10). Then, the orientation matrix is rotated into the position before final rotation

$$\mathbf{O} = \mathbf{R}_f'\mathbf{O}_R, \tag{7}$$

where $\mathbf{O_R}$ is the orientation matrix specifying the orientations of the final principal strains after the final rotation. Then, using equation (10) the $\mathbf{U_f}$ matrix can easily be calculated.

The relationship between the AMS and strain, though not known in all its aspects, for the majority of deformational magnetic fabrics can, to a first approximation, be described satisfactorily as follows (see, for example, Hrouda 1982, Borradaile 1988, Rathore 1988)

$$L = (S_1/S_2)^a, \quad F = (S_2/S_3)^a, \quad P = (S_1/S_3)^a, \quad (8)$$

where $L = k_1/k_2$, $F = k_2/k_3$ and $P = k_1/k_3$ are the magnetic lineation, magnetic foliation and degree of magnetic anisotropy, respectively. $k_1 \geq k_2 \geq k_3$ are the principal susceptibilities and $S_1 \geq S_2 \geq S_3$ are the principal stretches. The value of a varies according to lithology, the carriers of magnetism in a rock, and the orienting mechanisms of magnetic minerals during straining. It ranges approximately from 0.02 to 0.2, being 0.05 on average (Hrouda 1990).

Assuming constant volume strain and using the principal susceptibilities normalized by their geometric mean, the relationship can be simplified to

$$K_i = S_i^a, \qquad K_i = k_i(k_1k_2k_3)^{1/3}, \qquad i = 1,2,3. \quad (9)$$

The stretch matrix can be expressed in diagonal form as follows,

$$\mathbf{s} = \mathbf{OSO}', \quad (10)$$

where $\mathbf{s}$ is the stretch matrix, $\mathbf{S}$ is the diagonal form of the $\mathbf{s}$ matrix and $\mathbf{O}$ is the matrix specifying the orientation of the $\mathbf{S}$ matrix. The $\mathbf{S}$ matrix is defined

$$\begin{gathered} S_{11} = S_1, \qquad S_{22} = S_2, \qquad S_{33} = S_3 \\ S_{12} = S_{21} = S_{23} = S_{13} = S_{31} = 0. \end{gathered} \quad (11)$$

Using equations (9)–(11) one can calculate the stretch matrices from the AMS and these can be used in the calculation of equations (4) and (7). If the a factor in equation (9) is not known for the rock investigated it is not possible to calculate the tectonic stretch from equations (4). Nevertheless, using the normalized susceptibility tensors instead of the stretch tensors, the tectonic components of the AMS can be calculated provided that the relationship of AMS to strain is of the type defined in equations (8) and (9).

EXAMPLES

For an illustration of how the method works, some examples were selected from the Lower Carboniferous flysch (Culm) rocks of the Nízký Jeseník Mountains of the NE Bohemian massif. These rocks are characterized by the gradual development of structures from an almost undeformed sedimentary state, through spaced cleavage and slaty cleavage, to metamorphic schistosity (see for example Hrouda 1976, 1978, 1979a). In the eastern part of the Nízký Jeseník Mountains, one can observe bedding (sometimes gently folded) and, rarely, also sedimentary current structures in outcrop. In the central part of the mountains, spaced cleavage is also present, and in the western part the presence of asymmetric NW-vergent folds in bedding and extremely well developed slaty cleavage is characteristic. This slaty cleavage is much more conspicuous than bedding (which can only be identified by laborious searching for lithological boundaries between greywackes and slates), is almost perfectly planar and penetrative and keeps constant orientation over large areas regardless of position in the fold. In the NW-dipping fold limbs the slaty cleavage cuts the bedding at a high angle, while in the SE-dipping folds limbs it cuts bedding at a low angle, and in some places is even parallel to bedding.

Detailed analyses by Dvořák & Hrouda (1972) and Hrouda (1976, 1978, 1979a,b, 1981) have shown that in the eastern part of the region the deformation ended with the buckle folding of bedding and the development of thrust sheets, while in the western part the buckle folding was followed by the development of slaty cleavage, which was associated with ductile deformation mostly represented by strong shortening perpendicular to the cleavage.

