Kinematics of a sigmoidal fold and vertical axis rotation in the east of the Zagros–Makran syntaxis (southern Iran):
Paleomagnetic, magnetic fabric and microtectonic approaches

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Received 3 September 2004; received in revised form 18 July 2005; accepted 18 August 2005

Abstract

The Zagros Simple Fold Belt is characterized by elongated, curved, or sigmoidal folds. The trend of these structures together with the structural style, change suddenly across the Zagros–Makran syntaxis which separates the continental collision domain of Zagros from the oceanic subduction one in Makran. This work focuses on the Minab anticline, outcropping in the easternmost part of Zagros. In order to understand the kinematics of a sigmoidal fold and underscore possible vertical axis rotations in the eastern side of the syntaxis, we performed a joint study of magnetic fabric, microtectonics and paleomagnetism of the northern termination of this fold. The two limbs have been sampled (7 sites, 134 samples) along three cross-sections corresponding to three different orientations of the fold axis. The rocks are weakly deformed fine-grained Mio-Pliocene reddish siltstones. The shortening directions deduced from both magnetic fabric analysis and microtectonic observations are consistent with each other, they are horizontal and roughly perpendicular to the local fold axis, following the torsion of the fold hinge line, and indicating a tectonic origin of the magnetic fabric. Rockmagnetic analyses (thermomagnetic curves, hysteresis loops) point to the presence of magnetite in the PSD and MD ranges as the main magnetic carriers, together with a minor contribution from hematite.

Apart from a post-tilting sub-actual VRM and/or CRM (component A), paleomagnetic analyses yield mainly two pre-tilting magnetization components: Component B is carried by magnetite, spanning the intermediate to high unblocking temperature range (300 °C ≤ T_{ubs} ≤ 580 °C). Component C has unblocking temperatures characteristic of hematite (580 °C ≤ T_{ubs} ≤ 680 °C). Both are ante-folding, based on positive reversal and fold tests, inside each of the cross-section but also for the three sections together. However, because component C is biased by some inclination flattening, only component B is taken into account afterwards. After full tectonic correction, the site mean direction of component B calculated for 6 out of the 7 sites is not statistically different from the individual site mean directions. The intersection direction determined by small circle analysis in these six sites is: D=18.0° I=33.7° k=482 alpha_{95}=2.6°. Compared to the Mio-Pliocene direction expected in this area for Africa, it corresponds to an overall clockwise rotation of the whole structure of some 18.3° ± 5.1°, consistent with the regional geodynamic context.

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0040-1951/$ - see front matter © 2005 Elsevier B.V. All rights reserved.
doi:10.1016/j.tecto.2005.08.024

TECTO-07540; No of Pages 21
The absence of differential paleomagnetic rotation from one cross-section to the other, and the fact that the magnetic lineation together with the shortening direction deduced from microtectonic analysis follows the fold curvature demonstrate that the torsion of this fold is not secondary but rather coeval with the fold amplification. The torsion of the fold would stem from underlying inherited Mesozoic structures and the rheological response, during the deformation, of two incompetent décollement layers undergoing differential velocities and a change of thickness, in a constant stress field direction.

Keywords: Paleomagnetism; Vertical axis rotations; Magnetic fabric; Microtectonics; Sigmoidal fold; Zagros–Makran syntaxis; Iran

1. Introduction

Studies have been carried out in order to understand the formation of curved folds in arcuate thrust belts. They are based on geometric considerations (Cosgrove and Ameen, 2000; Sattarzadeh et al., 2000), strain patterns and displacement vector field’s analyses (Hindle and Burkhard, 1999) or paleomagnetic recordings (Gray and Stamatakis, 1997). The formation of a sigmoidal (or S-shaped) fold is commonly associated with a blind shear along a basement wrench fault and to the subsequent formation of an echelon folds in the overlying sedimentary cover, as analogue models have produced (Odonne and Vialon, 1983; Richards et al., 1991). The development of a sigmoidal fold has been compared to the formation of sigmoidal tension gashes in a shear zone (Price and Cosgrove, 1990; Sattarzadeh et al., 2000). A sigmoidal fold may also generate from the interference of two growing neighbour folds: If buckle folds are formed parallel to each other in response to a regional shortening, their coeval amplification may degenerate into the coalescence of their facing periclinal terminations, if the hinge lines of the two interfering folds are offset by less than half the wavelength of the folds (Price and Cosgrove, 1990). A longer fold is thus formed with a deflection of its hinge line giving way to an S-shaped fold.

The purpose of this work is twofold: First, document a possible vertical axis rotation in the eastern part of the Zagros–Makran syntaxis and explain it in the geodynamical context of this transfer zone, which separates the continental collision domain of Zagros from the oceanic subduction one in Makran. Secondly, understand the kinematics of a sigmoidal fold, in particular the development of its curvature during the folding process. For this purpose, we performed a paleomagnetic study, associated to microtectonic and magnetic fabric analyses, on a sigmoidal fold displaying the characteristic S-shape and outcropping in the eastern part of the Zagros–Makran syntaxis. Microtectonics provides information about the finite strain field in the studied area. Magnetic fabric, through the anisotropy of magnetic susceptibility (AMS), reflects the statistically preferred orientation of grains and/or crystal lattices of all the minerals, which contribute to the magnetic susceptibility (essentially ferri- and paramagnetic minerals). In deformed areas, it is believed to record states of the finite strain imprinted on the rock. Finally, paleomagnetism should be able to precise the relative chronology between the remanent magnetization acquisition time and the folding process, as well as provide information on the deformation itself through, namely, possible vertical axis rotations related to the syntaxis formation.

2. Geological setting

The Zagros fold and thrust belt (Fig. 1a) represents the currently deforming cover of the Arabian plate which has been colliding in an oblique convergence, from the NW toward the SE, with the Eurasian and central Iranian blocks since the late Cretaceous abdication of oceanic crust slivers of ophiolites (Berberian and King, 1981; Alavi, 1994; Sattarzadeh et al., 2000). According to Dewey et al. (1973), McCall and Kidd (1982), Stoneley (1981), Dercourt et al. (1993), Ravaut et al. (1997), the collision occurred since the Miocene.

