Structural setting of a transpressive shear zone: Insights from geological mapping, quartz petrofabric and kinematic vorticity analysis in NE Sardinia (Italy)

Graziani, R

http://hdl.handle.net/10026.1/18770

10.1017/S0016756820000138

Geological Magazine

All content in PEARL is protected by copyright law. Author manuscripts are made available in accordance with publisher policies. Please cite only the published version using the details provided on the item record or document. In the absence of an open licence (e.g. Creative Commons), permissions for further reuse of content should be sought from the publisher or author.
Structural setting of a transpressive shear zone: insights from geological mapping, quartz petrofabric and kinematic vorticity analysis in NE Sardinia (Italy)

Riccardo Graziani1, Chiara Montomoli2,3,4, Salvatore Iaccarino2, Luca Menegon4,5, Laura Nania6,7 and Rodolfo Carosi2

1Earth and Environmental Sciences, IKBSAS, University of British Columbia Okanagan, 3333 University Way, Kelowna, BC V1V 1V7, Canada; 2Dipartimento di Scienze della Terra, Università di Torino, via Valperga Caluso 35, 10125, Torino, Italy; 3IGG-CNR via Moruzzi 6, Pisa, Italy; 4The Njord Centre, Department of Geosciences, University of Oslo, P.O. Box 1048 Blindern, Norway; 5School of Geography, Earth and Environmental Sciences, Plymouth University, Plymouth PL4 8AA, UK; 6Dottorato Regionale in Scienze della Terra Pegaso, Università di Firenze, via La Pira 4, 50121, Firenze, Italy and 7Dipartimento di Scienze della Terra, Università di Pisa, via Santa Maria, 53, 56126, Pisa, Italy

Abstract

The Posada–Asinara Line is a crustal-scale transpressive shear zone affecting the Variscan basement in northern Sardinia during Late Carboniferous time. We investigated a structural transect of the Posada–Asinara Line (Baronie) with the aid of geological mapping and structural analysis. N-verging F2 isoclinal folds with associated mylonitic foliation (S2) are the main deformation features developed during the Posada–Asinara Line activity (D2). The mineral assemblages and microstructures suggest that the Posada–Asinara Line was affected by a retrograde metamorphic path. This is also confirmed by quartz microstructures, where subgrain rotation recrystallization superimposes on grain boundary migration recrystallization. Crystallographic preferred orientation data, obtained using electron backscatter diffraction, allowed analysis of quartz slip systems and estimation of the deformation temperature, vorticity of flow and rheological parameters (flow stress and strain rate) during the Posada–Asinara Line activity. Quartz deformation temperatures of 400 ± 50 °C have been estimated along a transect perpendicular to the Posada–Asinara Line, in agreement with the syn-kinematic post-metamorphic peak mineral assemblages and the late microstructures of quartz. The D2 phase can be subdivided in two events: an early D2early phase, related to the metamorphic peak and low kinematic vorticity (pure shear dominated), and a late D2late phase characterized by a lower metamorphic grade and an increased kinematic vorticity (simple shear dominated). Palaeopiezometry and strain rate estimates associated with the D2late deformation event showed an intensity gradient increasing towards the core of the shear zone. The D2early deformation developed under peak temperature conditions, while the D2late event was active at shallower structural levels.

1. Introduction

Collisional type orogens are often characterized by the presence of crustal-scale shear zones driving and affecting the tectono-metamorphic evolution of the inner portion of the belts (Fossen & Cavalcante, 2017). Such crustal-scale shear zones can show different kinematics, from normal-sense (e.g. South Tibetan Detachment System; Burchfiel et al. 1992) to thrust-sense (e.g. Main Central Thrust; Searle et al. 2008 and Higher Himalayan Discontinuity; Montomoli et al. 2013, 2015) up to transtension and transpression (Goscombe et al. 2005). Regardless, their long-lasting tectonic history (several Ma) is able to have a deep impact on the P–T–t paths of the metamorphic rocks and their exhumation (Carosi et al. 2018 and references therein).

Transpressive tectonics, at the regional scale, can result from different factors such as an oblique convergence (e.g. Coast Mountains; Depine et al. 2011) or the irregular shape of the continental margins (e.g. Armorican Massif; Gébelin et al. 2009). Regardless, the occurrence of transpression coeval with or subsequent to the continental collision deeply affects the evolution of the orogen with respect to the frontal collisional setting.

During the last 30 years, many theoretical, modelling and fieldwork studies have been carried out in order to characterize transpressional tectonics in complex oblique collisional events (e.g. Sanderson & Marchini, 1984; Tikoff & Fossen, 1993; Tikoff & Teyssier, 1994; Fossen & Tikoff, 1998; Schulmann et al. 2003). Occurrences of transpressional tectonics are often related to collisional belts where transpression represents an evolution of the nappe stacking and crustal...
thickening (Matte et al. 1998; Carosi & Oggiano, 2002; Carosi & Palmeri, 2002; Carosi et al. 2004). Transpressional tectonics also has a profound impact on the metamorphic architecture of an orogenic belt, the latter presenting substantial differences in terms of the tectono-metamorphic evolution compared to perpendicular collisional environments (Thompson et al. 1997; Carosi & Palmeri, 2002; Goscombe et al. 2003, 2005). Syn-collisional transpression has been investigated in first-order regional-scale shear zones within the main European crystalline basements such as the South-Armorican Shear Zone in Brittany (Gébelin et al. 2009); the central sector of the Maures-Tanneron massif in southern France (Corsini & Rolland, 2009; Schneider et al. 2014); the central sector of the Maures-Tanneron massif in southern France (Corsini & Rolland, 2009; Schneider et al. 2014). Syn-collisional transpression has been investigated in first-order regional-scale shear zones within the main European crystalline basements such as the South-Armorican Shear Zone in Brittany (Gébelin et al. 2009); the central sector of the Maures-Tanneron massif in southern France (Corsini & Rolland, 2009; Schneider et al. 2014); the central sector of the Maures-Tanneron massif in southern France (Corsini & Rolland, 2009; Schneider et al. 2014). Syn-collisional transpression has been investigated in first-order regional-scale shear zones within the main European crystalline basements such as the South-Armorican Shear Zone in Brittany (Gébelin et al. 2009); the central sector of the Maures-Tanneron massif in southern France (Corsini & Rolland, 2009; Schneider et al. 2014); the central sector of the Maures-Tanneron massif in southern France (Corsini & Rolland, 2009; Schneider et al. 2014). Syn-collisional transpression has been investigated in first-order regional-scale shear zones within the main European crystalline basements such as the South-Armorican Shear Zone in Brittany (Gébelin et al. 2009); the central sector of the Maures-Tanneron massif in southern France (Corsini & Rolland, 2009; Schneider et al. 2014); the central sector of the Maures-Tanneron massif in southern France (Corsini & Rolland, 2009; Schneider et al. 2014). Syn-collisional transpression has been investigated in first-order regional-scale shear zones within the main European crystalline basements such as the South-Armorican Shear Zone in Brittany (Gébelin et al. 2009); the central sector of the Maures-Tanneron massif in southern France (Corsini & Rolland, 2009; Schneider et al. 2014); the central sector of the Maures-Tanneron massif in southern France (Corsini & Rolland, 2009; Schneider et al. 2014). Syn-collisional transpression has been investigated in first-order regional-scale shear zones within the main European crystalline basements such as the South-Armorican Shear Zone in Brittany (Gébelin et al. 2009); the central sector of the Maures-Tanneron massif in southern France (Corsini & Rolland, 2009; Schneider et al. 2014); the central sector of the Maures-Tanneron massif in southern France (Corsini & Rolland, 2009; Schneider et al. 2014). Syn-collisional transpression has been investigated in first-order regional-scale shear zones within the main European crystalline basements such as the South-Armorican Shear Zone in Brittany (Gébelin et al. 2009); the central sector of the Maures-Tanneron massif in southern France (Corsini & Rolland, 2009; Schneider et al. 2014); the central sector of the Maures-Tanneron massif in southern France (Corsini & Rolland, 2009; Schneider et al. 2014). Syn-collisional transpression has been investigated in first-order regional-scale shear zones within the main European crystalline basements such as the South-Armorican Shear Zone in Brittany (Gébelin et al. 2009); the central sector of the Maures-Tanneron massif in southern France (Corsini & Rolland, 2009; Schneider et al. 2014); the central sector of the Maures-Tanneron massif in southern France (Corsini & Rolland, 2009; Schneider et al. 2014). Syn-collisional transpression has been investigated in first-order regional-scale shear zones within the main European crystalline basements such as the South-Armorican Shear Zone in Brittany (Gébelin et al. 2009); the central sector of the Maures-Tanneron massif in southern France (Corsini & Rolland, 2009; Schneider et al. 2014); the central sector of the Maures-Tanneron massif in southern France (Corsini & Rolland, 2009; Schneider et al. 2014). Syn-collisional transpression has been investigated in first-order regional-scale shear zones within the main European crystalline basements such as the South-Armorican Shear Zone in Brittany (Gébelin et al. 2009); the central sector of the Maures-Tanneron massif in southern France (Corsini & Rolland, 2009; Schneider et al. 2014); the central sector of the Maures-Tanneron massif in southern France (Corsini & Rolland, 2009; Schneider et al. 2014).