For demonstration of the method, one outcrop in the eastern part of the Nízký Jeseník Mountains and two exposures in the western part were selected. The results are summarized in Table 1 and Figs. 1–5. In the table, the type of magnetic fabric (sedimentary, final, tectonic), type of decomposition, the values of the c, P and T parameters, as well as the trend and plunge of the rotation axis and the angle of rotation for each exposure, are presented. The parameters characterize the strain intensity, degree of anisotropy and shape of the suscepti-

Table 1. Strain and AMS parameters of selected outcrops of the Nízký Jeseník Mountains

Exposure	c	P	T	Axis*	Angle†	Strain	Type of decomposition
Tošovce	1.99	1.035	0.50			Primary	
	5.11	1.085	0.73	210/0	30	Final	
	2.80	1.053	0.68	26/2	32	Tectonic	Right
'Cleavage'	2.19	1.040	0.85			Primary	
	10.18	1.123	0.05	44/1	90	Final	
	8.90	1.115	0.52	58/4	75	Tectonic	Left
'Bedding'	2.19	1.040	0.85			Primary	
	18.80	1.158	0.69	44/1	34	Final	
	8.85	1.115	0.59	56/1	36	Tectonic	Left

*This represents the trend and plunge (in degrees), respectively, of the axis of rigid body rotation.
†This represents the amount of rotation (in degrees).

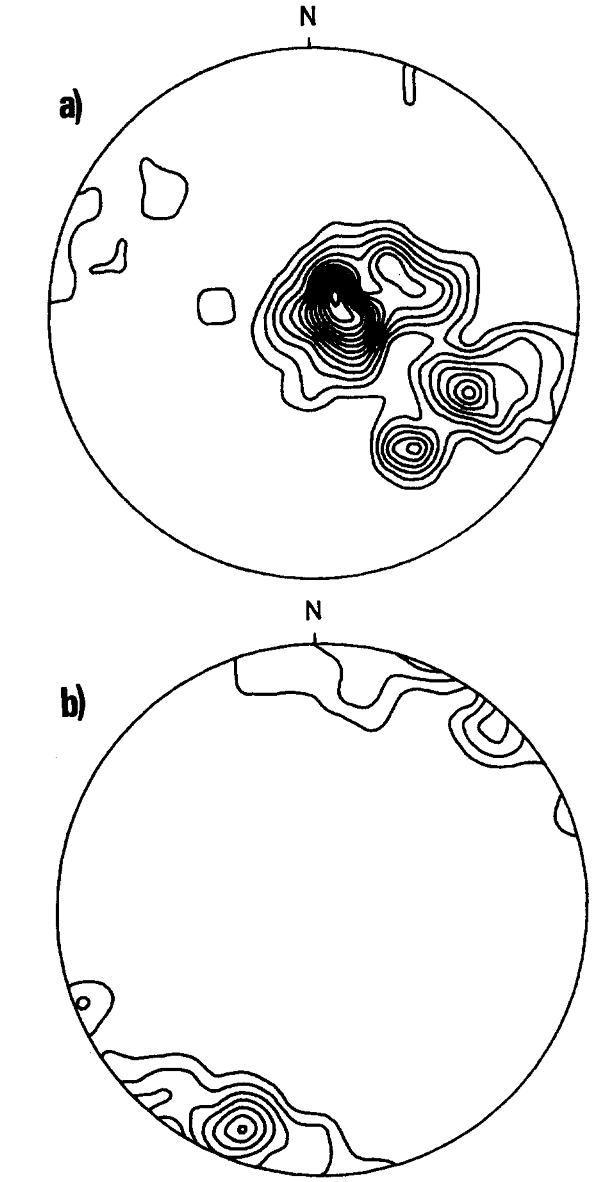

Fig. 1. Contoured plots of (a) magnetic foliation poles and (b) magnetic lineations in the Hradec–Kyjovice Formation. Equal-area lower-hemisphere projections in geographic co-ordinates. (a) The lowest contour is 0.5%, with steps at 2% intervals; (b) the lowest contour is 1%, with steps at 2% intervals.

