Post collisional transpressive tectonics in northern Sardinia (Italy)

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Abstract: The aim of this work is to review and synthetize the geological and structural analysis performed in the Variscan Basement of Northern Sardinian during the last ten years and to add new preliminary data on the Anglona–SW Gallura area. A transpressive crustal-scale deformation (D2), is documented in the Variscan Basement of northern Sardinia. A shear deformation parallel to the belt, overprinting previous D1 structures related to a top-to-the S and SW nappe stacking, has been recognized. The L2 stretching lineation points to an orogen-parallel stretching and to a general change in the tectonic transport from D1 to D2 deformation phases. D1 phase developed during initial frontal collision whereas D2 deformation was characterized by dextral shearing. In this sector of the Variscan belt exhumation is due to continuing compression with an increasing component of horizontal displacement developed in a regime of decreasing pressure. The D2 transpressional deformation enhanced telescoping of the Barrovian isogrades and the exhumation of the low- to medium-grade metamorphic rocks. The overall change of the shortening direction in a large sector of an orogenic belt with the occurrence of increasing orogen-parallel displacement, may be regarded as a general mechanism affecting the exhumation of rocks and preventing the overthickened and thermally softened collisional crust from undergoing a diffused gravitational collapse. The rotation of nearly 90° of the tectonic transport in Sardinia during collisional and post-collisional stages could be related to paleoposition of the Corsica-Sardinia block, close to southeast France and northeast Spain, and to the development of the Ibero-Armorican arc.

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Introduction

Collisional type orogens are characterized by crustal thickening due to overthrusting and nappe development. The following tectonic evolution of the orogen and its overthickened crust is controlled by several factors, such as the progressive cessation of the driving forces, the change of the kinematics of the involved tectonic plates or extensional tectonics. As the tectonic compressive forces tend to diminish, the overthickened crust at high-temperature tends to collapse inducing the development of widespread extensional structures superimposed over compressional ones. However, large-scale plate kinematics often plays a first-order role in determining the fate of an orogen. For example extensional tectonics in the Caledonides affecting the belt soon after the development of compressive structures is attributed to a change from converging to diverging movement induced by far plate tectonic movement (Rey et al., 1997).

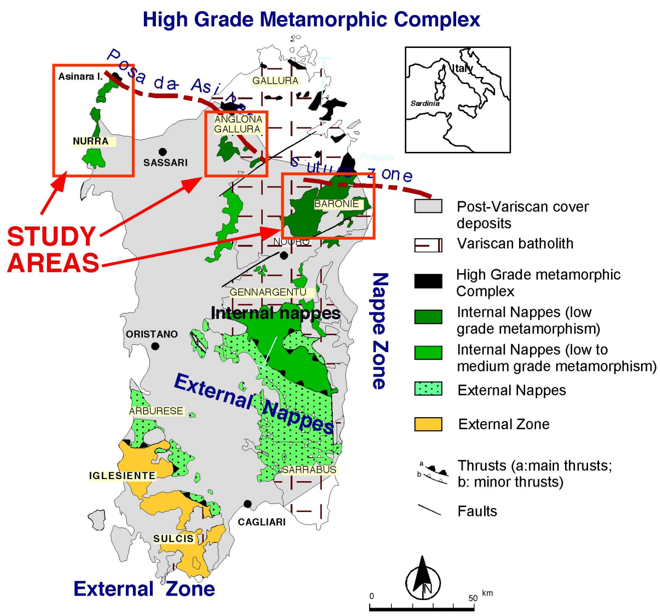
In the last twenty years an increasing number of evidence led to the recognition of extensional structures in many orogens (Dewey, 1988; Platt, 1986) such as the Variscan belt (Eisbacher et al., 1989; Van Den Driessche & Brun, 1992; Malavieille et al., 1990; Rey et al., 1992; Costa & Rey, 1995; Gardien et al., 1997). Metamorphic core complexes have been recognized in several parts of the Variscan belt and extensional tectonics is regarded as an efficient mechanism of exhumation of metamorphic rocks in the later tectonic history (Malavieille et al., 1990; Carmignani et al., 1993, 1994; Rey et al., 1997; Gardien et al., 1997).

Compression and/or transpression are often active after the thickening stage and could play an important role in the tectonic evolution and in the exhumation (Matte et al., 1998; Carosi & Palmeri, 2002; Carosi et al., 2004). The continuing compression and the change in the direction of convergence may originate transpressional deformation that deeply affects the thermal evolution of orogens (Thompson et al., 1997a, b).

However, there are increasing examples of metamorphic terrains in which transcurrent or transpressional shear zones overprint a previous D1 deformation responsible of crustal stacking (Carosi & Palmeri, 2002; Little et al., 2002; Vassallo & Wilson, 2002; Goscombe et al., 2003; Konopásek et al. 2005)

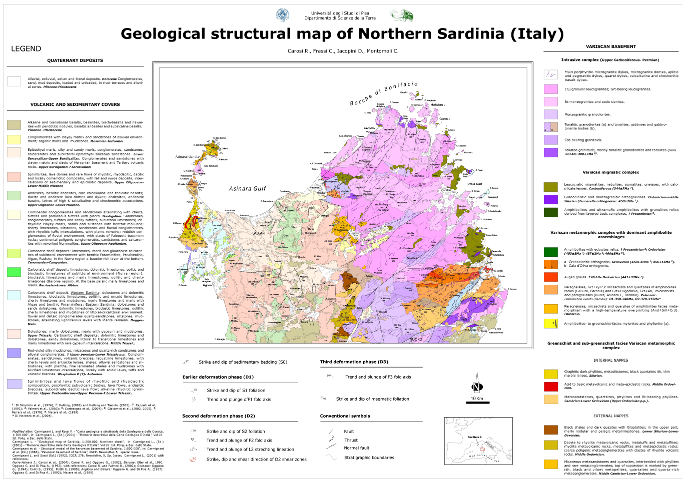
This paper is based on structural investigations in three key sectors of the Variscan basement of northern Sardinia, located respectively in the NW (Nurra and Asinara Island transect), in the centre (SW Gallura and Anglona) and in the NE (Baronie transect) of the island (Figs. 1 and 2). The aim of the paper is to show the syn- and post-thickening mechanisms which controlled the tectonic-metamorphic evolution and the exhumation of this part of the belt. The thickening-related structures are well documented all over the island but the following tectonic history remains unclear.

Figure 1. Tectonic sketch-map of the Variscan belt



Tectonic sketch-map of the Variscan belt in Sardinia and position of the study areas. 1: Post-Variscan cover deposits; 2: Variscan batholith; 3: High Grade Metamorphic Complex; 4: Internal nappes (low- to medium-grade metamorphism); 5: Internal nappes (low-grade metamorphism); 6: External nappes; 7: External Zone; 8: Thrusts (a: main thrusts; b: minor thrusts); 9: Faults; 10: Posada-Asinara Line (PAL).