In this work we investigated an area in NE Sardinia (Fig. 1) where a portion of the PAL, one of the first-order transpressive shear zones of the Variscan Belt in southern Europe, crops out (Corsini & Rolland, 2009; Carosi et al. 2015; Simonetti et al. 2018). This shear zone has been studied by many authors with respect to the structural settings (e.g. Elter et al. 1990), metamorphic evolution, and dynamics and kinematics of flow (e.g. Carosi & Palmeri, 2002; Carosi et al. 2005; Iacopini et al. 2008; Frassi et al. 2009; Cruciani et al. 2015). These authors investigated the bulk deformation linked to the transpressive monoclinic flow (Iacopini et al. 2008) mainly focusing on the initial conditions of the shearing event at medium T conditions (520–630 °C; Carosi & Palmeri, 2002). The aim of this paper is to fill in the gap concerning the geological history of the long-lasting late transpressive tectonics, constraining the structural evolution and flow kinematics during the latest increments of shear deformation related to the PAL activity.

We present the results of a new structural study on mylonites along the PAL, metamorphosed under medium-grade conditions,
coupled with microstructural analyses including the study of quartz crystallographic preferred orientation (CPO). Quartz petrofabric data allowed us to infer the deformation temperatures and the kinematic vorticity of the flow. Moreover we estimated for the first time the flow stresses and the strain rate along a N–S profile of the PAL, applying published recrystallized grain-size palaeopiezometers for quartz and wet-quartzite flow.

Combining our structural and petrofabric data with the existing published dataset, we propose an updated model of the PAL tectonic evolution and its role in the exhumation/extrusion of metamorphic complexes in orogenic belts.

2. Geological setting
2.a. Overview of the Variscan Belt in Sardinia

The Variscan Belt in Sardinia (Italy) is one of the most complete trans-ects of the Southern Variscan Domain (Matte, 1986; Carmignani et al. 1994, 2001; Corsini & Rolland, 2009; Cruciani et al. 2015) characterized by a general lack of strong Alpine–Apenninic reworking. The current position of Sardinia derives from a 30° anticlockwise rotation of the Corsica–Sardinia microplate, related to the opening of the Balearic basin (Alvarez, 1972; Montgny et al. 1981; Deino et al. 2001; Gattacceca, 2001) in the Oligocene Period. Variscan Sardinia consists of metasedimentary and metaigneous rocks of the northern margin of Gondwana, affected by Palaeozoic deformation from the Devonian to Carboniferous periods, related to the collision between Gondwana, Armorica and Laurussia (Carmignani et al. 1982, 1994, 2001). During the collision, the Gondwanan margin was the lower plate subducting under Armorica and the peri-Gondwanan terranes, accommodating most of the deformation (Matte, 1986). The Variscan Orogeny is responsible for the main deformation and metamorphic features of Sardinia, consisting of a S- to SW-verging stacking of tectonic units with an increasing metamorphic grade moving from the SW to NE (Carmignani et al. 1982, 1994; Franceschelli et al. 1982).

The Sardinian basement represents the eastern portion of the inner Ibero-Armorican indenter (Matte, 1986; Carosi & Palmeri, 2002; Cruciani et al. 2015), and it can be subdivided into three areas characterized by a different tectono-metamorphic evolution (Carmignani et al. 1982, 1994, 2001; Cruciani et al. 2015). Along a S to N structural profile, the three areas are (Fig. 1): (i) The foreland area (External Zone), restricted to SW Sardinia, composed of a poly-deformed Palaeozoic sequence showing very low- to low-grade metamorphism (e.g. Carmignani et al. 1994; Franceschelli et al. 2017 and references therein). (ii) A SW-verging nappe stack in the central area constituted by Palaeozoic metasedimentary and metagneous rocks with low- to medium-grade metamorphism (Carmignani et al. 1994; Carosi & Pertusati, 1990; Carosi et al. 1991; Montomoli et al. 2018). This area is subdivided into two sectors: the External and the Internal Nappe zones. The Internal Nappe Zone has also been subdivided into the ‘Medium-Grade Metamorphic Complex’ and the ‘Low-Grade Metamorphic Complex’ characterized, respectively, by medium- and low-grade metamorphic imprints. (iii) The Inner Zone, or High-Grade Metamorphic Complex, in the northernmost sector of the island, represented by high-grade metamorphic rocks (Franceschelli et al. 1982, 2005; Cruciani et al. 2015).

This architecture is intruded by late- to post-Variscan granitoids (Ghezzo et al. 1979; Macera et al. 1989; Casini et al. 2015).

