Anisotropy of magnetic susceptibility (AMS) records synsedimentary deformation kinematics at Pico del Aguila anticline, Pyrenees, Spain

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Abstract: Pico del Aguila anticline is a transverse décollement fold located at the Pyrenean thrust front. The anticline is a synsedimentary structure buried during growth by delta front mudstones and sands of the Eocene Arguis and Belsué-Atares formations. Both the anisotropy of magnetic susceptibility measured at 77 K and 294 K and the anisotropy of anhysteretic remanence show that susceptibility is dominated by paramagnetic clay minerals and can be used as a proxy for depositional and tectonic fabric orientations. In general, the maximum and intermediate principal susceptibilities (k1 and k2) of the AMS lie in bedding and the minimum principal susceptibility (k3) is oriented nearly normal to bedding. Layer-parallel shortening (LPS) produced a c. north–south-trending magnetic intersection lineation in bedding on anticline limbs and in the adjacent Belsué and Arguis synclines by deforming the depositional and diagenetic compaction fabric. The degree of magnetic anisotropy is higher along axial surfaces than on limbs. At the anticline hinge, oblate magnetic ellipsoids with an east–west-aligned lineation and a bedding-parallel magnetic foliation demonstrate the overprinting of the LPS magnetic fabric during the emplacement of the underlying thrust sheet. AMS data record fold kinematics characterized by constant-length limb rotation about pinned hinges and are compatible with kinematics recorded by growth strata geometries. This study emphasizes that AMS is a very sensitive measure of depositional, compaction and tectonic fabrics in marine clastic rocks in the diagenetic realm.

Supplementary material: Data tables including specimen locations, orientation, and AMS matrix elements for new samples and AMS data from Pueyo et al. (1997). Bulk susceptibility at 77 K and 294 K for representative specimens is available at http://www.geolsoc.org.uk/SUP18842

Studies of deformation allow assessment of rheology, strain history and kinematics that are necessary prerequisites for conducting dynamic studies, incrementally balancing cross-sections or generating palaeogeographical reconstructions. The location of pin lines during folding creates distinct strain patterns that can be used to test predictions from various kinematic models of folding (fixed v. migrating hinges, flexural v. buckle folding; e.g. Ramsay 1967; Dahlstrom 1990; Fisher & Anastasio 1994; Anastasio et al. 1997). Numerous studies have used finite or incremental strain data or mesoscopic fabric variation as a function of structural position to elucidate deformation kinematics (e.g. Cloos 1947; Ramsay & Huber 1983; Fischer et al. 1992). However, in orogenic forelands where deformation occurs at shallow depths and low temperatures, ductile penetrative deformation features are subtle and brittle structures may be sparse. Anisotropy of magnetic susceptibility (AMS) has been successfully used as a proxy for rock strain in sedimentary rocks in regions where other deformation markers were not available (Averbuch et al. 1992; Borradaile & Henry 1997; Parés 2004). In general, comparative studies from siliciclastic rocks show good agreement between the relative magnitude and orientation of penetrative rock strain determined by traditional geometric methods of strain analysis and AMS patterns, whereas only the principal directions of strain were comparable in carbonate rocks (e.g. Borradaile & Tarling 1981; Borradaile 1988; Averbuch et al. 1992; Mattei et al. 1997; Pueyo et al. 1997; Sagnotti et al. 1998; Parés & van der Pluijm 2003; Borradaile & Jackson 2004; Larrausoña et al. 2004; Latta & Anastasio 2007; Burmeister et al. 2009).

AMS is a second-rank tensor that characterizes a material’s magnetization response to an applied magnetic field (e.g. Nye 1957; Tarling & Hrouda...
The tensor can be represented by an ellipsoid, with greatest principal axis $k_1$, intermediate axis $k_2$ and shortest axis $k_3$, representing the magnetic susceptibility contributions of a rock’s constituent magnetic minerals. Crystallography and grain shape control magnetic mineral properties. For oblate paramagnetic clay minerals, the crystallographic axes also correspond to the susceptibility tensor axes, that is, magnetic axes in such crystals conform to the density distributions of mineral lattice planes obtained by x-ray goniometry (Richter et al. 1993; Martín-Hernández & Hirt 2003; Schmidt et al. 2009). Natural processes such as current deposition, diagenesis and lithification, or tectonic deformation can all contribute to the measured AMS of a sample. Because we know that the individual particle anisotropies of phyllosilicate minerals are oblate, we can use the AMS fabrics to measure their preferred orientation (e.g. March 1932; Richter et al. 1993).

