The magnetic anisotropy of rocks: principles, techniques and geodynamic applications in the Italian peninsula

Aldo Winkler, Laura Alfonsi, Fabio Florindo, Leonardo Sagnotti and Fabio Speranza (*)
Istituto Nazionale di Geofisica, Roma, Italy

Abstract
Magnetic anisotropy studies have recently come to the forefront as accurate, fast and inexpensive methods in the investigation of the rock fabric. In this paper we summarize the physical principles and the experimental techniques commonly used to resolve the Anisotropy of Magnetic Susceptibility (AMS) and the Anisotropy of Anhysteretic Remanence (AAR) tensors, and we give a description of the parameters which usually describe the magnetic anisotropy properties of a rock. A synthetic review of the magnetic fabric studies carried out on sedimentary rocks of the Italian peninsula is also given, discussing the potentiality of this technique in geodynamic studies.

Key words *magnetic susceptibility – magnetic remanence – anisotropy – fabric*

1. Introduction

The magnetization $\vec{M}$ of a rock induced by a weak magnetic field $\vec{H}$ (*e.g.*, less than 1 mT) is approximately a linear function of the inducing field;

$$\vec{M} = \hat{k}\vec{H},$$

$\hat{k}$ being a symmetric tensor of the second rank, called the susceptibility tensor;

$$\hat{k} = \begin{pmatrix} k_{11} & k_{12} & k_{13} \\ k_{21} & k_{22} & k_{23} \\ k_{31} & k_{32} & k_{33} \end{pmatrix} \quad \begin{cases} k_{ij} = k_{ji} \\ i, j = 1, 2, 3 \end{cases}. \tag{1.1}$$

Generally, the non-diagonal elements of the tensor are not zero, but a Cartesian co-ordinates basis can be found in which:

$$\hat{k} = \begin{pmatrix} k_1 & 0 & 0 \\ 0 & k_2 & 0 \\ 0 & 0 & k_3 \end{pmatrix}. \tag{1.2}$$

The eigenvalues $k_1 \geq k_2 \geq k_3$ are called principal susceptibilities and their directions (eigenvectors) are the principal susceptibility axes.

(*) CEE Contract EV5V-CT94-0464

Mailing address: Dr. Aldo Winkler, Istituto Nazionale di Geofisica, Via di Vigna Murata 605, 00143 Roma, Italy; e-mail: winkler@ing750.ingrm.it

The Anisotropy of Magnetic Susceptibility (AMS) is often geometrically represented by an ellipsoid, called the magnitude ellipsoid, in which the semi-axes are proportional to the AMS tensor eigenvalues. According to the eigenvalues, several parameters are used to quantify the anisotropy degree and the shape of the AMS ellipsoid (table I). Ellipsoid shapes are classified according to:

$k_1 > k_2 > k_3$ – the AMS ellipsoid is triaxial;

$k_3 \ll k_2 \approx k_1$ – the AMS ellipsoid has an oblate shape (fig. 1a);

$k_1 \approx k_2 \approx k_3$ – isotropic susceptibility, the AMS ellipsoid is a sphere (fig. 1b);

$k_1 \gg k_2 \approx k_3$ – the AMS ellipsoid has a prolate shape (fig. 1c).

Each mineral has a definite magnetic susceptibility, which ranges from very low negative values (of the order of 10^{-5} SI for the volume susceptibility, the susceptibility expressed in a unit volume basis) for diamagnetism (*e.g.*, quartz, calcite), to positive values which are low in paramagnetism (10^{-4}-10^{-5} SI; *e.g.*, clay minerals) and high in ferromagnetism (*e.g.*, magnetite, hematite). Thus, the overall low-field susceptibility of a rock is a summation of the contribution from all mineral species composing the rock, weighted according to their relative abundance and susceptibilities (*e.g.*, Hrouda, 1982; Jackson, 1991).

Two principal mechanisms control the magnetic anisotropy of rocks (*e.g.*, Halgedahl, 1989; O'Reilly, 1989): 1) lattice alignment of crystals with magnetocrystalline anisotropy; 2) shape alignment of ferromagnetic grains. Magnetocrystalline anisotropy is an intrinsic property of matter and arises from the coupling between the spin and the orbital motion of the electron; it governs the orientation of the spin axes with respect to the crystal structure to determine «easy» and «hard» directions of magnetization in a crystal lattice. The shape anisotropy mostly occurs in ferromagnetic minerals with high intrinsic susceptibility and low degree of magnetocrystalline anisotropy (*e.g.*, magnetite, maghaemite), where just a slight deviation from the isotropic shape produces notable magnetic anisotropies, in order to minimize the magnetostatic energy. Concerning the shape anisotropy, the effective susceptibility k_e along a direction i of a ferromagnetic grain, is given by:

$$k_e = \frac{k_i}{1 + N_i\, k_i}\,, \tag{1.3}$$

where N_i is called demagnetization factor, which minimizes in the direction of the long axis of the grain, and k_i is the intrinsic susceptibility of the grain in the i direction; see fig. 2 for an example in a disc-shaped grain.

Table I. Most commonly used AMS parameters.

Parameter name	Parameter symbol	Parameter formula	Author
Anisotropy degree	P	k_1/k_3	Nagata (1961)
Corrected anisotropy degree	P'	$\exp\sqrt{\{2[(\eta_1-\eta)^2+(\eta_2-\eta)^2+(\eta_3-\eta)^2]\}}$	Jelinek (1981)
Magnetic lineation	L	k_1/k_2	Balsley and Buddington (1960)
Magnetic foliation	F	k_2/k_3	Stacey *et al.* (1960) Balsley and Buddington (1960)
Ellipsoid shape	T	$2(\eta_2-\eta_3)/(\eta_1-\eta_3)-1$	Jelinek (1981)
Mean susceptibility	k_m	$(k_1+k_2+k_3)/3$	Nagata (1961)

$\eta_1 = \ln k_1$, $\eta_2 = \ln k_2$, $\eta_3 = \ln k_3$, $\eta = (\eta_1+\eta_2+\eta_3)/3$.

