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Variscan metamorphism in Sardinia, Italy: review and discussion

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Abstract: The Sardinian segment of the Variscan chain is divided into four tectono-metamorphic zones from NE to SW: 1) Inner zone, with medium- to high-grade metamorphic rocks and migmatites; 2) Internal Nappe zone, with low- to medium-grade rocks; 3) External Nappe zone, with low-grade metamorphic rocks; 4) External zone, with very low- to low-grade rocks. Metabasite lenses, with eclogite and granulite relics, occur in the migmatites.

Pre-Variscan ages are attributed to the E-W trending folds below the Sardic unconformity (SW Sardinia). Five Variscan deformation phases have been distinguished, the principal ones being: D_1, compressional, with SW-vergent folds and S_1 schistosity; D_2, extensional, with S_2 schistosity transposing S_1. Later, a composite network of post-D_3, syn-D_4 shear zones was active from an older HT/LP shear event to a younger MT-LT/LP one. The late D_5 phase produced kilometric flexures with an axis parallel to the orogenic trend.

For Inner zone sequences, a complete prograde sequence has been recognised, from the chlorite zone to the sillimanite + K-feldspar one. The Nappe zone shows decreasing metamorphic grade from bottom to top of the Nappe pile, as well as in each single unit. The Variscan P-T-t path is the clockwise loop typical of continental collision. For the Barrovian stage, thermobarometers yielded a P_{peak} of ≈1.2 GPa (end of D_1) and T_{peak} of ≈750°C (early D_2). For HP/HT metamorphic relics, calculations indicated: eclogite stage, T ≈ 550°-700°C, P ≈ 1.3-1.7 GPa; granulite stage, T ≈ 650°-900°C, P ≈ 0.8-1.2 GPa; amphibolite stage: T ≈ 550°-650°C, P ≈ 0.3-0.7 GPa. Pre-Variscan igneous rocks consist of Lower to Middle Paleozoic metabasites with MORB or WPB affinity and Middle Ordovician calc-alkaline intrusive and effusive rocks.

The Corsica-Sardinia microplate belongs to the southern passive margin of the Hun Superterrane of the Swiss authors. The beginning of Variscan continental collision is marked by a probable eclogite-producing HP event. Ages of 355-335 Ma and 335-320 Ma for the first two Variscan metamorphic phases and 310-280 Ma and 290-280Ma for late-post orogenic Variscan magmatism in Sardinia match those of the same events in the northern Hun Superterrane active leading margin.

The Variscan Corsica-Sardinia batholith is made up of late-orogenic calc-alkaline intrusive rocks and post orogenic leucogranites. The greater volumes (surface area≈ 12000 Km^2) and long gestation times of the Corsica-Sardinia batholith as compared to those of the northern Hun Superterrane batholiths is attributed to the collision between the southern Hun Superterrane margin, including Corsica-Sardinia, and the huge Gondwana Supercontinent.

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Introduction

After pioneering work by Charrier (1957) on metamorphism in northern Sardinia, in the last 35 years, several new, detailed papers have made great contributions to knowledge of the tectono-metamorphic evolution of Sardinia during Variscan orogeny (among others: Arthaud, 1970; Ricci, 1972; Di Simplicio et al., 1974; Miller et al., 1976; Ghezzo and Ricci, 1977; Conti et al., 1978; Carmignani et al., 1979, 1982a,b; Ghezzo et al., 1979, 1982; Franceschelli et al., 1982a,b; Tucci, 1983; Elter et al., 1986; Oggiano and Di Pisa, 1988; Franceschelli et al., 1989, 1990; Carosi et al., 1990; Carosi and Pertusati, 1990; Elter et al., 1990; Franceschelli et al., 1991; Franceschelli et al., 1992; Musumeci, 1992; Ricci, 1992; Carmignani et al., 1994; Eltrudis and Franceschelli, 1995; Eltrudis et al., 1995; Franceschelli et al., 1998; Elter et al., 1999; Carmignani et al., 2001; Cruciani et al., 2001; Franceschelli et al., 2002; Cortesogno et al., 2004; Di Vincenzo et al., 2004; Elter et al., 2004; Palmeri et al., 2004; Giacomini et al., 2005a). This paper attempts to carefully review all contributions on Variscan metamorphism in Sardinia and insert the complex and still highly debatable tectono-metamorphic evolution of the Corsica-Sardinia microplate into the general framework of the Variscan chain.

