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This is a pre print version of the following article:

Original Citation:

Availability:

This version is available http://hdl.handle.net/2318/1660571

since 2018-02-23T11:09:39Z

Published version:

DOI:10.1007/s00531-018-1593-y

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1 2 3	Kinematic and geochronological constraints on shear deformation in the Ferriere-Mollières shear zone (Argentera-Mercantour Massif, Western Alps): implications for the evolution of the Southern European Variscan Belt
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36 Kinematic and geochronological constraints on shear deformation in the Ferriere-Mollières shear zone

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system of dextral shear zones of the southern portion of the EVSZ.

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49 Abstract

50 In the Western Alps a steeply dipping km-scale shear zone (the Ferriere-Mollières shear zone) cross-cuts Variscan 51 migmatites in the Argentera-Mercantour External Crystalline Massif.

52 Structural analysis joined with kinematic vorticity and finite strain analyses allowed to recognize a high-temperature 53 deformation associated to dextral transpression characterized by a variation in the percentage of pure shear and simple 54 shear along a deformation gradient. U-Th-Pb dating of syn-kinematic monazites was performed on mylonites. The oldest ~340 Ma ages were obtained in protomylonites whereas ages of ~ 320 Ma were found in mylonites from the core 55 56 of the shear zone. These ages indicate that the Ferriere-Mollières shear zone is a still preserved Variscan shear zone. 57 Ages of ~320 Ma obtained in this work are in agreement with ages of the dextral transpressional shear zones occurring 58 in the Maures-Tanneron Massif and Corsica-Sardinia. However, transpression in the Argentera-Mercantour Massif 59 started earlier than in other sectors of the southern Variscan Belt. This is possibly caused by the curvature of the belt 60 triggering the progressive migration of shear deformation. Our data allow a correlation between the Argentera-61 Mercantour Massif and other segments of the Southern European Variscan Belt, in particular with Maures-Tanneron 62 Massif and Corsica-Sardinia and contribute to fill a gap in the age of activity and in the kinematics of the flow of the

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Key words: Transpression, Argentera-Mercantour Massif, Mylonites, East Variscan Shear Zone, Monazite

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69 1. Introduction

The European Variscan Belt is the result of a Devonian–Carboniferous continent-continent collision (Arthaud and Matte, 1977; Burg and Matte, 1978; Tollmann, 1982; Matte, 1986a,b; Franke, 1989) between Laurentia-Baltica and Gondwana (Matte, 1986a, 2001). Between these two continents, small microplates existed, defined essentially on the basis of palaeomagnetism and palaeobiostratigraphy (Scotese and McKerrow, 1990; Franke et al., 2017, with references therein). The main microplates are known as Avalonia and Armorica (Matte, 2001). They broke away from Gondwana during the early Palaeozoic, prior to docking against Baltica and Laurentia before the Devonian-Carboniferous collision (Matte, 2001).

The Architecture of the Variscan Belt is well defined in eastern, central and western Europe (Matte, 1986b, 2001; Fluck
et al., 1991; Ballévre et al., 2009; Faure et al. 2009; Skrzypek et al., 2012) where the Saxothuringian Zone
(southwestern Iberian Massif, northern Armorican Massif, Vosges and Bohemian Massif) and the Moldanubian Zone
(southern Iberian Massif, central and southern Vosges, Massif Central, southern Armorican Massif and southern
Bohemian Massif) are recognized (Fig. 1a).

In contrast, the structural arrangement of the SE segment of the European Variscan Belt is less clear because of tectonic
 reworking during Alpine Orogeny.

Some authors proposed that the European Variscan Belt is characterized by a composite orocline, developed because of indentation tectonics (Matte and Ribeiro, 1975; Matte, 1986a,b), made by two main branches: the well-known western Ibero–Armorican arc (Matte and Ribeiro, 1975; Brun and Burg, 1982; Dias and Ribeiro, 1995; Dias et al., 2016; Fernández-Lozano et al., 2016) and the eastern branch (Matte, 2001; Bellot, 2005), delimited by a system of regionalscale dextral, transpressive shear zones known as the East Variscan Shear Zone (EVSZ; Corsini and Rolland, 2009; Carosi et al., 2012; Padovano et al., 2012, 2014). The EVSZ is actually less understood, expecially in the sectors that are now part of the Alps.

Although during the Carboniferous extensive transpressive shear deformation is recognized in several fragments
belonging to the Southern European Variscan Belt (Matte, 1986a,b; Carosi and Palmeri, 2002; Iacopini et al., 2008;
Frassi et al., 2009; Carosi et al., 2012; Corsini and Rolland, 2009; Guillot et al., 2009; Schneider et al., 2014), the
impact of this deformation on the arrangement of the belt needs further constraints.

95 A deeper knowledge of Variscan deformation is of great importance for enhancing correlations between the fragments 96 of the belt in the Mediterranean area since these correlations are mostly based on lithological and stratigraphic affinities 97 and paleomagnetic data. In particular, the correlation between Corsica-Sardinia Block, Maures-Tanneron Massif and 98 Variscan basement of the Western Alps (External Crystalline Massifs) is still debated. According to some authors these 99 three sectors should have been in continuity during Variscan orogenesis (Rollet et al., 2002; Rosenbaum et al., 2002; Advokaat et al., 2014), while according to Stampfli et al. (2002), Turco et al. (2012), von Raumer et al. (2013), the
Corsica-Sardinia Block was located in a more westerly position connected to Iberia and therefore, away from southern
France and the future Alpine External Crystalline Massifs.

103 In the Western Alps, the Variscan External Crystalline Massifs are cross-cut by km-scale shear zones whose age of 104 activity and deformation regime are not entirely clear. It is also not clear whether these shear zones have been fully 105 reactivated during the Alpine orogeny, or if evidence of Variscan deformation is still preserved.

A useful method to date shear zone activity, especially when developed under high-temperature conditions, is the U-ThPb geochronology on syn-kinematic monazite (Williams et al., 2007). Reliable results were obtained in the Himalayan
belt in dating both the activity of regional–scale shear zones (Carosi et al., 2010; Montomoli et al., 2013, 2015; Cottle et
al., 2015; Iaccarino et al., 2015, 2017; Carosi et al., 2016a) and metamorphic events (Khon et al., 2005; Khon, 2008;
Larson and Cottle, 2015). In high-temperature shear zones, this method is more reliable than the ⁴⁰Ar/³⁹Ar method
because argon isotopes are often mobile during deformation (Dunlap et al., 1991; Mulch and Cosca, 2004; Villa et al.,
1997, 2014; Villa, 2015; Challandes et al., 2003, 2008; Sanchez et al., 2011).

113 In the present work we focus on the northern sector of the Ferriere-Mollières shear zone (FMSZ; Faure-Muret, 1955; 114 Malaroda et al., 1970; Compagnoni et al., 2010) located in the Argentera-Mercantour Massif, a km-scale shear zone in 115 the Alpine External Crystalline Massifs. The FMSZ is constituted by mylonites developed from Variscan migmatites 116 and leucogranites (Musumeci and Colombo, 2002; Carosi et al., 2016b). Conflicting deformation ages have been 117 proposed: Musumeci and Colombo (2002) obtained a cooling age for mylonitic leucogranite of 327 ± 3 Ma (whole rock 118 Rb/Sr ages on magmatic muscovite grains) interpreted as the younger limit of the FMSZ activity. Corsini et al. (2004) 119 and Sanchez et al. (2011) proposed deformation ages of ~ 22 and ~ 20 Ma (40 Ar/ 39 Ar on phengites from mylonitic 120 micaschists and from ultramylonites of the Argentera granite), suggesting a reactivation of the Variscan shear zone (e.g. 121 Valetta shear zone) and formation of new shear zones (e.g Fremamorta-Colle del Sabbione shear zone) during Alpine 122 orogeny under greenschist metamorphic conditions.

This paper aims to clarify both the kinematics of the flow and the age of activity of the FMSZ in order to check if Variscan deformation structures are still preserved and if dextral shear deformation affecting the Alpine External Crystalline Massifs can be linked to the activity of other similar transpressional shear zones in the Southern European Variscan Belt. To verify this hypothesis, we carried out a kinematic vorticity analysis and a U-Th-Pb geochronological study on syn-kinematic monazites, combined with structural and microstructural analyses on different types of mylonites recently mapped by Carosi et al. (2016b).

130 2. Geological Setting of the Argentera-Mercantour Massif

The External Crystalline Massifs in the Western Alps are, together with the Corsica-Sardinia Block and the Maures Tanneron Massif, fragments of the Variscan Belt well-preserved within the Mediterranean area.

The Argentera-Mercantour Massif is constituted by the Gesso-Stura-Vésubie (GSV) and the Tinée (TMC) metamorphic complexes (Malaroda et al., 1970; Compagnoni et al., 2010) which are separated by the FMSZ (Fig. 1b). The GSV complex is made of migmatitic gneiss derived from Late Ordovician granitoids and migmatitic paragneiss. Ferrando et al. (2008) and Compagnoni et al. (2010) recognized a metamorphic evolution characterized by: 1) HP metamorphic peak; 2) initial decompression stage; 3) HT–MP amphibolite-facies metamorphism; 4) LT–LP amphibolite-facies metamorphism.