In order to test the sensitivity of AMS to establish variations in penetrative deformation in low strain and shallow crustal environmental conditions to determine fold kinematics, we present results from Pico del Aguila anticline, a submarine, synsedimentary fold that developed near the South Pyrenean mountain front. Pico del Aguila was chosen as the location for this study because deformation conditions and fold kinematics are independently known from previous studies. The anticline formed near the marine sediment surface in the Palaeogene wedge-top basin of the southern Pyrenees (Fig. 1). Among the numerous studies of Pico del Aguila anticline, Puigdefabregas (1975) and more recently Castelltort et al. (2003) studied the stratigraphy of the growth strata, Hogan & Burbank (1996) reconstructed fold burial, Kodama et al. (2010) used rock magnetic cyclostratigraphy to determine the age of the growth strata, Anastasio & Holl (2001) studied deformation patterns and thermal conditions, and Millán et al. (1994), Poblet & Hardy (1995) and Castelltort et al. (2003) analysed growth strata geometries to explore fold kinematics. By comparing the AMS fabrics as a function of

Fig. 1. (a) Location (inset) and simplified geological map of the central External Sierras and Pico del Aguila anticline (Anastasio 1987). (b) Interpretive vertical cross-section of the southern Pyrenees across the Jaca wedge-top basin and External Sierras to the Ebro foreland Basin in the south. Location of the cross-section shown in Figure 2a and area of Figure 3 shown.
structural position and stratigraphic level in the growth strata to predicted strain patterns, we evaluate the ability of AMS to resolve kinematics in the diagenetic realm.

**Geological setting**

The Pyrenees Mountains are a dual-vergent mountain belt that formed during the Late Cretaceous—early Miocene as a result of Iberian and Eurasian plate convergence (Roest & Srivastava 1991; Muñoz 1992). The External Sierras are the Spanish foothills that separate the Jaca wedge-top basin from the Ebro foreland basin to the south along the emergent South Pyrenean thrust front (Fig. 1). The Guarga thrust sheet transported the Jaca basin southwards a minimum of 19.4 km during the Middle Tertiary in the west central part of the orogen (Anastasio 1987) with displacement on the Guarga thrust sheet decreasing westwards (DePaor & Anastasio, 1987; Millán 2006). A series of transverse detachment folds, with a décollement level within the Triassic evaporites, developed above the ductile evaporitic mudstone of the Keuper facies of the Pont de Suert Formation which acts as the décollement and is the base of a c. 1 km thick pre-growth stratigraphy dominated by Triassic–Middle Eocene carbonate rocks. The growth strata are up to 1.5 km thick and shallow in dip to horizontal in the basal Campodarbe Formation. The growth sequence includes deposits of the Arguis and Belsué-Atarès formations, which are of Middle Eocene–Oligocene in age. The Arguis Formation is composed of beds of marly siltstones and sandstones, some of which are bioturbated, and rare interbedded limestones. It represents a large-scale prograding/retrograding flood-dominated delta and silty shelf sequence deposited from east to west, burying the active Pico del Aguila anticline and the other transverse anticlines of the External Sierras. The Belsué-Atarès Formation, which is dominated by coarser sediment deposited on the delta plain, overlies the Arguis Formation and finally the mostly post-folding Campodarbe Formation, which represents Oligocene–Miocene-aged fluvial molasses (Puigdefàbregas 1975; Castelltort et al. 2003; Fig. 2). Folding at Pico del Aguila began in the Middle Eocene (Anastasio 1992; Millán et al. 1994; Kodama et al. 2010; Rodríguez-Pintó et al. 2012) during the deposition of the upper part of the Guara Formation, a foraminiferal carbonate bank that forms the resistant topography of Pico del Aguila. Most of the folding occurred during Arguis Formation deposition. A latest Lutetian–early Priabonian age of the Arguis Formation is supported by multispecies biostratigraphy (Canudo 1990; Canudo et al. 1991) and magnetostratigraphy (Hogan 1993; Hogan & Burbank 1996; Pueyo et al. 2002; Kodama et al. 2010; Rodríguez-Pintó et al. 2012).