The present seismicity of the Zagros is generally explained by the reactivation of old basement normal faults as high angle thrusts (Jackson et al., 1981). In such a tectonic framework, the folding of the Mesozoic sedimentary cover is related to a reverse motion along these blind basement faults locally displaced by transverse sets of deep-seated active strike-slip faults, and to the rheological behaviour of the sedimentary sequence which overlies the Precambrian basement (Berberian, 1995; Berberian and King, 1981; Sattarzadeh et al., 2000; Hessami et al., 2001). In large areas of the Zagros fold belt, the infra-Cambrian Hormuz evaporite series constitutes an obvious décollement layer. But the general morphology of this belt is further complicated by the occurrence within the stratigraphic pile of subsidiary décollement levels responsi-
ble for disharmonic folding: the upper Cretaceous Gurpi marl, the Oligocene evaporitic Gasharan Formation (Falcon, 1969) and the Miocene Mishan marl (Molinaro et al., 2004). As a result of the above tectonics, elongated forced folds commonly displaying curved or sigmoidal shapes are observed (Sattarzadeh et al., 2000). They are characteristic of the mountain belt, extending from the north of the Persian Gulf to the Zagros–Makran syntaxis in its southeastern termination. However, east of the Zagros–Makran syntaxis the structural style changes significantly both in strike and style. The folds are more elongated than in the Zagros (Fig. 1b) and are interpreted by Molinaro et al. (2004) as fault propagation folds. Relating strictly the fold wavelength to the depth of the décollement, the latter would no longer lie within the infra-Cambrian evaporitic level but in one of the stratigraphically higher horizons.

The sigmoidal Minab fold lies in the eastern part of the Zagros–Makran transition zone, immediately to the west and roughly parallel to the dextral transpressive Zendan fault (Yamini Fard et al., 2003; Regard et al., 2004 and Fig. 1b). This fault is a major structure (Byrne et al., 1992) which separates the Zagros continental foreland belt from the western strand of the Makran accretionary prism, resulting from the subduction of the Oman Sea oceanic crust beneath the central Iranian accreted blocks.

The Minab fold is an elongated tight anticline about 60 km long and 8 km wide, which exhibits a torsion of its northern termination (Fig. 1b, c). This fold has been studied in detail by Molinaro et al. (2004) who interpret this structure as a fault propagation fold (Suppe and Medwedeff, 1990), soling out into a décollement level at a depth of about 6.5 km, on the basis of balanced cross-sections. It is slightly asymmetric with a steep SW forelimb, locally close to vertical, and a gentler NE backlimb. South from the sampling area, the forelimb of the Minab anticline is cut through by the so-called Minab fault which corresponds to a secondary out-of-sequence thrust undulating parallel to the main fold strike and complicating the forelimb structures. In this central part of the fold the internal strain of the forelimb is relatively strong, as attested by a space cleavage and a pervasive drag folding. However, these deformation features decrease toward the north, when going away from the out-of-sequence thrust, that is toward the sigmoidal part of the fold, which is the object of the present study (Fig. 1c). Here, the layering is generally weakly deformed, displaying only an incipient set of

Fig. 1. (a) Schematic tectonic map of the Hormuz strait region from Molinaro et al. (2004) showing the Zagros–Makran syntaxis and its proposed structural relationship with the Omani promontory (see text for explanations). (b): Detailed geological map of the Minab anticline and thrust area from Molinaro et al. (2004). The insert represents Fig. 1c. (c): Sigmoidal northern termination of the Minab anticline and the paleomagnetic sampling sites.
jointing classically observed in folding, except at one site.

The rock sequence consists essentially in consolidated deltaic and inshore brown to reddish sandstones and siltstones of upper Miocene to lower Pliocene age (the Kheku sandstones of the Makran Unit, explanatory text of the Minab map 1:250,000, McCall et al. (1985), equivalent to the Agha Jari Formation of Zagros). These foredeep deposits have been folded after the late Miocene–early Pliocene (McCall and Kidd, 1982; Berberian and King, 1981; McCall, 1997) or in two stages during the Plio-Pleistocene (Molinaro et al., 2004), and before the Quaternary (Regard et al., 2004).

3. Sampling and field observations

In an attempt to avoid remagnetizations found to be widespread in the coarse-grained detrital facies of the Agha Jari Formation (Delaunay et al., 2002; Aubourg et al., accepted), we restricted the sampling to the finest reddish siltstone layers. We sampled the two limbs of the anticline along three cross-sections (7 sites, 134 samples) situated in the northern sigmoidal termination of the fold (Fig. 1c). Each section corresponds to a slightly different orientation of the bedding, from N170° in the northern part to N131° southward (Fig. 1c). Note that the out-of-sequence thrust, developed further south around the small town of Minab, dies out toward the north at about the latitude of our southernmost section (Section C, Fig. 1c). Consequently, our sites should not be perturbed by this secondary tectonic feature, except maybe at Site Z-51, the most south western one where the forelimb is more deformed. In each site, 12 to 25 samples were sampled in more than four different layers, except at site Z-112 due to bad outcropping.
conditions, and the dip and strikes measured individually for each sample or group of samples.

Microtectonic measurements of microfaults, fracture planes and striae, when available, were taken at all the paleomagnetic sites. Complementary observations were made in many other sites of the anticline to better constrain its geometry and infer the regional style and kinematics.

4. Experimental procedures

All the microtectonic features observed on the field (joints, fractures, faults, striae) have been reported on stereographic plots. When enough data were available, principal stress axes have been computed from striated fault planes by means of inversion algorithm of Carey (1979) or Angelier (1984), or Daisy (Salvini, 2002).