The TMC migmatites resulted from partial melting of metasediments. Eclogite relicts are reported by Faure-Muret (1955) and Malaroda et al. (1970). According to Compagnoni et al. (2010) the TMC shows a minor degree of melting than the migmatites of the GSV complex as testified by a smaller proportion of leucosomes. A greenschist facies imprint is mainly due to a system of ductile to brittle-ductile shear zones (Baietto et al., 2009) affecting the whole Argentera-Mercantour Massif (Corsini et al., 2004; Sanchez et al., 2011). The main Alpine faults of the Argentera-Mercantour Massif are the NW-SE Bersezio Fault Zone, in the central part, and the E-W Fremamorta-Colle Sabbione Shear Zone, in the southernmost part of the massif (Fig. 1b).

146 The FMSZ (also known as Valletta shear zone; Corsini et al., 2004) strikes NW-SE and extends for about 20 km (Fig. 147 1b) with a thickness between 100 m to the south-east and more than 1000 m to the north-west. The fault rocks in the 148 FMSZ resulted from the shearing of migmatites of the GSV and the TMC. The transition from non-sheared migmatites 149 to protomylonites, mylonites and ultramylonites can be observed along a deformation gradient towards the center of the 150 shear zone (Fig. 2; Carosi et al., 2016b). The rocks in the shear zone are medium-grained dark mylonitic schists with 151 biotite and white mica ± garnet. Mylonites and ultramylonites with chlorite and white mica also occur in the central part 152 of the shear zone while mylonites and protomylonites with biotite and sillimanite \pm garnet are present in the margins of 153 the shear zone. Because of the constant orientation of the mylonitic foliation, the mineral lineation and the sense of 154 shear in all the rocks inside the FMSZ, Carosi et al. (2016b) interpreted these parageneses as linked to an evolution of 155 the deformation under decreasing temperature.

- 157 3. Structural Analysis
- 158
- 159 3.1 Variscan shear zone

160 The main structural elements in the FMSZ are a mylonitic foliation defined by the preferred orientation of biotite + 161 white mica and a protomylonitic foliation with biotite + sillimanite in the external part of the shear zone, both striking 162 N100-140 and steeply dipping towards the northeast or the southwest (Fig. 2). A mineral lineation, defined by a 163 preferred alignement of stretched quartz and feldspar, trends N110-130 and dips 20° towards the northwest (Fig. 2).

164 In protomylonites the foliation is an anastomosed disjunctive cleavage (Passchier and Trouw, 2005) formed by 165 alternating coarse-grained layers of deformed quartz and feldspar and biotite- and sillimanite-rich layers (Fig. 3a,b). 166 Sillimanite occurs both in prismatic and fibrolitic habitus. In the mylonites, a transition can be observed from a 167 disjunctive cleavage with sub-parallel cleavage domains, formed by alternating layers of deformed quartz and biotite + white mica (Fig. 3c,d), to a continuous cleavage in the most intensely deformed ultramylonites rich in phyllosilicates 168 169 (Fig. 3e,f). The mylonitic foliation underwent a later gentle folding of uncertain age with axial planes moderately dipping toward the SW and NW-SE trending axes (Fig. 2). Kinematic indicators (S-C' fabric, micafish, mantled 170 171 porphyroclasts and quartz oblique foliation) indicate a strike-slip component of movement with a minor reverse, top-to-172 the-SE, component where the foliation dips northeast (Fig. 4a,b,c) and with a minor normal component where it dips 173 southwest. Quartz shows subgrain rotation recrystallization (Piazolo and Passchier, 2002; Stipp et al., 2002) and only 174 very minor effects of grain boundary migration in mylonites (Fig. 4d) and ultramylonites. Quartz in protomylonites is 175 mainly affected by grain boundary migration (Piazolo and Passchier, 2002; Stipp et al., 2002). Feldspar shows undulose 176 extinction due to ductile deformation (Fig. 4e) and does not show evidence of brittle deformation. The thickness of the 177 FMSZ progressively decreases from NW to SE. The maximum thickness of ~2 km is reached to the NW, in the Ferriere 178 village area (Fig. 1b). Thickness variations of the different types of mylonites, at a local scale, were also reported (in 179 agreement with Carosi et al., 2016b, Fig. 2). In the wall rocks of the FMSZ, in the GSV and TMC complex, Carosi et al. 180 (2016b) and Simonetti et al. (2017) recognized outcrop-scale open to tight symmetric upright folds with NW-SE 181 oriented axial plane and NW dipping axis deforming the main foliation of the migmatites.

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184 **3.2** Alpine Shear Zones

Shear zones cross-cutting the mylonites of the FMSZ (Fig. 2) and the migmatites of the two metamorphic complexes have been identified (Fig. 5a). These shear zones have metric to decametric thickness with lateral continuities up to several hundred meters. They consist of fine-grained mylonites or phyllonites with quartz and feldspar porphyroclasts in a fine-grained chlorite + white mica matrix. This paragenesis is indicative of greenschist facies metamorphism. Biotite relicts, indicating an older, higher temperature foliation, are still recognizable. The proportion of matrix is ~ 65% - 190 75%. The foliation strikes nearly E-W and dips at moderate to low angles toward the north with a north plunging191 mineral lineation.

The foliation is a spaced cleavage defined by the preferred orientation of chlorite and white mica. Kinematic indicators, mainly micafish and S-C and S-C' fabrics, indicate a top-to-the-S reverse sense of shear (Fig. 5b). C and C' planes are characterized by the presence of fine-grained chlorite. Along the S foliation biotite relicts are sometimes present (Fig. 5c). Quartz shows undulose extinction and subgrains, the latter indicating subgrain rotation recrystallization (Piazolo and Passchier, 2002; Stipp et al., 2002) as the dominant deformation mechanism. Feldspar shows undulose extinction and deformation lamellae while some fractured porphyroclasts are also present.

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199 4. Kinematics and vorticity of the flow of the FMSZ

Deformation in shear zones is often approximate to simple shear, especially in high-strain zones (Passchier 1987, 1991;
Iacopini et al., 2010, 2011). However, a significative component of pure shear has been detected in many regional-scale
shear zones around the world, under various metamorphic conditions (Carosi and Palmeri, 2002, Carosi et al., 2005;
Carosi et al., 2006, 2007; Festa et al., 2016; Goscombe et al., 2003, 2005; Iacopini et al., 2008; Kurz and Northrup,
2008; Larson and Godin, 2009; Law et al., 2004; Li et al., 2016; Nabavi et al., 2016; Sarkarinejad and Azizi, 2008;
Sarkarinejad et al., 2015; Wu et al., 2016; Xypolias and Kokkalas, 2006; Zang and Teyssier, 2013).

In order to characterize the type of flow and the deformation of the FMSZ, a kinematic vorticity analysis was performed on mylonites collected along three transects (A, B and C) perpendicular to the shear zone boundaries and parallel to the deformation gradient (Fig. 6). In addition, we performed a finite strain analysis in order to obtain the axial ratio of the finite strain ellipsoid (Rxz), which is indispensable for calculating the shortening perpendicular to the flow plane.

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211 **4.1 Methods**

212 Pure and simple shear can be expressed through the dimensionless mean kinematic vorticity number (Wm). Within a 213 ductile flow, the non-coaxial component of deformation, defined vorticity, can be normalized to the stretching along the 214 strain axes in order to obtain a dimensionless number that allows to compare the different types of flow. In this way it is possible to define the kinematic vorticity number as $Wk = W/|d^2-d^3|$ (Passchier, 1987). W represents the vorticity and 215 216 d2-d3 represents the difference of stretching along the intermediate and minimum principal strain axes. The mean 217 kinematic vorticity number Wm can be assumed equal to Wk because it represents an average value over the 218 deformation interval during which the structure or fabric formed (Xypolias, 2010; Fossen and Cavalcante, 2017). Wm is 219 therefore a measure of the proportion of pure and simple shear components: Wm = 0 for pure shear; Wm = 1 for simple

shear; 0<Wm<1 for general shear (Passchier, 1987). Simple and pure shear contribute equally to the flow for a value of
Wm = 0.71 (Law et al., 2004; Xypolias, 2010).

The kinematic vorticity analysis was carried out on thin sections oriented parallel to the XZ plane of the finite strainellipsoid.

224 It was performed using the S-C' method (Kurz and Northrup, 2008), which is based on the measurement of the 225 orientation of C' planes (Fig. 7a) with respect to the shear zone boundary. In any type of flow it is possible to recognize 226 two lines, defined as flow apophyses (A1 and A2), along which the particles do not undergo rotation. The component of 227 simple shear decreases with the angle between the two apophyses: the flow apophyses are orthogonal for pure shear 228 flow, and coincide with each other for simple shear flow. For general shear flow A1 and A2 form an acute angle in the 229 direction of the flow (Fig. 7a). The C' plane represents the bisector of the angle 2v between the A1 and A2 flow 230 apophyses (Fig. 7a; e.g. Kurz and Northrup, 2008). The vorticity number can be derived through the relation Wk =231 $\cos 2v$ (Kurz and Northrup, 2008) where v is the angle between C' and C planes that assumed are parallel to the flow 232 apophysis A2 (Kurz and Northrup, 2008; Gillam et al., 2013). Following the procedure proposed by Gillam et al. 233 (2013), we measured the v angle since the mylonitic foliation has a constant trend, sub-parallel to the well-defined and 234 mapped shear zone boundaries (Carosi et al., 2016b), in the studied transects.