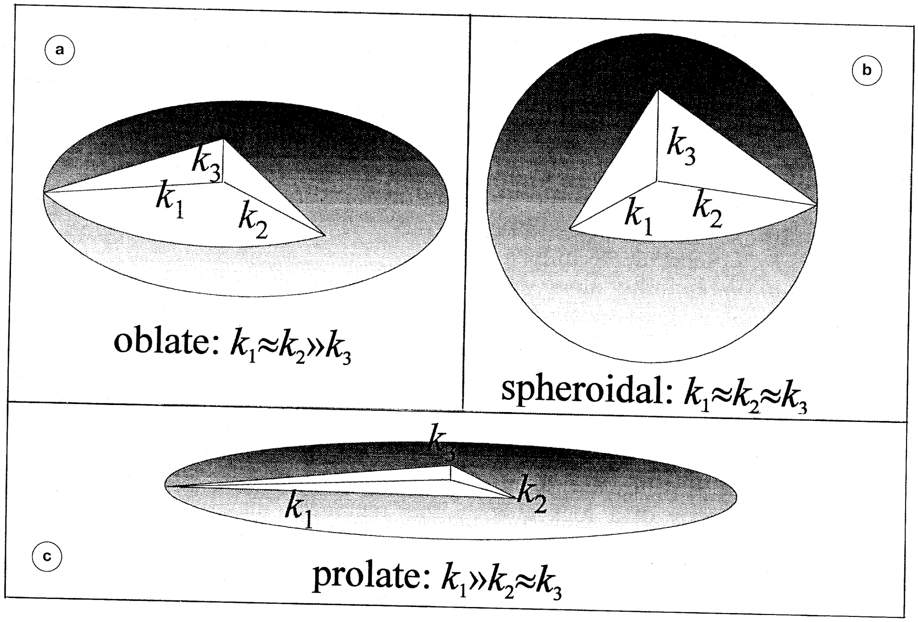

Fig. 1a-c. a) Oblate; b) isotropic and c) prolate AMS ellipsoids.

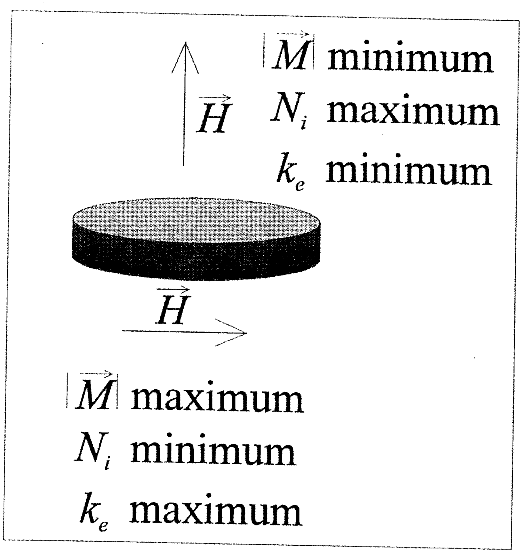

Fig. 2. Anisotropy of magnetic susceptibility on a disc-shaped grain; the effective magnetic susceptibility k_e has a maximum value if the magnetic field $\vec{H}$ is parallel to the horizontal long axis of the grain and has a minimum value if $\vec{H}$ is parallel to the vertical short axis.

Hence, the magnetic anisotropies of a rock result from the preferred orientation of the mineral shapes and/or the crystallographic axes. The magnetic fabric of a rock reflects the geometrical and spatial distribution of the constituting grains, in dependence of their intrinsic magnetic properties, degree of alignment and abundance. As a rule of thumb, the AMS of a rock is mostly controlled by the average preferred orientation of the paramagnetic and/or diamagnetic matrix, with a subordinate ferromagnetic contribution, when the mean susceptibility k_m is less than about $3\text{-}5\times10^{-4}$ SI and $P < 1.35$, whereas it reflects the average preferred orientation of the ferromagnetic fraction if k_m is of the order of 10^{-3} SI or higher with $P \gg 1.35$ (*e.g.*, Rochette, 1987; Tarling and Hrouda, 1993).

2. Methods

2.1. *Determination of AMS*

Several different techniques are employed to determine the AMS tensor of a rock sample (*e.g.*, Tarling and Hrouda, 1993); a standard specimen for magnetic anisotropy studies is a cylinder of 2.2 cm height and 2.5 cm diameter (fig. 3a). In the commercial device Kappa-bridge KLY-2, in which the magnetic field applied is about 0.1 mT, it is possible to measure magnetic susceptibility as low as 10^{-6} SI and anisotropy degree $P < 1.01$. The AMS tensor is determined by a procedure of 15 measurements of magnetic susceptibility along different directions (fig. 3b). A least-square best fit method is then used to calculate the six independent elements of the AMS tensor according to the method of Hext (1963), adapted to a 15 measurement cycle by Jelinek (1977); errors are estimated according to residuals (k_{measured}-$k_{\text{estimated}}$) for each measurement direction. The magnetic anisotropy of a rock at a given site is computed from the AMS data of a statistically significant set of independent specimens (at least 6 per site). The statistical treatment of AMS data from a set of specimens is achieved by a multivariate technique (Jelinek, 1978). This site-mean statistics has recently been checked and improved by means of Monte-Carlo simulation by Lienert (1991). AMS data for a set of samples are conventionally visualized in Schmidt equal-area projection, lower hemisphere, indicating k_1 axes by squares, k_2 axes by triangles and k_3 axes by circles (Ellwood *et al.*, 1988). Ellipses indicating 95% confidence regions are drawn around the mean principal axes; for an example see fig. 4.