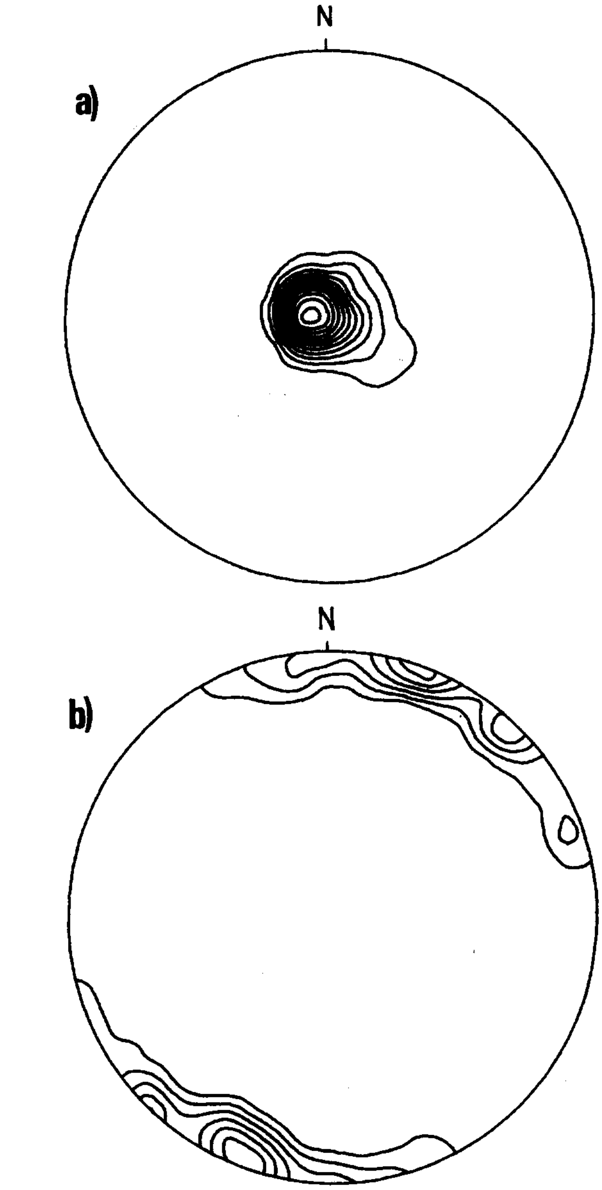

Fig. 2. Contoured plots of (a) magnetic foliation poles and (b) magnetic lineations in the Hradec–Kyjovice Formation in palaeogeographic co-ordinates. Equal-area lower-hemisphere projections. The lowest contour is 1%, with steps at 2% intervals.

bility ellipsoid (which is the same as that of the strain ellipsoid), respectively. They are defined as follows (Nagata 1961, Jelínek 1981)

$$c = S_1/S_3$$
$$P = k_1/k_3 \qquad (12)$$
$$T = 2\ln(k_2/k_3)/(\ln k_1/k_3) - 1.$$

If $0 < T \leq 1$ the magnetic fabric is planar, if $-1 \leq T < 0$ it is linear.

The figures present the following directional data: magnetic lineations and magnetic foliation poles, bedding poles, cleavage poles, bedding–cleavage intersections, fold axes and principal strain directions.

In the easternmost rocks (of the Hradec–Kyjovice Formation) of the Nízký Jeseník Mountains, the magnetic fabric is closely related to the sedimentary fabric. The magnetic foliation is nearly parallel to bedding and the pattern of the magnetic lineation is compatible with that of sedimentary lineations (Hrouda 1979b). The magnetic foliation poles create an imperfect girdle, oriented NW–SE (see Fig. 1a), while the magnetic lineations are mostly oriented NE–SW (Fig. 1b). However, in the so-called palaeogeographic co-ordinate system (i.e. that defined by horizontal bedding and geographical north) the magnetic foliation poles are more or less vertical (Fig. 2a) and the magnetic lineations mostly remain NE–SW (Fig. 2b). These facts, as well as the results of detailed fold analysis (Hrouda 1978), suggest that the folding in the Hradec–Kyjovice Formation was mostly represented by the buckling of strata without strong internal straining, which would be detectable by the AMS. During this folding not only the bedding but also the magnetic fabric rotated, and the buckle folding

