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Method for estimating ductile horizontal strain from magnetic fabrics in poorly consolidated clay-rich sediments

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ABSTRACT

We propose a simple method based on the normalized eigenvalues of the tensor of anisotropy of magnetic susceptibility (AMS) to estimate horizontal ductile strain in detrital clay-rich sedimentary rocks subjected to layer parallel shortening (LPS). This method is mainly based on two assumptions: (1) The magnetic carriers are contacting clay-mineral platelets, and (2) The platelets respond to LPS by crenulation and formation of a zone axis of unchanged magnetic susceptibility perpendicular to the plane of deformation. We further simplify the problem by considering, as the exact distribution of particles orientations is unknown, that the AMS tensor arises from a single equivalent clay platelet, of which the angle with the bedding plane directly reflects the amount of horizontal strain. The proposed method is applied to data from samples retrieved during Integrated Ocean Drilling Program (IODP) Expeditions 315 and 316 of the Nankai Trough Seismogenic Zone Experiment (NanTroSEIZE) complex drilling project. Sample-wise finite strain values assist AMS data interpretation, which is traditionally based on the AMS ellipsoid shape parameter, and allow comparison with strain estimates obtained by other means. In the Nankai Accretionary Prism, AMS-derived strain values range from 0 to 25% and suggest substantial mechanical compartmentalization with inter-unit strain contrasts larger than 5 percentage points, including the penetrated shallow portion of a megasplay fault. At a local scale, AMS strain also appears to be affected by thrusting, what can be interpreted as a record of a ductile strain component away for the brittle zone or as an 'interseismic' ductile strain within otherwise brecciated material.

1. Introduction

The anisotropy of magnetic susceptibility (AMS) was long proposed as a potential tool to probe ductile strain, based on the observation that the AMS fabric may in certain conditions directly reflect the rock fabric. This relationship, which basically hinges on the assumption that the magnetic signal is primarily carried by elongated particles rotating according to a prescribed strain field, was investigated empirically (e.g. Borradaile, 1991; Borradaile and Henry, 1997; Borradaile and Jackson, 2010; Hirt et al., 1995), experimentally (Borradaile and Alford, 1987, 1988; Borradaile and Puumala, 1989; Owens and Rutter, 1978), and theoretically (Jezek and Hrouda, 2002; Hrouda, 1993; Hrouda and Jezek, 1999; Tarling and Hrouda, 1993). Most models describing that relationship are based on the March theory (Fernandez, 1981; Henry and Hrouda, 1989; March, 1932; Owens, 1974), but a variety of approaches have been put forth, some requiring various input parameters such as the different carriers of the magnetic signal, their initial distribution and rheology, the macroscopic strain field and the type of interactions between the particles and surrounding matrix (Hrouda, 1993; Jezek and Hrouda, 2002). In essence, however, the AMS data set for a given sample only consists of three independent values which can either be described as the three eigenvalues of the AMS tensor, or as an average susceptibility K_m and, respectively, the anisotropy (P) and shape (T) parameters. If one desires to derive information on strain based solely on AMS data, the simplest possible model should be sought so as to avoid under-determination of the problem, even though the range of applicability of the model might be limited as a result.

Pares et al. (1999) identified accretionary Prism Sediments as ideal candidates for examining the AMS/strain relationship because of their high porosity, high water content and low degree of cohesion, which overall allow particles to slide and rotate easily under the influence of tectonic strains. Within the larger context of mudrock consolidation and diagenesis, and as far as fabric development is concerned, such sediments can be considered as belonging to the compaction-dominated domain, in contrast with indurated shales where smectite to illite transition and/or authigenic growth of phyllosilicates under stress are thought to be responsible for very high levels of preferred orientation (Haines et al., 2009). Accretionary systems are a particular case of active