2.2. *Determination of magnetic remanence anisotropy*

If a rock specimen is sequentially magnetized in several different directions and after each step the remanent magnetization is measured and then cleaned, the directional variability of the remanences defines the anisotropy of the remanent magnetization, which, as the AMS, can be symbolically expressed as a second rank tensor (*e.g.*, Jackson, 1991). The Anisotropy of Anhysteretic Remanence (AAR) is the most commonly studied anisotropy of remanence; an Anhysteretic Remanent Magnetization (ARM) is acquired by a sample when it is simultaneously exposed to an alternating magnetic field with decreasing amplitude (usually with a maximum peak in the range

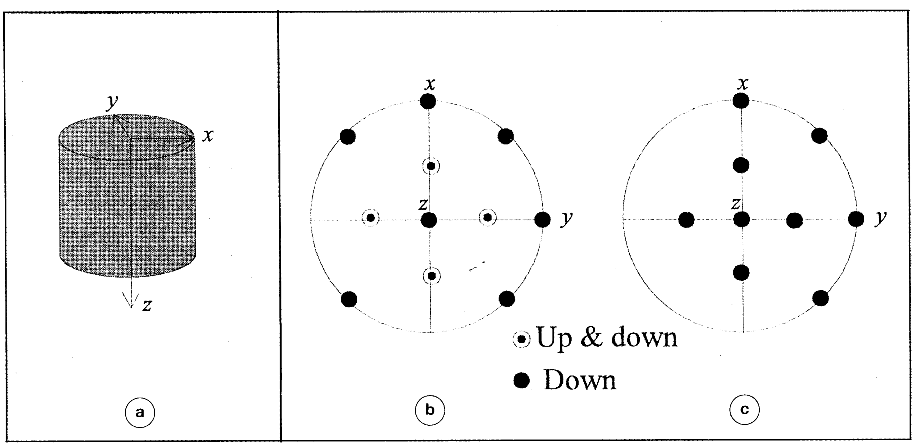

Fig. 3a-c. a) Standard (2.2 cm × 2.5 cm) palaeomagnetic specimen into its Cartesian reference; b) stereographic projection of the 15 measurement vectors for the AMS tensor definition, according to Jelinek (1977); c) stereographic projection of the 9 measurements directions for the AAR tensor definition, according to Girdler (1961).

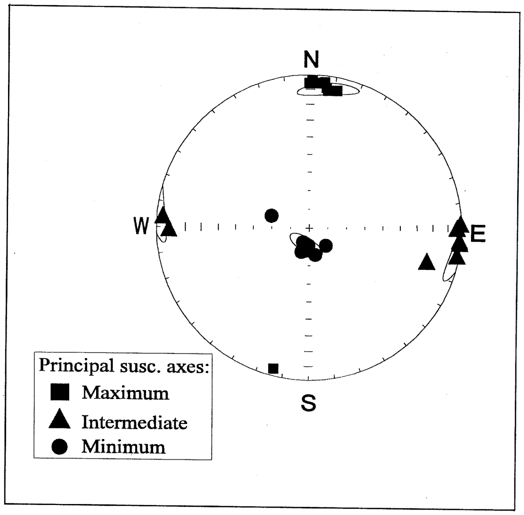

Fig. 4. An example of AMS data presentation: Schmidt equal-area projection, lower hemisphere, of the principal AMS axes from 8 samples collected at a site; the ellipses indicate 95% confidence regions around the mean principal axes.

10-100 mT) and a steady magnetic field (about 0.1 mT). In the method described by Mc Cabe *et al.* (1985) the sample acquires nine different anhysteretic remanences along as many axes (fig. 3c) and a best-fit tensor is then computed according to Girdler (1961), providing three ARM principal axes, with AAR parameters similar to those defined for the AMS.

Whilst the AMS tensor results from all the minerals (ferro, dia and paramagnetic) composing a rock, the AAR tensor is pertinent only to the ferromagnetic fraction. Knowledge of the ferromagnetic fraction anisotropy is particularly helpful in palaeomagnetism: large values of anisotropy degrees are symptomatic for bad remanence data, possibly indicating secondary magnetizations due to re-crystallization and/or large deflections of the magnetization vector. New methods based on the determination of ARM anisotropy degree provide a valuable correction for palaeomagnetic data of sediments affected by inclination error due to ferromagnetic particles with long axes lying in the bedding plane; for a review see Jackson *et al.* (1991).

3. Magnetic fabric of sedimentary rocks

A sedimentary rock acquires a primary magnetic fabric during its formation and diagenesis; this means that it is possible to consider a pure sedimentary fabric when the rock has undergone only depositional diagenetic and compactional processes, not influenced by tectonic events.

Many factors control the deposition of grains in a sediment: the first factor is the Earth's gravity field, which drives the particle's longest axes parallel to the depositional surface, the second one is the water current, which tries to orientate the longest axes of the grains parallel or perpendicular to the flow direction; the third factor, affecting only the ferromagnetic fraction, is the geomagnetic field, which rotates the longest axes of ferromagnetic grains parallel to the local magnetic meridian. For ferromagnetic grains, the influence of such factors depends on the grain size, the gravity field being more effective on the large Multi-Domain (MD) particles and the magnetic field more effective on the small Single Domain (SD) particles (*e.g.*, Verosub, 1977). When the deposition occurs in still water, the dominant factor is the gravity field, which makes all the platy grains lie in the depositional plane, with the long axes randomly aligned in the bedding plane. A foliated fabric is then emphasized by compaction. Thus, a typical pure sedimentary-compactional AMS fabric consists of oblate ellipsoids, with minimum susceptibility axes perpendicular to the bedding plane and maximum-intermediate susceptibility axes, corresponding to the long axes of the grains, scattered in the bedding plane (for a review on sedimentary AMS fabrics see Tarling and Hrouda, 1993). A magnetic lineation develops, mostly in sandy and coarse-grained sediments, only when a current flow is capable of producing alignment of the longest grain axes (Hamilton and Rees, 1970). If the currents are low (< 1 cm/s) the long axes of prolate grains are aligned parallel to the flow direction, producing a magnetic lineation parallel to the direction of the current; the susceptibility fabric is still oblate in the bedding plane due to the low imbrication angle. As the currents get stronger, the magnetic foliation plane is tilted by 5°-20° away from the bedding plane. The prolate shape grains are in this latter case more stable if they lie with their long axes perpendicular to the current direction; thus, the magnetic lineation sets perpendicular to the flow direction.