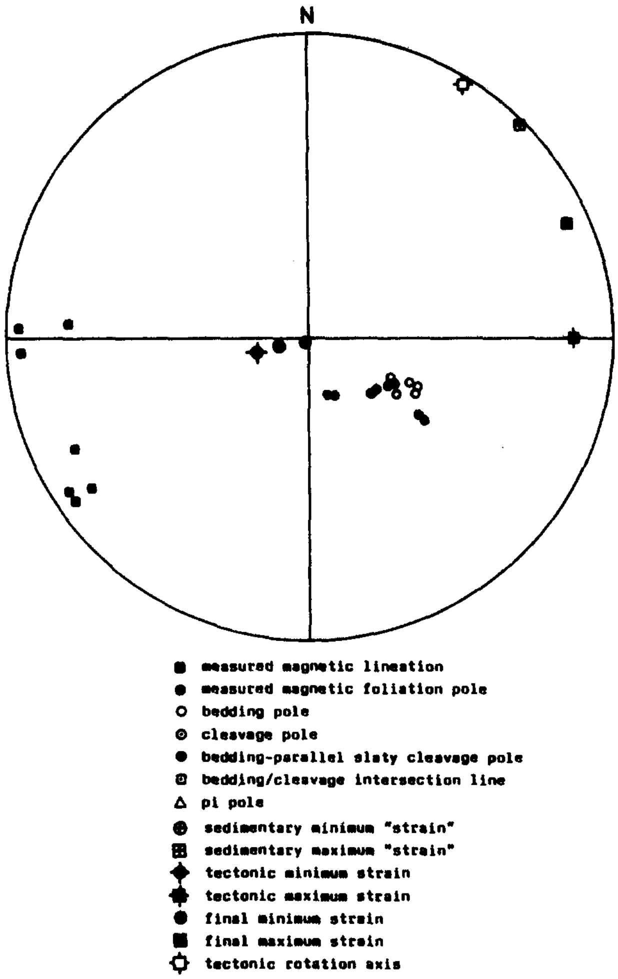

Fig. 3. Orientations of principal susceptibilities and principal strains in the Tošovce location. Equal-area lower-hemisphere projection in geographic co-ordinates.

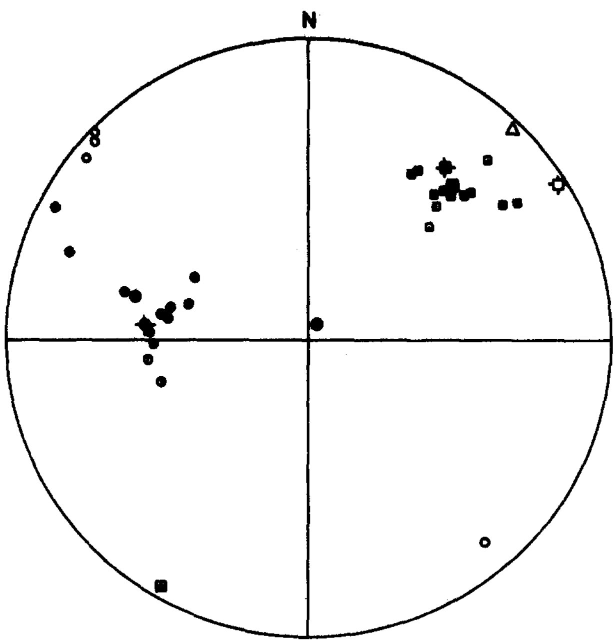

Fig. 4. Orientations of principal susceptibilities and principal strains in the Pasecký žleb valley exposure with slaty cleavage at a high angle to bedding. Equal-area lower-hemisphere projection in geographic co-ordinates. For legend see Fig. 3.