The Pico del Aguila anticline, which plunges at 30° to the north, is bounded to the west by the Arguis syncline and to the east by the Belsué syncline. The External Sierras formed under peak burial temperatures of c. 55–60°C based on vitrinite reflectance data (Holl & Anastasio 1995a) and a maximum burial depth of c. 3 km based on subsidence reconstruction by Hogan & Burbank (1996). Deformation at Pico del Aguila included both halotectonic folding and thrust sheet emplacement (Anastasio 1992). The north–south-aligned fold axis was monoclinal folded about an east–west-aligned axis causing an oblique cross-section of the fold to be exposed in map view. The anticline affords excellent access to all structural and stratigraphic positions in the growth strata, which were sampled for AMS analysis.

**Methods**

Growth bed attitudes were defined from individual bedding measurements from referenced locations, bed mapping with real-time kinematic GPS positioning and bed tracing from referenced 1:5000 digital elevation models. Figure 2 shows the down-plunge projection of these growth strata on a plane perpendicular to the fold axis at the northern end of the structure. The resolution of the spatial position on growth horizons depends on GPS receiver resolution, bed attitude variation, compass precision and projection errors resulting from non-cylindrical folding and fold axis variation. In this study, the traces of growth strata are accurate at the decimetre scale. The traces of fold axial surfaces were determined from deflections in the growth strata. The general thinning and shallowing of the growth beds onto the anticline is consistent with limb rotation about pinned hinges during folding (e.g. Ford et al. 1997). The growth strata suggest limb length has been constant during folding and the position of no interlayer shear (pin line) jumped hinge-wards on the east-dipping limb during fold development. A décollement fold with pinned hinges is expected to experience a change in décollement depth during
fold tightening (Epard & Groshong 1995; Homza & Wallace 1995), easily accommodated within the evaporitic strata coring the fold.

To determine fold kinematics, standard 1 inch oriented cores were collected from 49 sites around Pico del Aguila from the Arguis and Belsué-Atarés formations, which display a growth relationship with the anticline (Fig. 3). Sample sites were chosen from mudstone and fine sand beds at all structural and stratigraphic positions around the anticline, including additional samples from beds sampled by Pueyo et al. (1997). At each site, at least four independent cores were drilled and oriented. Multiple cylinders were trimmed from each core to obtain standard 11 cm$^3$ specimens. A total of 6–14 samples from 27 sites were measured for AMS using the KLY-3's Kappabridge at Lehigh University and applying the 15 directional susceptibilities scheme of Jelinek (1978). Ten additional sites (AA, AD, AE, kin1-kin7) were measured in the paleomagnetic laboratory at the University of Michigan and data from 12 sites reported in Pueyo et al. (1997) are also reconsidered in this analysis.

Low-field room temperature (294 K) bulk susceptibility was measured from all specimens. Additional rock magnetic experiments (see Kodama et al. 2010 for further details) were conducted to determine the magnetic mineralogy of representative samples, including the magnetic mineralogy contributing to the magnetostratigraphy based on thermal demagnetization of three orthogonal components of isothermal remanent magnetization (IRM; coercivity-unblocking temperature analysis, Lowrie 1990) and the relative contributions to remanence by different coercivity components using IRM acquisition modelling (e.g. Kruiver et al. 2001).

Fifty specimens from eight sites were selected to evaluate paramagnetic mineral contributions to AMS as a function of grain size and structural position. These samples were measured for bulk susceptibility at low temperature (77 K) in liquid nitrogen in a modified susceptibility meter at the paleomagnetic laboratory at the University of Michigan, according to the method of Parés & van der Pluijm (2003), in order to enhance the paramagnetic fabric following the Curie–Weiss law. Thirty samples from five sites chosen to represent grain size variations in the Arguis syncline were measured to determine the anisotropy of anhysteretic remanence (AAR). AAR was imparted using a GSD-5 Schonstedt alternating field demagnetizer and measured with a 2G Enterprises Inc. superconducting magnetometer, both at Lehigh University. Each specimen was given an ARM in nine different orientations. A partial ARM was applied with peak alternating field between 0 and 50 mT with 0.097 mT constant
DC field to activate magnetite phases and exclude higher coercivity minerals. Data were analysed using the PMAG software by Tauxe (2002).