If a sedimentary rock undergoes deformation under a compressional regime, the sedimentary fabric is partly overprinted by a tectonic fabric, which becomes more and more important as the deformation goes on (*e.g.*, Graham, 1966; Kligfield *et al.*, 1981; Hrouda, 1982; Borradaile, 1988; Lowrie, 1989). A simple speculative diagram of magnetic fabric *vs.* deformation was first proposed by Graham (1966). The starting point (fig. 5a) represents a primary sedimentary fabric, with k_3 axes parallel to the bedding pole and k_2-k_1 axes scattered in the bedding plane. In the first stage of deformation the k_3 direction remains perpendicular to the bedding plane, and the k_1 direction is reoriented perpendicular to the maximum shortening direction in the bedding plane (fig. 5b). As the deformation goes on, the AMS ellipsoid becomes prolate with k_1 direction still perpendicular to the compressional axis (fig. 5c); for higher deformation the k_3 direction becomes parallel to the shortening direction, and the k_1 direction remains perpendicular to the shortening axis (fig. 5d); the ellipsoid gradually returns to an oblate shape. Further, with the development of cleavage planes, the k_3 axis becomes perpendicular to the cleavage plane, and k_1 lies along the stretching direction in the cleavage plane (fig. 5e,f).

The influence of tectonic deformation on the sedimentary-compactional AMS tensor in extensional regime is less known; the first data were reported from Plio-Pleistocene clays from the Tyrrhenian extensional margin in Central Italy and will be discussed in the next section.

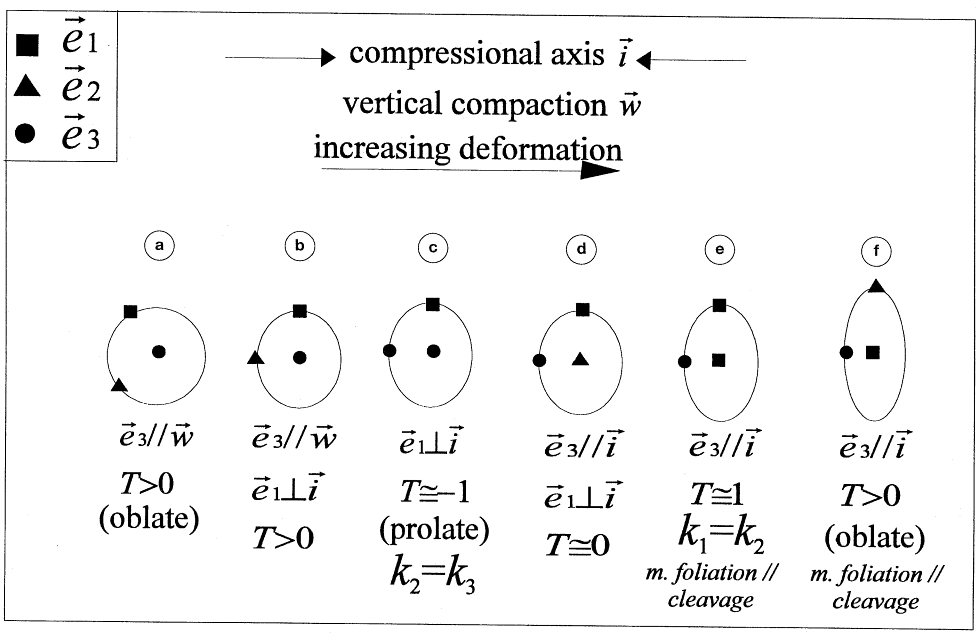

Fig. 5. Speculative diagram of the change of the AMS ellipsoid during progressive deformation of a sedimentary rock with initially pure depositional-compactional fabric (a). The plane of the figure is parallel to the bedding plane and the compression is purely horizontal; the eigenvectors related to k_1, k_2 and k_3 are indicated by $\vec{e}_1$, $\vec{e}_2$, $\vec{e}_3$ respectively. Modified after Graham (1966).

4. A synthetic review of magnetic anisotropy studies on sedimentary sequences of peninsular Italy

The magnetic anisotropy studies carried out in the past few years in peninsular Italy showed that the magnetic fabric analyses of fine-grained sediments can provide new insights into the empirical investigation of fabric to strain relationships during the very first stages of deformation. The available data allowed the reconstruction of the strain pattern in sedimentary sequences that almost lack visible strain markers. The method was shown to be effective in a variety of geological structures both in compressional and extensional settings.

Lowrie and Hirt (1987) first published AMS data from the late Cretaceous - early Tertiary Scaglia Formation of the Umbria Apennines. The authors showed that the magnetic anisotropy pattern in the Scaglia is compatible with the earliest stages of tectonic modification of an original compactional sedimentary fabric. The magnetic foliation was produced by vertical compaction during diagenesis. The magnetic lineation, less developed than the foliation, is close to the local fold axis direction and was produced by horizontal tectonic forces at an early stage of deformation in the late Tertiary, before the development of folds, but when the Scaglia was already indurated.

Sagnotti and Speranza (1993) carried out a first AMS study aimed at the definition of the internal strain affecting Plio-Pleistocene sediments in the Apennines. The results indicated a close consistency between the overall orientation of magnetic lineations found in the mildly deformed upper Pliocene - lower Pleistocene clays from the Sant'Arcangelo basin (Lucania) and the compressional directions indicated by meso-structural analysis in adjacent coarse-grained sediments (fig. 6). The magnetic fabric of these clays was interpreted as the result of a limited tectonic overprint related to compressional episodes on a primary, sedimentary-compactional, fabric. The origin of the magnetic lineation is supposed to be early diagenetic, due to syn-sedimentary tectonics acting on the sediment not yet lithified.

