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Distinctive diamagnetic fabrics in dolostones evolved at fault cores, the Dead Sea Transform

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ABSTRACT

We resolve the anisotropy of magnetic susceptibility (AMS) axes along fault planes, cores and damage zones in rocks that crop out next to the Dead Sea Transform (DST) plate boundary. We measured 261 samples of mainly diamagnetic dolostones that were collected from 15 stations. To test the possible effect of the iron content on the AMS we analyzed the Fe concentrations of the samples in different rock phases. Dolostones with mean magnetic susceptibility value lower than -4×10^{-6} SI and iron content less than ~1000 ppm are suitable for diamagnetic AMS-based strain analysis. The dolostones along fault planes display AMS fabrics that significantly deviate from the primary "sedimentary fabric". The characteristics of these fabrics include well-grouped, sub-horizontal, minimum principal AMS axes (k_3) and sub-vertical magnetic foliations commonly defined by maximum and intermediate principal AMS axes (k_1 and k_2 axes, respectively). These fabrics are distinctive along fault planes located tens of kilometers apart, with strikes ranging between NNW-SSE and NNE-SSW and different senses of motion. The obtained magnetic foliations (k_1–k_2) are sub-parallel (within ~20°) to the fault planes. Based on rock magnetic and geochemical analyses, we interpret the AMS fabrics as the product of both shape and crystallographic anisotropy of the dolostones. Preferred shape alignment evolves due to mechanical rotation of subordinate particles and rock fragments at the fault core. Preferred crystallographic orientation results from elevated frictional heating (>300 °C) during faulting, which enhances c-axes alignment in the cement-supported dolomite breccia due to crystal-plastic processes. The penetrative deformation within fault zones resulted from the local, fault-related strain field and does not reflect the regional strain field. The analyzed AMS fabrics together with fault-plane kinematics provide valuable information on faulting characteristics in the uppermost crust.

1. Introduction

A fault zone in the upper crust is a tabular body consisting of a principal slip surface surrounded by a volume of deformed rock that is spatially and genetically related to the faulting process. Once a fault slip surface is established in the rock it represents a mechanically discontinuous weak zone. Consequently, it is prone to recurrence failure, resulting from stress builds up until it exceeds the frictional resistance of the fault during seismic events. Between failure events, displacement may accumulate at a nearly steady rate due to aseismic creep (Scholz, 2002). Hence, the distribution of total displacement along a large fault is the sum of displacements contributed by numerous individual slip motions (either seismic or aseismic). The slip motions are commonly expressed by kinematic indicators such as slickenside lineations and grooves (termed hereafter striations), which form on the fault surface and commonly reveal the last events of the deformation history (e.g., Pan et al., 2014). Based on these kinematic indicators, the local or remote strain field at the time of deformation might be reconstructed by analyzing the orientation of single faults or fault populations, respectively (e.g., Fossen, 2010).

Characterizing fault zones is important for understanding the mechanism of fault propagation during brittle fracturing and seismic events. The architecture of fault zone includes high-strain fault core enveloped by a damage zone (e.g., Caine et al., 1996). The fault core typically consists of ground-down fragments of the host-rock (i.e., fault gouge) and may consist of cataclasite or

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ultracataclasite (e.g. Sibson, 1977). Plastic deformation, enhanced by the elevated frictional heating (>300 °C; e.g., Sibson, 1975) may affect the rocks at the fault core during slip. Brittle deformation, mainly expressed in the damage zone, is characterized by fractures density that is higher than in the surrounding host rock (Chester and Logan, 1986). The mechanical behavior of the faults strongly depends on the interplay between brittle and plastic processes. While these deformation processes are both the product of fault-related strain field, their kinematics has seldom been analyzed and compared along individual faults. Moreover, the transition between plastic deformation at the fault core and brittle deformation at the damage zone is not well known due to lack of kinematic indicators, and is complicated by the very fine-grained fault gouge. A magnetic-based measurement of fabrics may help in understanding the kinematics at the fault plane and core and the anisotropy of the magnetic susceptibility (AMS) of rocks adjacent to slip surfaces.

The magnetic susceptibility of rocks results from the combined contribution of ferromagnetic, paramagnetic, and diamagnetic grains. The preferred crystallographic orientation of minerals and grain shape alignment give rise to the AMS of rocks (Rochette et al., 1992; Tarling and Hrouda, 1993). Grain alignment by plastic crystal sliding processes and recrystallization produce a preferred crystallographic orientation and a strong correlation of the AMS axial orientations with mineral orientations. Preferred shape alignment of particles due to mechanical rotation can also produce a good correlation between AMS axial orientations and the physical fabric stored in the rock (Borradaile and Henry, 1997). Hence, AMS axial orientations have structurally significant meaning, because they are strongly related to the deformation processes that aligned the minerals and particles in the rock. This is essentially the basics of correlating the orientations of AMS and strain axes (e.g., Mattei et al., 1997; Cifelli et al., 2004, 2013; Borradaile and Jackson, 2010; Aubourg et al., 2010).

Carbonate rocks are abundant in many sedimentary environments and tectonic settings, but their AMS has not been extensively examined due to their weak diamagnetic response. With the advance in laboratory instruments, the AMS of mono-mineralic calcite-bearing rocks were accurately and reproducibly measured (Owens and Rutter, 1978; de Wall et al., 2000; Evans et al., 2003; Hamilton et al., 2004; Schmidt et al., 2006; Almqvist et al., 2010, 2011; Levi and Weinberger, 2011). In this study, we extend the use of diamagnetic AMS as a petrofabric tool for dolomite-bearing rocks. The AMS fabrics were analyzed along fault planes, cores and damage zones next to the Dead Sea Transform (DST) plate boundary in order to reveal the micro-structures that have been developed at the DST strands during the faulting process. We compare the directions of the AMS axes with fault-plane orientations and their kinematic solutions in order to examine the effect of the local strain fields during faulting. We also characterize the distribution of Fe content in the different rock phases (i.e., carbonate, detritus, Fe-oxides), in order to assess its possible effect on the net AMS. We show that the obtained AMS fabrics are distinctive and provide valuable information on the penetrative deformation during faulting along major faults.