In order to relate the macro deformation estimated by means of microtectonics to the physical properties of the rock deformation, we carried out a magnetic fabric analysis of all the samples by determining the AMS (anisotropy of magnetic susceptibility) on a Kappabridge KLY-3S.

We also examined a polished thin section of a reddish sandstone representative of these sandy rocks, but with coarser grains than the siltstone samples measured here (Aubourg et al., 2004). As the source area of these detrital sediments has not changed significantly during the deposition of the whole Agha Jari Formation, this sample provides qualitative information on the composition of the finer siltstones and on the nature of the magnetic phases.

To further investigate the size and the nature of the magnetic carriers, we performed rockmagnetic experiments on two samples in each site: hysteresis cycles were recorded on a Micromag Vibrating Sample Magnetometer and thermomagnetic curves ($K$–$T$ curves) were run in an inducting field of 300 A/m (0.38 mT) in Argon atmosphere, using a Kappabridge KLY-3 coupled with the CS-3 furnace. Few extra specimens were also heated in air in order to compare the results obtained in Argon and in air and better understand the
chemical composition changes occurring upon thermal demagnetization treatments.

Finally, a paleomagnetic study was performed in the hope of recovering the primary magnetization component. AF demagnetizations were carried out up to 160 mT on a laboratory-built demagnetizer described in Legoff (1985). Stepwise thermal demagnetizations were performed up to 680 °C in a non-inductive furnace in which the residual field is less than 10 nT. The remanent magnetization was measured on a JR-5 spinner magnetometer settled in a shielded amagnetic room whose maximum residual field reaches several hundreds of nanoteslas. The susceptibility was monitored at room temperature on a Bartington MS2 device after each heating step, in order to detect any possible changes in the magnetic mineral composition. Principal Component Analysis (Kirschvink, 1980) and, when necessary, combination of directions and remagnetized circles (McFadden and McElhinny, 1988) were used to determine the direction of the different magnetization components and the mean characteristic directions (ChRMs) were calculated by means of the Fisher’s (1953) statistics.

5. Results

5.1. Microscope examination

Microscope observation of a polished thin section in transmitted light shows that the rock is a volcanosedimentary sandstone in which the detrital minerals come from different sources: one source is the crystalline basement with eroded and altered minerals (rounded quartz with sometimes rolling extinction, sericitized Na and K feldspars, smeared zoned plagioclases, hornblende, micas, epidote . . . ) which indicate a fluviatile transport probably from the metamorphic and volcano-plutonic rocks outcropping in the Inner Makran.

The more important source is constituted by volcanic minerals usually fresh and angular either coming from acid volcanism (fragments of ignimbrite, quartz splinters, plagioclases) or from andesitic volcanism (lapillis and fragments of basic rocks wrapped in a matrix). These minerals testify to an Eolian transport probably in relation to explosive andesitic volcanism of the Makran volcanic arc.

Observation in reflected light shows quite numerous opaque minerals belonging to different populations. The largest ones (50–100 μm) are inherited from the basement and/or volcanic rocks. They occur as sub-automorph crystals with high temperature exsolutions, either smeared (grains from the basement) or with fresh trellis type ilmenite and titanomagnetite lamellae (grains of volcanic origin). The smallest magnetic minerals (<1 μm to few micrometers) are either small equant crystallites of iron oxides concentrated in some areas in the matrix or frambooidal opaque oxides of sedimentary origin. Both look fresh and homogeneous but they are too small to be really identified. Ferrous hydroxides and hematite are also probably present in the matrix.

5.2. Rock-magnetic experiments

Experiments of rock-magnetism reveal generally homogeneous magnetic properties throughout all the sites. Thermomagnetic analyses performed under
argon provide rather simple curves with only one Curie point between 564 and 588 °C, characteristic of non-substituted magnetite. The curves are either close to reversible or more often slightly to clearly irreversible (Fig. 2). In both cases, the Curie point of the heating and of the cooling curves is not significantly different. After the thermal cycle in argon, we note that the originally brown to reddish crushed samples have become grey, whereas they become more reddish in the few experiments carried out in air. Similarly, for irreversible thermomagnetic curves, the cooling curves have higher magnetization than the heating one when the experiment is performed in argon. It is the opposite in air. Also, upon heating in air the Curie point in the cooling curve is slightly but significantly higher than in the heating one. These observations suggest that, under argon atmosphere, the samples are somewhat reduced, leading to the formation of new magnetite, while they are oxidized in air leading to the conversion of some magnetite into hematite and/or maghemite.

Hysteresis loops show only one magnetic phase with relatively low coercivities. The magnetization increases sharply with the applied field and saturates between 0.3 and maximum 0.5 mT after removal of a small paramagnetic contribution. Hysteresis ratios \( \frac{H_{cr}}{H_c} \) and \( \frac{J_{rs}}{J_s} \) are well clustered in the diagram of Day et al. (1977). According to the theoretical study of Dunlop (2002a) and its application to natural rocks (Dunlop, 2002b), these values plot in the PSD range for natural magnetites (Fig. 3). They are also in the field of the theoretical curves calculated for a mixture of SD+MD magnetite including about 70% of soft MD fraction.

5.3. Magnetic fabric analysis

The magnetic fabric analysis of the different Cenozoic Formations outcropping in the Fars Arc (Eastern Zagros) is reported in Bakhtari et al. (1998). Results concerning more specifically the Mio-Pliocene Agha Jari Formation, and also the Minab fold, are reported in Aubourg et al. (2004). In the Minab anticline, the average bulk susceptibility \( K_m \) is quite high for sediment and varies little throughout the sites \( (10^{-3} \leq K \leq 3 \times 10^{-3} \text{ SI units}) \) in agreement with a roughly constant lithology. Fig. 4a illustrates the variation of the shape of the anisotropy ellipsoid as a function of the percentage of anisotropy using the \( T \) and \( P \) parameters defined by Jelinek (1981). The anisotropy is generally weak \((1.02 \leq P \leq 1.08)\) and the shape factor, \( T \), indicates that the susceptibility ellipsoids are oblate in all the sites (Fig. 4a).