Scheepers and Langereis (1994) compared the AMS data of early Pleistocene sediments from different structures distributed in the Apulia-Gargano foreland, the Bradano unit, the Southern Apennines and Calabria. They confirmed the influence of tectonic deformation on the development of a magnetic lineation in fine-grained sediments. No magnetic lineation was found in the almost completely undeformed sediments from the Bradano unit and the Apulia-Gargano foreland. On the contrary, a well defined magnetic lineation was found in the deformed sequences from the Apennines and Calabria. Comparison with palaeomagnetically determined rotations induced the authors to postulate that the magnetic lineation was due to a middle Pleistocene compressional phase, after that differential block rotations took place in the Tyrrhenian arc.

Further integrated analyses of AMS and palaeomagnetic data from compressional structures of the Central Apennines (Mattei *et al.*, 1995) and the Marche Romagna Apennines (Speranza *et al.*, 1997; Sagnotti *et al.*, 1996) were recently reported. In these studies, the comparison of palaeomagnetic declination deviations and magnetic lineation deviations (reflecting the structural trends variation) indicated an oroclinal mechanism for the development of the present chain. In agreement with the evidence from the adjacent Scaglia Formation, it was concluded that the magnetic lineation was acquired during the early stage of deformation and passively rotated during the following tectonic phases. This magnetic lineation is recorded in sedimentary sequences that were deposited in almost straight foredeeps and were subsequently incorporated in the chain and rotated in the present-day configuration.

Averbuch *et al.* (1995) joined the AMS study to detailed meso-structural analysis in the Montagna dei Fiori structure. This study illustrates the potentiality of the AMS analyses to structural reconstruction; the authors concluded that layer parallel shortening influenced the magnetic fabric of the fine-grained levels of the Messinian Laga Formation before that folding and thrust emplacement occurred.

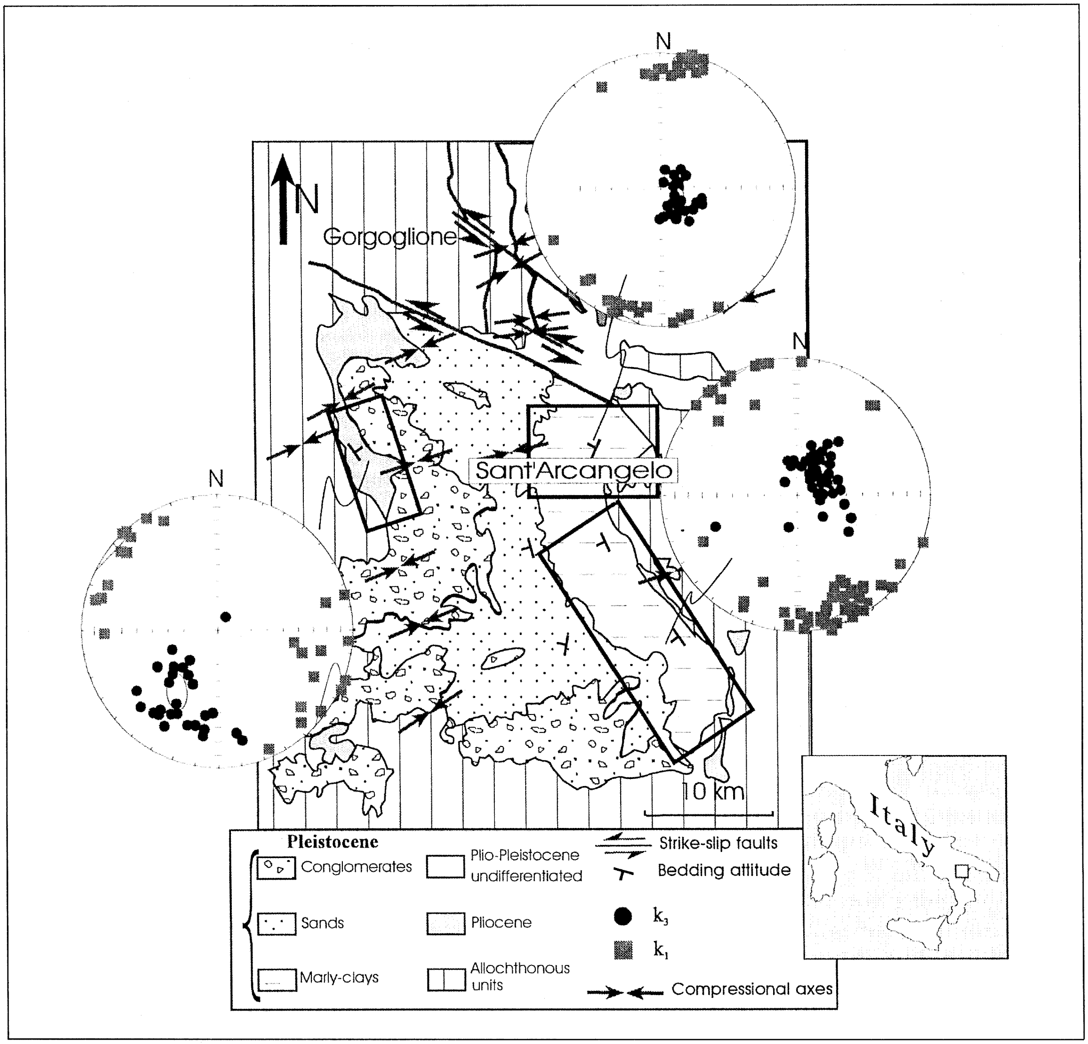

Fig. 6. The Sant'Arcangelo basin, Southern Italy: the three equal-area plots represent the *in-situ* AMS data from clayey sediments lacking classical structural markers. The magnetic lineation closely follows the bedding strike, which results sub-perpendicular to the local compressional axes (as determined from other lithotypes by Hippolyte *et al.*, 1991).