Figure 2. Geological structural map of the Variscan basement in northern Sardinia



Modified after Carmignani et al., (1986), Carmignani & Rossi, (2001), and Carmignani et al., (2001). The main structural elements of the D1, D2 and D3 deformation phases have been reported.

A general change in the direction of tectonic transport from perpendicular to parallel to the belt during collisional and post collisional tectonics is described and a transpressional tectonic model is proposed as the main mechanism of exhumation of large sectors of the belt before the onset of the later extensional collapse at higher structural levels (Carosi & Oggiano, 2002; Carosi & Palmeri, 2002).

Geological setting

The Sardinian basement is one of the best preserved segments of the Southern European Variscan belt in the Mediterranean area, as in this area it has not been deformed significantly during the later Alpine deformation (Carmignani et al., 2001). The Sardinian Variscan basement is characterized by several deformation phases that testify the long evolution of Gondwanan and Avalonian margins (Stampfli et al., 2002) from the rifting and evolution of the Rheic Ocean (Cambrian) to the continental collision and final collapse (Upper Carboniferous). The basement of Sardinia is composed by Carboniferous magmatic and sedimentary rocks in the south and south-western part of the island and by metasedimentary Cambro-Lower Carboniferous sequences in the central northern areas, with an increasing in metamorphic grade and intensity deformation moving toward the inner zone of chain (from south-south west to north/north-east) (Carmignani et al., 1994; 2001 and references therein).

The collisional structural frame is broadly characterized by three main tectono-metamorphic complexes (Figs. 1, 2 and 3; Carmignani et al., 1982; 1994; 2001):

1. A thrust and fold belt foreland consisting of a sedimentary succession ranging in age from upper (?) Vendian to lower Carboniferous which crops out in the southwestern part of the island (External zone: Iglesiente and Sulcis);
2. A SW verging nappe stack which equilibrated mostly under greenschist facies conditions, consisting of a Palaeozoic sedimentary sequence bearing a thick continental arc-related volcanic suite (the so called "Nappe zone");
3. An inner zone ("Inner" or "Axial zone") characterised by medium- to high-grade metamorphic rocks that consists of two different metamorphic complexes: a) A medium grade, chiefly metapelitic complex, consisting of micaschists and paragneisses bearing ky ±stau±grt (Franceschelli et al., 1982; mineral abbreviation according to Kretz, 1983) and including quartzites and grt-bearing amphibolites boudins with N-MORB chemical affinity (Cappelli et al., 1992; Oggiano & Di Pisa, 1992; Cortesogno et al., 2004). In some areas eclogitic bodies are well preserved (e.g. South-West Gallura) (Eclogite B: Cortesogno et al., 2004). b) A polymetamorphic high-grade complex (HGMC) made up of anatexites and metatexites that developed from the northern part of the island to Corsica. Decametric amphibolitic bodies were recognized in this complex. They testify the complex history of this sector of the chain as the eclogitic paragneisses relics (Eclogite B: Cortesogno et al., 2004) registered a strong re-equilibration under the granulite facies (Miller et al., 1976; Ghezzo et al., 1979, 1982; Di Pisa et al., 1992; Franceschelli et al., 1998, 2002; Cortesogno et al., 2004; Giacomini et al., 2005) and afterward under medium pressure and temperature conditions (amphibolite facies).

U/Pb dating on zircons from the eclogitic bodies constrained the evolution of the HGMC. (Cortesogno et al., 2004; Giacomini et al., 2005). According to these authors the emplacement of mafic protolith is dated at ~460 Ma whereas the eclogitic peak could be placed at 403±4 Ma (Cortesogno et al., 2004). A different age of resetting of the U/Pb system in zircon at 352±3 Ma was instead interpreted from Giacomini et al. (2005), as the age of amphibolitic metamorphism.

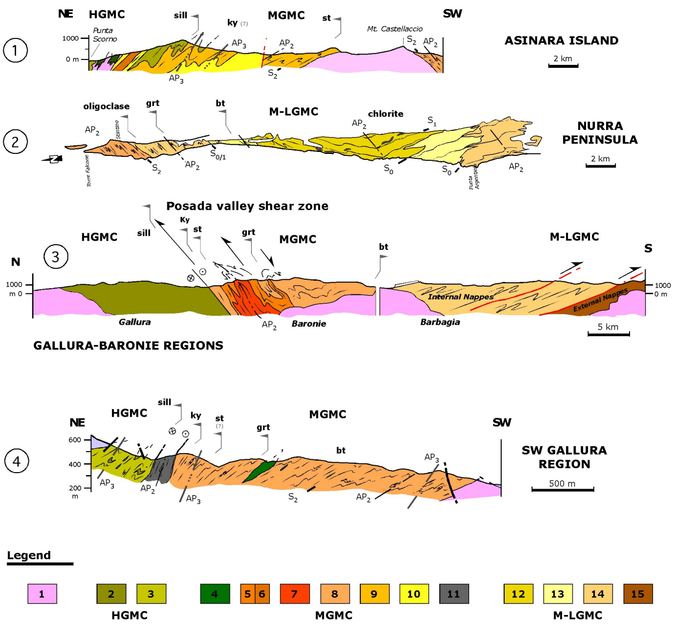
The contact between these two metamorphic complexes is well exposed along the Posada Valley (Figs. 2 and 3) (Elter et al., 1990) as well as in Southern Gallura and Asinara island (Oggiano & Di Pisa, 1992; Carosi et al., 2004), and it corresponds to a wide transpressive shear belt (Carosi & Palmeri, 2002), which is interpreted by several authors as a Hercynian suture zone (Posada-Asinara Line, PAL: Cappelli et al., 1992; Carmignani et al., 1994), between the Gondwana and Armorica continents affected by Late Variscan shear zones (Elter et al., 1990).

The three study transects are portions of the Internal Nappes (Inner zone: Carmignani et al., 1994) (Figs. 1, 2, and 3), translated toward the SW during the main collisional event (D1) and located just south of the Posada-Asinara suture. Micaschists and porphyroblastic paragneisses intruded by granodioritic orthogneisses and granitic augen gneisses (458±31 Ma and 441±33 Ma, respectively, Ferrara et al., 1978; ~457 Ma, Helbing & Tiepolo, 2005) crop out in the southern regions of study areas, whereas migmatites and migmatitic gneisses with bodies of amphibolites in the northern areas (Figs. 2 and 3).