Results

The eigenvalues and eigenvectors of the AMS tensor \(k_1 \geq k_2 \geq k_3\) describe the magnetic fabric that can be related to the mineral fabric. The AMS of sites from the Arguis syncline, west limb, hinge and east limb of the Pico del Aguila anticline and Belsue syncline are shown in Figure 4a–e. Overall, the magnetic fabric is well defined with \(k_1\) and \(k_2\) in the plane of bedding and \(k_3\) parallel to the pole to bedding. At all sites, from all structural positions except the hinge area, \(k_1\) plunges shallowly to the north (c. 350° ± 20°) nearly parallel to the Pico del Aguila hinge line. In contrast, in the hinge region \(k_1\) axes plunge shallowly to the east or west (Fig. 4c).

The shape and anisotropy of the site-averaged AMS data are presented on a Jelínek (1981) plot in Figure 5. Triaxial fabric ellipsoids characterize nearly all sites. The degree of anisotropy \((P_j)\) is <1.06 for all samples, consistent with the lack of visible macroscopic tectonic foliation in the field. A stronger oblate AMS fabric characterizes most sites along the axial surfaces and adjacent to the mechanically stiff Guara Formation, which suggests these samples have higher rock strains. The lowest anisotropies \((P_j < 1.03)\) are associated with triaxial prolate fabrics at sites in the interior of the Arguis and Belsué synclines.

The mean bulk susceptibility of all samples is \(144 \times 10^{-6}\) SI, suggestive of a dominantly paramagnetic susceptibility (e.g. Pueyo-Anchuela et al. 2013; Fig. 6a). The distribution of all bulk susceptibility measurements is tightly clustered about the mean with a slight high side tail to the distribution. Low-temperature AMS measurements include specimens from all structural positions and examples of each lithology (Fig. 6b). The average low-temperature bulk susceptibility enhancement for the eight sites was 2.9, despite the inclusion of two coarse-grained Belsué–Atarés Formation sandstones (PdA17 and PdA27) which exhibit the lowest paramagnetic contribution to the bulk susceptibility. AAR measurements from representative samples are shown in Figure 6c. The AAR measurements isolate the anisotropy of remanence-carrying ferromagnetic grains in a sample and can be compared to the orientations of principal susceptibilities from nearby Arguis syncline samples AA, AD, AE, shown in Figures 4 and 7 in geographical and bedding-corrected reference frames, respectively. Comparison of the AMS and AAR from the same sites shows different orientations for \(k_1\) axes: WSW for AAR and NNW for AMS. \(k_2\) axes are in bedding while \(k_3\) axes are normal to bedding for both AAR and AMS measurements.

Discussion

Magnetic mineralogy

The use of magnetic fabric to characterize tectonic strain requires knowledge of magnetic mineralogy and pre-tectonic fabric. The AMS of a sample is a measure of the total magnetic fabric due to the preferred orientation of all diamagnetic, ferromagnetic and paramagnetic grains in a sample. The individual
Fig. 4. Lower hemisphere stereographic projections of principal magnetic susceptibilities: $k_1$ squares; $k_2$ triangles; $k_3$ circles shown in geographical coordinates. Bedding shown by great circle and the solid dot, which is the pole to bedding. Sample location and bedding attitude shown on map: (a) Arguis syncline; (b) Pico del Aguila west limb; (c) Pico del Aguila hinge; (d) Pico del Aguila east limb; (e) Belsué syncline.
(c) Pico del Aguila Hinge

(d) Pico del Aguila East Limb

Fig. 4. Continued.
grain anisotropy of ferromagnetic magnetite grains is typically controlled by grain shape and alignment, whereas paramagnetic phyllosilicate grains have magnetic susceptibility principal axes parallel to the crystallographic axes. The Arguis Formation includes variable amounts of diamagnetic quartz and carbonate grains; however, these minerals will contribute little to the AMS because of their weak negative susceptibilities (Rochette 1987). All magnetic susceptibility measurements in this study were positive.