Sagnotti *et al.* (1994) pointed out a close relationship between the magnetic lineation and the stretching direction in the neo-autochthonous sequences from the extensional Tyrrhenian margin of Central Italy. The consistency between magnetic lineation and maximum stretching trends was also confirmed at a local scale. The authors concluded that the stretching processes that were able to modify the original magnetic fabric predated significant normal-fault-related tilting. The magnetic lineations are generally orthogonal to the main trend of the tectonic structures and parallel to the dip of bedding in roll-over antiform with rotational normal faults (fig. 7). The process of tectonic overprint of the primary fabric was supposed to work as follows: 1) extensional tectonics produce a well defined magnetic lineation in the bedding planes, when sediments are still almost sub-horizontal, parallel to the maximum stretching direction; 2) movements on rotational normal faults tilt the bedding and originate roll-over structures; then the magnetic fabric is not further overprinted and is passively tilted together with the strata.

Winkler and Sagnotti (1994) compared the AMS with the ARM anisotropy data from the upper Pliocene marly clays exposed at the Tiny quarry section in Valle Ricca, Central Italy (fig. 8a). They showed that while both k_3 and the minimum ARM axis are parallel to the bedding pole (fig. 8b,c), the maximum ARM axis is deviated 45° away from the maximum stretching direction (that is instead parallel to k_1, as indicated by the contour plot of poles to joint planes, fig. 8d). This suggests that only the clayey matrix fabric was slightly influenced by the stress that acted on these sediments and underlines that clay sequences with low ferromagnetic content are particularly suitable for the empirical investigation of AMS fabric to strain relationships in sediments at the very first stages of deformation.

Alfonsi and Sagnotti (1996) studied the AMS of Plio-Pleistocene sediments from the Fucino intermontane basin of the Central Apennines. The studied sequences crop out in the footwall of the seismogenetic structure responsible for the destructive 1915 Avezzano earthquake. The results showed a variable magnetic fabric; a distinct magnetic lineation, which is thought to be of tectonic origin, was found in the most anisotropic sites. The relationships of the magnetic fabric of these sediments with the main structural elements are however not fully understood.

In particular cases, the magnetic anisotropy studies indicated that the magnetic fabric of a

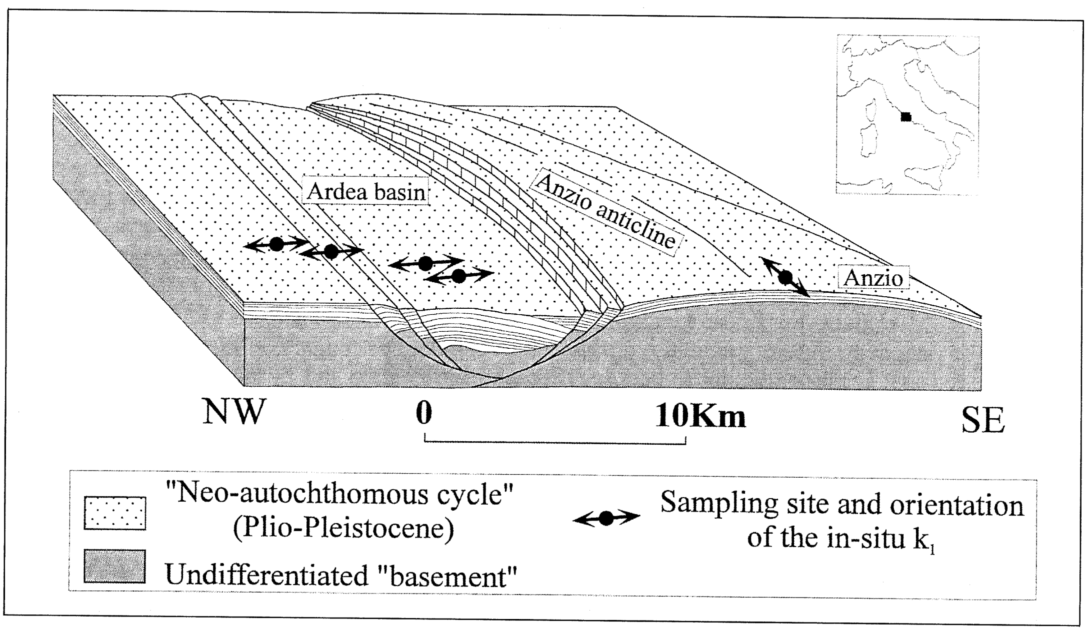

Fig. 7. Schematic block-diagram for the Ardea basin; arrows indicate mean magnetic lineations for the 5 sampled sites.