Structural pattern

The Variscan basement in Sardinia is characterized by a polyphase deformation and metamorphic history. In northern Sardinia the main pervasive foliation at the regional scale is the S2 foliation that increases in intensity moving from south to north (Franceschelli et al., 1982; 1989; Carmignani et al., 1994). The oldest S1 foliation is clearly observable only at the outcrop scale in the southern portion of the Internal Nappes whereas going northward, it was progressively transposed by the D2 deformation phase (Fig. 3) and it is recognizable only as relics at the microscopic scale. Subsequently, two systems of later folds (F3 and F4) affected the S2 foliation.

Figure 3. Geological cross section



Geological cross section (1-4) throughout the Variscan basement of northern Sardinia in the three study areas. Same legend as in Figure 1. 1: Asinara Island, 2: Nurra region; 3: Baronie region; 4: SW Gallura region; HGMC: High Grade Metamorphic Complex; MGMC: Medium Grade Metamorphic Complex: L-MGMC: Low to Medium Grade Metamorphic Complex (Modified after Carmignani et al. (1979), Di Pisa & Oggiano, (1992), Carosi & Oggiano, (2002), Carosi & Palmeri, (2002), and Carosi et al., (2004)).

Strain pattern of D1 deformation

The D1 collisional event is well-recorded in the low-grade metamorphic rocks of the southern portion of the Nurra-Asinara transect, where it is associated with the development of an S1 foliation, axial plane of SW facing folds and shear zones (Carmignani et al., 1979; Franceschelli et al., 1990; Simpson, 1999; Carosi & Oggiano 2002; Montomoli, 2003; Carosi et al., 2004) (Figs. 2 and 3).

Moving to the north, D2 strain increases and completely transposes S1 foliation and related structures (cross-section n. 2 in Fig. 3). In southern Nurra, F1 folds are meter to decimeter in size and show variable opening angles (from 30° - 40° up to isoclinal). They show thickened hinges and stretched limbs and commonly belong to class 2 of Ramsay (1967).

A low grade S1 foliation, with syn-kinematic crystallization of quartz, muscovite, paragonite, chlorite and oxides (Carmignani et al., 1979) develops parallel to F1 axial planes..

Late D1 ductile/brittle shallowly dipping shear zones have been recognized in phyllites and quartzites (Simpson, 1999; Montomoli, 2003). Shear planes strike from N130 to N150°E and dip 30-40° to the NE (Fig. 2). C-S fabric indicates a top-to-the S and SE sense of shear. Late D1 shear zones are affected by S2 crenulation cleavage and they are characterized by abundant sigmoidal quartz veins. According to Simpson (1999), quartz veins are due to dehydration reactions during prograde metamorphism. During non-coaxial deformation they assume a sigmoidal shape.

Fluid inclusions trapped in quartz veins along secondary trails oriented both at high and low angles with respect to the shear zone boundaries, highlight the presence of aqueous carbonic fluids (Montomoli, 2003). Bulk composition and densities, calculated on the basis of microthermometric and Raman data, point out to two main types of fluid inclusions: a first type is characterized by density values ranging between 0.84 and 0.87 g/cm3 while a second type has density values between 0.88 and 0.92 g/cm3 (Montomoli, 2003). The computed isochores for both fluid inclusion types do not match the P-T peak conditions proposed for the area by Franceschelli et al., (1990), suggesting that fluid inclusions have been trapped or at least re-equilibrated during a retrograde metamorphic pattern. In particular, during their trapping, the host rocks experienced a lowering pressure with respect to the peak metamorphic conditions estimated for the D1 tectonic phase (Franceschelli et al., 1989, 1990).

The decreasing pressure values at the time of late D1 shear zones activity suggest that during crustal stacking the nappe pile continued to undergo overall compression causing the development of shear zones and allowing the exhumation of the hanging wall rocks.

Going towards the northernmost part of the study sections, D2 deformation shows a gradual strain increase up to the complete transposition of sedimentary bedding and S1 foliation (Fig. 3). S1 foliation is recognized only at the microscopic scale in D2 microlithons or as inclusion trails in post-D1 Barrovian porphyroblasts such as: plagioclase, biotite, garnet, staurolite and kyanite.

Mineral crystallized along the S1 foliation allowed to constrain the P-T conditions during prograde metamorphism. Moreover, syn-D1 celadonite-rich white mica, inside plagioclase in the garnet zone, allowed Di Vincenzo et al., (2004), to date the S1 foliation in northern Sardinia at 330-340 Ma.

Strain pattern of D2 deformation

This tectonic phase is characterized by a heterogeneous deformation affecting the previous syn-collisional fabric. D2 deformation is partitioned into domains alternating prevailing folding and shearing deformation (Carmignani et al., 1979; Simpson, 1999; Carosi & Oggiano, 2002; Carosi et al., 2004). This is particularly well recognizable along the Nurra-Asinara transect.

In low strain areas, to the south, D2 deformation is characterized by a crenulation cleavage axial planar of F2 folds. Moving northward, approaching the Posada-Asinara Line strain is predominantly characterized by non-coaxial deformation giving rise to the development of protomylonites, mylonites and phyllonites in which folding is only a secondary effect.

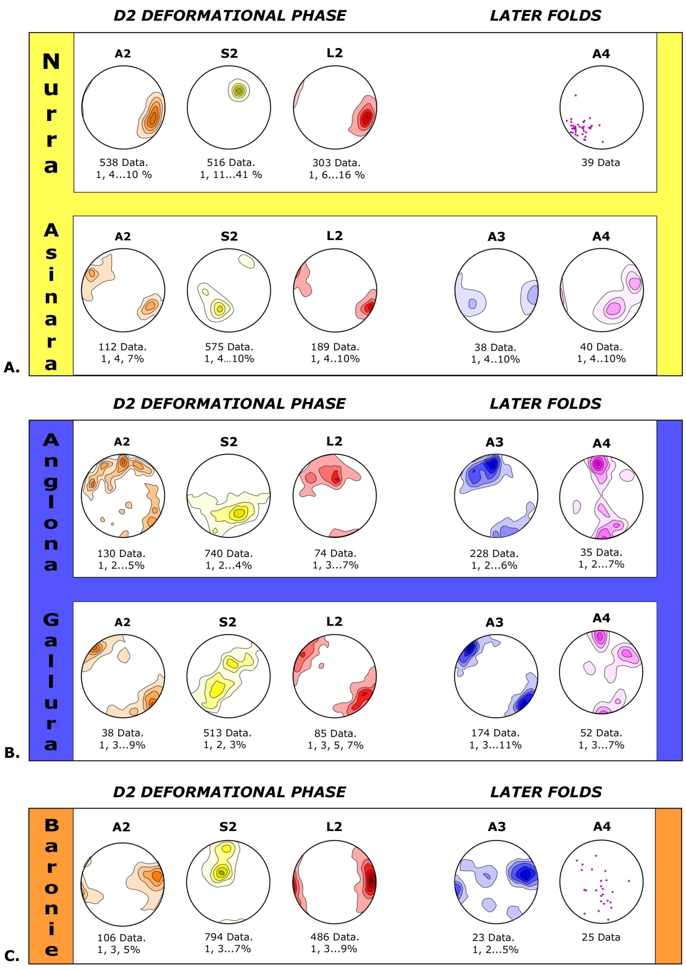
The F2 folds are well-detectable in the southern portion of the Nurra peninsula and in the southern part of the Baronie area. The Nurra-Asinara transect offers good outcrops showing the progressive development of the D2 deformation (Figs. 2 and 3) (Carmignani et al., 1979; Simpson, 1999; Carosi & Oggiano, 2002; Carosi et al., 2004).