2.b. Northern Sardinia

Northern Sardinia is divided into areas with different metamorphic signatures (Fig. 1): the Internal Nappe Zone represented by the Low- and Medium-Grade Metamorphic complexes (L–MGMC) and the Inner Zone represented by the High-Grade Metamorphic Complex (HGM) (Carmignani et al. 2015 and references therein). The HGM is located in the northernmost part of the island and consists of migmatic, orthogneisses, amphibolites, minor paragneisses and micaschists, calciludic and Carboniferous granitoids (Carmignani et al. 1994, 2001; Casini et al. 2012). The HGM experienced an early high-pressure (HP) eclogitic stage (700–740 °C and >1.5 GPa; see Cruciani et al. 2015 for a review) followed by a HP granulitic event (660–730 °C, 0.75–0.9 GPa; Miller et al. 1976; Ghezzo et al. 1979; Franceschelli et al. 1982; Di Pisa et al. 1993; Cruciani et al. 2015) as suggested by granulitized relics of eclogite-facies metamabite embedded within the migmatite. Metamorphic ages are available only for the granulitic event that is constrained at ~350 Ma (Giacomini et al. 2005) or younger at ~330 Ma (Palmer et al. 2004). Southwards, the L–MGMC made up of lower Palaeozoic metasedimentary rocks intruded by Ordovician metagranitic bodies, crops out (Carmignani et al. 1994, 2001). The metasedimentary rocks consist of greenschist- to amphibolite-facies micaschists and paragneisses characterized by an increasing Barrovian metamorphic grade from south to north (Franceschelli et al. 1982, 1989; Cruciani et al. 2015). The metamorphic zoning in the L–MGMC comprises, moving northward, a garnet + albitite zone, garnet + oligoclase zone, staurolite + garnet + biotite zone, kyanite zone and sillimanite zone (Franceschelli et al. 1982, 1989; Carmignani et al. 2001; Cruciani et al. 2015). Close to the sheared contact between the two metamorphic complexes, the Barrovian isograds are very tight (Carosi & Palmeri, 2002; Carosi et al. 2005). The boundary between the HGM and L–MGMC is represented by the PAL. The PAL (Fig. 1) is a nearly 150 km long and 10–15 km wide dextral transpressive shear belt crossing the whole of northern Sardinia from Asinara island in the NW to the Posada River Valley in the NE (Cruciani et al. 2015). This structure is variably oriented (Fig. 1) from E–W to NW–SE and steeply dips to the south in the NE part (Carosi & Palmeri, 2002). The PAL was originally interpreted as a strike-slip shear zone active in Late Carboniferous time (Elder et al. 1990), and later as a wide transpressional shear zone (Carosi & Palmeri, 2002; Iacopini et al. 2008; Carosi et al. 2009; Frassi et al. 2009). Cappelli et al. (1992) interpreted the PAL as an oceanic suture zone between Armorica and Gondwana because of the occurrence of amphibolitic boudins aligned along it. The amphibolites show a normal mid-ocean ridge basalt (N-MORB) geochemical signature and locally preserve relics of eclogite facies (Oggiano & Di Pisa, 1992; Cortesogno et al. 2004). Recent geochemical constraints attributed the HGM to the northern margin of Gondwana, confirming the PAL as an intracontinental orogen-parallel shear zone (Giacomini et al. 2006). In situ Ar–Ar dating of white mica (Di Vincenzo et al. 2004) and U–(Th)–Pb dating of monazite (Carosi et al. 2012) constrained the age of the transpressional shearing to between ~300 Ma and 320 Ma. These authors recognized the presence of two main deformation phases associated with the Variscan tectonic evolution: a regional D1 phase associated with the collisional stage recognizable all across the whole of Sardinia, and a later orogen-parallel D2 phase linked to the transpressive activity. Based on pseudosection calculations, a prograde HP metamorphic signature (1.8 GPa and 460–500 °C) has been recognized by Cruciani et al. (2013) for the D1, allowing the linking of...
this phase to prograde metamorphism related to underthrusting and nappe stacking. The age of the D1 is \( \sim 330-340 \text{ Ma} \) (Di Vincenzo et al. 2004; Carosi et al. 2012). The transition between D1 and D2 represents a change in metamorphic conditions from high pressure to lower pressure with increasing temperature during decompression (Carosi & Palmeri, 2002; Cruciani et al. 2013, 2015) followed by decompression and cooling.

2.c. Study area (Baronie Region)

The study area is located within the northern sector of the L–MGMC and comprises part of the garnet + plagioclase zone and the garnet + staurolite + biotite zone, a few kilometres south of the boundary with the HGMG (Fig. 2) where a detailed geological and structural map (at the scale of 1:5000) was compiled.

The structural setting is dominated by N–NE-verging regional-scale isoclinal folds with axial planes striking N20°–90° and dipping 20–80° to the SE (Fig. 2b). In the core of the antiforms an orthogneiss body crops out with a protolith age of 456 ± 14 Ma (Helbing & Tiepolo, 2005). It is composed of granodioritic orthogneiss (Fig. 3a) at the base, surrounded by granitic augen orthogneiss (Fig. 3b). The northern area and the cores of the synforms expose Pre-Cambrian (?) to Cambrian micaschist (Fig. 3c, d) with paragneiss and quartzitic lenses (Carmignani et al. 1994). In the southern area the micaschist is affected by a low- (garnet + albite) to medium- (garnet + elioclase) grade metamorphism. The medium-grade metamorphic conditions in the micaschist, cropping out in the northern area, are evident from the garnet + staurolite + biotite metamorphic assemblage.

3. Deformation history and structural analysis

On the basis of geological mapping, as well as of detailed meso- and microstructural analyses, five ductile deformation phases, followed by brittle tectonics, have been identified. All microstructural observations have been conducted on thin-sections cut perpendicular to the main foliation and parallel to the mineral lineation to better approximate the XZ section of the finite strain ellipsoid.

The first deformation phase is recognizable mainly at the microscale and it is poorly expressed at the mesoscale. It is represented by a relict foliation, S1, within D2 microlithon domains and by internal foliations, Si, within garnet, staurolite and plagioclase intertectonic porphyroblasts. This phase is related to the syn-kinematic growth of white mica, biotite, feldspar and garnet in the micaschist.

In the study area the prominent deformation event is the second deformation phase (D2), associated with the development of isoclinal folds (F2). The F2 folds have similar geometry (Ramsay, 1967); they are non-cylindrical, trending N080° of isoclinal folds (F2). The F2 folds have similar geometry (Ramsay, 1967). Garnet in the micaschist.

To the syn-kinematic growth of white mica, biotite, feldspar and garnet in the micaschist, S2 varies from a continuous to a spaced foliation. In domains of spaced foliation, the S2 is characterized by anastomosing biotite- and white mica-bearing lepidoblastic layers alternated with lenticular microthlinit domains (Fig. 4a, b), where S1 relics are present.

The D2 mylonitic fabric is well recognizable at all scales. Shear sense indicators have been observed only on sections parallel to the XZ plane of the finite strain ellipsoid (i.e. perpendicular to the S2 foliation and parallel to mineral lineation) and they have not been detected on YZ and XY sections.

S-C–C′ structures, mica fishes (mainly of groups 1, 2 and 3 according to the classification of ten Grotenhuis et al. 2003), asymmetric strain shadows around porphyroblasts and oblique foliations point to a dextral sense of shear consistent with the top-to-the-W/NW direction of tectonic transport associated with the PAL (Fig. 4a–c). Elongation of feldspar, garnet and staurolite crystals on the S2 planes form a sub-horizontal L2 mineral/grain and aggregate lineation (average plunge 0–10° to the E).

The M2 metamorphism is related to the syn-kinematic growth of the main mineral assemblages (biotite + white mica + garnet + plagioclase + quartz in the southern sector and biotite + white mica + garnet + staurolite + quartz in the northern sector). A late syn-kinematic growth of chlorite on the late shearing planes (mainly C′ planes and shear bands), in the strain shadows around garnet (Fig. 4d, e) and filling fractures in garnet and staurolite has been recognized. This retrogression is present in the studied transect and highlights a progressive decrease in metamorphic grade during D2, starting from lower amphibolite to greenschist facies. Interlobate to ameboidal grain boundaries in quartz ribbons are consistent with grain boundary migration (GBM) recrystallization (Stipp et al. 2002a, b; Law, 2014).

The presence of finer equigranular quartz grains along the grain boundaries suggests superposition of subgrain rotation (SGR) recrystallization, which heterogeneously reworks and overprints the GBM microstructures in all the studied samples (Fig. 5a–c). Although SGR occurs in all of the study area, in the N–S transect the intensity of SGR varies from incipient in the southern sector (Fig. 5a) to pervasive in the northern sector, where core and mantle structures (Fig. 5b) or nearly complete recrystallization (Fig. 5c) are present.

The D3 phase is characterized by a heterogeneous deformation localized in E–W-trending lenticular domains developed at the hectometric scale. Here an S3 crenulation cleavage (Fig. 3e, f), linked to centimetric- to hectometric-scale F3 similar folds (Ramsay, 1967) occurs (Fig. 4f). F3 axial planes (Ap3) and S3 foliation are parallel to S2 mylonitic foliation. F3 folds have a non-cylindrical geometry, and the plunge of the A3 axes varies from sub-vertical to a few degrees to the east. The parallelism between the S2 and S3 foliation planes, and the scattering of the A3 axes along the S2 great circle (Fig. 2a) on the stereographic projection, point to a similar geometry and kinematics of deformation during the D2 and D3 phases. At the microscale, the S3 is a crenulation cleavage characterized by pressure solution and by a shape preferred orientation (SPO) of quartz aggregates (Fig. 5d). Pressure solution and quartz deformation mechanisms acting during the D3 phase suggest shallower conditions with respect to the D2 phase.