5.3.1. Magnetic foliation

In all the sites but Z-115, the magnetic foliation coincides with or is close to the bedding plane, which means that the post-depositional compaction has been preserved. This is illustrated in Fig. 4b (particularly Section A) where the \( K_{min} \) axes of the anisotropy ellipsoids gather around the vertical direction after tilt correction, like the joint planes observed by Molinaro et al. (2004). In details however, based on the confidence ellipses around \( K_{min} \) the magnetic foliation is really parallel to the bedding plane in one site only (Z-114, Fig. 4b). Elsewhere (Z-110, Z-111, Z-51, Z-113), and also in three other sites of coarse-grained sandstones sampled in both limbs of Section C (Z-53, Z-54 and Z-59, Aubourg et al., 2004), the magnetic foliation is slightly oblique to the bedding, always in the dip direction. At the scale of the fold, we observe a smearing of the \( K_{min} \) in a direction parallel to the shortening direction (Aubourg et al., 2004), consistent with a magnetic fabric intermediate between sedimentary and tectonic, i.e., with a progressive loss of the sedimentary foliation.

5.3.2. Magnetic lineation \( (K_{max}) \)

As can be expected in a propagation fold, the magnetic lineation lies in the bedding plane, is horizontal

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Fig. 3. Hysteresis parameters according to Day et al. (1977) and Dunlop (2002a). \( J_r \) is the saturation remanent magnetization, \( J_s \) the saturation induced magnetization, \( H_{cr} \) the remanent coercive force and \( H_c \) the induced coercive force.
and parallel to the local fold axis within 15° in most sites (Z-110, Z-111, Z-114, Z-51, Z-113) (Fig. 4b). This lineation probably represents the usually termed Layer Parallel Shortening (LPS) imprinted in the early stage of the compression, before the onset of folding, as it was first supposed by Graham (1966), later supported by laboratory experiments (Borradaile, 1988) and commonly observed in compressional geological context (Kligfield et al., 1981; Averbuch et al., 1992; Aubourg et al., 1997; Frizon de Lamotte et al., 1997; Mattei et al., 1997; Bakhtari et al., 1998). However, two sites behave differently: At Site Z-112, the magnetic lineation is more than 40° away from the local fold axis and at Site Z-115, it is perpendicular to the fold axis and strongly oblique to the bedding plane. The AMS pattern provided by Site Z-112 might not be representative of this area, because we could only sample two different levels of silt, and most samples have been taken in one of these layers. At Site Z-115, the tectonic fabric is more pronounced than in the other sites (Aubourg et al., 2004). Here, the AMS reference frame is offset relative to the bedding reference one, but fits better the kinematic planes (Fig. 5). The magnetic lineation may represent the transport direction (Aubourg et al., 2004) imprinted to the rock during a deformation sequence that includes prefolding layer-parallel shortening.

Fig. 4. Anisotropy of the magnetic susceptibility (AMS) (a) — site mean AMS parameters according to Jelinek (1981): $T$ is the shape factor, $P'$ the corrected anisotropy degree. (b) — Equal area plots of the site mean principal directions in each section, before and after tilt correction. Squares: $K_{\text{max}}$; circles: $K_{\text{min}}$; crosses: bedding poles.

Fig. 5. Schematic cross-sections and equal area projections of the microtectonic observations around the paleomagnetic sampling sites together with the mean directions $K_{\text{max}}$ and $K_{\text{min}}$ of the AMS ellipsoid at each site. The pair of opposite black triangle indicates the shortening direction deduced from microtectonic analyses (see text). (a): Synthetic cross-section along a northern profile in the Minab fold (paleomagnetic Sections A and B). (b): Corresponding plots of microtectonic and "in situ" AMS data. (c): Cross-section along Section C and corresponding microtectonic and AMS plots.
ing overprinted by flexural slip during the fold amplification (Kodama, 1988; Stamatakos and Kodama, 1991a,b). In Section A, the mean $K_{\text{max}}$ declination, $162 \pm 16^\circ$, is not significantly different from the mean fold axis strike ($N169^\circ$) in geographic coordinates. In Section B, Site Z-114 only can be used because...
in Z115, \( K_{\text{max}} \) is not parallel but perpendicular to the fold axis. The declination of \( K_{\text{max}} \) (159 ± 14°) is the same as the local strike (N158°). In Section C, the mean \( K_{\text{max}} \) declination (129 ± 10°) is also very similar to the fold axis strike (N131°).

5.4. Microtectonic analyses

The fracture pattern in a fold (Price, 1966; Ruhland, 1973; Marrett and Allmendinger, 1990; Jamison, 1999) has been tentatively used as representative of the bulk deformation resulting from the regional and local stress field (Price and Cosgrove, 1990; Cosgrove and Ameen, 2000). In the Minab fold, Molinaro et al. (2004) observe a well-developed set of joints perpendicular to bedding. In agreement with Artaud (1969), they interpret these features as pre-folding “natural jointing” resulting from the strain layer parallel shortening. The different sets of fracture planes observed at and all around the paleomagnetic sampling sites are reported in Fig. 5 and placed on the corresponding field profiles of the fold along each one of the sections. They allow movement (or kinematic) planes (Artaud, 1969; Marrett and Allmendinger, 1990) to be identified graphically: these planes are defined as the best fitting planes of the measured slip systems. A slip system is defined by the slip direction (striae) and the normal to the slip plane (fault plane). A movement plane is orthogonal to the fault and contains the striae. The slip systems are the kinematic elements associated with the evolution of the fold, up to its finite state. The movement planes reported for the whole fault systems define principal kinematic directions which, in most of the known examples, fit with enough accuracy the regional principal direction of deformation. The stereographic presentation of the microtectonics (Fig. 5) has been treated with the software of Duyester (2000) and Salvini (2002). In the Minab anticline, the kinematic planes are mostly sub-vertical planes trending from ENE–WSW in the north (Section A and B) to NE–SW in the south (Section C). These planes are generally confined within the “error bar” of the inversion methods using joints and striae to calculate the principal directions (\( \sigma_1 \), \( \sigma_2 \) and \( \sigma_3 \)) of the paleostress tensor. Assuming that the strain tensor and the stress tensor are linearly related, we define a shortening direction (\( \varepsilon_{33} \)) in each site. The shortening directions deduced from the microtectonic analyses are consistently horizontal and roughly perpendicular to the local fold axis at all the paleomagnetic sites (thick black arrows in Fig. 5). It means that this direction follows the torsion of the fold hinge line. Along Section C, the direction of the \( \sigma_1 \) stress axis calculated by Regard et al. (2004) is N40° plunge 14°, in agreement with the horizontal direction, \( \varepsilon_{33} \), N52° and N45°, given at sites Z-51 and Z-113, respectively (Fig. 5).