D2 structures

The D2 deformation is characterized in the Nurra- Asinara and Anglona-Gallura zone by F2 folds with E-W to NW-SE trending axes (Figs. 4) and S to SW moderately to steeply dipping axial planes.

In the Nurra area, the F2 folds have centimetric to decimetric size and a well-developed S2 axial plane foliation that represents the main planar element recognized in the area. The F2 fold vergence is toward the north, which is opposite to the vergence of F1 folds (Fig. 3). In the Nurra area the F2 similar fold geometry changes from open (in the south) to isoclinal (in the north), with interlimb angles decreasing from 60° to 1-2°, whereas the fold geometry (Ramsay, 1967) varies from class 1C to class 3 (see Carosi et al., 2003). Sheath folds have been detected in central and northern Nurra (Carosi & Oggiano, 2002). In central Asinara Island F2 fold geometry varies from class 1C to 2. A2 axes trend mostly NW-SE plunging to the SE in the Nurra area and to the NW in the Asinara Island (Fig. 4a).

Figure 4. Stereographic projections



Stereographic projections (Schmidt equal area projection, lower hemisphere) of main structural elements for the study areas (a: Nurra-Asinara zone; b: Anglona-SW Gallura zones; c: Baronie zone). A2: axes of F2 folds; S2: schistosity; L2: stretching lineation. A3 and A4 are referred to later fold axes.

In the Tula-Erula area (Anglona region; Fig. 2), the D2 deformation phase is highlighted by rare F2 isoclinals folds with axis scattering from NW-SE to NNE-SSW with a maximum in the N-S direction (Fig 4b). Their plunge is less than 30 degree mainly to NNW (Fig. 4b).

In the Badesi-Giagazzu area (SW Gallura region; Fig. 2) the F2 axis orientation shows a NW-SE trend, plunging 10-20° toward both the SE and the NW (Fig. 4b). In this area from SW to NE, it is possible to observe a strain increase causing different F2 fold geometries. The F2 folds are chevron and kink in the SW region, whereas in the phyllonitic zone (toward NE), they become isoclinals (Fig. 5) with parallel limbs.

Figure 5. F2 tight folds in phyllonites

F2 tight folds in phyllonites from SW Gallura close to the boundary between the Medium Grade Metamorphic Complex and the High Grade Metamorphic Complex.

In the Baronie zone the F2 fold axes are still parallel to the stretching lineation but show a different trend with a mean ENE-WSW strike direction. F2 folds have moderately dipping axial plane (Fig. 4C).

In the northwestern areas the strike of S2 foliation ranges from NW-SE to WNW-ESE, dipping 45-50° both to the NE and the SW (Fig. 4). In the Nurra-Asinara island, within the phyllite complex, the S2 foliation is a spaced crenulation cleavage and is composed by white mica, chlorite and biotite. In paragneiss and micaschist it is defined by quartz, muscovite, biotite, chlorite, oxides, garnet and albite/oligoclase porphyroblasts and show a transition to a mylonitic foliation defined by ribbon quartz.

In the Badesi-Giagazzu area (SW Gallura region; Fig. 2), the S2 foliation shows a transition from an F2 axial plane foliation, in the SW area, to a S2 mylonitic foliation in the phyllonites level. In this area S2 strike from NE-SW to NW-SE and dips to the North.

In the Baronie zone, the S2 foliation shows NE-SW direction, which is moderately to steeply dipping toward the NW (Fig 4c).

In the all the northwestern and central area, the L2 object lineation (sensu Piazolo & Passchier, 2001 and reference therein) is represented both by grain and stretching lineations. It is well defined in the gneisses and quartzites and it trends N100-N170 (Fig. 4), parallel to F2 fold axes and plunges a few degrees toward the NW and the SE. In the Baronie zone it mostly trends E-W (Fig 4c).

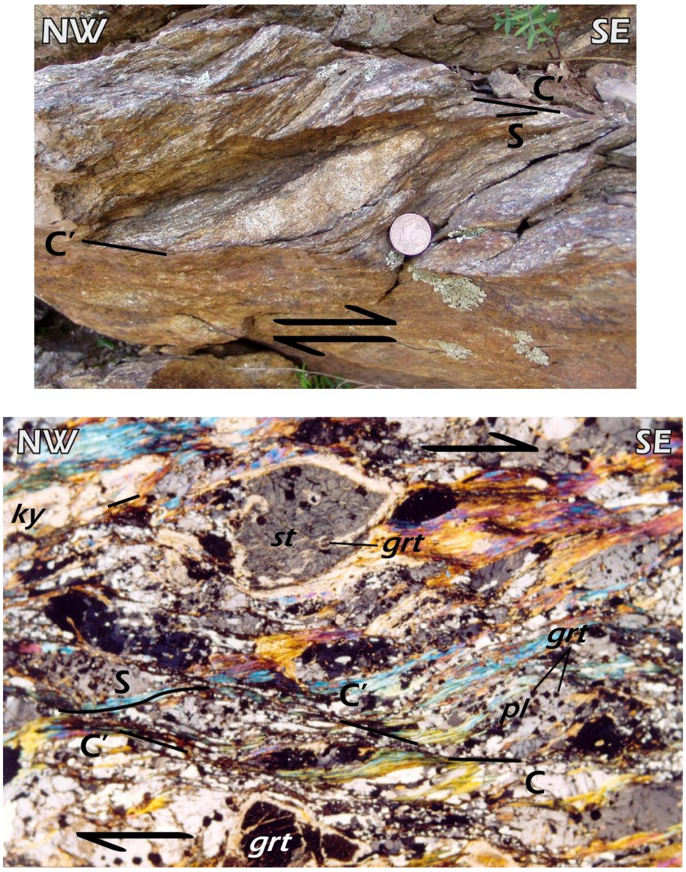
D2 shear zones

In fine-grained and porphyroblastic paragneiss, mantled porphyroclasts with sigmoidal and delta-type geometry, mica-fish, shear band cleavage and symmetric - structures have been observed in sections parallel to the lineation and orthogonal to the foliation plane.