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fold and thrust belts forming underwater and made up of high porosity sediments (typically in the range 30%–70%). In such environments, compressional fabrics mainly result from lithostatic pressure and tectonic horizontal stress. In compressive systems, if mechanical decoupling between the substratum and upper units occurs early, subhorizontal pure-shear shortening (also called 'layer parallel shortening' or LPS) may be recorded with AMS (Averbuch et al., 1992; Frizon de Lamotte et al., 2002; Robion et al., 2007). Strain analysis indicates that LPS can be accommodated by internal shortening with magnitudes up to 20–30% prior to strain localization on major thrust systems (Koyi et al., 2003; Mitra, 1994; Protzman and Mitra, 1990; Sans et al., 2003; Yonkee and Weil, 2010), although the strain values are often less than 15% (e.g. Couzens et al., 1993; Evans and Dunne, 1991). More specifically, studies on accretionary wedges have shown mechanical decoupling at the décollement level in the toe region or in the trench (Henry et al., 2003; Housen, 1997; Morgan and Karig, 1995a,b; Owens, 1993) and have also indentified LPS fabrics acquired at a very early stage after deposition and incorporation of the sediments into accretionary wedges (e.g. Housen, 1997; Owens, 1993).

The goal of this study is to propose a simple method for quantifying low to moderate subhorizontal ductile strain based solely on AMS data in poorly consolidated clay rich sediments. The method is first detailed by building on the popular March approach and addressing the challenge associated with a non-randomly oriented initial fabric. Then, the method is applied to the Nankai accretionary prism using samples retrieved during the Nankai Trough Seismogenic Zone Experiment (NanTroSEIZE) IODP expeditions 315 and 316 along the Kumano transect. This transect is very attractive as it presents a typical architecture with an inner and outer wedge separated by a transition zone marked by the presence of an out of sequence thrust described as a megasplay fault (Kimura et al., 2007; Moore et al., 2009). AMS data measured onshore are used to estimate horizontal shortening in and around the megasplay fault zone, and in the prism toe. Values of shortening calculated sample-wise are analyzed and compared to existing estimates of ductile strain in the same region.

2. Strain model

2.1. March model vs. collapse model

Most models linking the AMS to finite strain are based on the March theory (March, 1932). They describe the progressive change in orientation of anisotropic particles as a function of strain, without change of particle shape (e.g. Borradaile, 1991; Fernandez, 1981; Henry and Hrouda, 1989; Hrouda, 1993; Owens, 1974) and can be applied to a majority of rocks, even those bearing triaxial magnetic grains (Hrouda and Jezek, 1999). The March model is based on the two following assumptions: 1) the initial rock fabric (i.e. at zero strain) is equivalent to a cumulate of lines or planes (representing anisotropic magnetic minerals) with a random orientation and 2) these imaginary lines or planes rotate as prescribed by the finite strain tensor. As noticed by Hrouda (1993), although this model is not strictly physical, it actually corresponds to an extreme case of the more physically grounded viscous model which describes the rotation of rigid particles embedded in a viscous matrix (Ghosh and Ramberg, 1976; Jeffery, 1922). The viscous model reduces to the March or 'line/plane' one in the case of particles of infinite aspect ratio.