As each element in a flow tends to rotate, including C' planes, it is necessary to consider the largest value for v, among those measured, to estimate the initial amplitude of the angle during formation of the C' planes. Gillam et al. (2013) proposed to use an average value of the angle v for low-strain rocks or in cases where C' planes developed in a late stage of the deformation history and nucleated in a stable orientation, as in this case they underwent

little or no rotation after inception. For highly strained rocks deformed in long lasting shear zones the maximum value
of v is preferable because, as demonstrated by Kurz and Northrup (2008), the average value is not representative of the
original angle of nucleation of C' planes.

Finite strain analysis has been performed on the XZ section of oriented samples applying the center-to-center method (Fry, 1979). The analysis was performed on samples in which an adequate amount of feldspar porphyroclasts with a similar grain size are present. Feldspar porphyroclasts show a homogeneous distribution in the samples used for the analysis and therefore are in agreement with the fundamental assumption of the method (see Genier and Epard, 2007 for a critical review). The center-to-center analysis was carried out using the software EllipseFit 3.2 (Vollmer, 2005).

247 The data obtained from strain and vorticity analyses were combined using the equations proposed by Wallis et al.

248 (1993) and Law et al. (2004) to obtain the percentages of shortening and stretching of the shear zone (Fig. 7b).

249 In order to check the kinematics and the type of deformation of the shear zone we calculated the angles θ , between the

250 maximum Instantaneous Stretching Axis (ISA_{max}) in the horizontal plane and the shear zone boundary, as this parameter

appears to be fundamental to distinguish between transpression and transtension, according to Fossen and Tikoff (1993)

252 and Fossen et al. (1994). θ angles larger than 45° are indicative of transtensive deformation while θ angles smaller than

 45° are indicative of transpressional deformation.

254 The calculation was performed using the formula proposed by Xypolias (2010) according to which $Wk = \sin 2\theta$ and

- 255 consequently $\theta = (\arcsin Wk)/2$. Results were checked with the software Strain Calculator 3.2 (Holcombe, 2009).
- 256

257 4.2 Results of vorticity and strain analyses

Results of vorticity and strain analyses are reported in table 1, polar histograms used to derive the angle v are reported in figure 8 and Fry graphs of the XZ section are reported in figure 9. In each of the three transects the estimated Wk values are very similar taking into account the degree of deformation of the samples. In protomylonites (samples: ARG 144, ARG 143, ARG 19) Wk varies from 0.32 to 0.34. In mylonites (samples: ARG 45, ARG 21a, ARG 68, ARG 116, ARG 117, ARG 115) Wk varies between 0.37 and 0.69. In ultramylonites (samples: ARG 35, ARG 101, ARG 37, ARG RCV, ARG 41) Wk is variable between 0.81 and 0.89. It is therefore possible to observe a change in the value of Wk along the deformation gradient toward the center of the shear zone.

Finite deformation analysis, combined with Wk values, allowed to obtain shortening and stretching values quite concordant in all the studied samples: shortening varies from 19% to 37% with an average value of 27%; stretching ranges from 23% to 59% with an average value of 36%. θ varies from 9° in protomylonites to 31° in ultramylonites.

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270 5. Geochronology

In-situ U-Th-Pb analyses were performed with LA-ICP-MS on monazites in five samples (ARG 19, ARG 21, ARG 27,

ARG 29, ARG 48b) collected along the deformation gradient of the shear zone along the crest of Costabella del Piz

273 (Fig. 6c). Sample ARG 19 is a protomylonite derived from biotite- and sillimanite-bearing migmatite. Samples ARG21,

ARG 27, ARG 29 and ARG 48b are mylonites derived from biotite- and white mica-bearing migmatites.

Dating was performed following the analytical procedure suggested by Montomoli et al. (2013). First, minerals with high relief and high birefringence were located using the optical microscope. Subsequently the samples were polished and C-coated to be inspected using a Scanning Electron Microscope (JEOL JSM IT300LV), at the University of Torino (Italy), which allows to distinguish monazite from zircon. On grains selected for dating (examples reported in figure 10), quantitative chemical analyses, reported in Online Resource 1 (spots reported in figure 11), and compositional maps (Fig. 11) were acquired by using electronic microprobe (JEOL 8200 Super Probe), at the University of Milano (Italy), in order to highlight any compositional zoning. 282 Monazite grains were analyzed in situ by laser-ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) 283 on 30-µm-thick sections at the CNR-Istituto di Geoscienze e Georisorse U.O. Pavia (Italy) using an Ar-F 193-nm excimer laser (GeolLas 102 from Micro-Las) coupled with a magnetic sector ICP-MS (Element I from Thermo-284 285 Finnigan). The full description of the analytical procedures is reported in Paquette and Tiepolo (2007) and Tiepolo 286 (2003). Single analyses were performed by a one-minute acquisition of the background signal followed by recording, 287 for at least 30 seconds, the ablation signal of the masses: ²⁰²Hg, ²⁰⁴(Hg+Pb), ²⁰⁶Pb, ²⁰⁷Pb, ²⁰⁸Pb, ²³²Th and ²³⁸U. The 288 presence of common Pb was evaluated in each analysis on the basis of the net signal of ²⁰⁴Pb (i.e. subtracted for the interference of ²⁰⁴Hg and background). None of the samples gave ²⁰⁴Pb counts above the background level. However 289 290 the relatively high Hg signal in the gas blank does not exclude the effective presence of common Pb in the analyzed 291 monazites. Analytical conditions were 10 µm diameter of spot size, 8 J/cm² of energy density, and 3 Hz of repetition 292 rate. Time-resolved signals were carefully inspected to verify the presence of perturbations related to inclusions, 293 fractures or mixing of different age domains. Laser-induced elemental fractionation and mass bias were corrected using 294 matrix-matched external monazite standard (Moacir monazite: Cruz et al., 1996; Seydoux-Guillaume et al., 2002a,b) 295 considering the values, re-calibrated for isotopic disequilibrium, reported by Gasquet et al. (2010). Eight to nine 296 external standards were analyzed in each analytical run and only those close to the reference values (at least 4 in each 297 run) were considered in order to reduce errors related to the standard reproducibility (Table 2). The relative standard 298 deviation of the analyses was mostly within 2% - 4%. External standards and unknowns were integrated over the same 299 time intervals to ensure the efficient correction of fractionation effects. Data reduction was carried out with the 300 GLITTER® software (van Achterbergh et al., 2001). In order to better estimate the uncertainty affecting the ²⁰⁶Pb/²³⁸U, 301 ²⁰⁷Pb/²³⁵U and ²⁰⁸Pb/²³²Th isotope ratios, the external reproducibility of the standard was propagated relative to 302 individual uncertainties for the isotope ratios. This procedure was carried out for each analytical run as reported in 303 Horstwood et al. (2003). After this error propagation each analysis is accurate within the quoted errors. Data processing 304 and plotting was done with the macro ISOPLOT/Ex (Ludwig, 2003).

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306 5.1 Characterization of the Monazites

307 Monazite belongs to the family of orthophosphates with general formula A(PO₄) and is an anhydrous rare earth

308 phosphate (Williams et al., 2007). Monazite is ideal for recording the progress of geological processes, because with its

309 extremely variable composition it reflects the chemical and physical changes of the host rocks and preserves the age of

- crystallization or growth over long geological periods (Williams et al., 2007). It is ideal for the U-Th-Pb method
- because it can have large concentrations of U and Th but does not incorporate Pb during its formation.

312 In systems in which the monazite is associated with garnet it is possible to observe partition phenomena of Y and HREE 313 between these two phases (Williams et al., 2007). After xenotime, garnet is the main mineral that can accommodate 314 HREE and Y in metamorphic rocks and its growth or its destabilization strongly influence the distribution of Y and 315 HREE in metamorphic monazites (Pyle and Spear, 1999; Pyle et al., 2001): during prograde metamorphism Y tends to 316 enter preferentially in garnet while during retro-metamorphism Y tends to enter preferentially in monazite (Pyle and 317 Spear, 1999; Pyle et al., 2001; Williams et al., 2007). The microstructural study, combined with the analysis of the 318 variation of the composition of monazites in deformed rocks, allows to understand in which moment of the deformation 319 history the two phases have grown.

Monazites were analyzed for microstructural position (examples of monazites located in different structural positions are reported in figure 10) and the Y content and zoning highlighted by the compositional maps acquired by electron microprobe. In both protomylonite (ARG 19) and mylonites (ARG 21, ARG 27, ARG 29, ARD 48b) no xenotime crystals have been detected and cores of the monazites have often a lower Y content with respect to the rims. Some monazites have asymmetrical shapes. Monazites in mylonites show irregular and frayed edges with asymmetric rims constituted by allanite ± apatite often developed in extensional sites around the crystals.