As a whole, AMS and microtectonic patterns are consistent with each other and throughout the Minab anticline, indicating a good correspondence between the principal directions of deformation and magnetic anisotropy (Fig. 5). The directions of \( K_{\text{max}} \) and \( \sigma_1 \), deduced from respectively AMS and microtectonics, roughly follow the S-shape of the hinge line, suggesting a corresponding S-shape of the strain field in this area.

5.5. Paleomagnetic study

AF demagnetization proved to be noisy and inefficient to separate the magnetization components due to the occurrence of a high coercivity fraction. Thermal treatment was much more successful and was therefore applied to most samples.

5.5.1. General observation

The demagnetization curves (\( J_{\text{NRM}} \) versus \( T \)) display different pattern between sites as well as within sites but several common features can be outlined:

- In many samples a fast drop of the intensity is recorded in the lowest unblocking temperatures (Fig. 6a), corresponding to the removal of a dominant secondary component close to the present dipole field direction. Afterwards, the remaining NRM can be low (about 20% of its initial intensity), so that the primary signal is sometimes hard to extract. Conversely, in the majority of “good” samples, the NRM (Natural Remanent Magnetization) intensity decreases slowly upon heating because the ChRM (Characteristic Remanent Magnetization) is relatively larger and thus can be more precisely determined (Fig. 6b).
- Normal (Fig. 6a) and reversed (Fig. 6b) polarities have been recorded, either in different sites (normal polarity at sites Z-114, Z-115, Z-51 and reversed at site Z-111) or inside the same sites but in different layers (Z-110, Z-112, Z-113).
- The susceptibility monitored at room \( T \) after each heating step reveals that chemical changes occur in the rock throughout the whole heating treatment. However, the remanent magnetization does not seem to be affected because its changes in direction are not correlated with those of the susceptibility values. Also, the residual field inside the oven is
Fig. 6. Thermal demagnetization of different types of samples: orthogonal plots, change of the NRM intensity and the magnetic susceptibility. The name of the components isolated is indicated on the orthogonal plots for each sample. (a) and (b): typical samples with respectively normal and reversed polarities. (c): the only sample displaying a large A’ remagnetization component.
lower than 10 nT, which limits the possibility of parasitic TRM acquisitions.

- The samples carry three main magnetization components (A, B and C), the latest being isolated in only a limited number of samples in each site. When unblocking temperatures of two neighbour components overlap, great circles were used and combined with the ChRM according to McFadden and McElhinny (1988). When both polarities were recorded inside a single site, the reversal test of McFadden and McElhinny (1990) was performed. Finally all the reversed polarities were converted to normal ones before computing Fisher statistics (Fisher, 1953). The results are summarized in Table 1.

5.5.2. Magnetization component

5.5.2.1. Component A. With the lowest unblocking temperature ranges ($T_{ubs} \leq 250 ^\circ C$), this component is either a viscous (VRM) or chemical (CRM) remanence of recent origin, as attested by its average in situ direction ($D_{g}=0.2^\circ; I_{g}=46.0^\circ$ $k=220$ $a_{95}=1.1^\circ$ $n=70$ samples), similar to the present dipole field direction in the studied area ($D=0^\circ; I=\pm 46.1^\circ$; Table 1) and its negative fold test.

<table>
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<th>Site</th>
<th>Polarity</th>
<th>$n$</th>
<th>$D_g$</th>
<th>$I_g$</th>
<th>$k$</th>
<th>$a_{95}$</th>
<th>$D_s$</th>
<th>$I_s$</th>
<th>$k$</th>
<th>$a_{95}$</th>
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<td>220</td>
<td>1.1</td>
<td>346.1</td>
<td>30.9</td>
<td>3</td>
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<td>35.6</td>
<td>3</td>
<td>37.9</td>
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<tr>
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<td>± 46.1</td>
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<td></td>
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<td></td>
</tr>
<tr>
<td>Component A': remagnetization</td>
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<td>43.5</td>
<td>160</td>
<td>6.1</td>
<td>266.8</td>
<td>32.0</td>
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<td>8.0</td>
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<tr>
<td>Component B: magnetite</td>
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— $n$: number of samples; $D$: Declination; $I$: Inclination; subscripts “g” and “s” refer to geographic (“in situ”) and stratigraphic (full tilt correction), respectively. $k$: precision parameter; $a_{95}$: 95% confident angle (Fisher, 1953).
5.5.2.2. Component B (magnetite ChRM). Spanning the intermediate unblocking temperature ranges (300°C ≤ \(T_{\text{ubs}}\) ≤ 580°C) it is the largest and most significant ChRM, present in most samples of all the sites, except at Site 51 where only the component with the highest unblocking temperatures (ChRM C) could be isolated. In the principal component analysis, we chose a MAD cutoff angle of 10°, but it is generally better than 5°. According to rockmagnetic data, the magnetic mineral is most probably magnetite. Microscope observation together with Curie temperature suggests that this component is carried by the authigenic fraction of magnetite.