The S2 mylonitic foliation is generally steeply dipping and has been folded by later folds so that it dips to the NE in the Baronie and Nurra section, whereas it dips to the SW in Anglona-Gallura and Asinara Island (Figs. 2 and 3). In the Nurra peninsula, between Isola dei Porri and Punta Falcone, sheath folds have been observed associated with top to the NW shear sense indicators. In Asinara Island up to the Cala d'Oliva orthogneiss, the D2 microstructures are poorly preserved because of a HT/LP overprint, and only locally micafish and shear band cleavage have been observed, confirming a top-to-the NW shear sense. In the Baronie and SW Gallura areas, mylonites with top-to-the SE sense of shear are the prominent D2 features.

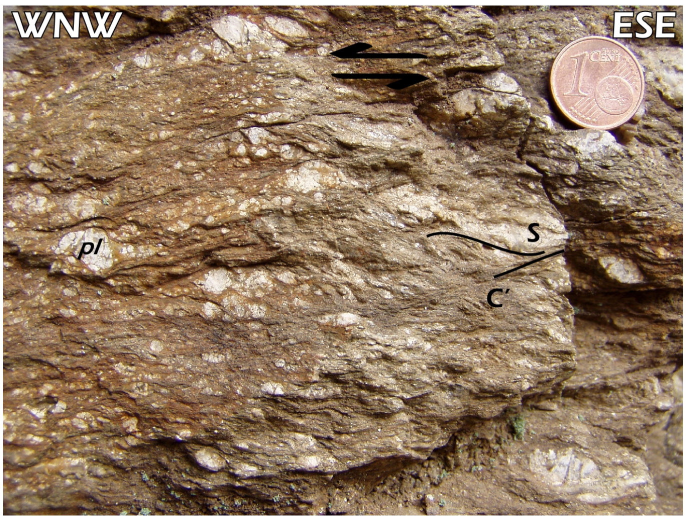
However, in the SW Gallura NNW-SSE striking shear zones with a sinistral sense (top-to-the NNW) of shear have been recently detected in the migmatites and migmatite gneisses of the High Grade Metamorphic Complex (Fig. 7) (Carosi et al., 2005). The relation between dextral and sinistral shear zone is not clear in the field but they point to a more complex shear zone evolution in this sector of the belt.

Figure 6. S-C' fabric in mylonites from SW Gallura



S-C' fabric in mylonites from SW Gallura (MGMC); dextral sense of shear: a) outcrop view; b) Shear deformation affects Barrovian mineral assemblage in the MGMC (CP, field of view nearly 2 mm).

Figure 7. Mylonites in the gneisses of the HGMC, SW Gallura.



S-C fabric and rotated porphyroclasts point to a sinistral sense of shear.

In the metatexitic complex (HGMC) in the Asinara Island, S2 foliation is steeply dipping and strikes between N160° and N180° (as indicated by a maximum in S2 plot within Fig. 4a), whereas in the central and northern Asinara, the mean direction is of N 110-120° (Fig. 4a). In the central area within the andalusite, sillimanite bearing micaschist complexes, the foliation becomes a mylonitic fabric defined by both ribbon quartz and microlithons indicating a top-to-the-SE sense of shear.

Vorticity analysis

Several authors showed that deformation within strongly deformed ductile rocks (considered as deformable continuum medium) could be approximated as a steady state flow system (Ramberg, 1975; Passchier, 1988; Simpson & De Paor, 1993). In this case deformation could be decomposed in two components: stretching rate (which includes rates of volume changes) and vorticity (instantaneous rotation rate). The relationship between the contributions of vorticity and stretching rate to the flow is referred to as the degree of non-coaxiality which can be expressed in terms of a vorticity numbers (Wm=0 in pure shear and Wm=1 in simple shear). Studies of the degree of non-coaxiality are commonly referred to as vorticity analyses and several examples exist in geological literature (Passchier, 1988; Wallis, 1995; Xypolias & Koukouvelas, 2001; Carosi & Palmeri, 2002; Holcombe & Little, 2001; Law et al., 2004).

In order to unravel the kinematic processes acting during the D2 deformation history vorticity analyses, using two different methods, have been carried out in the study areas.

Analyses have been performed on pre/syn-D2 plagioclase crystals and weakly rigidly deformed garnets from the porphyroblastic paragneisses and micaschists of the Low- Medium Grade Metamorphic Complex. The two methods (Passchier, 1988; Wallis, 1995; Holcombe & Little, 2001) are based on the theoretical mathematical consideration of Jeffery (1922), and on their relative applications to ductile rocks in 2-D and 3-D system made respectively by Gosh & Ramberg (1976) and Passchier (1988). The porphyroblasts are considered as rigid bodies immersed in a matrix behaving as a general Newtonian steady state and homogenous flow. The vorticity vector is assumed to be oriented perpendicular to the maximum and minimum principal axes of finite strain inducing a fabric with a monoclinic symmetry. As stated both by Ramberg (1975), and Passchier (1988), such methods find an ideal application within mylonitic shear zones but a good field context could be represented by strongly deformed and sheared ductile rocks where no strong structural anisotropies (C-S or C' fabric) are observed. As rocks system are not homogeneous and not necessary steady state and the boundaries conditions assumed by all these methods permit us to have only an approximate kinematic description of the deformation history. This implies that the measured flow parameters can only broadly indicate the type of tectonic regime acting during the study deformation event analysed and the reliability of these results strongly depend on the sample chosen to perform the vorticity measurement. The two different techniques used here differ principally on the different parameter type that can be measured in order to define the vorticity number. The two methods employ data collected on sections cut perpendicular to the foliation and parallel to the stretching lineations.

Method 1 (Gosh & Ramberg, 1976; Passchier, 1988)

This method is based on measuring the orientation of the long axis of rigid porphyroclasts with respect to the flow, the aspect ratio of rigid porphyroclasts rotating in an homogeneously deforming matrix and on finding a critical aspect ratio (Gosh & Ramberg, 1976; Passchier, 1988; Wallis, 1992) below which porphyroclasts continuously rotate until they achieve a stable orientation. This method assumes no mechanical interaction between the porphyroclasts and no slipping component along the matrix-clast contact. The flow system is considered to merge to the principal flow apophysis direction that could be well approximated by the recrystallized tails (Passchier, 1988) or by the flow matrix (Wallis, 1992). This method has been applied by Carosi & Palmeri, (2002), and Carosi et al., (2003) in the paragneiss and micaschist complexes cropping out respectively in the NE (Posada valley) and NW (Central-Northern Nurra) areas of axial zone using plagioclase and garnet porphyroclasts. The different vorticity values (Wm) estimates show the following results:

- in the central part of Nurra, Wm shows mean value of 0.60-0.70 whereas in the northern area the mean value are in between 0.30-0.50 (Fig. 8a)
- in the Posada valley, Wm varies from 0.52 to 0,73 (Fig. 8b)
- in the SW Gallura region, Wm confirm the non-coaxial regime (with vorticity values of 0.60-0.75) of

deformation with increasing simple shearing approaching the most strained areas (Carosi et al., 2005; Frassi, unpublished data).