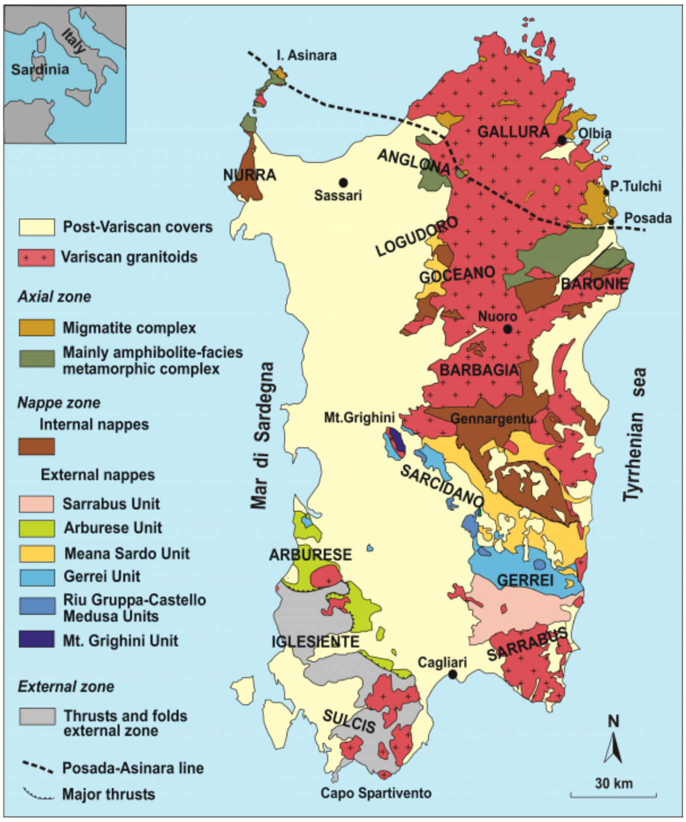
Geological and Tectonic Framework

Regional tectonic zonation

The metamorphic basement of Sardinia is a fragment of the southern European Variscan Belt and represents the continuation of the Maures, Montagne Noire and Pyrénées, France (Arthaud and Matte, 1966; Westphal et al., 1976; Arthaud and Matte, 1977; Ricci and Sabatini, 1978).

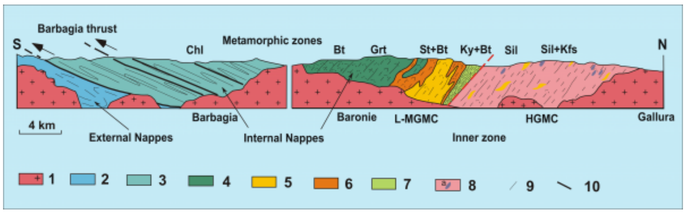
The Sardinian Variscan Belt (Figure 1 , 2) is generally divided into four tectono-metamorphic zones with decreasing grade from NE to SW: 1) Inner zone or "Axial zone", with medium- to high-grade metamorphic rocks and migmatites (northern Sardinia) of Precambrian (?) to lower Paleozoic age; 2) Internal Nappe zone (central-northern Sardinia), with crystalline units consisting of low- to medium-grade metamorphic rocks; 3) External Nappe zone (central-southern Sardinia), with low-grade metamorphic rocks; 4) External zone (southern Sardinia, Iglesiente, Sulcis), with very low- to low-grade metamorphic rocks. The Internal and External Nappe zones consist of Paleozoic metasedimentary and metavolcanic successions.

Figure 1. Tectonic sketch map



Tectonic sketch map of the Variscan Belt in Sardinia (after Carmignani et al., 2001, modified).

Figure 2. Geological sketch cross-section



Geological sketch cross-section throughout the Variscan basement of Sardinia from Barbagia to the Gallura region (after Carmignani et al., 1989 and Carosi and Palmeri, 2002, modified). 1: Variscan granitoids; 2: External Nappes; 3: Internal Nappes; 4: phyllites, micaschist and gneiss; 5: orthogneiss, 6: augen gneiss; 7: mainly mylonite and amphibolite lenses; 8: migmatite with lenses of metabasite with eclogite facies relics; 9: regional schistosity; 10: thrusts. The regional distribution of some index minerals in pelitic rocks from Barbagia to Gallura is also shown. Mineral abbreviations according to Kretz (1983).

The Inner zone or "Axial zone" includes two metamorphic complexes: the High Grade Metamorphic Complex (HGMC) of North Sardinia, consisting of gneisses and migmatites with a metamorphic grade reaching the sillimanite + K-feldspar isograd. Bodies and lenses of mainly Ordovician granitic-granodioritic orthogneisses (Di Simplicio et al., 1974; Ferrara et al., 1978), mafic and ultramafic metamorphic rocks showing eclogite to granulite relics and calcsilicate nodules are frequently embedded within the HGMC.

The second metamorphic complex in the axial zone is made up of low- to medium-grade metamorphic rocks (L-MGMC), largely outcropping adjacent to the southern side of the Posada-Asinara Line (Figure 1 , 2), in central-southern Asinara, northern Nurra, Anglona and the northern Baronie regions. The Posada-Asinara line is considered by Cappelli et al. (1992) and Carmignani et al. (2001) as a "South Hercynian Suture Zone" between the Armorica and Gondwana plate margin.