In the sites of mixed polarities, the reversal tests was always negative before untilting and positive after tilt correction: For Site Z-110 (8 samples of normal polarities, N, and 11 of reverse polarities, R, see Table 1), this test after bedding correction is positive of class B (\(\gamma_0 = 7.2° < \gamma_c = 9.0°\)), for Site Z-112 (1 N and 11 R), the test is positive of class C (\(\gamma_0 = 2.3° < \gamma_c = 19.6°\)), for Site Z-113 (6 N and 11 R) the test is again positive and of class C (\(\gamma_0 = 9.8° < \gamma_c = 11.3°\)).

The ChRM B directions are first examined in the three cross-sections separately. According to the suggestion of Cairanne et al. (2002), we first checked whether they evolve along small circles close to each other upon untilting of the strata, in each section with different orientations of the fold axis (Fig. 7). They do for the sites of Section A (Z-110-111 and 112) and Section B (Z-114 and 115), but in Section C, the ChRM of Sites Z-51 and Z-113 evolve on two parallel but very distinct small circles. The magnetization direction always shows a best grouping of the directions after total tilt correction than before (Fig. 7.

Fig. 7. Site mean directions of the B magnetization component (magnetite), except at Site Z-51 where the direction of component C (hematite) is represented, in the three cross-sections separately and together.
Table 1). The fold test (McFadden and Jones, 1981; Bazhenov and Shipunov, 1991; Fisher and Hall, 1990) is positive for 100% of untilting for the two sites of Section B. It is also true for sites Z-110 and Z-111 in Section A (Fig. 7), only the direction of Site Z-112 sampled close to the hinge of the fold appears to be slightly over-corrected with respect to the two previous ones, suggesting a syn-folding acquisition. Most likely, the ChRM isolated in this site does not perfectly record the true local magnetization direction, just like the AMS lineation is offset relative to the one of the other sites, possibly because the surface of sampling was not spread enough to cancel out errors from different origins.

5.5.2.3. Component C (hematite ChRM). Component C is seen in the highest temperature range (580 °C ≤ T_{ubs} ≤ 680 °C), where the remaining magnetization is generally weak relative to the initial NRM so that it is easily overprinted by parasitic magnetizations acquired during heating. Consequently, it is generally very noisy and scattered (MAD cutoff angle 18°), as shown by the large confidence angles associated to the average directions in most sites (Fig. 9). Component C could be determined on a smaller number of samples than component B, in each site except at site Z-51 where it seems to be the only really definable component (Table 1), mostly deduced from great circle analyses. Its polarity always mimics those of component B and the direction of both vectors is close to each other, but with a slightly lower inclination. This suggests that components B and C are roughly coeval. Note that in cross-section C, the direction of component C at Site Z-51 is again significantly offset relative to the same component of the counterpart site Z-113.

The magnetic carrier is hematite, and more likely originates from the pigment responsible for the reddish colour of the formation, rather than from the few large detrital grains observed in the rocks, as suggested by the high unblocking fields inferred from AF demagnetization tests.

5.5.2.4. Component A'. Observed in the unblocking temperature range of 140–300 °C, this component could be determined in only 5 samples from Site Z-51 (Fig. 6c, Table 1). However, its presence can be

Fig. 8. Small circles analysis of component B (Shipunov, 1997) showing the evolution of the site mean directions from no tilt correction (light grey squares) to 100% tilt correction (dark grey diamonds). The star represents the direction of the best intersection.
inferred in several other samples characterized by large overlap of the unblocking temperature spectra of components A’ with B and/or C. Component A’ is a post-folding remagnetization, as shown in Aubourg et al. (accepted) who observed it in coarse-grained samples of two neighbour sites from the same Formation. The $T_{\text{ubs}}$ overlap of components A’ and C up to the highest temperatures, even if the NRM decay looks linear (Dinare`s-Turell and McClelland, 1991), can explain why the mean ChRM C direction of Site Z-51 is deviated westward relative to the ChRMs determined in the six other sites (Fig. 7). Because it is observed in the coarsest grained samples, component A’ most likely does not represent a true paleofield direction and thus will not be considered any further.

5.5.2.5. Synthetic results concerning components B and C at the scale of the Minab anticline. Fig. 7 shows that after tilt correction the mean directions of the B component are not significantly different from one section to the other, in all the sites but Z-51 (Figs. 7 and 10). Consequently, the directions of component B can be compared in the six coherent sites and a mean direction calculated. After 100% untilting this direction is: $D=18.2^\circ$ $I=32.4^\circ$ $k=257$ $\alpha_{95}=4.2^\circ$. This result is confirmed by the small circle analysis of Shipunov (1997) (Fig. 8), where the best grouping of the directions is obtained when the beds are restored to the horizontal (total tilt correction) for all the sites except for Z-112. The average component B direction calculated for the six sites by intersection of the small circles is not statistically different from the previous one: $D=18.0^\circ$ $I=33.7^\circ$ $k=482$ $\alpha_{95}=2.6^\circ$ (Table 1). Component B is thus a pre-folding magnetization. As for component B, component C does not display significant directional differences between the three cross-sections, except at Site Z-51, and the fold test follows the same pattern (Fig. 9). Again, the ChRM of Site Z-51 is discordant with the ChRM of the six other sites, due to a marked difference in declination after tilt correction (Fig. 9). The average direction calculated from the six coherent sites after full tectonic correction is: $D=19.9^\circ$ $I=25.0^\circ$ $k=81$ $\alpha_{95}=7.5^\circ$. It becomes: $D=19.4^\circ$ $I=28.0^\circ$ $k=172$ $\alpha_{95}=4.4^\circ$ when the intersection of the small circles is calculated. This is not statistically different from the component B average direction, although the inclination is weaker by about $6^\circ$. Component C is also ante-folding (Table 1).