326

327 5.2 U-Th-Pb results

328 U-Pb data (Table 3) provide mainly discordant ages, with eight concordant U-Pb ages ranging from 348 to 297 Ma (Fig. 12). Except for ARG 19 protomylonite, ²⁰⁶Pb /²³⁸U and ²⁰⁸Pb/²³²Th data yield more concordant ages (Table 3, Fig. 12). 329 Both ²⁰⁶Pb /²³⁸U and ²⁰⁸Pb/²³²Th ages range mainly from 340 to 320 Ma and define a broad rejuvenation trend from the 330 331 least deformed sample (ARG 19 protomylonite) to highly sheared samples in the core of the shear zone (Fig. 12). 332 We did not note a systematic correlation between ages and monazite chemical domains. The oldest U-Th-Pb ages were 333 obtained from a monazite grain included in garnet (>400 Ma; Mnz 59, sample ARG 48b) and a monazite grain included in a deformed white mica (206 Pb/ 238 U age of 374 ± 6 Ma; Mnz 9, ARG 29), suggesting a textural control and inheritance 334 335 of the obtained ages.

336

337

338 6. Discussion

339 The FMSZ is a steeply dipping shear zone with a main right-lateral reverse top-to-the-SE sense of shear, whereas

340 locally it shows a normal sense of shear, depending on the dip direction of the mylonitic foliation. The FMSZ

341 developed under conditions of decreasing temperature, starting from the HT amphibolite facies to at least the LT

342 amphibolite facies, as suggested by the presence of sillimanite + biotite and biotite + white mica along the

343 protomylonitic and mylonitic foliations, respectively. The geometry of the foliation and mineral lineation as well as the

344 kinematic indicators are concordant in the protomylonites and in the mylonites. Temperature during Alpine deformation

was 375 ± 30 °C (greenschist facies metamorphism; Corsini et al., 2004; Sanchez et al., 2010) and, as a consequence,

346 the amphibolite-facies metamorphism may be inferred to be pre-Alpine.

347 The results allowed us to define for the first time the deformation regime and the finite strain of the FMSZ.

348 The kinematic vorticity data obtained by the S-C' method (Kurz and Northrup, 2008) allowed us to quantify the

deformation, characterized by a non-coaxial flow, in terms of percentage of pure and simple shear components.

350 Samples from the less deformed lithotypes, localized in the external portion of the shear zone, record a deformation

dominated by pure shear. The amount of simple shear increases towards the center of the shear zone, from ~ 24 % in the

352 protomylonites to ~ 62% in the ultramylonites (Fig. 13a). Finite strain analysis yields average shortening and stretching

values of 27 % and 36 %, respectively.

354 Taking into account the type of flow, the finite strain and the angles θ between the maximum horizontal ISA and the 355 shear zone boundaries (Fig. 13b), we can recognize a change from a pure shear dominated transpression to a simple 356 shear dominated transpression (Fig. 13c) according to the models proposed by Fossen and Tikoff (1993) and by Fossen 357 et al. (1994). The attitude of the mylonitic foliation is sub-vertical along the 20 km length of the shear zone, in 358 accordance with a transpressive shear zone setting (Fossen et al., 1994; Fossen, 2016). In the case of transtensional 359 deformation as suggested by Musumeci & Colombo (2002), the attitude of mylonitic foliation is expected to be sub-360 horizontal due to the main sub-vertical shortening direction. Iacopini et al. (2008), discussing the orientations of the 361 elongation lineation, suggest two kinds of transpressive shear zones: in the case of a sub-vertical lineation the shear 362 zone was subjected to vertical extrusion while in the case of a sub-horizontal lineation extrusion was horizontal. The 363 elongation lineation within the FMSZ is gently plunging, thus it would be compatible with a sub-horizontal direction of 364 extrusion. Extrusion and shear zones with a component of pure shear are generally subjected to strain compatibility 365 problems (Hudleston, 1999) and it is not trivial to explain how the flattening component of the deformation is 366 accommodated.

To fix this problem Ramsay and Huber (1987) and Hudleston (1999) suggested that the flattening component can be obtained in a simple shear regime by volume loss in the deforming material. In this way the strain compatibility is maintained. However, this is not the case of the FMSZ because there is no evidence of structures suggesting major volume loss in the deformed rocks and the kinematic vorticity analysis revealed a major component of pure shear during deformation.

Anyway, according to Fossen (2016) strain compatibility problems are also overcome if the wall rocks are deformed bythe same amount of coaxial strain as the shear zone. The wall rocks of the FMSZ, i.e. the GSV and the TMC complexes

are deformed by open to tight upright folds with axial planes parallel to the mylonitic foliation in the FMSZ. According

to Carosi et al. (2016b) and Simonetti et al. (2017) these upright folds accommodate the component of shortening

perpendicular to the FMSZ. According to this, the wall rocks and the shear zone could stretch together with no straincompatibility problems.

378 The Argentera-Mercantour Massif was involved in the Alpine orogeny at ~22 Ma (Corsini et al., 2004, Sanchez et al.,

2011). Most of the Alpine deformation is concentrated in the sedimentary covers, detached from the metamorphic

basement along gypsum and limestone breccias levels (Malaroda et al., 1970), in which several folding phases are

381 recognizable (d'Atri et al., 2016; Barale et al., 2016). Evidence of Alpine deformation in the basement are brittle/ductile

382 strike-slip faults and top-to-the-S reverse shear zones developed under greenschist facies conditions during the lower

383 Miocene (Corsini et al., 2004; Baietto et al., 2009). Variscan HT foliation and mylonitic foliation in the FMSZ all along

its length are always subvertical and are not affected by strong subsequent folding.

385 U-Th-Pb geochronology on monazites provides no Alpine ages showing that the shear zone was not significantly

reactivated during the Alpine orogeny. U-Pb data are discordant whereas the ²⁰⁶Pb/²³⁸U and ²⁰⁸Pb/²³²Th ages show a

better concordance. As recently demonstrated by Erickson et al. (2015), the perturbation of the U-Th-Pb systematics of

388 monazite can be caused by intracrystalline deformation. We interpret the obtained geochronological data combining

textural observations, monazite chemistry and age distribution. The oldest U-Th-Pb ages were obtained from monazite

grains included in garnet (Fig. 10a) and within a deformed white mica (Fig. 10b). These grains probably escaped

391 subsequent deformation and other events perturbating the U-Th-Pb systematics. They can be interpreted as inherited

ages from older tectono-metamorphic events as previously documented by other authors (Monié and Maluski 1983;

393 Rubatto et al. 2001; Corsini et al. 2004).

The ²⁰⁶Pb /²³⁸U and ²⁰⁸Pb/²³²Th ages range mainly from about 340 to 320 Ma. The oldest ²⁰⁶Pb /²³⁸U ages of this range
were obtained in monazites along the protomylonitic foliation (Fig. 10c) close to the margin of the shear zone.

396 This age, according to the syn-kinematic mineral assemblage and to the main deformation mechanism of quartz and

397 feldspar, suggests that Variscan shear deformation in the Argentera-Mercantour Massif started at the metamorphic peak

398 (dated at ~340 Ma in the GSV complex, Compagnoni et al., 2010; Rubatto et al., 2010) or shortly after it during the first

partial melting event and triggered the exhumation of the GSV complex.

400 Both ²⁰⁶Pb/²³⁸U and ²⁰⁸Pb/²³²Th ages become younger along the deformation gradient (Fig. 14) and monazite grains

401 along the mylonitic foliation in the core of the shear zone (Fig. 10d) show $^{206}Pb/^{238}U$ and $^{208}Pb/^{232}Th$ ages of $\sim 320 \pm 6$

402 Ma. This broad correlation between ages and deformation gradient suggests a progressive concentration of deformation

403 into the center of the shear zone.

404 This is in agreement with the type II shear zone growth-model proposed by Fossen (2016). Here deformation localizes 405 in the central part because of strain softening. Thus, the margins become inactive and preserve the features acquired 406 during the early stages of shearing while the active part gets progressively thinner (Fig. 15) and records the final stages 407 of shearing. The result is a shear zone with a deformation gradient toward the central part along which, as in the case of 408 the FMSZ, it is possible to observe variations in temperature, age of deformation and deformation regime (Fig. 13d). 409 The occurrence of allanite ± apatite rims around monazites in mylonites is indicative of the breakdown of monazite 410 during retrograde metamorphism. Since allanite and apatite grew asymmetrically in extensional sites around the 411 monazite crystals, in agreement with the orientation of the instantaneous stretching axis during dextral shearing, it is 412 possible to state that this reaction took place synkinematically during retrograde metamorphism (Gibson et al., 2004) 413 and testify the further decrease in temperature during a late stage of shearing.