Kodama et al. (2010) report that the magnetic mineralogy of the remanence-carrying grains in the Arguis Formation is mostly detrital magnetite with subsidiary magnetic sulphide, likely pyrrhotite, based on rock magnetic experiments. Because ferromagnetic minerals have high susceptibility, they can dominate the total fabric despite low concentration. The AAR isolates the ferromagnetic fabric and can therefore aid in the understanding of the AMS fabric. Comparison of the AAR (Fig. 6c) and AMS fabric from the same sites (Fig. 4c in geographical and Fig. 7a in stratigraphic reference frames, respectively) show that the principal axes differ in orientation, suggesting the ferromagnetic grains are not dominating the AMS fabric. Both AAR and AMS fabrics from the Arguis syncline samples display a similar compaction fabric with the least principal susceptibility ($k_3$) axes nearly perpendicular to bedding. However, the AAR $k_1$ axes
orientations are close to the trend of the basin axis, suggesting the magnetite grains could record a depositional fabric, and the AMS $k_1$ axes orientation of the paramagnetic phyllosilicate grains are approximately fold-axis parallel, suggesting they record a tectonic fabric.

Low-temperature AMS measurements were used to enhance the paramagnetic grain contribution to the AMS signal since paramagnetic susceptibility is enhanced as temperature is lowered, according to the Curie–Weiss Law. Susceptibility of a pure paramagnetic material at 77 K is expected to increase by a factor of 3.8 times the room temperature measurement (e.g. Schultz-Krutisch & Heller 1985; Ihmle et al. 1989). Figure 6 shows the comparison of low- and room-temperature susceptibility measurements. The increase of magnetic susceptibility at cold temperature as compared to room temperature measurements suggests the paramagnetic minerals dominate the AMS fabrics of the mudstones and sandstones. This is consistent with expectations of phyllosilicate mineral-rich rocks from analogous rocks elsewhere in the Pyrenean realm (e.g. Parés & van der Pluijm 2002; Pueyo-Anchuela et al. 2013). Because the AMS resides mostly on the paramagnetic phyllosilicate fraction,
we can use the AMS as a proxy for clay grain preferred orientation.

Magnetic fabric

Sedimentary rocks such as shale, mudstone, and marl, dominated by phyllosilicate minerals, are excellent for AMS studies because of the predominant oblate individual grain anisotropy which adjusts readily during lithification and any subsequent deformation. As grains reorient in response to deposition, compaction or tectonic strain, the magnetic fabric will continuously adjust. During deposition by water currents or simply by settling, the phyllosilicate grains will orient with their basal planes parallel or slightly imbricated to the depositional surface. Because the intermediate and maximum axes are nearly equal in magnitude they will be randomly oriented in bedding, but the minimum axes will tend to be vertical (e.g. Martín-Hernández & Hirt 2003). A weak lineation may form if the imbrication caused by deposition from flowing water yields a weak intersection lineation parallel to flow (e.g. Aubourg et al. 1991).

Depositional-diagenetic fabric and layer-parallel shortening strain

At all sites $k_3$ is nearly orthogonal to bedding. Some variation in $k_3$ clustering and orientation relative to the pole to bedding does exist, but the dispersion has no preferred bias and is therefore interpreted to record sediment deposition and compaction (Fig. 7a). Compaction during dewatering and lithification will amplify the initial oblate depositional fabric prior to the superposition of any tectonic fabric, and compaction fabrics only should produce oblate fabrics with low anisotropy that are not observed at Pico del Aguila (Fig. 8). The maximum and intermediate susceptibility axes are clustered in bedding rather than widely distributed in the bedding plane, consistent with some post-depositional tectonic strain (e.g. Kligfield et al. 1981, 1983; Kodama 2012; Figs 4 & 8).
The growth strata on Pico del Aguila have developed bedding fissility but lack a pervasive tectonic foliation including disjunctive or pencil cleavage. The lack of a mesoscopic fabric confirms the low level of finite strain in the rocks (e.g. Ramsay & Huber 1983). At nearly all sites the fabric shows a well-organized magnetic lineation (Fig. 4); however, the lineation magnitude is weak overall. At only a few sites, for example PdA1, PdA7 and PdA16, $k_1$ and $k_2$ have orientations distributed in bedding indicative of pencil cleaved rocks but below the strain threshold to develop a macroscopic fabric (e.g. Parés 2004). Kligfield et al. (1981) describe similar AMS tectonic lineations produced at low strains from siltstones in the Alps Maritime, France, as does Hrouda et al. (2009) in samples from the western Carpathians.