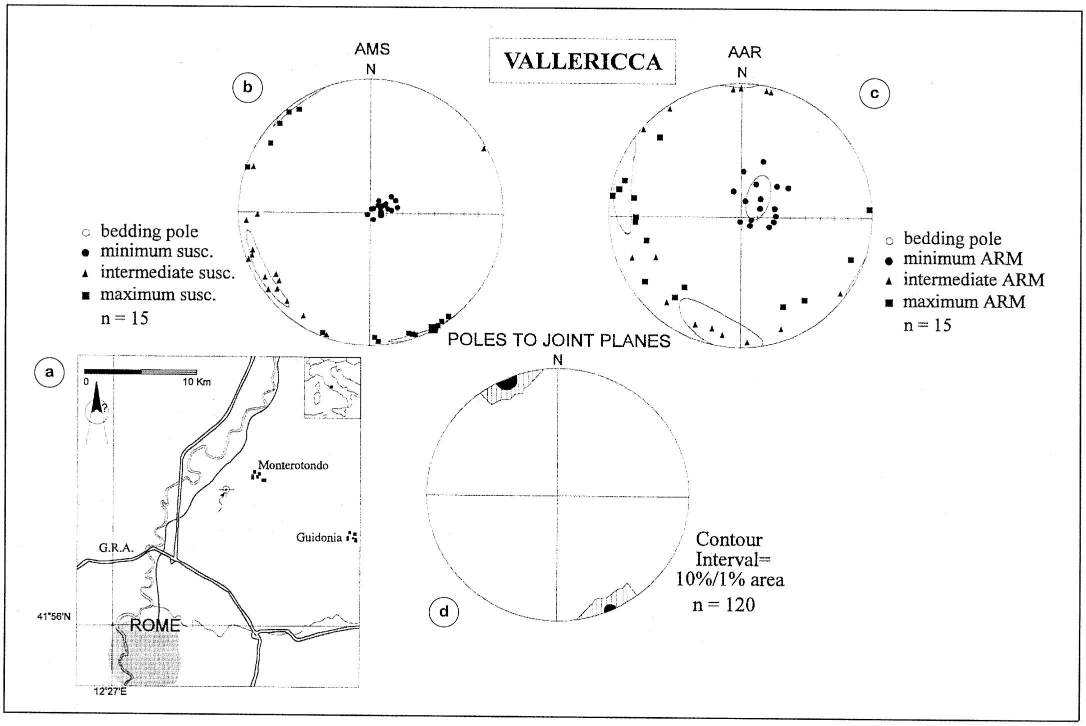

Fig. 8a-d. a) Location of the Vallericca section; Schmidt equal-area projection of AMS (b) and AAR (c) principal axes; the ellipses indicate 95% confidence regions around the mean principal axes. d) Contour plot of poles to joint planes at the Tiny quarry.

sediment may be almost completely controlled by a peculiar mineralogical composition and therefore it may not be directly related to sedimentary processes, compaction, strain and tectonics. In a Messinian marly sequence near Livorno the magnetic fabric gradually passes from inverse (*e.g.*, k_1 perpendicular to bedding) to normal (*e.g.*, k_3 perpendicular to bedding), due to a variable content of iron-bearing carbonates in the rock matrix (Winkler *et al.*, 1996).

The presence of large amounts of antiferromagnetic goethite was found to almost balance the normal fabric due to the paramagnetic clay minerals in a 40 m thick interval of the Pleistocene sequence near Montalto di Castro, producing an almost isotropic susceptibility fabric (Lehman *et al.*, 1996).

Anomalous magnetic fabrics were found in the Triassic «Calcari con Selce» limestones of the Lagonegro basin (Gialanella *et al.*, 1994). Inverse magnetic fabrics were found both from AMS and AAR analyses, which are respectively due to magnetite in the superparamagnetic (SP) state and in the stable SD state. The magnetic anisotropy data can explain the overstep characteristic remanent magnetization directions found with palaeomagnetic analyses. In these limestones paramagnetic minerals are virtually absent; both SP and SD magnetites have long axes subperpendicular to bedding and they probably suggest internal deformation of the limestones. Alternatively, magnetite may be due to chemical remagnetization activated before or during folding by the main Tertiary deformative events.

5. Conclusions

The anisotropy of magnetic susceptibility study is becoming a common tool for the determination of the rock fabric; its accurate applicability to practically every kind of rock and soft sediment, and the high sensitivity and repeatability of the measurements, allow fabric determination even in rocks lacking classical mesoscopic structural markers. The timely operation (no more than 15 min per specimen) is valuable for a statistical approach in the determination of the petrofabric, classical petrographic methods, such as micro-probe analysis and *X*-rays diffraction, being more elaborate and time-consuming. This technique, combined with the more recent analysis of the anisotropy of anhysteretic remanence, is also valuable to better constrain palaeomagnetic data, supplying a precise determination of bedding attitudes (when not directly detectable in the field), and a valuable correction for palaeomagnetic data affected by inclination errors. The AMS studies of Neogene and Quaternary sediments of the Italian peninsula provided an accurate estimation of the strain patterns in weakly deformed sequences from a variety of structural settings.

Acknowledgements

The authors thank E. Boschi and A. Meloni for their encouragement and R. Lanza for his careful review.

REFERENCES

ALFONSI, L. and L. SAGNOTTI (1996): Magnetic fabric of Plio-Pleistocene clayey sediments from the footwall of the Avezzano earthquake fault (Central Apennines, Italy), *Il Quaternario*, **9**, 145-154.

AVERBUCH, O., M. MATTEI, C. KISSEL, D. FRIZON DE LAMOTTE and F. SPERANZA (1995): Cinématique des déformations au sein d'un systeme chevauchant aveugle: l'exemple de la «Montagna dei Fiori» (front des Apennins Centraux, Italie), *Bull. Soc. Géol. France*, **166** (5), 451-461.

BALSLEY, J.R. and A.F. BUDDINGTON (1960): Magnetic susceptibility anisotropy and fabric of some Adirondack granites and orthogneisses, *Am. J. Sci.*, **258-A**, 6-20.

BORRADAILE, G.J. (1988): Magnetic fabrics, petrofabric and strain, *Tectonophysics*, **156**, 1-20.

ELLWOOD, B.B., F. HROUDA and J.J. WAGNER (1988): Symposia on magnetic fabric: introductory comments, *Phys. Earth Planet. Inter.*, **51**, 249-252.

GIALANELLA, P.R., F. HELLER and A. INCORONATO (1994): Rock magnetism of deformed upper Triassic limestones from the Lagonegro Basin (Southern Apennines, Italy), *Geophys. Res. Lett.*, **21**, 2665-2668.

GIRDLER, R.W. (1961): The measurement and computation of anisotropy of magnetic susceptibility in rocks, *Geophys. J. R. Astron. Soc.*, **5**, 34-44.

GRAHAM, J.W. (1966): Significance of magnetic anisotropy in Appalachian sedimentary rocks, in *The Earth Beneath the Continents*, edited by J.S. STEINHART and T.J. SMITH, *Am. Geophys. Un. Geophys. Monogr.*, **10**, 627-648.

HALGEDAHL, S.L. (1989): Magnetic domains, in *The Encyclopedia of Solid Earth Geophysics*, edited by D. JONES (Van Nostrand Reinhold Company, New York), 706-721.

HAMILTON, N. and A.I. REES (1970): The use of magnetic fabric in palaeocurrent estimation, in *Palaeogeophysics*, edited by S.K. RUNCORN (Academic Press, London), 445-464.