The D4 deformation event is detectable only at the map scale by the occurrence of gentle kilometric folds. The F4 folds cause the variation of S2 strike from N080°–90° up to N000°–010° (Fig. 2a). The F4 axial planes (Ap4) strike about N135°/55° NE while the A4 axes trend N105°, plunging 35° to the SE. No axial plane foliation is observed parallel to the F4 axial planes.
The D5 deformation phase is characterized by meso- to mapscale folds. The F5 folds have a gentle to close geometry with a sub-horizontal axial plane and N88°-trending axes slightly plunging to the east. The variation in the dip of the S2 mylonitic foliation is due to later F5 folds producing type 3 interference structures when overprinting the F2 folds (Fig. 2b). Foliations and lineations related to the D5 phase have not been detected in the study area. This observation, coupled with the absence of mineral growth or

![Geological map of the study area.](https://www.cambridge.org/core/fig/2a)

![N–S geological cross-section of the studied area.](https://www.cambridge.org/core/fig/2b)
related microstructures, suggests that D5 took place at shallow structural levels. The D5 phase has been associated with the late exhumation related to the final collapse of the Sardinian Variscan Belt (Carmignani et al. 1994).

Evidence of brittle deformation is poorly represented in the study area. In the southeastern portion of the study area (Fig. 2a), conjugated normal faults have been detected. They strike N045° and N155° and steeply dip, respectively, to the NW and NE (Fig. 2a).

4. EBSD analysis and quartz petrofabric
4.a. Methods

Of the 39 rock specimens (Fig. 6) used for the microstructural analysis, ten representative samples of the main lithotypes were selected to investigate the rheological behaviour of quartz during the deformation. On these samples we performed image analyses using the software ImageJ (ver: 1.47v by Wayne Rasband).

For each specimen, several statistics images away from quartz ribbon areas (Fig. 7) were processed to estimate the amount of quartz in the matrix. All the analysed samples show a modal abundance of quartz in the matrix ranging between 24 % and 81 % (Fig. 8; Table 1). Following a microstructural analysis on the complete dataset and the modal estimates, three quartz-rich samples were selected for the CPO study.

These samples contain polycrystalline quartz ribbons recrystallized during the D2 phase (Fig. 8). The analysed samples were selected at different distances from the high-strain zone (Fig. 7) within the three main lithotypes occurring in the study area: SCB006R from the garnet + plagioclase zone (southern area); OGO026R from the augen orthogneiss zone (central area); and SGR031R from the garnet + staurolite + biotite zone (northern area) (Fig. 6). The electron backscatter diffraction (EBSD) analysis was performed on selected areas representative of the SGR recrystallization domains, related to the late D2 phase, in order to better constrain the late-D2 deformation event (Fig. 8).

EBSD analysis was performed on carbon-coated polished thin-sections using a JEOL 6610 scanning electron microscope (SEM) at the Plymouth University Electron Microscopy Centre, with the following working conditions: acceleration voltage of 20 kV, high vacuum and 70° sample tilt. EBSD patterns were acquired on rectangular grids ~6–9 mm² in size (Fig. 8), with an electron beam step-size of 3.5–5.0 μm.

The bulk CPO data have been represented with pole figures of the main crystallographic elements of quartz: c axis <0001>, a axes {11–20} and m planes {10–10}. The orientation data have been plotted as one point per grain. The EBSD results have also been shown as inverse pole figure (IPF) crystallographic maps to visualize the spatial distribution of the CPO domains (Fig. 9).

---

**Fig. 3.** (Colour online) Main lithotypes in the study area.
(a) Biotite-rich granodiortic orthogneiss (coin for scale is 2.3 cm diameter). (b) Augen orthogneiss with K-feldspar porphyroclasts (pencil tip for scale is 3 cm long). (c) Garnet + plagioclase-bearing micaschist (hammer for scale is 32 cm long). (d) Garnet + staurolite + biotite-bearing micaschist (fingernail for scale is ~1.2 cm wide). In some specific areas, S3 crenulation cleavage is affecting, respectively, (e) orthogneiss (compass for scale; visible upper side is ~3.2 cm long) and (f) micaschist (hammer for scale is 30 cm long).
Fig. 4. (Colour online) Microstructures associated with the main deformation events present in the study area from the (a–c) early D2 stage to the (d–e) late D2 stage and (f) D3 phase. (a) S-C fabric in garnet + plagioclase-bearing micaschist with S planes composed of biotite and white mica. Dextral sense of shear, corresponding to a top-to-the-W and -NW sense of shear in the field (mylonitic foliation steeply dips to the south) (crossed nicols). (b) Biotite and white mica foliation fish in garnet + staurolite + biotite-bearing micaschist (crossed nicols). (c) Mica fishes (group 1 and 2) in a quartz-rich matrix (ten Grotenhuis et al. 2003) (crossed nicols). (d) Garnet porphyroclast in garnet + plagioclase-bearing micaschist with syn-D2late growth of chlorite in fractures and strain shadows (parallel nicols). (e) C’ plane with syn-kinematic growth of chlorite in garnet + plagioclase + biotite-bearing micaschist (parallel nicols). (f) F3 centimetre fold in the granodioritic orthogneiss. Furthermore, it is also possible to note how the D3 event is associated with pressure solution as the main deformation mechanisms (parallel nicols). Mineral abbreviations: Grt – garnet; Qtz – quartz; Wm – white mica; Chl – chlorite; Bt – biotite.

Fig. 5. (Colour online) General overview of syn-D2 quartz microstructures recognized in the study area along samples collected at different distances from the high strain zone of the PAL (i.e. boundary between L–MGMC and HGMC). (a) Southern sector of the study area: quartz microstructures are dominated by GBM recrystallization with incipient SGR. Sb represents the oblique foliation due to the shape preferred orientation (SPO) of quartz aggregates (crossed nicols). (b) Central sector of the study area: SGR microstructures are more developed (crossed nicols). (c) In the northern sector of the study area, close to the high strain zone, the SGR process is pervasive and completely obliterates GBM microstructures (crossed nicols). (d) Plastic deformation in quartz within the hinge zone of an F3 fold where the SPO of quartz crystals is parallel to the S3 foliation (crossed nicols).
4.b Crystallographic preferred orientation data

The distribution of the quartz c axes is similar for the three analysed samples. Both SCB006R and SGR031R have a type-1 crossed girdle transitional to single girdle distribution (Fig. 9a, c) (Lister & Hobbs, 1980; Schmid & Casey, 1986; Passchier & Trouw, 2005, p. 104; Toy et al. 2008). OGO026R presents an incomplete c-axes distribution owing to the larger grain size compared to the other samples. Type-1 crossed girdle distributions suggest a plane strain deformation (Lister & Hobbs, 1980; Schmid & Casey, 1986). The asymmetry of the distribution points to a non-coaxial regime (Law, 1990; Passchier & Trouw, 2005, p. 105) with a top-to-the-W sense of shear, in agreement with independent kinematic indicators (see above) and consistent with the D2 phase. Despite the incomplete data, a type-1 crossed girdle distribution is still recognizable.

