

# Magnetic characterisation of folded aeolian sandstones: Interpretation of magnetic fabrics in diamagnetic rocks

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## ABSTRACT

This study provides an original example of exploitation of Anisotropy of Magnetic Susceptibility (AMS) for rocks with weak magnetic susceptibility. Within the upper Weber Sandstone at Split Mountain, Utah, 430 cores from 31 sites were collected for magnetic characterization. The magnetic susceptibility ranges from  $-10$  to  $10 \mu\text{SI}$ , indicating a mostly diamagnetic matrix, with degree of anisotropy up to 1.6. Specific treatment of magnetic susceptibility allows using diamagnetic data. The fabrics are fairly clustered and triaxial. Sedimentary magnetic fabrics show a foliation plane parallel to the lamina of the sand dunes, without defined lineation. Apart from sedimentary fabrics ( $<30\%$ ), most of the sites display intermediate to tectonic fabrics related to variable degree of strain ( $>70\%$ ). Magnetic fabric patterns averaged for sites distributed on the anticline are well defined in sub-groups related to the major structural domains of the anticline. The fracture network at Split Mountain is composed of a dominant N120 set and a secondary N035 set. A scenario of strain record is proposed based on the correlation of (1) fracture sets orientation, (2) diagenetic cementation, (3) paleostresses and (4) distribution of magnetic susceptibility anisotropy. Following the Sevier orogeny and N120 fracture set emplacement, the N035 fracture network and AMS signal were recorded during the Laramide Layer Parallel Shortening phase, with local deviation along pre-existing structures, and recorded a partitioning of the strain during early folding, with a maximum horizontal stress axis perpendicular to the fold bounding faults within the fold.

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## 1. Introduction

Geological investigation of fractured strata in folded structure aims at (1) providing physical explanation for the formation mechanisms of the fracture and (2) developing methods to predict their spatial distribution and orientation (Stearns and Friedman, 1972). This coupled matrix/fracture network depends at least on the stress and strain loading path, the folding geometry and kinematics as well (provided you can derive the stress field from the kinematic), the pre-existing tectonic and sedimentary structures and the rock mechanical properties (Stearns, 1964; Stearns and Friedman, 1972; Price and Cosgrove, 1990; Lisle, 1994; Salvini and Storti, 2001; Silliphant et al., 2002; Bourne, 2003; Guiton et al., 2003; Bergbauer and Pollard, 2004; Amrouch et al., 2009; Laubach et al., 2009).

In foreland settings of fold-and-thrust belts, the typical succession of deformation events includes: (i) vertical compaction; (ii) Layer Parallel Shortening (LPS); (iii) folding, and (iv) post folding relaxation (Engelder and Geiser, 1979; Averbuch et al., 1992; Mitra, 1994; Frizon de Lamotte et al., 1997; Evans et al., 2003; Roure et al., 2005). Each

event contributes to the finite strain and involves different mechanisms, among which are pressure solution (cleavage and stylolitization), grain boundary sliding and cataclasis at the grain scale, and both joints/veins emplacement and faulting at the macro scale (Elmore et al., 2006). A proper quantitative estimate of this deformation is a major objective in structural geology, particularly when studying the strain record of the rocks.

- The pre-fold phase of diffuse and quasi homogeneous at the fold scale deformation is recorded by markers of Layer Parallel Shortening (LPS) and involves remobilization of basinal fluids. This record has been largely recognized in different fold-and-thrust belts (Alvarez et al., 1978; Engelder and Geiser, 1979; Evans et al., 2003; Roure et al., 2005). It is an important deformation component as it can account for up to 20% of the total shortening in some fold thrust belts (Mitra, 1994; Sans et al., 1999; Koyi and Sans, 2003). Several authors have demonstrated that LPS strain is well recorded by magnetic fabric (depending on the mineralogical content of the rock), which are mainly developed prior to folding (Bakhtari et al., 1998; Frizon de Lamotte et al., 2003; Robion et al., 2007; Amrouch et al., 2010a).
- The folding phase is characterized at the fold scale by a more heterogeneous deformation, which may be localized along active

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faults and hinges connecting various reservoirs bodies and allowing the injection of exotic fluids. The micromechanisms activated during the LPS phase in the matrix may still be active during the folding phase with variable re-orientation with respect to bedding due to the bed tilting (see review in [Louis et al., 2006](#); [Robion et al., 2007](#)).