HEXT, G.R. (1963): The estimation of second-order tensors, with related tests and designs, *Biometrika*, **50**, 353-373.

HIPPOLYTE, J.C., J. ANGELIER, F. ROURE and C. MULLER (1991): Géométrie et mécanisme de formation d'un bassin «piggyback»: le bassin de Sant'Arcangelo (Italie Méridionale), *C. R. Acad. Sci. Paris*, Ser. II, **312**, 1373-1378.

HROUDA, F. (1982): Magnetic anisotropy of rocks and its application in geology and geophysics, *Geophys. Surv.*, **5**, 37-82.

JACKSON, M. (1991): Anisotropy of magnetic remanence: a brief review of mineralogical sources, physical origins, and geological applications, and comparison with susceptibility anisotropy, *Pure Appl. Geophys.*, **136**, 1-28.

JACKSON, M., S.K. BANERJEE, J. MARVIN, R. LU and W. GRUBER (1991): Detrital remanence, inclination errors and anisotropy of anhysteretic remanence: quantitative model and experimental results, *Geophys. J. Int.*, **104**, 95-103.

JELINEK, V. (1977): *The Statistical Theory of Measuring Anisotropy of Magnetic Susceptibility of Rocks and its Application*, Geofizika Brno, pp. 88.

JELINEK, V. (1978): Statistical processing of anisotropy of magnetic susceptibility on groups of specimens, *Studia Geophys. Geod.*, **22**, 50-62.

JELINEK, V. (1981): Characterization of magnetic fabric of a rocks, *Tectonophysics*, **79**, 563-567.

KLIGFIELD, R., W.H. OWENS and W. LOWRIE (1981): Magnetic susceptibility anisotropy, strain, and progressive deformation in Permian sediments from the Maritime Alps (France), *Earth Planet. Sci. Lett.*, **55**, 181-189.

LEHMAN, B., L. SAGNOTTI, A. WINKLER and C. LO CASCIO (1996): Magnetic mineralogy changes in the Pleistocene marine sequence of Montalto di Castro (Central Italy) and influences on the magnetic anisotropy, *Geophys. J. Int.*, **127**, 529-541.

LIENERT, B.R. (1991): Monte Carlo simulation of errors in the anisotropy of magnetic susceptibility: a second rank tensor, *J. Geophys. Res.*, **96**, 19539-19544.

LOWRIE, W. (1989): Magnetic analysis of rock fabric, in *The Encyclopedia of Solid Earth Geophysics*, edited by D. JONES (Van Nostrand Reinhold Company, New York), 698-706.

LOWRIE, W. and A.M. HIRT (1987): Anisotropy of magnetic susceptibility in the Scaglia Rossa pelagic limestone, *Earth Planet. Sci. Lett.*, **82**, 349-356.

MATTEI, M., R. FUNICIELLO and C. KISSEL (1995): Paleomagnetism and structural evidence for Neogene block rotations in the Central Apennines, Italy, *J. Geophys. Res.*, **100**, 17863-17883.

MC CABE, C., M. JACKSON and B.B. ELLWOOD (1985): Magnetic anisotropy in the Trenton limestones: Results of a new technique, anisotropy of anhysteretic susceptibility, *Geophys. Res. Lett.*, **12**, 333-336.

NAGATA, T. (1961): *Rock Magnetism* (Maruzen, Tokyo), pp. 350.

O'REILLY, W., (1989): Magnetic properties of minerals, in *The Encyclopedia of Solid Earth Geophysics*, edited by D. JONES (Van Nostrand Reinhold Company, New York), 721-730.

ROCHETTE, P. (1987): Magnetic susceptibility of the rock matrix related to magnetic fabric studies, *J. Struct. Geol.*, **9**, 1015-1020.

SAGNOTTI, L. and F. SPERANZA (1993): Magnetic fabric analysis of the Plio-Pleistocene clayey units of the Sant'Arcangelo basin, Southern Italy, *Phys. Earth Planet. Inter.*, **77**, 165-176.

SAGNOTTI, L., C. FACCENNA, R. FUNICIELLO and M. MATTEI (1994): Magnetic fabric and structural setting of Plio-Pleistocene clayey units in an extensional regime: the Tyrrhenian margin of Central Italy, *J. Struc. Geol.*, **16**, 1243-1257.

SAGNOTTI, L., F. SPERANZA, A. WINKLER and M. MATTEI (1996). Magnetic fabric of clayey sediments from the external Northern Apennines (Italy), *Annales Geophysicae*, **14** (I), C153.

SCHEEPERS, P. and C.G. LANGEREIS (1994): Magnetic fabric of Pleistocene clays from the Tyrrhenian arc: A magnetic lineation induced in the final stage of the middle Pleistocene compressive event, *Tectonics.*, **13** (5), 1190-1200.

SPERANZA, F., L. SAGNOTTI and M. MATTEI (1997): Tectonics of the Umbria-Marche-Romagna Arc (Central-Northern Apennines, Italy): new paleomagnetic constraints, *J. Geophys. Res.*, **102**, 3153-3166.

STACEY, F.D., G. JOPLIN and J. LINDSAY (1960): Magnetic anisotropy and fabric of some foliated rocks from S.E. Australia, *Geofiz. Pura Appl.*, **47**, 30-40.

TARLING, D.H. and F. HROUDA (1993): *The Magnetic Anisotropy of rocks* (Chapman and Hall, London), pp. 217.

VEROSUB, K.L. (1977), Depositional and post-depositional processes in the magnetization of sediments, *Rev. Geophys. Space Phys.*, **15**, 129-143.

WINKLER, A. and L. SAGNOTTI (1994): Anisotropies of anhysteretic remanence and magnetic susceptibility of marly clays from Central Italy, *Annali di Geofisica*, **37**, 1027-1035.

WINKLER, A., F. FLORINDO, L. SAGNOTTI and G. SARTI (1996), Inverse to normal magnetic fabric transition in an upper Miocene marly sequence from Tuscany, Italy, *Geophys. Res. Lett.*, **23**, 909-912.