The pole figures and the IPF maps are consistent with the dominant activity of the rhomb <a> slip system, with a lesser contribution of prism <a> and basal <a> (Fig. 9a, b) (Toy et al. 2008; Fazio et al. 2017; Hunter et al. 2018). These data, under typical geological conditions for H2O content and geological strain rate, suggest a deformation temperature in the upper greenschist facies (e.g. Passchier & Trouw, 2005, p. 57; Toy et al. 2008). Deformation temperatures (Td) have been estimated using the relationship between Td and the opening angle (OA) of the c-axes distribution (Fig. 9c) (Kruhl, 1998; Morgan & Law, 2004; Law, 2014) with the most recent, pressure sensitive, calibration proposed by Faleiros et al. (2016):

$$ T_d = 410.44 \ln(OA) + 14.22P - 1272 $$

To estimate the Td with this relationship, an external pressure constraint (P) is necessary. In this work, the P constraint has been obtained from the data of Carosi & Palmeri (2002), who estimated a pressure of 0.7 GPa for the D2 peak. Moreover, considering that the late recrystallized areas developed under retrograde greenschist facies, the rocks deformed during the late D2 phase were likely under lower pressure conditions compared to the peak conditions. For this reason, a more realistic P of 0.5 GPa has been also assumed (based on P–T paths of Carosi & Palmeri, 2002) for the calculations. The estimated temperatures at the corresponding P of 0.5 GPa for SCB006R and SGR031R are 400 ± 50 °C and 390 ± 50 °C, respectively. The estimations for OGO026R were not possible owing to the incomplete nature of the c-axes distribution. These temperatures estimated using 0.5 GPa as the pressure constraint do not present a significant deviation (less than 30 °C) from those obtained at the corresponding pressure of 0.7 GPa. These data do not show large variations from the Td values obtained by the original Kruhl (1998), as modified by Morgan & Law (2004), calibration. The results obtained by the different calibrations and pressure values are well within the error range of the method (± 50°; see also Law, 2014). The Td data are consistent with the estimated greenschist-facies conditions, and indicate a homogeneous deformation temperature in the study area along the N–S transect, as also supported by the late syn-D2 greenschist minerals.
4.c. Kinematic vorticity data

The kinematic vorticity represents the magnitude of the kinematic vector, which the material lines tend to rotate around during flow deformation (Xypolias, 2010 and references therein). The kinematic vorticity number (indicated as Wn, for single deformation increments) is identified by the cosine of the angle between the two flow apophyses. Kinematic vorticity is an estimate of the relative contribution of pure and simple shear components during the flow.

Quartz CPO data have been used to estimate the components of simple and pure shear of the flow kinematic during the late D2 deformation increments (Wallis, 1995; Law et al., 2004, 2010, 2011, 2013; Xypolias, 2010). This estimate has been performed by calculating the sectional kinematic vorticity number (Wn) on recrystallized quartz domains using the $\beta/\delta$ method (Fig. 10) proposed by Xypolias (2010) (see also Wallis, 1995), where:

$$Wn = \sin[2(\delta + \beta)]$$

In this equation, $\beta$ is the angle between the main foliation and the plane normal to the quartz c-axes distribution (Figs 9c, 10b), while $\delta$ is the highest angle between the main foliation and the SPO foliation of quartz aggregates (Fig. 10a, b). $\delta$ has been derived from a range of angles measured via optical microscopy (Fig. 5a).

This method is based on the relationship between the simple shear component and the angle between the flow apophysis, $A_2$, highlighted by the orientation of the quartz CPO and the instantaneous stretching axis, ISA$_2$, represented by the oblique foliation. Showing the main flow elements in the Mohr space (Fig. 9c), the Wn can be obtained by the sine of the angle between $A_2$ and ISA$_2$. The kinematic vorticity estimations (Fig. 11) provided $Wn = 0.99-1.00$ for SCB006R, $Wn = 0.91-1.00$ for OGO026R and $Wn = 0.99-1.00$ for SGR031R. The obtained Wn values point to very high components of simple shear for all samples.

4.d. Palaeopiezometry and strain rate

Following the pioneering analysis of Twiss (1977) it has been suggested that the grain size of recrystallized grains is a primary function of the applied flow stress, representing the theoretical base of palaeopiezometry (Behr & Platt, 2011, 2013; Menegon et al., 2011; Boutonnet et al., 2013). The grain-size distribution of the analysed quartz aggregates (Fig. 12a) was derived for each sample with the aid of the EBSD data. These distributions have been used to estimate the flow stress acting during the recrystallization process...
related to the late D2 activity of the PAL. For this purpose, we used the recrystallized grain-size palaeopiezometer for quartz proposed by Stipp & Tullis (2003) for the recrystallization regime 2/3, where:

$$D = 3631 \Delta \sigma^{1.26}$$

With this relationship, the flow stress ($\Delta \sigma$) can be obtained from the average recrystallized grain diameter (D). The calculated values were: $D = 34 \pm 16 \mu m$ and $\sigma = 41 \pm 18$ MPa for SCB006R; $D = 25 \pm 6 \mu m$ and $\sigma = 53 \pm 11$ MPa for OGO026R; and $D = 16 \pm 4 \mu m$ and $\sigma = 74 \pm 16$ MPa for SGR031R, showing an increase in flow stress moving from the southern to the northern sectors of the study area, along a N–S transect of the PAL. The flow stress data have been used to calculate the strain rate ($\dot{\varepsilon}$) with the wet-quartzite flow law:

$$\dot{\varepsilon} = A \Delta \sigma^n (f_{H_2O})^m e^{-Q/RT}$$

For the strain rate estimation, a temperature constraint (T) is necessary. The T values used in this work have been obtained from the opening angle of the c-axes distribution (Law, 2014) as described above. A water fugacity ($f_{H_2O}$) of 12.25 MPa was calculated using the water fugacity coefficient listed in Tödheide (1972) for $T = 400 \ ^\circ C$ and $P = 0.5$ GPa. Different experimental calibrations for the wet-quartzite flow law have been proposed in the literature (see Table 2, where: $A$, $n$ and $m$ are experimentally calculated parameters that change for each calibration and $R$ is the ideal gas constant), and they led to dissimilar strain rate estimations (Menegon et al. 2011; Boutonnet et al. 2013; Montomoli et al. 2018).

The results of the different calibrations used in this paper are summarized in Figure 12b. Strain rate estimations cover a wide range, spanning from $10^{-16}$ to $10^{-11}$ s$^{-1}$ (including the uncertainties due to the propagation of the error on temperature and flow stress). Although the strain rates estimated using different flow laws are different, they coherently indicate an increasing trend of $\varepsilon$ moving from south to north (Fig. 12b). Among the different calibrations for the quartzite flow law, the one proposed by Hirth et al. (2001) has been selected since, according to Behr & Platt (2013, 2014), it is considered the most realistic (see also Boutonnet et al. 2013 for a discussion on this topic). The estimated strain rate values are in the range of $10^{-13}$ to $10^{-12}$ s$^{-1}$ (Fig. 12b).

Comparable strain rate results have been reported for the mid crustal shear zone from different areas such as the Betic Cordillera of southern Spain (Behr & Platt, 2013), the Rodope Massif in Greece (Fazio et al. 2018), the Kabilo–Calabride crystalline basement in southern Italy (Ortolano et al. 2020) and the
Table 1. Modal abundance of quartz along the study transect (see Fig. 6 for sample locations)

| Sample      | Structural distance (km)* | Lithotype                                    | Unit                                         | %Qtz |
|-------------|---------------------------|----------------------------------------------|----------------------------------------------|------|
| Ky          | 1.59                      | micaschist                                   | kyanite-bearing micaschist                   | 24.7 |
| PRQ034C     | 2.15                      | paragneiss                                   | garnet + staurolite + biotite-bearing micaschist | 42.5 |
| OGG030      | 3.05                      | orthogneiss                                  | granodioritic orthogneiss                    | 39.4 |
| PRQ022      | 3.65                      | paragneiss                                   | garnet + plagioclase-bearing micaschist      | 81.1 |
| OCCHI-2     | 4.96                      | orthogneiss                                  | augen orthogneiss                           | 75.2 |
| OCCHI-4     | 4.99                      | orthogneiss                                  | augen orthogneiss                           | 79.7 |
| OCCHI-6     | 5.01                      | orthogneiss                                  | augen orthogneiss                           | 58.3 |
| SCB003      | 5.93                      | micaschist                                   | garnet + plagioclase + biotite-bearing micaschist | 72.9 |
| 12-6-3-6b   | 8.65                      | micaschist                                   | garnet + plagioclase + biotite-bearing micaschist | 34.6 |
| 12-6-3-4    | 9.63                      | micaschist                                   | garnet + plagioclase + biotite-bearing micaschist | 56.0 |

* The structural distance refers to the relative position with respect to the boundary between the HGMC and L-MGMC.
Chelmos Shear Zone in the External Hellenides (Xypolias & Koukouvelas, 2001).