Figure 8. Table of the mean vorticity number

Nurra region

Sample	Wm
6-02	0.3-0.5
7-02	0.3-0.5
7-02-1s	0.4-0.5
7-02-2s	0.4-0.5
86994	0.6-0.7
86997	0.6-0.7
12699	0.6-0.7

A.

Baronie region

Sample	Wm
C12	0.62-0.82
SC25	0.50-0.70
SC27	0.54-0.74
CAL1A	0.63-0.83
SC23	0.56-0.76
C15	0.42-0.62

B.

SW Gallura region

Sample	Wm
9541p	0.61-0.70
10941	0.69-0.76
161231	0.70-0.76
121232	0.68-0.71
81232	0.60-0.69
2453p	0.76-0.78
22547	0.76-0.81

C.

Table of the mean vorticity number (Wm) measured in the Northern sector of the Sardinian Basement according to Method 1 (see text). a: Nurra region; b: Baronie region and c: SW Gallura region.

Figure 9. Recumbent fold

Recumbent fold with shallowly dipping axial planes affecting the S2 mylonitic foliation in L-MGMC, SW Gallura.

Method 2 (Holcombe & Little, 2001)

This method uses porphyroblasts that have overgrown the external shear fabric and contain rotated relict of that fabric as an internal foliation. The internal foliation, represented by helicitic inclusion trails, should be a syn-tectonic foliation. In these rigid porphyroblasts the degree of rotation is measured directly from the orientation of the internal fabric relative to the external foliation. For steady-state monoclinic deformation the amount of internal rotation is a function of pure shear, simple shear, vorticity parameter Wm as well as on the initial orientation of the porphyroblast long axes relative to the shear direction. The vorticity number could be inferred from plots of the orientation of the long axes of each porphyroblast versus the orientation of its internal foliation. This method has been applied by Iacopini, (2005), in the NW transect within rotated garnet with helicitic inclusion continuously linked to the D2 principal schistosity. The Wm value ranges from south to north from 0.60-0.70 to 0.40-0.50, respectively, and it reasonably fits with other values obtained from Method 1.

Comparison of the data and discussion about the vorticity of the flow

The two methods indicate that in the eastern section, a pure shear component varies between 45-65% of the total recorded deformation, whereas in the NW regions the pure shear component ranges from 58% to 70 % (from south to north) of the recorded deformation. These broad estimations of the vorticity flow indicate that a significant amount of pure shear, varying in space, was linked to the D2 transpressive event. The development of closed to isoclinal folds with shortening increasing to the north is in agreement with vorticity estimation. Assuming a volume constant deformation, the amount of shortening induced by this event seems to have been accommodated and redistributed partly by a lateral horizontal extrusion as shown by the sub-horizontal stretching lineation observed all over northern Nurra and partly by a vertical extrusion in the northern HGMC (northern Asinara Island), where the stretching lineations are down dip (Carosi et al., 2004).

Late folding

The S2 foliation is affected by later open to tight folds (F3 and F4). F3 folds have sub-vertical axial planes and mainly NW-SE trending axes and a poorly developed axial plane crenulation cleavage. In the Anglona region the F3 fold axes show a NNW-SSE trend plunging 20-25° toward north (Fig. 4). These folds are mainly caused by a nearly N-S continuing compression during or soon after the main transpressive deformation event (D2). The F3 folds show an axial trend weakly deviating from the F2 axes folds and to L2 stretching lineation (Figs. 4a and b). Analogously to F2 folds the F3 folds are more developed towards the N-NE. In Asinara island, moving northward, up to the Stretti zone, the F2 fold axes are weakly affected by F3 folds that show a more regular trend of N90-100° plunging 35-40° to both the NE and SW (Fig. 4 a).

In both the Nurra-Asinara island and the Anglona-SW Gallura zones, the F3 folds produce interference pattern of both the type 0 and 3 (Ramsay & Huber, 1987). F4 have kink geometry with relatively steeply dipping axial plane. Their axes show variable trends and plunges in the different areas (Fig. 4). Rare recumbent folds have been observed with sub-horizontal or shallowly dipping axial planes (Fig. 9). They are associated with shallowly dipping normal shear planes with a broadly top-to-the south sense of shear, with development of cataclasites and asymmetric folds. Top-down to the south ductile/brittle to brittle shear zones and folds with horizontal axial plane accommodate the extensional collapse of the belt at upper crustal levels, after the thickening stage (Carmignani et al., 1994).

Relations between metamorphism and deformation

The relations between metamorphism and deformation show very similar characteristics all over the studied transects although detailed studies on metamorphism and P-T-t paths are available only for the NW (Carmignani et al., 1979; Franceschelli et al., 1990; Ricci, 1992; Simpson, 1999; Carosi & Oggiano, 2002; Carosi et al., 2004) and the NE areas (Franceschelli et al., 1982; 1989; Ricci, 1992; Carosi & Palmeri, 2002; Di Vincenzo et al., 2004; Ricci et al., 2004). In the Anglona and Gallura areas, relatively small amount of data are available (Oggiano & Di Pisa, 1992; Ricci, 1992). Where S2 is the most prominent fabric in the sheared rocks, the S1 foliation consists of aligned inclusion trails within grt, pl, st and ky Barrovian porphyroblasts, which define a foliation made of quartz, flakes of phengite, biotite, chlorite, albite, almandine garnets and ilmenite. Syn-D1 garnet within albite porphyroblasts shows an increase of MgO from core to rim and a CaO content lower with respect to the core of garnets that grow along S2 foliation in the same thin section (Franceschelli et al., 1982; Carosi & Palmeri, 2002). The latter suggests a growth zoning and an increase of pressure from the growth of syn-D1 garnet to syn-D2 garnet of the albite-garnet zone. The growth of the Barrovian index minerals was post-D1 deformation phase and partly during the early stages of the D2 deformation phase (Franceschelli et al., 1982, 1989; Carosi & Palmeri, 2002; Di Vincenzo et al., 2004; Ricci et al., 2004). Occasionally, plagioclase porhyroblasts have the S1 foliation truncated by the external S2 foliation. In other cases, plagioclase shows a sigmoidal internal foliation continuing to the S2 foliation, suggesting a syn-D2 growth of plagioclases during the D2 deformation. The S2 foliation is defined by white mica, biotite, chlorite, quartz, small size albite and/or oligoclase and garnet. White mica has a phengitic composition in the core of the large flakes of the garnet zone and is a muscovite near the rim or along the S2 foliation of the higher grade zones.