414 In the Alpine External Crystalline Massifs Variscan shear deformation has never been characterized in terms of 415 vorticity of the flow, and its age, considered to be between ~320 Ma and ~300 Ma (Bussy and Von Raumer, 1993; 416 Genier et al., 2008; Guillot et al., 2009) or Late Visean (Musumeci and Colombo, 2002), was always inferred relying on 417 the age of emplacement of sin-kynematic granites. Ages of ~340 Ma obtained in this work are in agreement with the 418 age proposed by Musumeci and Colombo (2002). Ages of ~320 Ma are consistent with both magmatic ages present in 419 literature and ages of the dextral transpressional deformation in the Maures-Tanneron Massif (Schneider et al., 2014) 420 and Corsica-Sardinia (Carosi and Palmeri, 2002; Di Vincenzo et al., 2004; Iacopini et al., 2008), dated at ~320-310 Ma 421 (U-Th-Pb on monazites, Carosi et al., 2012), that affects migmatitic crust with strong tectono-metamorphic similarities 422 with the migmatites cropping out in the Argentera-Mercantour Massif. If we restore the counterclockwise Oligo-423 Miocenic rotation of the Corsica-Sardinia block (Vigliotti and Kent, 1990; Todesco and Vigliotti, 1993, Advokaat et al., 424 2014) and of the Western Alps (Thomas et al., 1999; Collombet et al., 2002; Maffione et al., 2008), the three studied 425 sectors lie in lateral continuity and show a similar structural and metamorphic evolution. Shear deformation recognized 426 in northern Sardinia has been attributed to a network of dextral shear zones belonging to the East Variscan Shear Zone 427 (EVSZ; Corsini and Rolland, 2009; Carosi et al., 2012; Schneider et al., 2014). The FMSZ can be considered as a 428 branch of EVSZ. In the Argentera-Mercantour Massif dextral shear deformation started at ~340 Ma under amphibolite-429 facies conditions. In the Maures-Tanneron Massif an anatectic post-collisional event is reported at ~340 Ma (Schneider 430 et al., 2014; Oliot et al., 2015) while in northern Sardinia collision-related MP - MT metamorphism has been recognized 431 at 344 ± 7 Ma (Rb-Sr method; Ferrara et al., 1978), although older ages of the thickening stage also occur in Sardinia 432 and Corsica (between 380 and 340 Ma; Oliot et al., 2015). Although the different sectors of the southern European 433 Variscan Belt record a similar tectono-metamorphic evolution (Corsini and Rolland, 2009), the age of shearing along 434 the EVSZ is diachronous, and this is the first time that an age of ~340 Ma is obtained for amphibolite-facies mylonitic

435 deformation linked to this regional-scale system of shear zones. The different sectors of the belt are affected by orogen

436 parallel dextral transpression (Carosi and Palmeri, 2002) at different time because of the progressive change of

437 orientation of the belt with respect to the regional stress field (Carosi et al., 2012). This is also confirmed by the

438 increase of the simple shear component of the deformation recognized both in Sardinia (Carosi and Palmeri, 2002), and,

439 for the first time, in the Argentera-Mercantour Massif. Our data are therefore in agreement with the occurrence of a

440 shear belt in the eastern sector of the Variscan Belt (Fig. 16).

441 Our new structural data support the model proposed by Rollet et al. (2002), Rosenbaum et al. (2002) and Advokaat et

442 al. (2014) of the Corsica-Sardinia Block connected to southern France and in continuity with the Western Alps since all

these sectors of the Variscan Belt show structural and lithological similarities, in contrast to reconstructions of the

Variscan framework which show Corsica and/or Sardinia in a more westerly position connected to Spain (e.g., Fig. 8 in
von Raumer et al., 2013).

446 In the western sector of the Iberian Massif, syn-anatectic sinistral transpressive deformation is recognized and dated at 447 ~340 Ma using monazite and zircon (Pereira et al., 2008; 2010), therefore contemporary with the onset of shear 448 deformation in the External Crystalline Massifs. Shearing in this sector of the belt is interpreted to be due to the 449 formation of the Ibero-Armorican Arc. The presence of a contemporary system of shear zones with opposite kinematics 450 in the western and eastern sector of the belt suggests that, as proposed by Matte (1986a,b), Garcia-Navarro and 451 Fernandez (2004) and Pereira et al. (2008), the European Variscan Belt formed and evolved in a manner analogous to 452 the indentation model proposed by Tapponier and Molnar (1977) and Tapponier et al. (1982). Our data are in agreement 453 with a composite orocline model already proposed by Matte (1986a,b, 2001), Corsini and Rolland (2009), Guillot et al. 454 (2009) and Carosi et al. (2012) with a western arc, still preserved, and an eastern arc that is truncated by the EVSZ. 455 Our data support that the FMSZ is not a minor shear zone confined to one tectonic unit, but it is part of a large-scale 456 system of shear zones that, because of its long-lasting activity, length and thickness, may have played a primary role in 457 the exhumation of migmatitic rocks during Variscan time. Goscombe and Gray (2009) demonstrate that maximum 458 stretching directions do not necessarily correlate with the flow vectors experienced during orogenesis. Although the 459 stretching lineation in the FMSZ is shallowly inclined, the material flow may have had a large vertical component. 460

461

462 **7.** Conclusive remarks

463 The data obtained in this work allow us to establish that:

- 464 the FMSZ is a dextral transpressional shear zone characterized by a non-coaxial flow with a component of pure shear
 465 and a component of simple shear, the latter increasing toward the central and most strained part of the shear zone;
- + the FMSZ initiated during high-temperature amphibolite facies metamorphism and evidence of it is preserved in
 protomylonites and mylonites;
- Alpine deformation in the Argentera-Mercantour Massif occurred as local reactivation of Variscan structure and as
 localized reverse top-to-the-S shear zones, developed under greenschist facies conditions, cross-cutting the FMSZ
- 470 mylonites;
- 471 shear deformation started at ~ 340 Ma, near the metamorphic peak recorded in the GSV complex or shortly after it,
- 472 and progressively migrated toward the core of the shear zone in a time span of ~10-20 Ma. In the same time interval a
- 473 gradual change in the deformation regime is observed;
- 474 microstructural and geochronological analysis constrain a type II evolution for this shear zone that can be a good
- 475 example of strain softening within a regional-scale shear zone.
- 476 the FMSZ is a sector of a system of regional-scale dextral shear zones occurring in the eastern sector of the Variscan
 477 Belt (EVSZ) at 320-310 Ma;
- 478 shear deformation did not begin synchronously in all the part of the EVSZ but initiated earlier in its NE part and then
 479 propagated to the SW.
- 480

481 Software

- 482 The data processing of the geochronological study was performed using the software GLITTER® (Macquarie Research
- 483 Ltd 2001; Van Achterbergh et alii, 2001). The finite strain analysis was performed using the software EllipseFit 3.2 by
- 484 Vollmer (2015). Calculation of the θ angle was performed using the software Strain Calculator 3.2 by Holcombe

485 (2009).

486

487 Acknowledgements

- 488 Research supported by funds from Torino University (Ricerca Locale 2014, 2015), Pisa University (PRA 2016) and
- 489 PRIN 2016 (prot. 2015EC9PJ5_004; resp. R Carosi and C. Montomoli). We thank Michel Corsini and an anonymous
- 490 reviewer for their careful revision that significantly improved the quality of the manuscript.

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- 797

798 Figure captions

- **Figure 1.** A) Distribution of the Variscan units in Europe at the present day. S = Sardinia; C = Corsica; MTM =
- 800 Maures-Tanneron Massif; MC = Massif Central; IM = Iberian Massif; AM = Armorican Massif; RH= Rheno-
- 801 Hercynian; BM = Bohemian Massif; V = Vosges. Blue circle indicates the location of the Argentera Massif (modified

- 802 after Compagnoni et al., 2010); B) Geological sketch map of the Argentera-Mercantour Massif. FMSZ: Ferriere-
- 803 Mollières shear zone; FCSZ: Fremamorta-Colle del Sabbione shear zone; BF: Bersezio fault; VLS: Valle Stura
- 804 Leucogranite; ACG: Argentera Central Granite (modified after Compagnoni et al., 2010).

Figure 2. Geological map of the northwestern portion of the Ferriere-Mollières shear zone (modified after Carosi et al.,