5. Discussion

5.a. Structural evolution

Our field, meso- and microstructural data document a complex polyphase tectono-metamorphic evolution of the Sardinian Variscan Belt after the collisional stage. The PAL activity is confirmed to be related to the D2 phase, during which micaschist and orthogneiss are deformed under amphibolite-facies conditions, coeval with a non-coaxial dextral transpressive shearing (Carosi & Palmeri, 2002). The syn-kinematic growth of chlorite, in strain shadows and along the C’ planes (Fig. 4e), is consistent with a metamorphic retrogression towards greenschist facies during the evolution of the late D2 phase. The syn-kinematic growth of chlorite during the D2 phase, at the expense of biotite and garnet, supports the presence of H2O-rich fluids during this phase.

The D3 phase developed heterogeneously in the study area. The parallelism of the structural elements between the D2 and D3 deformation phases allows the D3 to be interpreted as an evolution of the D2 phase linked to the latest deformation increments of the PAL. The D2–D3 transition could be related to a strain hardening due to shallower metamorphic conditions with strain localization and deformation concentrated in parallel crenulated domains (Fig. 13).

The absence of metamorphic assemblages related to the D4 and D5 phases points to a further T decrease associated with the deformation of the L–MGMC at shallower structural levels.
5.b. Quartz deformation

The important role of quartz in governing the rheology of the analysed rocks is evidenced by the modal abundance of this phase. The modal estimations show a quartz percentage range of 24 to 81% (Table 1), reaching the minimum abundance (Fig. 7) to be considered the strain-supporting phase in each specimen (~20% according to Handy, 1990). A quartz-dominated rheology allowed us to consider the CPO analysis, performed on quartz ribbons, reliable at the larger scale.

At fluid-present conditions and at geological strain rates (Law, 2014), microstructures in quartz related to GBM recrystallization are generally indicative of dislocation creep deformation under amphibolite-facies metamorphic conditions (Stipp et al. 2002a, b). The transition from GBM to SGR can be related to several factors:

### Table 2. Available calibrations for the wet-quartzite flow law and associated experimentally derived flow law parameters*

| Flow law calibration | Q (kJ/mol) | A (MPa⁻¹s⁻¹) | m  | n  |
|----------------------|------------|--------------|----|----|
| Paterson & Luan (1990) | 135        | 6.50E-08     | 0  | 3  |
| Luan & Paterson (1992) | 152        | 4.00E-10     | 0  | 4  |
| Gleason & Tullis (1995) | 223        | 0.00011      | 0  | 4  |
| Hirth et al. (2001) | 135        | 6.31E-12     | 1  | 4  |
| Gleason & Tullis (1995) (corrected by Holyoke & Kronenberg, 2010) | 223 | 0.00051 | 0 | 4 |

* Q - activation energy; A - material parameter; m - water fugacity exponent; n - stress exponent.

![Grain-size distribution](image)

**Fig. 12.** (a) Quartz grain-size distributions for the selected sample. See Figure 6 for sample locations. The grain-size intervals used for palaeopiezometry have been picked out from the total distribution, selecting the D2 new grains formed by SGR, which represent the finest population of grains for each sample. (b) Results of strain rate (s⁻¹) estimations from the analysed samples using the different quartz flow law calibrations in the dislocation creep regime. A consistent trend of increasing strain rates towards the N (i.e. from sample SCB006R to sample SGR031R) is evident.
a variation in deformation temperature, a decrease in water content or an increase in strain rate (Stipp et al. 2002a,b; Menegon et al. 2008, 2011; Law, 2014 and references therein). On the other hand, assuming no significant variations in water content and strain rate, SGR recrystallization (Fig. 5a–c) suggests a dislocation creep regime under a lower deformation temperature with respect to GBM (Stipp et al. 2002a,b; Passchier & Trouw, 2005; Law, 2014). The dextral sense of shear, testified to by the orientation of the oblique foliation, highlighted by quartz grains recrystallized by SGR fits well with the dominant sense of shear during the activity of the PAL. The synkinematic growth of retrograde assemblages, including chlorite along the C' planes (Fig. 4c, e), suggests a continuation of the PAL activity from higher to lower temperatures. This transition is testified to also by the quartz c-axis opening angle that points to lower temperatures compared to the peak conditions associated with the D2 (see Carosi & Palmeri, 2002; Di Vincenzo et al. 2004). For these reasons we infer that the superimposition of SGR is related mainly to a syn-shearing temperature decrease. In this framework, according to the structural and metamorphic evidence discussed above, a complex (and long-lasting) evolution of the D2 phase has been inferred. The D2 deformation phase is subdivided in two stages: an early D2early stage associated with the thermal D2 peak, and a late D2late stage as part of the retrograde exhumation path with decreasing temperatures. The variation in quartz dislocation creep regime from GBM to SGR is linked to a temperature decrease, and the acquisition of CPO data on the areas affected by SGR recrystallization allowed the constraining of the conditions of the D2late deformation increments. The temperature of nearly 400 °C, derived from the c-axes opening angle, is consistent with the greenschist-facies metamorphism and suggests a thermal homogenization during the late stages of D2.

The kinematic vorticity analysis performed on recrystallized quartz aggregates resulted in high Wn values (0.91–1.00) indicative of a simple shear dominated flow (Fig. 11). Previous studies (Carosi & Palmeri, 2002; Carosi et al. 2005) pointed out a pure shear dominated flow in the same study area (Wm = 0.30–0.70), estimated using the stable porphyroclasts method (Passchier, 1987; Wallis et al. 1993; Xypolias, 2010). A possible explanation for these different results could be found in the different methods applied to the estimation of the kinematic vorticity values. It is necessary to take into account the different strain memory of the quartz recrystallization with respect to the porphyroclasts (Wallis, 1995; Xypolias, 2009, 2010). Carosi & Palmeri (2002) and Carosi et al. (2005) applied the stable porphyroclasts method using K-feldspar and plagioclase porphyroclasts, crystallized before the D2 phase and experiencing a rigid-clast behaviour during the PAL shearing. Competency contrast could be responsible for strain partitioning (Goodwin & Tikoff, 2002), so that some minerals, such as, for example, K-feldspar, with a different viscosity than quartz, could partition different kinematic (coaxial versus non-coaxial) and rheological (brittle versus viscous) components of the bulk deformation. On the other hand, quartz microstructures and CPO have a shorter strain memory (Xypolias, 2009, 2010) and better record the late deformation increments, especially in our case study, where the later SGR recrystallization areas were selected for the EBSD analysis. From the comparison of the data acquired by the two different methods, likely related to different stages of the transpression, it is possible to better reconstruct the kinematic evolution of the PAL, which was dominated by pure shear during the early deformation stages and by simple shear in the late deformation increments (Fig. 13).

Other transpressive systems, showing a bulk non-coaxial deformation, with a deformation regime evolving from a pure shear dominated transpression to a simple shear dominated transpression, have been documented by Carreras et al. (2010).

5.c. Consideration of the exhumation mechanism

Shear zones are common features in deforming rocks and occur at all scales from the millimetre scale to the kilometre scale (Fossen & Cavalcante, 2017). Whereas most of the studies have focused on metre- to hectometre-scale shear zones, comparatively fewer studies are available for regional-scale shear zones and on the consequences on the P–T–t paths of the deformed rocks. In the Himalayas, a regional-scale (thrust-sense) shear zone running for more than 1000 km along-strike, the High Himalayan Discontinuity (Montomoli et al. 2013, 2015), affected the tectonic and metamorphic evolution of the metamorphic core of the belt for more than 10–15 Ma (Carosi et al. 2018).