- The post-fold processes are marked by the exhumation of the folded structure and the possible interaction with meteoric fluids. During exhumation, new fracture set due to decompression could develop by release of residual stresses ([Price and Cosgrove, 1990](#); [Bergbauer and Pollard, 2004](#); [Bellahsen et al., 2006](#)).

The anisotropy of low field magnetic susceptibility (AMS) is a physical property of the rock defined by the ratio of the induced magnetization to the applied external field. It is used now as a standard tool for petrofabric and structural studies to characterize the matrix deformation related to textural distribution of grain, grain shape, porosity, and microfracturing (e.g. [Kissel et al., 1986](#); [Hrouda, 1991](#); [Averbuch et al., 1992](#); [Borradaile and Henry, 1997](#)). In sedimentary basins, deformed at low temperature (i.e. below 120 °C), this technique is classically used to complete the mesoscale structure analysis (joints, tension gashes, stylolites, microfaults, and cleavage). However, a correlation between strain and AMS is not straightforward, due to (i) the inherent heterogeneity of strain at the grain scale; (ii) the fact that AMS signal is the summation of the dia- and paramagnetic contribution of the rock matrix, plus a few highly susceptible ferromagnetic minerals; (iii) the fact that the anisotropy arises both from shape and crystallographic preferred orientations. It is thus generally recommended to try to separate the various source of the signal, i.e. the magnetic fabric related to ferromagnetic s.l. ([Borradaile and Henry, 1997](#)).

Nevertheless, AMS studies have been proven successful in deciphering the strain acquisition mechanisms for both sandstones (e.g. [Saint Bezar et al., 2001](#)) and para to diamagnetic limestones lithology (e.g. [Evans et al., 2003](#)), allowing the description of a general AMS fabric acquisition path related to strain. During the pre-fold LPS deformation, AMS fabric evolves from a sedimentary fabric to an intermediate and to a tectonic fabric ([Fig. 1](#), see further in the paper), in limestones, sandstones, shales, and silts ([Graham, 1966](#); [Averbuch et al., 1992](#); [Aubourg et al., 1997](#); [Sagnotti et al., 1998](#); [Parès et al., 1999](#); [Evans et al., 2003](#)). Both in extensional ([Sagnotti et al., 1994](#); [Mattei et al., 1997](#)) and compressional strain regime ([Kissel et al., 1986](#); [Lowrie and Hirt, 1987](#); [Averbuch et al., 1992](#); [Sagnotti et al., 1998](#); [Sans et al., 1999](#); [Amrouch et al., 2010b](#)), most studies emphasized that magnetic fabrics were mostly acquired in response to pre-fold loading. During folding, the state of strain, and particularly the imposed shear in the flanks and distribution of extensional and compressional zones in the hinge of the fold (e.g. [Price and Cosgrove, 1990](#)), depends on the fold kinematics, and could be recorded locally by the AMS fabric ([Frizon de Lamotte et al., 1997, 2003](#); [Grelaud et al., 2000](#); [Saint Bezar et al., 2001](#)). However, this relative timing of fabric acquisition during buckling or forced folding evolution cannot always be determined, and some authors demonstrated late record of