- 806 2016b). Rectangles indicate sampling areas: transect A; transect B; transect C. Details of the selected areas are shown in
- figure 6 and cross sections are shown in figure 14. Equal angle, lower hemisphere projections are also shown: 1)
- 808 Mineral lineation (red), poles to mylonitic foliation (black) and poles to ultramylonitic foliation (yellow triangles) in
- mylonitic schist with biotite and white mica. 2) Mineral lineation (red) and poles to mylonitic foliation (black) in
 mylonitic gneiss with biotite and white mica. 3) Axial plane poles (blue) and axes (red) of the post-shearing folds. 4)
- 811 Lineation (red) and mylonitic foliation poles (black) in the alpine shear zones.
- Figure 3. A) protomylonites at the outcrop-scale; B) anastomosing disjunctive cleavage in protomylonites. It is possible
 to recognize K-feldspar, plagioclase and quartz in a medium grained biotite and sillimanite matrix with some white
 mica (plane-polarized light); C) mylonites at the outcrop-scale; D) disjunctive cleavage with sub-parallel cleavage
 domains in mylonites. Quartz, K-feldspar and plagioclase porphyroclasts in a fine grained biotite and white mica matrix
 are recognizable (plane-polarized light); E) ultramylonites at the outcrop-scale; F) continuous cleavage marked by white
 mica and chlorite in ultramylonites (plane-polarized light).
- Figure 4. Micrographs (crossed polars) of the kinematic indicators showing a top-to-the-SE sense of shear in the
 FMSZ. The short and the long borders of each micrograph are parallel to the Z and X axes of the strain ellipsoid,
 respectively. A) white mica fish and S-C' fabric; B) K-feldspar σ-type porphyroclast (yellow), C' plane (green) and
 oblique foliation (orange) in quartz level; C) S-C' fabric. D) Subgrain rotation and recrystallization in quartz (crossed
 polars); E) Undulose extinction in K-feldspar porphyroclasts (crossed polars).
- Figure 5. Low-angle Alpine shear zones. A) S-C fabric at the outcrop scale; B) S-C and S-C' fabric in thin section
 (plane-polarized light); C) Deformed foliation with biotite relicts (plane-polarized light). The short and the long borders
 of each micrograph are parallel to the Z and X axes of the strain ellipsoid, respectively. All kinematic indicators show a
 reverse top-to-the-S sense of shear.
- Figure 6. Detailed geological maps of the three studied transects along which samples were collected: A) transect A; B)
 transect B; C) transect C. Samples used for kinematic vorticity and strain analysis are indicated with red stars, whereas
 samples used for the geochronological study are indicated with yellow stars. Legend as in Figure 1.
- 831 Figure 7. A) To the right, micrograph (crossed polars) showing C' planes in mylonite of the FMSZ (the short and the 832 long borders of the micrograph are parallel to Z and X axes of the finite strain ellipsoid, respectively). To the left, 833 schematic diagram of possible microstructures developed under general flow conditions in a mylonite with dextral 834 sense of shear and their appearance in thin section. The flow apophyses A1 and A2 are indicated in red and green, 835 respectively. The angle 2v between the two apophyses (yellow) is bisected by the C' shear plane (dotted line). The 836 equation to calculate Wk is also shown (modified after Kurz and Northrup, 2008). B) Relation to calculate the 837 shortening value S perpendicular to the flow plane on the basis of the values of both the mean kinematic vorticity 838 number Wm, that can be assumed equal to Wk, and the axial ratio of the strain ellipse measured on the XZ section 839 (modified after Law et al., 2004).
- Figure 8. Polar histograms used to derive the angle v and to calculate kinematic vorticity with the S-C' method (Kurz and Northrup, 2008) on field oriented samples. A1 = flow apophysis 1; A2 = flow apophysis 2. Dashed line represent the bisector of the angle between A1 and A2, green bars represent the number of data. All histograms are shown with the same sense of shear (red arrows).
- 844

- Figure 9. Fry graphs used in the strain analysis obtained with the center-to-center method on the XZ sections; n =
 number of centers used for the Fry analysis. A) ARG 143; B) ARG 144; C) ARG 21a; D) ARG 68; E) ARG 117; F)
 ARG 19; G) ARG 21; H) ARG 27; I) ARG 29. Axial ratios Rxz are reported in table 1.
- 848
- Figure 10. Back-scattered electron pictures of monazite grains in different microstructural position: A) Mnz 59, sample
 ARG 48b, included in garnet; B) Mnz 29, sample ARG 29, included in a white mica fish; C) Monazite with asymmetric

shape along the main foliation marked by biotite in protomylonte (Mnz 14, sample ARG19); D) Monazite along themain foliation in mylonite (Mnz 52, sample ARG 27).

Figure 11. Back-scattered electron images of the analyzed monazites and related compositional maps of Y made by
electron microprobe. Red dots represent the spots of the quantitative chemical analyses (data are reported in Online
Resource 1).

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Figure 12. Distribution of ²⁰⁶Pb/²³⁸U (A) and ²⁰⁸Pb/²³²Th (B) ages for different samples along the deformation
gradient. Black vertical bars indicate concordant ages, black horizontal bar contains the averaged data (MSWD=Mean
Square Weighted Deviates; prob.= probability). Geochronological data define a broad trend (grey area) reflecting the
deformation degree.

861

862 Figure 13. A) Percentage of pure shear (PS) and simple shear (SS) in relation to the calculated maximum and minimum 863 Wk values; B) simplified sketch showing orientation of the instantaneous flow elements and their angular relationships 864 in a dextral shear zone with simultaneous pure shear and simple shear, θ is the angle between ISA_{max} and the shear zone 865 boundary (modified after Xypolias, 2010); C) diagram showing relationship between the orientation of the maximum 866 Istantaneous Stretching Axis (ISA_{max}) with respect to the shear zone boundary (angle θ) related to the kinematic 867 vorticity number Wk (modified after Fossen and Tikoff, 1993; Fossen et al., 1994) Wk value for which simple shear = 868 pure shear is 0,71 according to Law et al., 2004 and Xypolias, 2010. The distribution of the samples shows a variation 869 from a pure shear dominated transpression to a simple shear dominated transpression linked to the increase of the 870 vorticity number. Orange dashed line represents the theoretical trend of the angle θ . It is possible to observe that the distribution of the values of angle θ of the studied samples is in good agreement with the theoretical curve; D) variation 871 872 of the vorticity number in relation with the distance of the sample from the center of the shear zone. The distribution of 873 points shows increasing Wk values (i.e. increasing of simple shear component of deformation) toward the central part 874 of the shear zone (trend: blue dashed line).

Figure 14. Parallel cross sections along the deformation gradient (modified after Carosi et al., 2016b). The position of
 the analyzed samples and the relative ²⁰⁶Pb/²³⁸U (red) and ²⁰⁸Pb/²³²Th (blue) ages are reported. Traces of the geological
 cross sections and the legend are reported in figure 2.

Figure 15. Evolution of the Ferriere-Molliéres Shear Zone according to a type II growth model (Fossen, 2010) and
relationship between deformation gradient and age of shear deformation: the margins become progressively inactive
and preserve the features acquired during the older stages of shearing while the active part gets progressively thinner
and records the younger stages of shear zone activity.

Figure 16. Reconstruction of the Variscan Belt during Permian (modified after Guillot and Ménot, 2009; Carosi et al.,
2012). Age of shearing in the eastern sector of the Variscan Belt are reported. The East Variscan Shear Zone (EVSZ)
affects the Argentera-Mercantour Massif (AR) at ~ 340 Ma and then propagated southward affecting the Maures-

- 885 Tanneron Massif (MTM) and Corsica and Sardinia (SA) at ~ 320 Ma (Carosi et al., 2012). BM = Bohemian Massif.
- **Table 1.** Kinematic vorticity numbers obtained by the S-C' method and strain data of samples from the three transects. N = number of C' plane; v *max* = mesured maximum angle between C' plane and S plane; 2v = calculated angle between apophysis A1 and apophysis A2; Wk = vorticity number; Angle θ = angle between the ISA_{max} and the shear zone boundary; n = number of centers used for the Fry analysis. Shortening and Stretching values, deformation and position of the sample are also reported.
- **Table 2.** Standards analyzed for the geochronological study at the beginning and at the end of each run.
- Table 3. Geochronological data obtained from the monazites of the samples ARG 19, ARG 21, ARG 27, ARG 29 andARG 48b.
- 895 Online Resource 1. Chemical quantitative analyses of the monazites of the sample ARG 19, ARG 21, ARG 27, ARG896 29 and ARG 48b.
- 897











Fig.4









928 Fig.6





























Fig.8







Fig.10



938 Fig.11



940 Fig.12











948 Fig.16

SAMPLE	N	v max	2ν	Wk	Angle θ	n	Rxz	Shortening (S)	Stretching (1/S)	Deformation	Transect	
ARG 144	42	36°	72°	0,31	9°	67	1,88	0,74	1,35	Protomylonite	Α	
ARG 143	29	36°	72°	0,31	9°	56	1,99	0,72	1,38	Protomylonite	Α	
ARG45	20	33°	66°	0,41	12°	-	-	-	-	Mylonite	A	
ARG 21a	27	27°	54°	0,59	18°	45	2,1	0,74	1,35	Mylonite	A	
ARG 35	8	13°	26°	0,89	31°	-	-	-	-	Ultramylonite	A	
ARG 101	7	17°	34°	0,83	28°	-	-	-	-	Ultramylonite	В	
ARG 37	15	18°	36°	0,81	27°	-	-	-	-	Ultramylonite	В	
ARG RCV	9	17°	34°	0,83	28°	-	-	-	-	Ultramylonite	В	
ARG 68	18	31°	62°	0,47	14°	35	1,87	0,76	1,31	Mylonite	В	
ARG 116	31	33°	66°	0,41	12°	-	-	-	-	Mylonite	В	
ARG 117	25	34°	68°	0,37	11°	61	2,05	0,71	1,39	Mylonite	В	
ARG 115	20	32°	64°	0,44	13°	-	-	-	-	Mylonite	В	
ARG 19	16	35°	70°	0,34	10°	113	1,71	0,78	1,28	Protomylonite	C	
ARG 21	36	26°	52°	0,61	19°	99	2,85	0,63	1,59	Mylonite	C	
ARG 27	14	25°	50°	0,64	20°	99	1,75	0,81	1,23	Mylonite	C	
ARG 48B	17	23°	46°	0,69	21°	-	-	-	-	Mylonite	С	
ARG 29	13	24°	48°	0,66	20°	148	2,46	0,72	1,38	Mylonite	C	
ARG 41	11	15°	30°	0,86	30°	-	-	-	-	Ultramylonite	С	