Analogously, the PAL is an example of a (transpressive) crustal-scale shear zone that affected the inner portion of the Variscan Belt for several hundred kilometres (from Sardinia–Corsica and the Maures massif up to the External Massifs of the Alps: Corsini & Cavalcante, 2017). Whereas most of the studies have focused on metre- to kilometre-scale shear zones, comparatively fewer studies are available for regional-scale shear zones and on the consequences on the P–T–t paths of the deformed rocks. In the Himalayas, a regional-scale (thrust-sense) shear zone running for more than 1000 km along-strike, the High Himalayan Discontinuity (Montomoli et al. 2013, 2015), affected the tectonic and metamorphic evolution of the metamorphic core of the belt for more than 10–15 Ma (Carosi et al. 2018). However, it is known that the deeper portions of the belt from medium-temperature up to low-temperature conditions, playing a primary role in the tectonic end-metamorphic evolution of the Variscan Belt. Considering the overall development of the PAL and its prosecution in the External Crystalline Massifs of the Alps, this first-order...
shear zone is a major zone of weakness in the crust, and it allowed further localization of the deformation during the first stages of the Alpine cycle (Bergomi et al. 2017; Balleure et al. 2018).

The PAL has a sub-horizontal or gently plunging (L2) stretching lineation and sub-vertical foliation, and the exhumation could be expected to be driven by a nearly horizontal extrusion (e.g. Tikoff & Fossen, 1993; Schulmann et al. 2003; Iacopini et al. 2008). On the other hand, mylonite in a transpressive zone could be affected by vertical movement of the rock flow. This case, where pure and simple shear are active coevally during deformation, is typically related to a vertical stretching lineation during the whole deformation history or caused by the flipping of an originally shallowly plunging lineation to a vertical attitude (Tikoff & Fossen, 1993; Schulmann et al. 2003; Iacopini et al. 2008). However, in simple shear dominated transpression, in the case of vertical extrusion, horizontal lineation can be preserved (Iacopini et al. 2008). It is worth noting that in the study sector of the PAL, the kinematic flow was dominated by simple shear only during the latest D2 incre-

As previously mentioned, the superposition intensity of SGR (D2late) microstructures over the older GBM (D2early) fabric in quartz aggregates increases moving from south to north in the study area, reaching the highest intensity approaching the core of the PAL. Considering that the deformation temperature during D2late was nearly homogeneous along the N–S study transect, as suggested by syn-mineralogic minerals and opening angle thermometry, the increase in SGR intensity cannot be attributed to a (significant) temperature variation. On the contrary, petrofabric data suggest that the strain rate spatially increases moving to the north from \(10^{-13} \text{s}^{-1}\) to \(10^{-12} \text{s}^{-1}\) according to the Hirth et al. (2001) flow law (Fig. 12b). Thus, we can infer that strain rate variations played an important role in the microstructural evolution of quartz during SGR recrystallization in this area (Hobbs, 1985; Passchier & Trouw, 2005).

The grain-size palaeopiezometry and strain rate estimates, performed on the analysed samples, can be compared with other data available in the literature for the Sardinian Variscan Belt (e.g. Casini et al. 2010; Casini & Funedda, 2014; Montomoli et al. 2018). According to these authors, the strain rates increase moving from south to north, ranging from \(10^{-16} \text{s}^{-1}\) to \(10^{-15} \text{s}^{-1}\) in the Foreland Zone to \(10^{-15} \text{s}^{-1}\) to \(10^{-13} \text{s}^{-1}\) in the Nappe Zone and \(10^{-13} \text{s}^{-1}\) to \(10^{-11} \text{s}^{-1}\) in the Inner Zone.

The strain rate proposed in this work is closely comparable to a typical geological strain rate (Pfiffner & Ramsay, 1982; Passchier & Trouw, 2005). Nevertheless, it is worthwhile to consider the discussion recently opened by Fagereng & Biggs (2019) where they postulate on the underestimation of \(10^{-15} \text{s}^{-1}\) as a typical strain rate value.

6. Conclusions

Data from geological mapping and structural analysis, at different scales, allowed the constraint of the tectonic evolution of the northern sector of the Variscan Belt in Sardinia within the L–MGMC close to the PAL. The PAL is regarded as an orogen-parallel transpressional shear zone that drove the exhumation of the Sardinian metamorphic complexes. In the northern L–MGMC, the shearing event is represented by the D2 phase, which started to be active close to the metamorphic ‘peak’, under amphibolite-facies conditions (D2early), and lasted up to greenschist-facies conditions during the D2late event. The transpressive tectonics related to the PAL continued during the D3 phase under even shallower crustal conditions. The shift in metamorphic conditions caused strain partitioning along the mylonitic belt, giving rise to shear zone-parallel discontinuous domains characterized by the folding of S2 foliation (Fig. 13).

Quartz petrofabrics, together with microstructural data, suggest that the transition from D2early to D2late has been characterized by:

1. a nearly thermal homogenization at \(\sim400^\circ\text{C}\) where a shift in the dynamic recrystallization mechanisms in quartz aggregates, from GBM to SGR, is documented. The SGR overprinting microstructures are incipiently developed in the southern area and gradually become more pervasive moving into the northern area;
2. an increase in the simple shear component during deformation, ranging from pure shear to simple shear dominated transpression (Fig. 13).

Comparing the structural analysis data with the available kinematic vorticity estimates, based on different vorticity gauges, it is possible to infer that the PAL, in the study area, led to a tectonic evolution characterized by horizontal extrusion or, alternatively, by a horizontal extrusion occurring during the D2early phase followed by a vertical extrusion coeval with the D2late phase. With the present data it is not possible to verify the latter hypothesis and to clarify the kinematics of the late mylonitic flow of the PAL, and further investigations are needed.

The flow stresses and the strain rates suggest an increase in these two parameters moving closer to the core of the PAL. This variation is in agreement with the presence of the superimposition of the SGR recrystallization mechanism in quartz in the northern sector.

The new data support a framework in which a single long-lasting, crustal-scale shear zone, once formed at depth, is able to continue to localize deformation and to drive the exhumation of the inner portions of the belt towards lower P–T conditions. This shear zone, continuing in other portions of the Southern Variscan Belt, was active until the end of the Variscan Orogeny and acted as a weak zone reactivated during the later Alpine tectonics.

Acknowledgements. The staff at the Plymouth University Electron Microscopy Centre is thanked for support during EBSD analysis. Funding: PRIN 2015 (University of Torino: R. Carosi and C. Montomoli); funds Ricerca Locale Università di Torino (ex-60%, R. Carosi and S. Iaccarino); PRA 20018_41 to C. Montomoli. Jorge Alonso-Henar and Eugenio Fazio are thanked for their careful reviews. We thank the editor, Prof. Lacombe for his very efficient handling of the paper.

Downloaded from https://www.cambridge.org/core. IP address: 92.238.190.127, on 14 Feb 2022 at 14:25:04, subject to the Cambridge Core terms of use, available at https://www.cambridge.org/core/terms . https://doi.org/10.1017/S0016756820000138
References

Alvarez W (1972) Rotation of the Corsica–Sardinia Microplate. Nature 235, 103–5. doi: 10.1038/physci235103a0.

Balleire M, Manzotti P and Dal Piaz GV (2018) Pre-Alpine (Variscan) inheritance: a key for the location of the future Valaisian Basin (Western Alps). Tectonics 37, 786–817. doi: 10.1002/2017TC004633.