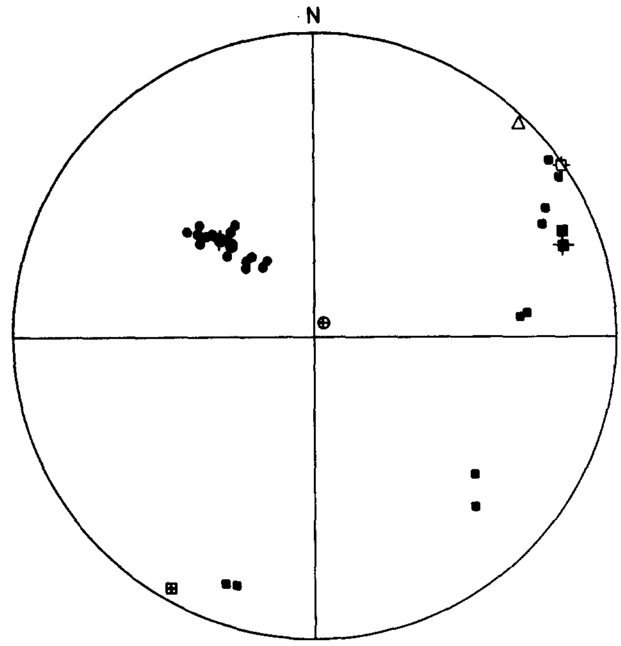

Fig. 5. Orientations of principal susceptibilities and principal strains in the Pasecký žleb valley exposure with bedding parallel to slaty cleavage. Equal-area lower-hemisphere projection in geographic co-ordinates. For legend see Fig. 3.

was the last deformation (rigid body rotation) that affected the magnetic fabric in an observable way.

To illustrate the use of the right polar decomposition, rocks of the Hradec–Kyjovice Formation at the locality of Tošovce were chosen. The results are summarized in Table 1 and Fig. 3. It is obvious that the degree of anisotropy at Tošovce is higher and the magnetic fabric is more oblate than in the undeformed flysch. (For example, in undeformed sandstones of the Flysch belt of the West Carpathians mean values of $P = 1.035$ and $T = 0.5$ are found, Hrouda 1991.) Also the magnetic foliations dip more steeply than usual after deposition. From this one can conclude that the rocks underwent some deformation. This deformation can be determined using our method. However, one needs to know the primary fabric. In our case, one can hardly obtain direct data on the primary magnetic fabric; one can only assume that, after deposition, the degree of anisotropy and the shape factor were similar to those of the undeformed flysch and that the magnetic foliation was roughly horizontal and the magnetic lineation was parallel to the current direction (these have been investigated sedimentologically by Kumpera 1984). As the measured final magnetic foliation is parallel to the bedding, it is likely that the rotational component of deformation (represented by buckle folding) that tilted the bedding took place later than the distortional component of deformation and thus the right polar decomposition should be used.

It can be seen in Table 1 and Fig. 3 that the calculated minimum tectonic stretch direction is nearly vertical and the maximum tectonic stretch direction is horizontal, oriented WNW–ESE. The strain ellipsoid is clearly of flattening type, the degree of anisotropy of the calculated tectonic deformation component is relatively high, the rotation axis is near the strike direction and the rotation angle is similar to the dip of bedding. It can be concluded that this deformation was probably represented by vertical shortening due to the weight of overlying strata. The W–E-oriented maximum tectonic stretch may represent weak stretching due to the thrusting.

To illustrate the use of the left polar decomposition, a pair of exposures in the Pasecký žleb valley in the western part of the Nízký Jeseník Mountains was selected. In the first exposure (denoted 'cleavage' in Table 1), where the slaty cleavage makes a high angle with bedding, the degree of anisotropy is relatively high, the magnetic fabric is triaxial with the magnetic foliation parallel to the cleavage and the magnetic lineation parallel to the bedding–cleavage intersection (Table 1, Fig. 4; for other details see Dvořák & Hrouda 1972). In the second exposure (denoted 'bedding' in Table 1), where the slaty cleavage is parallel to the bedding, the degree of magnetic anisotropy is very high, the magnetic fabric is oblate and the magnetic foliation is parallel to the bedding and slaty cleavage (Table 1, Fig. 5).

In the estimation of the strain associated with the formation of slaty cleavage, the primary AMS was obtained through computer matching of various primary AMS data in such a way that the AMS components due to the tectonic strain were similar in both the exposures investigated. This approach followed the idea that the strain affecting the two exposures occurring not far from each other should have been more or less the same, because it is associated with the generation of the slaty cleavage that is constant over large areas.

It is clear from Figs. 4 and 5 that the direction of minimum tectonic stretch in both exposures plots close to the cleavage pole and that the direction of maximum tectonic stretch plots fairly close to fold axis and/or the cleavage–bedding intersection. The rotation axis also plots near the fold aixs. The tectonic strain is relatively strong ($c = 9$) and of flattening type (Table 1). It can be concluded that this strain operated in association with the slaty cleavage development and was represented mostly by shortening perpendicular to the slaty cleavage.