Millimeter-size garnet porphyroblasts (3-4 mm) in the garnet and staurolite zones are idioblastic almandines and show snow-ball inclusion trails related to their rotation during D2. Occasionally they show an increasing grain size of quartz inclusions from core to rim, suggesting a temperature increase during their growth. This is chemically evidenced by the increase of XMg and the decrease of XMn from core to rim which is indicative of a prograde zoning (Tracy, 1983; Franceschelli et al., 1982; Carosi & Palmeri, 2002). The decrease of the XCa content from core to rim for the same syn-D2 garnet is indicative of a growth during decompression (Crawford, 1977; Carswell, 2000; Carosi & Palmeri, 2002; Ricci et al., 2004). Small size almandines are homogeneous but a decrease of MgO and an increase of MnO detected at the extreme edges indicate a reverse zoning (Grant and Veiblen, 1971). The sense of rotation of asymmetric inclusion trails on plagioclases and rotated garnets is the same as deduced from shear band cleavages and mica fish. Staurolite and kyanite porphyroblasts show evidence of pre- and, rarely, syn-D2 growth (Fig. 6b). They contain inclusions of rounded idioblastic prograde almandines, biotite and ilmenite.

Some grt, st and ky porphyroblasts are fractured, stretched and show asymmetric strain fringes, made of quartz, biotite and chlorite. The S2 schistosty in the staurolite and kyanite zones is marked by quartz, oligoclase, muscovite, biotite, and almandines (Fig. 6b) showing an increase of MnO and a decrease of MgO at the rim indicative of the reverse zoning. Chlorite is also present but is interpreted to be as a secondary phase.

Relations between metamorphism and deformation show that the Barrovian index minerals (bt, grt, st and ky) grew often as porhyroblasts pre- or during the early stages of D2 deformation. Matrix minerals are aligned along the S2 foliation and involved a diachronous development of the main metamorphic equilibria in the different metamorphic zones (Franceschelli et al, 1989; Ricci et al., 2004).

The relations between deformation and metamorphism and petrological data in the Nurra transect provide similar results as in the Baronie area (Carmignani et al., 1979; Franceschelli et al., 1990). The metamorphic peak was attained post-D1 and pre- up to syn D2. Syn-D1 phengitic white micas indicate higher pressures (nearly 7-8 Kbar) with respect to the D2 deformation phase

On the other hand the P-T-t evolution reported in Franceschelli et al. (1989), Carosi & Palmeri (2002), Di Vincenzo et al., (2004) and Ricci et al., (2004 with references therein), clearly shows that D2 deformation developed after metamorphic peak (pre to syn-D2) in a regime of decreasing pressure all over the Barrovian metamorphic zones.

Timing of transpressional deformation

Until recently, the age of wrench or transpressional tectonics in Sardinia was not very well constrained. Undeformed granitoids, cross-cutting the main structures, were dated at 290-310 Ma (Del Moro et al., 1975) constraining only the upper limit of the Variscan deformation events. This age is also confirmed from Permian basins that sealed the Variscan deformed complexes at 300 Ma. The lower limit age was deduced by the consideration that D2 deformation started during lower amphibolite facies metamorphism dated by the Rb/Sr whole-rock at 344±7 Ma in migmatites that underwent amphibolite facies metamorphism (Ferrara et al., 1978) and 330 Ma (Palmeri et al., 1997).

In situ 40Ar/39Ar laserprobe analysis on white mica allowed Di Vincenzo et al., (2004), to constrain the age of S1 and S2 foliation in the grt+bt zone of the Baronie area and to better constrain the age of collisional (D1) and transpressional (D2) deformation phases. Using microstructural relations, microchemical composition and age they have characterized a deformed celadonite-rich mica flakes that define the D1 phase (both inside garnet and plagioclase porphyroblasts and in the core of large white micas in the matrix) and a celadonite-poor white mica aligned along the S2 foliation. The resulting ages span from 345 Ma to 300-310 Ma from D1 to D2 (Di Vincenzo et al., 2004). D1 ages concentrate in the 330-340 Ma interval, whereas D2 ages have been constrained between 315 and 320 Ma (Di Vincenzo et al., 2004). Moreover, these results do not confirm the hypothesis proposed by Helbing & Tiepolo, (2005), regarding the existence of a pre-Variscan metamorphic basement represented by the micaschists and gneisses of the low- to medium grade metamorphic complex of Baronie.

Discussion

D1 deformation in the southern portions of the studied transects testifies to a thickening stage characterized by isoclinal folds overturned to the SW and ductile to brittle shear zones with the same sense of shear. D1 phase is responsible of the nappe stacking in central and southern Sardinia (Carmignani & Pertusati, 1979; Carmignani et al., 1982, 1994; Carosi et al., 2004).

F1 folds and a north dipping S1 foliation are the prominent features to the south of the studied areas. Recent studies on top-to-the SW ductile/brittle - shear zones in southern Nurra suggest that they accomplished exhumation of the sheared low grade metamorphic rock causing their decompression at the end of the D1 deformation phase (Montomoli, 2003).

D1 fabric has been progressively overprinted by a D2 deformation with strain increasing to the north. Vorticity analysis indicates that D2 deformation is characterized by both simple shear and pure shear acting contemporaneously in a transpressional regime.

D2 non-coaxial regional deformation in northern Sardinia is related to a crustal-scale oblique shear zone (Fig. 3). Shear sense indicators indicate a top-to northwest sense of shear in the western area whereas a top-to-the SE sense of shear in the eastern zone. In both cases they show a transport direction at a high angle with respect to the SW direction of tectonic transport during D1 (Carosi & Oggiano, 2002; Carosi & Palmeri, 2002). The attitude of the L2 stretching lineation implies both a dip-slip movement and a strike-slip movement.

A change in the attitude of the L2 stretching lineations is clearly observable in Asinara Island. In the southern and central portion of the island, L2 stretching lineation trends N090-N110, parallel to F2 fold axes and plunges a few degrees toward the NW and the SE. The L2 lineation switches from sub-horizontal, near Punta Gruzitta, to down-dip in the northern part of the island, whereas the S2 foliation maintains the same attitude. This geometry plays a key role for constraining the kinematic history of the D2 deformation phase.

This complex deformation pattern can be explained in two ways (Carosi et al., 2003; Carosi et al., 2004; Iacopini, 2005):

- D2 is a steady-state non coaxial deformation in which pure shear is partitioned in space and increases to the north. This could cause a rotation of the L2 stretching lineation from sub-horizontal to down-dip on the S2 foliation according to the transpressional model proposed by Tikoff & Teyssier, (1994), and Tikoff & Green, (1997).
- The switching of the L2 stretching lineation can be caused by a late-D2 ductile "thrusting" of the High Grade Metamorphic Complex onto the Low to Medium Grade Metamorphic Complex with a top-to-the SW sense of shear.