1.1. AMS of rocks and carbonates

The magnetic susceptibility indicates the capacity of materials to be magnetized in an applied magnetic field. The magnetic susceptibility (k) relates the applied magnetic field (H) to an induced magnetization (M) in a material by $M_i = k_{ij}H_j$. The shape of the AMS is described by the three principal values k_1, k_2, and k_3, which correspond to the maximum, intermediate and minimum magnetic susceptibility magnitudes, respectively (e.g., Borradaile and Jackson, 2004). There is a wide variety of parameters that have been used to describe the axial magnitude relationships of the susceptibility ellipsoid. The simplest expressions are the mean susceptibility, $k_m=(k_1+k_2+k_3)/3$, the anisotropy degree of AMS, $P = k_1/k_3$, the 'corrected anisotropy degree', P_j (for more details see Jelínek, 1981), the magnetic lineation, $L = k_1/k_2$, the magnetic foliation, $F = k_2/k_3$, and the shape of the susceptibility ellipsoid, $T = 2\ln(k_2/k_3)/\ln(k_1/k_3)-1$, measuring the range from prolate ($-1 < T < 0$) through sphere ($T = 0$) to oblate ($0 < T < 1$) ellipsoids (Jelínek, 1981). For diamagnetic rocks, the AMS parameters are calculated based on the absolute (unsigned) values of the principal susceptibility (Hrouda, 2004). These parameters are then $P = |k_3|/|k_1|$, $L = |k_3|/|k_2|$, $F = |k_2|/|k_1|$ and $T = 2\ln(|k_2|/|k_1|)/\ln(|k_3|/|k_1|)-1$.

During sedimentation, the evolved magnetic fabrics are typically characterized by bedding parallel and dispersed k_1 and k_2 axes, and k_3 axes that group perpendicular to bedding and parallel to the maximum shortening. In most deformed environments, the directions of the AMS axes show a fair to good correlation with the directions of the principal (finite) strain axes (e.g., Borradaile, 1987, 1988, 1991; Tarling and Hrouda, 1993; Borradaile and Henry, 1997; Parés et al., 1999; Latta and Anastasio, 2007; Soto et al., 2007; Mamtani and Sengupta, 2009; Hrouda et al., 2009; Burmeister et al., 2009; Hirt and Almqvist, 2012). No reliable correlation has been established between the magnitudes of the AMS axes and those of strain magnitudes (e.g., Parés and Van der Pluijm, 2004; Latta and Anastasio, 2007; Borradaile and Jackson, 2010) mainly because these relationships are complex, and may be governed by the percentage of the Fe and Mn-minerals in the rock (Rochette, 1988; Borradaile and Henry, 1997; Schmidt et al., 2006).

The chemical common unit of all carbonate minerals is the CO_3 ion, which shows a strong diamagnetic anisotropy (e.g., Schmidt et al., 2007). Pure dolomite has the composition of $[Mg,Ca](CO_3)_2$ in which cation layers of Mg and Ca alternate. Limited substitution of the Mg and Ca cations by Fe, Co, Pb, and Mn may occur. Pure dolomite is diamagnetic with mean range of susceptibility between -38×10^{-6} and 683×10^{-6} SI, and degree of anisotropy P between 1.45 and 4.96 (Voigt and Kinoshita, 1907; Stutzer et al., 1918; Rochette, 1988; Tarling and Hrouda, 1993). Schmidt et al. (2007) studied the AMS properties of dolomite crystals with iron content between 56 and 95,800 ppm, and showed that the mass magnetic susceptibility of dolomite varies over two orders of magnitude from -3.42×10^{-9} to 3.15×10^{-7} m^3/kg. The theoretically calculated mass magnetic susceptibility of pure dolomite is -4.804×10^{-9} m^3/kg, which is quite close to the lower measured value. Sass (1969) showed that a primary sedimentary fabric in dolostones is acquired during an early diagenetic stage of dolomitization. At that stage, individual rhombohedron dolomite crystals could rotate and adjust their position so that their maximum cross section tends to lie horizontally. Consequently, the dolomite c-axes is preferably aligned perpendicular to the bedding planes (Sass, 1969).

In view of the similar crystal structure of dolomite to calcite, it is worthwhile to review the AMS properties of the more intensively studied calcite. A pure calcite crystal is magnetically anisotropic with susceptibility values of $k_3 = -13.2 \times 10^{-6}$ SI along the c-axis, $k_1 = -12.09 \times 10^{-6}$ SI along the ab plane, $k_m = -12.47 \times 10^{-6}$ SI (Schmidt et al., 2006). Experiments of calcite-bearing rocks reveal that the redistribution of calcite c-axes is associated with applied stress or incremental strain (Casey et al., 1978); the c-axes are aligned parallel to the direction of maximum shortening. During sedimentation and compaction processes some of the c-axes are aligned parallel to the lithostatic pressure, depending on the pressure magnitude (Hrouda, 2004). The AMS parameters (P, L and F) of calcite-bearing rocks is expected to be lower than those of single calcite crystals, because perfect crystals alignments are rare,