Bergomi MA, Dal Piaz GV, Malusà MG, Monopoli B and Tunesi A (2017) The Grand St Bernard–Briainsonnais nappe system and the Paleozoic inheritance of the Western Alps unzipped by zircon U–Pb dating. Tectonics 36, 2950–72. doi: 10.1002/2017TC004621.

Behr WM and Platt JP (2011) A naturally constrained stress profile through the middle crust in an extensional terrane. Earth Planetary Science Letters 303, 181–92. doi: 10.1016/j.epsl.2010.11.044.

Behr WM and Platt JP (2013) Rheological evolution of a Mediterranean subduction complex. Journal of Structural Geology 54, 136–35. doi: 10.1016/j.jsg.2013.07.012.

Behr WM and Platt JP (2014) Brittle faults are weak, yet the ductile middle crust is strong: implications for lithospheric mechanics. Geophysical Research Letters 41, 8067–75. doi: 10.1002/2014GL061349.

Boutonnet E, Leloup PH, Sassier C, Garden V and Ricard Y (2013) Ductile strain rate measurements document long-term strain localization in the continental crust. Geology 41, 819–22. doi: 10.1130/G33723.1.

Burchfiel BC, Zhiliang C, Hodges KV, Yiping L, Royden LH, Changdong D and Jiene X (1992) The South Tibetan Detachment System, Himalayan Orogen: Extension Contemporaneous with and Parallel to Shortening in a Collisional Mountain Belt. Boulder, Colorado: Geological Society of America, Special Paper no. 269.

Cappelli B, Carmignani L, Castorina F, Di Pisa A, Oggiano G and Petrini R (2012) A balanced foreland–backarc tectonic model for the Western Alps around the North Calabrian Arc. Tectonophysics 538, 33–44. doi: 10.1016/j.tecto.2012.03.028.

Cappelli B, Carmignani L, Castorina F, Di Pisa A, Oggiano G and Petrini R (2013) Post-collisional evolution of the Variscan belt of northeastern Sardinia (Italy): implications for the exhumation of medium-pressure metamorphic rocks. Geological Magazine 150, 497–511. doi: 10.1017/S0016756812000676.

Carosi R and Pertusati PC (1990) Evolution strutturale delle unità tettoneciche nella Sardegna centro-meridionale. Bollettino della Società Geologica Italiana 109, 325–35.

Carosi R and Pertusati PC (2010) Evoluzione strutturale delle unità tettoneciche nella Sardegna centro-meridionale. Bollettino della Società Geologica Italiana 109, 325–35.

Carreras J, Czech DM, Druguet E and Hudleston PJ (2010) Structure and development of an anastomosing network of ductile shear zones. Journal of Structural Geology 32, 656–66. doi: 10.1016/j.jstrugeo.2010.03.013.

Casini L, Cuccuru S, Maino M, Oggiano G and Tiepolo M (2012) Emplacement of the Arzachena Pluton (Corsica–Sardinia Batholith) and the geodynamics of incoming Pangaea. Tectonophysics 544, 31–49. doi: 10.1016/j.tecto.2012.03.028.

Casini L, Cuccuru S, Puccini A, Oggiano G and Rossi P (2015) Evolution of the Corsica–Sardinia Batholith and late-orogenic shearing of the Variscides. Tectonophysics 646, 65–78.

Casini L and Funedda A (2014) Potential of pressure solution for strain localization in the Baccu Locci Shear Zone (Sardinia, Italy). Journal of Structural Geology 66, 188–204. doi: 10.1016/j.jstrugeo.2014.05.016.

Casini L, Funedda A and Oggiano G (2010) A balanced foreland–hindland deformation model for the Southern Variscan belt of Sardinia, Italy. Journal of Geophysical Research 115, B04318. doi: 10.1029/2009JB006553.

Corsi M and Rolland Y (2009) Late evolution of the southern European Variscan belt: exhumation of the lower crust in a context of oblique convergence. Comptes Rendus Geoscience 341, 214–23. doi: 10.1016/j.crte.2008.12.002.

Cortesogno L, Gaggero L, Oggiano G and Paquette JL (2004) Different tectono-thermal evolutionary paths in eclogitic rocks from the axial zone of the Variscan chain in Sardinia (Italy) compared with the Ligurian Alps. Offioli 29, 125–44. doi: 10.4454/Offioli.v29i2.210.

Cruciani G, Franceschelli M, Massonne HJ, Carosi R and Montomoli C (2013) Pressure temperature and deformational evolution of high pressure metapelites from Variscan NE Sardinia, Italy. Lithos 175–176, 272–84. doi: 10.1016/j. lithos.2013.05.001.

Cruciani G, Montomoli C, Carosi R, Franceschelli M and Puxeddu M (2015) Continental collision from two perspectives: a review of Variscan metamorphism and deformation in northern Sardinia. Periodico di Mineralogia 84, 657–99. doi: 10.2451/2015PM04555.

Deino A, Gattacceca J, Rizzo R and Montanari A (2001) 40Ar/39Ar dating and paleomagnetism of the Miocene volcanic succession of Monte Furru (Western Sardinia): implications for the rotation history of the Corsica-Sardinia Microplate. Geophysical Research Letters 28, 3373–6. doi: 10.1029/2001GL012941.

Depine GV, Andronicsi CL and Hollister LS (2011) Response of continental magmatic arcs to regional tectonic changes recorded by synorogenic plutons in the middle crust: an example from the Coast Mountains of British Columbia. Journal of Structural Geology 33, 1089–104. doi: 10.1016/j.jsg.2011.03.012.

Di Pisa A, Oggiano G and Talarico F (1993) Post collisional tectono-metamorphic evolution in the axial zone of the Hercynian belt in Sardinia: the example from the Asinara Island. Bulletin du Bureau de Recherches Géologiques et Minières 219, 216–17.

Di Vincenzo G, Carosi R and Palmeri R (2004) The relationship between tectono-metamorphic evolution and argon isotope records in white mica: constraints from in situ 40Ar/39Ar laser analysis of the Variscan basement of Sardinia. Journal of Petrology 45, 1013–43. doi: 10.1093/petrology/egh002.

Elter FM, Musumeci G and Pertusati PC (1990) Late Hercynian shear zones in Sardinia. Tectonophysics 176, 387–404. doi: 10.1016/0040-1951(90)90080-R.

Fagereng Å and Biggs J (2019) New perspectives on ‘geological strain rates’ calculated from both naturally deformed and actively deforming rocks. Journal of Structural Geology 125, 100–10. doi: 10.1016/j.jsg.2018.10.004.
Twiss RJ (1977) Theory and applicability of a recrystallized grain size paleopiezometer. In Stress in the Earth (ed. M Wyss), pp. 227–44. Contributions to Current Research in Geophysics (CCRG). Basel: Birkhäuser. doi:10.1007/978-3-0348-5745-1_13.

Wallis SR (1995) Vorticity analysis and recognition of ductile extension in the Sambagawa belt, SW Japan. Journal of Structural Geology 17, 1077–93. doi: 10.1016/0191-8141(95)00005-X.

Wallis SR, Platt JP and Knott SD (1993) Recognition of syn-convergence extension in accretionary wedges with examples from the Calabrian Arc and the Eastern Alps. American Journal of Science 293, 463–94. doi: 10.2475/ajs.293.5.463.

Xypolias P (2009) Some new aspects of kinematic vorticity analysis in naturally deformed quartzites. Journal of Structural Geology 31, 3–10. doi: 10.1016/j.jsg.2008.09.009.

Xypolias P (2010) Vorticity analysis in shear zones: a review of methods and applications. Journal of Structural Geology 32, 2072–92. doi: 10.1016/j.jsg.2010.08.009.

Xypolias P and Koukouvelas IK (2001) Kinematic vorticity and strain rate patterns associated with ductile extrusion in the Chelmos shear Zone (External Hellenides, Greece). Tectonophysics 338, 59–77. doi: 10.1016/S0040-1951(01)00125-1