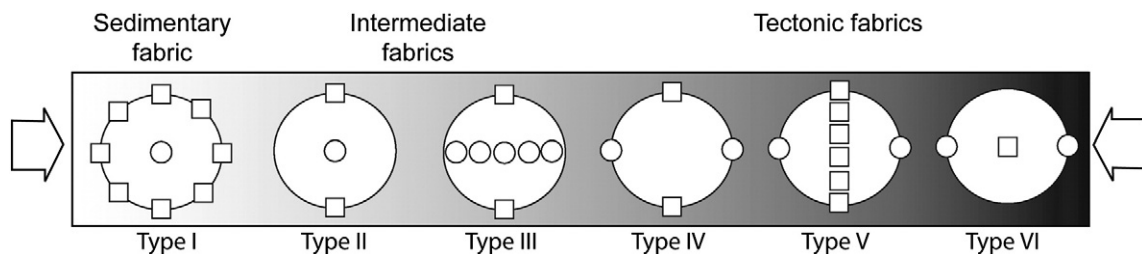
magnetic fabric. This was the case for rock subjected to metamorphism ([Housen and Van der Pluijm, 1991](#); [Robion et al., 1999](#)), and for carbonates rocks ([Jackson et al., 1989](#)). Imprint of a secondary AMS fabric is generally related either to the location of the site within a strongly strained area (shear zone, thrust zone, [Frizon de Lamotte et al., 2003](#); [Louis et al., 2006](#)) or to a secondary fluid circulation, allowing for complex dissolution/crystallization of susceptible minerals. Overprints by secondary “parasitic” fabric may be removed through heating of sample ([Souque et al., 2002](#), and references therein).

This paper aims at unraveling the history of strain acquisition and at linking the mesoscale and microscale mechanisms of deformation active during folding of para- to diamagnetic rocks. The underlying question is to what extent structural observations made at the microscopic scale in part of a fold may be relevant to the fold scale, and conversely. The chosen case study is the Split Mountain anticline (Utah, USA), a well-exposed Laramide basement-cored anticline showing weakly deformed rocks (e.g. [Kissel et al., 1986](#)). This work focuses on the description and characterization of small-scale deformation by combining various techniques of microstructural anisotropy analysis: Anisotropy of Magnetic Susceptibility, micro-tectonic and fracture record, and Fry strain analysis, together with petrographic data, in order to decipher the history of strain acquisition. Because the studied rocks are mostly diamagnetic, a specific treatment is applied to gain information. Our goal here is to link as much as possible the microscopic observations, the AMS data, which average the signal over several hundreds of grains, and the macroscopic deformation structures measured on the field. A coherent scenario of strain recording is proposed.

## 2. Geological setting

The Split Mountain anticline ([Fig. 2](#)) belongs to the Uinta Mountains, which formed the largest east–west trending arch uplifted during the Eocene Laramide orogeny. The outcrop is located within the Dinosaur National Monument in the Northeastern corner of Utah (USA). Following the middle Proterozoic Cheyenne orogeny, the Uinta area was characterized by the development of a rift zone responsible for roughly 26 km of north–south extension in the middle to late Proterozoic ([Stone, 1986; 1993](#)), accounting for the deposition and late South-westward tilting of the sandstones of the Uinta Mountain group. Marine limestone of the Madison and Round Valley formation are followed by the thick Weber formation (335 m thick), an aeolian sandstone succession of Pennsylvanian to Permian age, which is the main studied level. The thin limestone level of the Park City formation caps the Weber. These phosphatic fossil rich limestones are of Upper Permian age. This lower competent unit is separated from the Mesozoic competent rocks by the Moenkopi Triassic shale.

During the Mesozoic, the Utah area corresponded first to the NS oriented back bulge of the upper plate of the Farallon subduction (deposition of the shallow marine, sebkhas and coastal sediments of the Moenkopi to Morrison formation), then became the foreland of



**Fig. 1.** Schematic evolution of the magnetic fabric of sedimentary rocks submitted to shortening (arrows indicate shortening direction), from pure sedimentary fabric to tectonic fabrics (modified after [Averbuch et al., 1992](#); [Frizon de Lamotte et al., 1997](#)). The intensity of deformation increases to the right with the intensity of the grey-scale figured. The magnetic axes K1 (squares) and K3 (circles) are represented on stereograms following an equal angle projection on the lower hemisphere. See AMS and rock magnetisms section in the text for further explanations on sedimentary fabric, intermediary fabric and tectonic fabric patterns.

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