6. Discussion

Rockmagnetic and paleomagnetic investigations indicate that magnetite is the main magnetic carrier in the fine-grained siltstones presently studied. Hematite constitutes a minor magnetic population, the signature of which is only revealed by the remanent magnetization. As a matter of fact, hematite is not visible on the susceptibility versus temperature curves nor in the hysteresis loops, as also in the thermal demagnetization of three axes IRM (Lowrie’s, 1990 method) of coarse-grained samples of the Agha Jari Formation studied in Aubourg et al. (2004). Consequently, the AMS pattern observed in the Minab anticline is most likely due to the ferri-magnetic magnetite in the PSD–MD range. The susceptibility anisotropy of magnetite is controlled by the shape anisotropy of the grains (Hrouda, 1982). It reflects preferential linear or planar alignment of the long axes of these grains. The mag-

![Fig. 9. Site mean directions of the C magnetization component (hematite), in the three cross-sections together.](image-url)
netic fabric in the Minab anticline is dominantly planar (Fig. 4a, b). The magnetic foliation is either parallel to the bedding plane or slightly oblique to it, indicating that the original sedimentary features are preserved. At the scale of the anticline however, the smearing of the $K_{\text{min}}$ directions perpendicularly to the shortening direction is interpreted as an increase of the tectonic imprint with a progressive loss of the sedimentary foliation. This tectonic imprint, most likely related to the folding event is consistent with the development of a fault propagation fold proposed by Molinaro et al. (2004) and is supported by the alignment of the magnetic lineation along the local strike of the fold axis. On the whole, the AMS of the Minab anticline seems to be acquired mainly at an early stage of deformation (Layer Parallel Shortening, LPS) developed prior to folding.

Although the microtectonic measurements do not sample the same time interval and observation scale of tectonic schedule as the AMS does, it is worth noting that the local kinematic referential system deduced from microtectonic analysis is consistent with the AMS referential system. Both follow the local strike of the fold axis (Fig. 11) within few degrees, the magnetic lineation and the shortening direction being respectively parallel and perpendicular to the fold axis. However, from North to South of the sigmoidal termination, the amplitude of the variation of the shortening direction, determined either from microtectonics (~28°) or from AMS measurements (~32°), is slightly lower than the range of the corresponding bedding strikes (~39°).

Positive reversal and fold tests indicate that the magnetite ChRM (component B) is pre-folding. Because the rocks have an Upper Miocene to Early Pliocene age and the first folding phase most likely occurred after the Mio-Pliocene (Regard et al., 2004; Molinaro et al., 2004) the magnetite ChRM B is probably of primary origin. The same conclusion holds for the hematite ChRM C which mimics component B. The average direction of the hematite ChRM, although not statistically different from the direction of component B, exhibits a slightly shallower inclination at each site (mean $\Delta I = 6^\circ$). Anticipating that this weak flattening is an artefact, and therefore does not constitute a reliable recording of the paleomagnetic field, we will not take into account this direction, but rather consider the direction of the B component. However the above arguments suggest that both magnetite and hematite ChRM are

![Fig. 10. Site mean declinations as a function of the fold axis strike at each site. The grey shaded rectangles are the $\alpha_{0.05}$ confidence angles around the average observed declination (calculated without Site Z-51) and the expected declination. 18.3° ($\pm 5.1^\circ$) represents the amount of overall clockwise rotation of the studied area.](image)
primary magnetizations, but only the former one would have correctly recorded the paleo-field direction.

6.1. No between-site differential rotation

Excluding Site Z-51 where the ChRM is obliterated by component A', the paleomagnetic results show that there is no relative rotation from one section to the other (Figs. 7–11). It means that the magnetization is not correlated with the curvature of the fold axis as it should be expected if the sigmoidal shape of the hinge line had been generated by a late torsion related to an underlying blind fault, as it is inferred in most curved folds in the Zagros, for example (Sattarzadeh et al., 2000; Hessami et al., 2001). Because the acquisition of the remanent magnetization occurred prior to folding, the deflection of the hinge line in the Minab anticline termination must be contemporary with the fold amplification, as will be discussed later.

The mean direction of component B obtained in the six self-consistent sites is then compared to the Mio-Pliocene direction expected for the studied area. The Minab anticline being part of the Zagros accretionary prism, the expected direction has been calculated using 10 selected African poles from the GEOREF database for the time interval 11 to 4 My. Compared to this direction, \( D = 39.8^\circ \), \( I = 51.7^\circ \), \( k = 71 \), \( z_{95} = 5.8^\circ \), the component B overall mean direction indicates that the whole area under study has been clockwise rotated by \( 18.3^\circ \pm 5.1^\circ \) (error calculation of Demarest, 1983) relative to Africa, since the Mio-Pliocene (Figs. 8 and 10). The observed inclination is not statistically distinct from the expected one at the \( z_{95} \) level, although slight-
ly lower (Fig. 8). Indeed, the inclination flattening recorded by component B and calculated according to Demarest (1983) \((F=6.1^\circ \pm 6.0^\circ)\) might not be significant.

6.2. Paleomagnetic inclination shallowing for component C:

For component C, the difference of about 12\(^\circ\) between observed and expected inclination is significant. It cannot be attributed to a latitudinal displacement of more than 600 km since the early Pliocene, because it would imply a converging rate of about 12 cm/year between Arabia and Eurasia, which is at variance with the geodynamics of this region (Bayer et al., 2003; Masson et al., 2003; Vernant et al., 2004). The inclination shallowing is rather an inclination error recorded by the hematite, as it has been commonly observed, particularly in tertiary fine grained red beds (Thomas et al., 1993; Chauvin et al., 1996; Cogne et al., 1999; Dupont Nivet et al., 2002; Gilder et al., 2003; Tan et al., 2003) and in laboratory deposited samples containing hematite (Tan et al., 2002). The present study emphasizes that one must be very cautious when interpreting directional data provided by fine-grained red beds in which the primary magnetization is carried by the sole hematite.

6.3. Interpretation of the vertical axis rotation in the Minab anticline area

The ante-tectonic ChRM B isolated in the 6 coherent sites of the sigmoidal termination of the Minab anticline documents a final clockwise rotation of about 18\(^\circ\) of this area.