The $k_1$ orientation at Pico del Aguila is downdip in the plane of bedding except at the anticline hinge, where it is consistently shallower than bedding dip in plunge. Within the growth strata, $k_1$ is oriented $350 \pm 20^\circ$, parallel to the fold axis, except at anticline hinge area sites (Fig. 7). We interpret all the
clustered $k_1$ orientations at Pico del Aguila as a tectonic fabric. The dominant palaeocurrent direction along the southern Jaca Basin during the Eocene was to the west (Puigdefàbregas 1975), and was recovered by the AAR data (Fig. 6). In rocks dominated by phyllosilicate grains it is difficult to create a well-organized magnetic lineation by aligning grain crystallographic axes; however, an intersection lineation between slightly rotated clay grains orthogonal to a shortening direction has been observed previously (e.g. Henry 1997; Parés et al. 1999, 2007; Martín-Hernández & Ferré 2007; Yonkee & Weil 2010).

The c. north-trending $k_1$ axes recovered from the growth strata on the limbs of Pico del Aguila are attributed to layer-parallel shortening (LPS) which was subsequently passively reoriented by clockwise vertical-axis rotations during emplacement on the Guarga thrust sheet (Fig. 8b). The evidence for variable rotation magnitude has been recovered from remanence directions throughout the External Sierras (Pueyo et al. 2002; Mochales et al. 2010; Pueyo-Anchuela et al. 2012). Restoration of the rotation shows the LPS fabric is compatible with Pyrenean shortening directions (e.g. Anastasio & Holl 2001). Early LPS fabrics have also been observed in the Appalachians, USA (Nickelson 1979; Geiser & Engelder 1983; Gray & Mitra 1991) and in the Apennines, Italy (Tavernelli & Alvarez 2002). Hirt et al. (2004) interprets a similar strain history from mineral textures interpreted from an AMS study of the same area of the Appalachian fold–thrust belt where LPS was interpreted in the field.

### Table: AMS Fabric History

| AMS Fabric History | AMS Orientation (bedding corrected) | AMS Ellipticity (Jelinek Plot) | Deformation Regime |
|--------------------|-------------------------------------|--------------------------------|--------------------|
| Compaction and Layer-Parallel Shortening | ![Image](image1.png) | ![Image](image2.png) | ![Image](image3.png) |
| Folding | ![Image](image4.png) | ![Image](image5.png) | |
| Thrusting | ![Image](image6.png) | ![Image](image7.png) | |

**Fig. 8.** Deformation history of AMS fabric. Top: Depositional, compaction and layer-parallel shortening produced a fabric characterized by bedding-perpendicular $k_3$ axes and NW-trending $k_1$ axes, which were passively rotated clockwise throughout the structure during subsequent emplacement of the Guarga thrust sheet. The LPS fabric is closer to orthogonal to plate convergence (e.g. Roest & Srivastava 1991) and orogenic shortening directions (e.g. Holl & Anastasio 1995b) before subsequent clockwise vertical axis rotation of the Guarga thrust sheet. Middle: AMS fabric reorientation and anisotropy enhancement during fixed-hinge limb rotation accompanied detachment folding. Lower: Active fabric rotation with superposition of additional flattening strain in anticlinal hinge samples only during Guarga thrust sheet emplacement produced an east–west-trending $k_1$ axes orientation in the hinge area specimens. The evolution of the AMS fabric ellipsoid is shown on the associated Jelinek (1981) plots with ellipses (dashed: previous; solid: current) outlining the data distribution for each phase of the deformation history. The variation in block orientation relative to palaeo-north is to optimize the viewing perspective.
Fold kinematics

Growth triangles and constant bedding thickness supports rolling fold hinges during folding, while thinning and shallowing dip of growth strata support pinned fold hinges. Onlap and growth strata thinning towards the Pico del Aguila anticlinal hinge, along with shallow bedding dip upslope and constant fold limb length in the growth strata, support folding by limb rotation (spin) about a pinned anticlinal hinge (e.g. Riba 1976; Poblet & Hardy 1995; Salvini & Sorti 2002). The overall accumulation of strain as the sediments were lithified, folded and emplaced on the Guarga thrust sheet is faithfully recorded by the AMS (Fig. 8). The distribution of the AMS ellipsoids on a Jelínek (1981) plot and on a geological map of Pico del Aguila clearly show higher magnetic anisotropy \( (P_3) \) near fold axial surfaces (Fig. 5). With no difference in bulk susceptibility as a function of structural position or significant correlation between bulk susceptibility and the degree of anisotropy \( P_3 \) (Figs 5 & 6), there is little evidence to support a change in mineral composition or concentration between samples. A higher degree of magnetic fabric anisotropy along fold axial surfaces is compatible with pinned hinge fixed limb-length folding in agreement with growth strata geometries. Fixed limb-length detachment folding requires evacuation of material from the core and décollement depth migration during folding (e.g. Homza & Wallace 1995). The thrust fault and thinning of the Pont de Suert Formation in the fold core are consistent with the rotation of fixed-length limbs during folding (Poblet & McClay 1996; González-Mieres & Suppe 2006).