While uniaxial compaction (vertical shortening with no lateral strain) is the reference in consolidation studies, the use of the March model does not necessarily imply that such strain path has to be followed. Let's consider the simplest 2D case of a line in a homogeneously deforming square. Following Owens (1974), one can relate the coordinates of a line before and after an increment of strain through the finite strain tensor S as $\begin{pmatrix} x \\ y \end{pmatrix} = S \cdot \begin{pmatrix} x_0 \\ y_0 \end{pmatrix}$. For S, two end member cases will be considered here: uniaxial compaction, and pure shear with no volumetric change. Assuming that the vertical incremental strain is expressed as $e_v = dl / l_0$, the finite strain tensor writes $S_{UA} = \begin{pmatrix} 1 & 0 \\ 0 & 1 + e_v \end{pmatrix}$ for uniaxial compaction, and $S_{PS} = \begin{pmatrix} (1+e_v)^{-1} & 0 \\ 0 & 1 + e_v \end{pmatrix}$ for pure shear. Both cases are schematically represented in Fig. 1a–b where the coordinates of the upper right corner of the initial square are transformed. Throughout this process, we are in fact only interested in the change in the angle θ between the coordinate vector $\begin{pmatrix} x \\ y \end{pmatrix}$ and the horizontal axis. The finite strain tensor can easily be used to determine the new coordinates as a function of the initial ones and derive a relationship between initial and final values of θ as a function of e_v. Before doing so, we also consider the simple case of geometric collapse in Fig. 1c, where, without any assumption being made on the associated lateral strain, a rigid line is reoriented in reaction to compaction as it is mechanically pushed downward. In this case, the relationship between vertical strain and particle angle does not result from the definition of a deforming host material, but a macroscopic strain tensor should be obtainable given some constraints on the mechanical interaction between rotating particles (e.g. fixed vs. sliding contacts).

Using the expressions for the finite strain tensors and the geometry of Fig. 1a–b, it is easy to relate θ_0 and θ through e_v. In the case of uniaxial compaction, one writes $\tan\theta = (1 + e_v)\tan\theta_0$, and in the pure shear case, $\tan\theta = (1 + e_v)^2\tan\theta_0$. For the collapse in Fig. 1c, the relationship between the angles is $\sin\theta = (1 + e_v)\sin\theta_0$. Fig. 1d–f compares the evolution of the angle θ as a function of e_v for various values of starting angle θ_0. The first case to be considered is a random initial fabric which is typically represented with an average angle to horizontal of $\theta_0 = 45°$. For the present comparison, only one line with 45° initial angle is followed through compaction. As vertical compaction progresses, uniaxial compaction results in the slowest rotation (22° at $e_v = 0.6$), and pure shear in the fastest one (9° at $e_v = 0.6$), while the collapse follows an intermediate path (16° at $e_v = 0.6$). This suggests that the collapse is actually not a bad approach to uniaxial compaction, especially if a slight lateral thickening is allowed. But the most characteristic difference between the approaches is observed for an initial angle of 90° where the March model does not allow particles to rotate. Even at 75° initial angle, rotation is very slow in the uniaxial compaction case (56° at $e_v = 0.6$ and 69° at $e_v = 0.3$). Therefore, models are comparable for an initial angle of 45° but differ strongly at initial angles close to 90°. This exercise shows an apparent shortcoming of the March approach whereby particle rotation for non-random initial fabrics might be misrepresented. In the following, the collapse model will be used to account for and estimate the intensity of layer parallel shortening from anisotropy of magnetic susceptibility data.

2.2. Horizontal collapse and AMS fabric evolution

The magnetic susceptibility is measured through the magnetization M that is induced in a sample subjected to a low alternating field H. If the magnitude of the magnetization field H is on the same order as that of the Earth's magnetic field, then M and H are approximately linearly related according to $M_i = K_{ij} \times H_j$ where K_{ij} is a second-rank symmetric tensor representing the magnetic susceptibility (Daly, 1970; Nye, 1957) and summation over j is implicit. The eigenvalues of the magnetic susceptibility tensor K_{ij} are denoted by K_1, K_2 and K_3, with $K_1 \geq K_2 \geq K_3$.

The anisotropy of magnetic susceptibility (AMS) has provided important constraints on the evolution of layer parallel shortening in fold-and-thrust belts (Averbuch et al., 1992; Graham, 1996; Pares and van der Pluijm, 2002; Robion et al., 2007, 2012; Sans et al., 2003). AMS arises primarily from anisotropic attributes of the solid grains in the rock matrix such as the preferred crystallographic orientation of phyllosilicate and other tabular grains, and grain shape anisotropy of ferrimagnetic grains such as magnetite.