Clockwise vertical axis rotations in the eastern side of the syntaxis are indeed suggested by the GPS data of Bayer et al. (2003), although the associated errors are still large and need to be confirmed by additional measurements. This rotation together with the damping of the deformation from west to east across the N–S Zendan–Minab–Palami and Jiroft–Zabzevaran faults systems as calculated from GPS measurements (Bayer et al., 2003, in press; Masson et al., 2003), support the hypothesis of the indentation of the Oman peninsula into the Eurasian block (Kadinsky-Cade and Barazangi, 1982). An alternative interpretation is proposed by Molinaro et al., 2005, Fig. 13c, in which the clockwise rotation of this area is mostly accounted for by the south westward propagation of the Makran arc, as deduced from field work and the observation of maps and satellite images.

6.4. Possible mechanisms accounting for the torsion of the fold

This study shows that, in the northern termination of the Minab anticline, the shortening direction deduced from microtectonic and magnetic fabric analyses rotate progressively from respectively N73\(^\circ\) to N45\(^\circ\) and N86\(^\circ\) to N30\(^\circ\) from north to south in relation with the torsion of the fold. However, this is not accompanied by a corresponding vertical axis rotation of the rocks, as the ChRM directions remain constant, whatever the fold axis strike. The fold kinematics is here different from those prevailing in the Zagros where most of the curved folds are interpreted as drag folds formed in the sedimentary cover decoupled from the basement, in response to deep seated active transverse faults (Berberian, 1995; Talbot and Alavi, 1996; Sattarzadeh et al., 2000; Hessami et al., 2001). In such folds the torsion is inferred to be a late process and differential vertical axis rotations of the paleomagnetic vectors have indeed been observed (Aubourg et al., accepted).

In the studied area, Molinaro et al. (2005) propose that the Minab fold as others further south, are generated by the propagation of the Makran prism toward the S–SW. A mechanism of coalescence of two previous folds, as proposed in Price and Cosgrove (1990) is not appropriate here because Molinaro et al. (2004) have shown that the torsion of the Minab fold was a consequence of the out-of-sequence thrust (Minab thrust, Fig. 1b) breaking progressively the forelimb toward the SW, leaving behind the northern part of the early stage. Also, inherited structures in the Northern Oman–Hormuz Strait region constitute a heterogeneous obstacle to the fold propagation and probably contribute to the torsion of the Minab fold. The influence of the stacked Omanese nappes on the shape of the overlying Tertiary Basin in this region (Dunne et al., 1990; Michaelis and Pauken, 1990; Al-Lazki et al., 2002) has been discussed in detail in Molinaro et al. (2004) who suggest that these imbricated structures underlie the eastern side of the Zagros–Makran syntaxis, joining the Oman promontory to the contact zone between the Fars and Makran arcs (Fig. 1a). The NNE–SSW paleo-front of the Musandum Hagab thrust system continuing across the Hormuz strait (Fig. 1a) would represent a major tectonic limit (the Oman line) between the two distinct fold systems of the Zagros to the west and the Makran to the east. For these authors, two incompetent layers within the Tertiary Basin correspond to décollement levels which probably contribute efficiently to the development of a velocity gradient across the NE–SW mean slip direction of the Minab thrust system and thus to the torsion of the fold. This velocity
7. Conclusion

- The occurrence of both polarities and a positive fold test in six sites out of the seven sampled strongly suggests that the magnetite ChRM B is a pure and primary magnetization component.
- The hematite ChRM C is likely coeval with the magnetite ChRM. However, this component is less well defined than the magnetite one, and its inclination is systematically lower in all the sites. Once again, this suggests that hematite is more prone to inclination shallowing than magnetite and consequently is not always as reliable a paleomagnetic recorder.
- After tilt correction, the site mean ChRMUs carried by magnetite are not significantly different from each other, whatever the cross-section, that is irrespective of the fold axis trend.
- Compared to the Mio-Pliocene direction expected in this area for Africa, the magnetite ChRM direction corresponds to an overall coherent clockwise rotation of the whole structure of some $18.5 \pm 5.1^\circ$.
- Overall vertical axis clockwise rotations east of the Zagros–Makran syntaxis are in agreement with the hypothesis of the indentation of the segmented eastern corner of the Arabian plate into the Eurasian blocks. Within the regional geodynamics of the underthrusting of the Arabian plate beneath the Central Iran, the rotation of the Minab area would also stem from the interference between the eastward development of the Fars Arc, the northward motion of the subsiding inherited N–S Oman thrusts, and the propagation of the Makran prism toward the S–SW.
- The displacement directions inferred from microtectonic analysis and AMS pattern are corresponding with each other.
- They roughly follow the changes of the fold axis strike, suggesting that the torsion of the fold stems from several causes: the Makran propagation direction which is oblique to the strike of the fold, the inherited heterogeneous N–S Omanese structures which block the south westward development of the fold and the mechanics and geometry of the décollement layers within the sedimentary cover, which generate differential displacement rates in a constant stress field direction.
- Finally, the lack of differential paleomagnetic rotation with the change of the fold axis strike, the consistency between the shortening directions deduced from microtectonic and AMS analyses which follow the torsion of the fold hinge line, all support the fault propagation fold model proposed by Molinaro et al. (2004) for the Minab anticline. The sigmoidal torsion of this fold in map view can also be related to primary changes in the thickness of two décollement layers, owing to the possible reactivation of one of the Oman lateral branch lines. Thus, the sigmoidal torsion of this fold is not a late process occurring after completion of the anticline, but was rather initiated since the onset of the folding event.

Acknowledgements

This work is part of the cooperative research agreement between the INSU-CNRS (France), the Institute of Geophysics of Tehran (Iran) and the Geological Survey of Iran (GSI). Funding was provided by the “Intérieur de la Terre” program (IT, INSU-CNRS). We thank the Institute of Geophysics and the GSI for willing assistance during the fieldwork.

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