The primary AMS obtained by the above computer matching is characterized by a degreee of anisotropy that is slightly higher than that usual for undeformed flysch sediments, but slightly lower than that found in the least deformed rocks of the Nízký Jeseník Mountains (viz. the Hradec–Kyjovice Formation, in which the mean anisotropy degree is $P = 1.054$, Hrouda 1979a). This may indicate that the rocks of the western part of the Nízký Jeseník Mountains were also vertically shortened before folding and cleavage development, but not as intensely as were the rocks of the eastern part.

DISCUSSION AND CONCLUSIONS

The method for separation of the component of tectonic deformation from a complex fabric was developed above all for deformation indicated by magnetic anisotropy. However, the whole procedure is divided into two parts, one working with strains and the other with the relationship of AMS to strain. Consequently, the method can be used for the separation of the deformational component from a complex rock fabric investigated by any method.

In the method presented here the relationship of AMS to strain was considered to be very simple (equations 8–11). Even though expression (8) is satisfactory in most cases, it is not the only one which can be used in our method. If other relationships are developed (for example, the theoretical models proposed by Owens 1974), the method can work for these too.

Unlike the methods used until now (Goldstein 1980, Hirt *et al.* 1988, Park *et al.* 1988), the present method allows for possible rotations before or after tectonic distortion. Even though the distortional component can in principle be calculated, using other methods, by ignoring the rotations (for example, by including rotation in the primary AMS), the results are not as accurate as those provided by our method, because the observable orientation of bedding corresponds to the final and not to the pre-tectonic state.

The success of the method developed here depends to a great extent on the validity of the relationship of AMS to strain. This relationship has been investigated in many papers studying the correlation between AMS and strain determined using non-magnetic methods, and in some papers modelling this relationship theoretically and in some papers investigating rock analogs deformed in the laboratory. These investigations have been reviewed by the present author (Hrouda 1989). It has been shown that, despite various formulae for the relationship of AMS to strain introduced by various authors, the relationship that is most simple and satisfactory is that given by equation (8). It can be shown that the other suggested formulae can be converted into equation (8) (e.g. Hrouda 1982, 1991). It is fair to note that some scientists doubt the existence of any relationship of AMS to strain; this opinion is partially expressed in the paper by Borradaile (1988). The present author does not want to present here detailed arguments against this pessimism, but would only like to state that only two papers revealed no or even an inverse correlation between AMS and strain (Borradaile & Mothersill 1984, Borradaile & Tarling 1984). All the other papers revealed a relatively close correlation between the AMS and strain that can be described well by equation (8).

The value of a in equation (8) is not constant: it depends on the mineral carrying the AMS (Hrouda 1987, Rochette 1988, Henry & Hrouda 1989) and probably also on the lithology of the rock investigated. If the determination of the tectonic component of strain using our method is to be correct, then the value of a must be known. The best way of obtaining this value is to investigate the relationship of AMS to strain on a small group of pilot specimens (for example in the way used by Hirt *et al.* 1988). If one is satisfied with only the determination of the tectonic AMS component, the determination of the a value is not critical and the average value $a = 0.05$ can be used.

The last factor affecting accuracy in the separation of the pre-deformational and deformational AMS is the reliability with which the pre-deformational AMS is known. In principle, this AMS can be known reliably only in rocks showing a gradual transition from the undeformed into a strongly deformed state. In other cases, all one can use is the published data on primary AMS of various rock types, and, consequently, the separation is not accurate. Fortunately, the natural variation in primary AMS is not large (e.g. Hrouda 1982) and the error arising from use of inaccurate pre-deformational AMS is not large. In any case, the use of published AMS data gives rise to better results than ignoring the pre-deformation AMS entirely.

Acknowledgements—The writer would like to thank P. Hudleston for his constructive criticism of the manuscript and for his help in analysis of the polar decompositions. The Director General of the State Company Geofyzika Brno, is thanked for permission to publish this paper. The Technical Development Foundation of Geofyzika Brno supported the research financially.

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