The growth of Barrovian porphyroblasts took place either post-D1 or during the early stages of the D2 shear deformation (Franceschelli et al, 1982, 1989, 1990; Ricci et al., 2004 with references). Variable retrogressions have been recognized along the S2 foliation, so the early stages of its development during medium-grade conditions represent preserved relics. The sense of rotation of D2 porphyroblasts is consistent with the other D2 kinematic indicators.

The metamorphic peak is pre- and partly sin- D2 (Franceschelli et al. 1982, 1989, 1990) Evidences of D1 deformation is found within albite, oligoclase, staurolite and kyanite porphyroblasts. The composition and zoning of small-size garnets, together with the phengitic white mica included in the albites, indicates that D1 deformation was characterized by both increasing temperature and pressure. Mineralogical associations and composition of the main phases indicate that the D2 phase developed in a regime of increasing temperature and decreasing of pressure during the early stage. The latest D2 stages were characterized by decreasing both pressure and temperature (Carosi & Palmeri, 2002; Di Vincenzo et al., 2004; Ricci et al., 2004).

The increase of the metamorphic grade in northern Sardinia in a short distance is not consistent with Barrovian metamorphic gradients. D2 transpressional deformation was responsible for the observed condensed isograds (Franceschelli et al., 1982, 1989) by their stretching and shearing (Carosi & Palmeri, 2002).

Geometric, kinematic and petrological data suggest that the D2 deformation phase drove the decompression of medium-pressure rocks in a transpressive tectonic setting at 315-320 Ma (Carosi & Palmeri, 2002; Di Vincenzo et al., 2004). The crustal-scale transpression controlled the exhumation of medium pressure rocks up to upper crustal level, before undergoing the post-collisional extensional collapse at nearly 300 Ma (Carmignani et al, 1993).

Orogen-parallel tectonic transport in the southern Variscan belt

The rotation of the direction of tectonic transport by nearly 90° from D1 to D2 has been observed by Carosi &

Palmeri, (2002), in northeastern Sardinia. A similar tectonic behaviour has been described both in the northwest (Carosi & Oggiano, 2002) and in the south of Sardinia (Lünuburg & Lebit, 1998; Conti et al., 2001), suggesting that such kinematic events are clearly a common feature in the whole island.

The regional scale change in the direction of the tectonic transport in Sardinia can be interpreted as the expression of a change from frontal to oblique collision from D1 to D2. However, tectonic constraints for plate tectonic reconstruction at the end of the Paleozoic are too limited for the Corsica-Sardinia block and surrounding terrains and the tectonic scenario is still debated and poorly constrained (presence of a suture, numbers of oceans, plate involved) to test this hypothesis.

An alternative explanation, with the limitation for the lacking of detailed correlations between Sardinia and the other portions of the southern Variscan belt, has been proposed by Carosi & Palmeri, (2002), considering the position of the Corsica-Sardinia block before the Miocene counterclockwise rotation (Barnolas & Chiron, 1996 with references therein). According to Arthaud & Matte, (1977), Ricci & Sabatini, (1978), Matte, (1986) and Barnolas and Chiron, (1996), the northernmost part of Sardinia was connected to southeastern France during the upper Paleozoic. In this paleo-position, the activity of a transpressional shear zone in northern Sardinia could fit with the indentation between Armorica and Gondwana as proposed by Matte & Ribeiro, (1975), Brun & Burg, (1982), Ribeiro et al. (1995) and Dias & Ribeiro (1995). During the Carboniferous, a dextral transpression began to predominate in the Armorican branch of the arc, while in the southern branch southward thrusting predominated. Moreover, the indentation of the Cantabrian block (Dias & Ribeiro, 1995), during and after the thickening stage of the belt, caused tightening of the arc inducing an increasing strike-slip component of deformation in the limbs. This could have led to the generation of ductile transpressional shear zones while the crust was thermally weakened, because of the high temperatures after the D1 thickening stage. In some sectors of the Variscan belt, such as northern Sardinia, transpressional deformation affected medium pressure rocks and their exhumation history.

The change in tectonic transport by nearly 90°, from perpendicular to parallel to the belt, deeply influenced the fate of large portions of the thickened crust before its gravitational collapse.

The change from frontal collision to orogen-parallel displacement may be a common feature in the orogenic belts, in which indentation tectonics occurs or there is a change of relative kinematic of the plates. This change may play an important role in the equilibration, exhumation and the tectonic history of metamorphic rocks in orogenic belts.

Ricci et al., (2004), emphasized the higher exhumation rate of migmatites with respect to lower-grade rocks during the first stages of exhumation that can be attributed both to a decreasing strength of the lower crust and to the prevalent pure shear component in the developing transpressive regime.

The proposed tectonic evolution is able to explain an initial high rate of exhumation with dip-slip movement for the deepest rocks followed by slow exhumation rate up to higher structural levels with a major component of horizontal movement compatible with Barrovian metamorphism during decreasing pressure.

Conclusion

The studied sections across the metamorphic basement of northern Sardinia show evidence for the tectonic behaviour of a large part of an orogen, with a change from crustal thickening with tectonic transport perpendicular to the belt to a transpression with displacement parallel to the belt. D2 deformation in NE Sardinia has been related to the activity of a crustal-scale transpressional deformation, causing slow rate of exhumation and telescoped isograds in the low- to medium- grade metamorphic rocks.

Hitherto the exhumation of metamorphic rocks in the Variscan belt has been largely attributed to extensional tectonics and to the development of metamorphic core complexes. The D2 tectonic evolution of northern Sardinia, together with petrographic and geochronological data, suggests that metamorphic rocks were exhumed in different ways in different sectors of the Variscan belt. The exhumation of the metamorphic rocks of northern Sardinia during D2 was controlled by the activity of a crustal-scale transpressional deformation zone testifying to a progressive change from frontal (D1: 330-340 Ma;) to oblique convergence at 315-320 Ma (D2).

The exhumation started early during collision and during the development of amphibolite facies metamorphism. In the early stages, during frontal collision, exhumation was rapid and the velocity of exhumation decreased as collision became more and more oblique and the displacement of the metamorphic rocks switched in predominant horizontal component of movement.

The overall change of the displacement direction affecting a large sector of a belt, such as the exposed section in Sardinia, has been tentatively related to the progressive development of indentation tectonics of the Ibero-Armorican arc during the Upper Paleozoic.

The Variscan basement in Sardinia clearly shows how displacement parallel to the belt may deeply affect the tectono-metamorphic evolution of wide portions of an orogen.

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