There is little deflection of the minimum axes \( (k_3) \) of the AMS ellipsoid from the bedding pole. Penetrative bedding-parallel shear of opposing sense on the fold limbs with clockwise rotation of \( k_3 \) orientation on the east fold limb and counterclockwise on the west limb would be expected for kinematic dominated by pinned hinge flexural folding (e.g. Ramsay 1967; compare Fig. 4b, d). There is also little field evidence (e.g. shear veins, slickensides) of discrete layer-parallel shear on bedding planes; the AMS fabric is therefore interpreted to support buckle folding about pinned hinges consistent with the growth strata geometry.

Thrusted fabric

The AMS fabric varies by structural position but not stratigraphic position in the Pico del Aguila samples. Hinge zone sites are the exclusive structural position where \( k_1 \) orientation is dominantly shallow and east–west plunging rather than northwards plunging, north–south trending and in bedding. The anticlinal hinge and synclinal sites are similar in bedding orientation and strain history. At these locations the beds have not been rotated about a north–south axis during folding, and both of these structural positions were carried in a similar fashion on the Guarga thrust sheet. The orientation of the \( k_1 \) axis in hinge thrust sites is consistent with c. south-directed shortening during final thrust emplacement rather than recording folding strain. Emplacement of the Guarga thrust sheet continued after transverse fold décollement, which varied in timing along the strike of the External Sierras (DePaor & Anastasio 1987; Anastasio 1992; Millán et al. 2000). Transport up the frontal footwall ramp of the Guarga thrust sheet generated the northwards plunge of the Pico del Aguila fold and further strained sites between the thick stiff mechanical layers of the Guara Formation, a massive bedded carbonate, and the thick bedded sands and gravels of the overlying Campodarbe Formation, where the mudstone and sandstone dominated flysch is thinnest (Fig. 7b). Anticline hinge sites have high anisotropies and a \( k_1 \) axis oriented east–west rather than the north–south orientation that characterizes synclinal sites at the same stratigraphic level (Figs 4 & 5). Importantly, this means that AMS fabrics in the anticline hinge samples do not preserve the LPS fabric that defined the early intersection lineation in the other samples (e.g. Larrañaga et al. 2004). The phyllosilicate fabric in the growth strata surrounding Pico del Aguila are only reoriented where the units are thin as additional rock strain was superimposed during thrusting up the frontal footwall ramp of the Guarga thrust sheet.

Conclusion

The phyllosilicate depositional-compaction fabric surrounding the Pico del Aguila anticline was deformed by layer-parallel shortening oriented east–west (NE–SW in the pre-rotational reference frame), orthogonal to the fold axis orientation, to create a north–south-trending magnetic lineation in bedding on fold limbs and in the adjacent Belsué and Arguis synclines, parallel to the anticlinal fold axes. The magnetic lineation is likely a grain-scale intersection lineation produced by the paramagnetic phyllosilicate grains that dominate the magnetic susceptibility. The magnitude of the \( k_1 \) lineation varies little across the fold and throughout the growth strata, but varies instead by structural position. The minimum principal susceptibility axes \( (k_3) \) are perpendicular to bedding across the fold, suggesting that bedding-parallel flexural shear strain did not much deflect the paramagnetic grains. Near the anticlinal hinge, the magnetic lineation is fold axis perpendicular and thought to be the result of later flattening strain related to thrust sheet emplacement.
This study has shown the AMS was useful for resolving fold kinematics and to partition strain history at diagenetic conditions. We showed that the tectonic magnetic fabric developed early in the deformation history and, rather than locking in, the AMS fabric continued to evolve, recording all phases of the progressive deformation history.

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