- 950 Tab.1

	²⁰⁷ Pb/ ²⁰⁶ Pb			²⁰⁷ Pb/ ²³⁵ U			²⁰⁶ Pb/ ²³⁸ U			²⁰⁸ Pb/ ²³² Th			Ages							
Identifier Comments	Ratio	1s %	s % (Prop	Ratio	1s %	s % (Prop	Ratio	1s %	s % (Prop	Ratio	1s %	s % (Prop	²⁰⁷ Pb/ ²⁰⁶ Pb	1s abs	²⁰⁷ Pb/ ²³⁵ U	1s abs	²⁰⁶ Pb/ ²³⁸ U	1s abs	208Pb/232Th	1s abs
Ju08a001 begin	0,05630	0,06%	0,10%	0,64227	0,77%	1,14%	0,08274	0,10%	0,11%	0,02538	0,04%	0,04%	464	8	504	9	512	7	507	9
Ju08a003 begin	0,05708	0,06%	0,10%	0,65076	0,80%	1,18%	0,08270	0,11%	0,11%	0,02535	0,04%	0,04%	495	9	509	9	512	7	506	9
Ju08a004 begin	0,05629	0,06%	0,10%	0,63652	0,78%	1,15%	0,08203	0,10%	0,11%	0,02501	0,04%	0,04%	464	8	500	9	508	7	499	9
Ju08a015 end	0,05790	0,06%	0,10%	0,65608	0,86%	1,22%	0,08220	0,11%	0,12%	0,02522	0,04%	0,04%	526	9	512	10	509	7	503	9
Ju08a016 end	0,05609	0,06%	0,10%	0,64304	0,85%	1,20%	0,08316	0,11%	0,12%	0,02493	0,04%	0,04%	456	8	504	9	515	7	498	9
Ju08a017 end	0,05580	0,06%	0,10%	0,63342	0,84%	1,19%	0,08234	0,11%	0,12%	0,02561	0,04%	0,04%	444	8	498	9	510	7	511	9
Ju08b001 begin	0,05758	0,06%	0,14%	0,65908	0,87%	1,60%	0,08304	0,11%	0,14%	0,02527	0,04%	0,04%	514	12	514	12	514	9	504	9
Ju08b002 begin	0,05519	0,06%	0,13%	0,63537	0,85%	1,54%	0,08352	0,11%	0,14%	0,02553	0,04%	0,04%	420	10	499	12	517	9	510	9
Ju08b003 begin	0,05652	0,06%	0,14%	0,63603	0,85%	1,54%	0,08165	0,11%	0,14%	0,02510	0,04%	0,04%	473	11	500	12	506	9	501	8
Ju08b004 begin	0,05674	0,07%	0,14%	0,64005	0,86%	1,56%	0,08185	0,11%	0,14%	0,02507	0,04%	0,04%	481	12	502	12	507	9	500	8
Ju08b013 end	0,05737	0,07%	0,14%	0,64299	0,91%	1,59%	0,08135	0,11%	0,14%	0,02499	0,03%	0,04%	506	12	504	12	504	9	499	8
Ju08b014 end	0,05797	0,07%	0,14%	0,66486	0,95%	1,65%	0,08325	0,11%	0,15%	0,02562	0,04%	0,04%	529	13	518	13	515	9	511	9
Ju08b015 end	0,05482	0,07%	0,13%	0,62971	0,91%	1,57%	0,08337	0,11%	0,15%	0,02516	0,03%	0,04%	405	10	496	12	516	9	502	8
Ju08c002 begin	0,05705	0,06%	0,11%	0,64522	0,85%	1,45%	0,08209	0,11%	0,15%	0,02516	0,03%	0,04%	493	9	506	11	509	9	502	8
Ju08c003 begin	0,05590	0,06%	0,10%	0,63972	0,84%	1,43%	0,08303	0,11%	0,15%	0,02523	0,03%	0,04%	448	8	502	11	514	9	504	8
Ju08c004 begin	0,05766	0,07%	0,11%	0,65281	0,89%	1,48%	0,08217	0,11%	0,15%	0,02544	0,04%	0,04%	517	10	510	12	509	9	508	8
Ju08c013 end	0,05615	0,07%	0,11%	0,62536	0,88%	1,43%	0,08075	0,11%	0,15%	0,02484	0,03%	0,04%	458	9	493	11	501	9	496	8
Ju08c014 end	0,05742	0,07%	0,11%	0,65902	0,95%	1,53%	0,08325	0,11%	0,15%	0,02537	0,03%	0,04%	508	10	514	12	515	9	506	8
Ju08c015 end	0,05570	0,07%	0,11%	0,64087	0,93%	1,49%	0,08345	0,11%	0,15%	0,02541	0,04%	0,04%	440	9	503	12	517	9	507	8
No23a00 begin	0,05641	0,06%	0,07%	0,64883	0,85%	1,22%	0,08347	0,11%	0,14%	0,02531	0,00032	0,00037	469	6	508	10	517	9	505	7
No23a00:begin	0,05675	0,06%	0,07%	0,63471	0,85%	1,21%	0,08113	0,11%	0,14%	0,02501	0,00032	0,00037	482	6	499	10	503	9	499	7
No23a00 begin	0,05671	0,06%	0,07%	0,65188	0,87%	1,24%	0,08336	0,12%	0,15%	0,02557	0,00032	0,00037	480	6	510	10	516	9	510	7
No23a00!begin	0,0563	0,06%	0,07%	0,63964	0,86%	1,22%	0,08241	0,12%	0,15%	0,02515	0,00032	0,00037	464	6	502	10	510	9	502	7
No23a02 end	0,05692	0,06%	0,07%	0,65378	0,93%	1,29%	0,0833	0,12%	0,15%	0,02544	0,00032	0,00037	488	6	511	10	516	9	508	7
No23a02 end	0,057	0,06%	0,07%	0,65193	0,91%	1,27%	0,08295	0,12%	0,15%	0,02515	0,00032	0,00037	492	6	510	10	514	9	502	7
No23a02-end	0,05599	0,06%	0,07%	0,6331	0,89%	1,24%	0,082	0,12%	0,15%	0,02532	0,00032	0,00037	452	6	498	10	508	9	505	7
No23b00 begin	0,05627	0,06%	0,07%	0,63949	0,93%	1,03%	0,08245	0,12%	0,12%	0,02519	0,00032	0,00039	463	6	502	8	511	8	503	8
No23b00 begin	0,05716	0,06%	0,07%	0,64833	0,91%	1,02%	0,08229	0,12%	0,12%	0,0256	0,00033	0,0004	498	6	507	8	510	8	511	8
No23b00 begin	0,05618	0,06%	0,07%	0,64229	0,91%	1,02%	0,08294	0,12%	0,12%	0,02499	0,00032	0,00039	459	6	504	8	514	8	499	8
No23b00 begin	0,05676	0,06%	0,07%	0,64873	0,89%	1,01%	0,08291	0,12%	0,12%	0,02554	0,00032	0,00039	482	6	508	8	513	8	510	8
No23b02 end	0,05627	0,07%	0,08%	0,63878	0,95%	1,06%	0,08233	0,12%	0,13%	0,02532	0,00031	0,00038	463	6	502	8	510	8	505	8
No23b02 end	0,05664	0,06%	0,07%	0,64804	0,93%	1,04%	0,08297	0,12%	0,13%	0,02527	0,00031	0,00038	478	6	507	8	514	8	504	8

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Tab.2

 Data for Wetherill plot²

 1 S abs
 ²⁰²Pb/²⁰²U
 1 S abs
 ²⁰²Pb/²⁰²U
 1 S abs
 ²⁰²Pb/²⁰²U
 1 S abs

 0,00056
 0,37949
 0,00711
 0,05446
 0,00095
 0,9

 0,00058
 0,35984
 0,00057
 0,05349
 0,00097
 1,0

 0,00056
 0,35984
 0,00057
 0,05349
 0,00095
 0,9

 0,00056
 0,35984
 0,00057
 0,05349
 0,00095
 0,9

 0,00056
 0,35947
 0,00544
 0,00395
 1,0

 0,00056
 0,39997
 0,00750
 0,05426
 0,00097
 1,0

 0,00056
 0,0952
 0,07270
 0,05427
 0,00097
 1,0
 Ages³ bs 206Pb/²³⁸U 1s abs 208Pb/²³²Th 1s abs - 230 5 -Pb Concordant age
 207 Pb/208 Pb
 1 s
 abs
 207 Pb/205 U

 0,05060
 0,00060
 0,37949
 0,04850
 0,00058
 0,35984

 0,04718
 0,00054
 0,34671
 0,04671
 0,04671

 Sample
 Identifier
 Mn2#
 ipot locatio

 ARG 19
 No23a00
 Mn214
 core

 No23a00
 Mn214
 core

 No23a00
 Mn214
 rim

 No23a00
 Mn214
 rim

 No23a00
 Mn214
 rim
 A 208Pb/²³²Th 1s abs ²⁰⁷Pb/²⁰⁶Pb 1s abs ²⁰⁷Pb/²³⁵U 1s abs -4,6 **3,9** -8,3 **2,1** -11,1 **1,5** Zoning low-Y int-Y 2s abs 0,01639 0,00024 0,01651 0,00024 0,01672 0,00025 223 124 58 59 327 312 329 331 342 338 340 335 3,9 2,1 1,5 4,2 5,8 high-' 306 302 342 342 335 321 -10,9 -0,7 0,4 0,00023 high-\ 0,01602 rim 11 10 No23a01(Mnz14 high-Y 0,05304 0,01616 0,00023 331 344 324 341 core core core core core 344 No23a01 Mnz14 low-0,05362 0,00060 0,40053 0,00737 0,05427 0,00097 0,01596 0,00023 355 341 320 6,1 8,2 8,7 11,5 12,5 3,2 1,0 1,0 0,9 0,9
 No23a01
 Mnz14

 No23a01
 Mnz6

 No23a01
 Mnz3

 No23a01
 Mnz3

 No23a01
 Mnz3

 No23a01
 Mnz8

 No23a01
 Mnz8

 0.05362
 0.00060
 0.40053
 0.00737
 0.05427
 0.00647

 0.04672
 0.00052
 0.34731
 0.00641
 0.05405
 0.00097

 0.05366
 0.00075
 0.40986
 0.00853
 0.05523
 0.0011

 0.05191
 0.00073
 0.39204
 0.00837
 0.05546
 0.0010

 0.05102
 0.00073
 0.39204
 0.00837
 0.05544
 0.01015

 0.06046
 0.00089
 0.43421
 0.00937
 0.55249
 0.00102
 low-Y low-Y low-Y low-Y high-Y -12,1 0,6 -4,2 -6,1 10,6 35 357 237 0,01554 0,00022 303 349 334 336 366 339 347 348 356 327 312 0,01554 0,00022 0,01578 0,00024 0,01533 0,00023 0,01555 0,00023 0,01580 0,00024 316 308 348 12 0,9 0,9 201 620 312 317 rim 9 6 ARG 29 No23b00 high-Y low-Y low-Y
 0,04671
 0,00064
 0,33727
 0,00559
 0,05260
 0,00080

 0,04793
 0,00067
 0,35617
 0,00602
 0,05387
 0,00083
 0,01725 0,00028 0,01637 0,00027 295 309 316 330 346 -12,0 -4,6 Mnz5 core 0,9 34 96 54 14 23 2708 121 0 No23b00 Mnz5 Mnz2 rim/core 0,9 338 328 -9,3 -11,4 3,0 3,2 1,1 3,2 No23b00 0,04709 0,00054 0,36473 0,00547 0,05610 0,00085 0,01699 0,00027 352 341 core rim core core core rim 1,0 1,0 1,0 1,0 1,0 1,0 No23b00 Mnz2 nigh/int 0,04631 0,00056 0,33850 0,00520 0,05301 0,00081 0,01643 0,00026 -12,5 -12,4 60,3 -7,9 -6,5 1,0 296 303 942 305 311 323 333 341 374 329 331 329 330 303 333 330
 No23b00
 Mnz2

 No23b01
 Mnz9

 No23b01
 Mnz9

 No23b01
 Mnz11

 No23b01
 Mnz11

 No23b01
 Mnz11

 No23b01
 Mnz11

 No23b01
 Mnz11

 No23b01
 Mnz11
 0,00056 0,33850 0,00061 0,34793 0,00259 1,52869 0,00058 0,34997 0,00058 0,35799 0,00064 0,37450 high-Y low-Y low-Y low-Y 0,00026 0,00025 0,00025 0,00026 0.04648 0.00565 0.05427 0.0008 0.01645 0,04848 0,18607 0,04845 0,04926
 0,00585
 0,05427
 0,00085

 0,02448
 0,05974
 0,00093

 0,00538
 0,05234
 0,00080

 0,00544
 0,05268
 0,00081

 0,00578
 0,05085
 0,00078
 0,01645 0,01511 0,01663 0,01646 19,0 -1,4 0,3 -2,0 -4,1 -1,5 -3,1 11,6 160 345 0,05338 0,01626 0,00026 high-Y 320 326 core rim high-Y low-Y 0,04698 0,00072 0,33071 0,00590 0,05104 0,00080 0,9 0,01667 0,00027 48 290 301 290 365 321 334 -10,6 148 43 540 322 322 336 -7,1 -11,0 7,9 No23b01 Mnz20 0,04901 0,00074 0,34441 0,00616 0,05121 0,00082 0,9 0,01630 0,00026 327 core core high-\ low-Y No23b01; Mnz24 0.04687 0.00070 0.33064 0.00584 0.05121 0.00079 0,9 0.01655 0.00026 332 No23b01 Mnz24 0,05828 0,00114 0,43250 0,00934 0,05354 0,00088 0,8 0,01482 0,00023 297 11 ARG 48b Ju08c006 Mnz58 Ju08c007 Mnz59 Ju08c008 Mnz57 Ju08c009 Mnz56 Ju08c010 Mnz56 igh-Y/low 0.05115 0.00092 0.35022 0.00773 High-Y 0.05339 0.00112 0.54547 0.01353 igh-Y/low 0.035361 0.00100 0.39211 0.00893 Ledium+ 0.05584 0.00106 0.39324 0.00900 Low-Y 0.05269 0.00115 0.37788 0.00964 Core Core Core Core 0,01556 0,00025 0,02130 0,00035 **0,01636 0,00026** 0,01597 0,00026 248 345 312 461 333 0,04967 0,00090 312 0,1 0,8 4 -2,5 -4,4 **0,8** 4,7 -0,5 305 442 **336** 337 0,07418 0,00138 0,05303 0,00096 11 426 328 320 12 334 **0,8** 0,8 **355** 446 8 5 **1,5** 0,2 0,05105 0,00093 321 327 12 327 0,05204 0,00098 0,7 0,01557 0,00026 315 325 312 4,5 Ju08c011 Mnz55 Rim Medium-1 0,05513 0,00129 0,37930 0,01018 0,04995 0,00094 0,7 0,01562 0,00026 417 10 2 327 297 314 320 313 3,8 -7,7 0,3 -0,2 6 Low-Y 0,04840 0,00093 0,33991 0,00792 0,05091 0,00093 Ju08c012 Mnz55 Core 0,8 0,01599 0,00026 119 321 5 Core Core Core Core Core 321 287 317 318 48G21 Ju08a006 Mpz55 0.05031 0.00085 0.35590 0.00626 0.05132 0.00070 0.01602 0.00028 209 4 309 4 High-Y 0.8 323 297 318 325 6 -4,4 -0,6 -4,7 -7,2 -7,0 -2,4 0,4 3,3 0,3 2,1 1,1 Ju08a006 Mn255 Ju08a007 Mn230 Ju08a008 Mn226 Ju08a009 Mn27 0,05031 0,05196 0,05001 0,04879 0,00085 0,35590 0,00087 0,33730 0,00083 0,34893 0,00081 0,34771 0,05132 0,00070 0,04713 0,00064 0,05062 0,00069 0,05170 0,00071 0,01602 0,01430 0,01583 0,01587 0,00028 0,00025 0,00028 0,00028 High-Y High-Y High-Y 0,8 0,8 0,8 0,8 203 284 195 138 295 304 303 297 8 0.00594 0,00334 0,00612 0,00610 Ju08a010 Mnz7 High-Y 0,04893 0,00082 0,34800 0,00614 0,05161 0,00071 0,8 0,01600 0,00028 144 303 324 321 Ju08a011 Mnz16 Core Medium 0.05155 0.00087 0.36991 0.00661 0.05206 0.00072 0.8 0.01599 0.00028 266 320 327 321 2.0 Rim Rim Core High-Y High-Y High-Y -5,3 -7,5 -0,2 Ju08a012 Mnz16 0.04985 0.00085 0.35610 0.00638 0.05184 0.00072 0,8 0.01591 0.00028 188 309 326 319 2.1 Ju08a013 Mnz16 0.04849 0.00081 0 33799 0.00598 0.05056 0.00070 0.01573 0.00028 123 296 318 315 308 0,8 1,8 0,04988 4 Ju08a014 Mnz16 0,05254 0,00092 0,36114 0,00672 0,01536 0,00027 313 314 314 9 0,00070 Ju08b006 Mnz56 Ju08b007 Mnz58 High-Y Low-Y High-Y
 0,05027
 0,00124
 0,36285
 0,00912
 0,05220
 0,00093

 0,05897
 0,00139
 0,42181
 0,01010
 0,05192
 0,00090

 0,05226
 0,0123
 0,37062
 0,00886
 0,05143
 0,00092

 0,08657
 0,00204
 0,62877
 0,1503
 0,05273
 0,0092
 328 323 ARG27 Core 0,7 0,01609 0,00028 207 314 6 6 -4,4 8,7 1,6 4,7 13 7 32 566 297 1351 357 320 495 Core 0,7 0,01551 0,00026 326 323 331 311 323 11 Ju08b010 Mnz52 Rim-core 0,7 0,7 0,01550 0,00026 6 6 311 5 5 -1,0 33,1 3,8 8,6 8 12 Ju08b011 mnz52 Rim Low-Y 0,01509 0,00025 303 3 Discordance calculated as (1-(206Pb/238U age/207Pb/235U age))*100 4 Discordance calculated as (1-(208Pb/232Th age/206Pb/238U age))*100

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960 Tab.3