The Internal Nappe zone includes metasandstones, quartzites and phyllites outcropping in the Monti del Gennargentu ("Postgotlandiano "Auct. p.p"), Goceano and Nurra. Near the area of contact with metavolcanic rocks, metaconglomerates outcrop in Goceano, Barbagia and Baronie. Micaschists and paragneisses of the internal Units outcrop in Barbagia and Baronie. Internal Nappes also include acidic metavolcanic rocks ("Porfiroidi"), with related metarkoses and microcline-bearing quartzites, as well as mafic and intermediate (andesite) metavolcanic rocks of probable Middle-Ordovician age. As regards the Upper Ordovician to Lower Carboniferous sequences, internal nappes are characterised by Siluro-Devonian graphite-rich phyllites, marble and calcschist intercalations and metadolerites and metagabbros attributed to the Lower Carboniferous. Sequences in the southern part of Nurra and Baronie can easily be correlated with the Cambro-Ordovician successions of the External zone.

The External Nappe zone is characterised by Middle Ordovician magmatism, which produced the thick metavolcanic sequences occurring above Cambro-Ordovician metapsammites. The Variscan sedimentary cycle started at the beginning of the Upper Ordovician (Caradoc), with marine transgression on a previously-destroyed volcanic arc. Silurian black phyllites show a more widespread, homogeneous distribution, indicative of an extended pelagic environment, while Devonian platform carbonate gradually decreases in thickness towards the internal areas of the chain. The deepest units of the External Nappe zone are the Castello Medusa, Riu Gruppa and Monte Grighini Units, respectively reaching the biotite and medium-grade zones. The Monte Grighini represents the core of greatest tectonic culmination in the Nappe zones: the Flumendosa Antiform. On its southern side, from bottom to top, are the Gerrei and Genn'Argiolas units; on the northern side, the Nappe sequence, from bottom to top, is comprised of the Gerrei and Meana Sardo Units and the Barbagia Low-Grade Metamorphic Complex.

The External zone corresponds to the oldest, southwestern part of Sardinia. In southern Sulcis, the Cambro-Ordovician sequence of northern Sulcis-Iglesiente overlies a metamorphic complex including, from bottom to top, Monte Filau orthogneisses, Monte Settiballas micaschists and the Bithia formation (Fm). The latter is made up of alternating metapsammites and metapelites, with metabasite and carbonate intercalations (Junker and Schneider, 1980, 1983). Some authors (e.g. Minzoni, 1981) consider the Settiballas micaschists part of a Precambrian basement. A Precambrian to Early Cambrian age is also attributed by Naud (1979) to the Bithia Fm, owing to its position below the Early Cambrian Nebida group (Junker and Schneider, 1980, 1983; Cocozza, 1980) and to the appearance of schistosity unknown in overlying Cambrian formations (Carosi et al., 1995). Monte Filau orthogneisses, on the contrary, are lower Ordovician granitoid (Delaperrière and Lancelot, 1989) rocks intruded within Settiballas micaschists.

The Cambrian-Early Ordovician sequence is made up, from bottom to top, of the Nebida Group, Gonnesa Group and Iglesias Group (Pillola, 1991). The Nebida Group is a deltaic terrigenous sequence (metasandstone and metasiltite), with rare interlayered lagonal oolitic limestones containing Early Cambrian Archeocyaths.

The Gonnesa Group is mainly composed of metadolostones at the base and grey stratified metalimestones at the top. The Gonnesa Group is interpreted as a thick platform carbonate succession (arid tidal flat environment) (Boni and Cocozza, 1978; Boni and Gandin, 1980; Barca et al., 1987) showing gradual heteropic transition to nodular limestones (extensional deep basins).

The base of the Iglesias Group is made up of marly nodular crystalline limestone (Campo Pisano Fm) and the top of phyllites and metasiltites interbedded with minor massive fine-grained metasandstones (Cabitza Fm). This framework and the lack of Neoproterozoic to Early Ordovician calc-alkaline rocks indicates that Sardinia belonged to a passive continental margin from the Late Neoproterozoic to the Early Ordovician.

Cambrian-Early Ordovician sequences are separated from the overlying Upper Ordovician-lower Carboniferous sedimentary cycle by sharp angular unconformity (Sardic phase) and overlain by a transgressive (Martini et al., 1991) polygenic metaconglomerate, the Puddinga Auct.

Variscan Deformations

Northern Sardinia

A first systematic attempt at reconstructing the deformation phases of the Variscan chain in Northern Sardinia was published by Carmignani et al. (1979, 1982a) and Elter et al. (1986). Three main deformation phases were distinguished: D_1, D_2 and D_3. The first phase, recognisable in the southern part of northern Sardinia, is characterised by southwards recumbent kilometre-sized isoclinal folds with penetrative axial plane schistosity S_1. The second deformation phase D_2 produced folds with E-W trending axes. In the southern part of northern Sardinia, D_2 folds are open and characterised by strain-slip S_2 schistosity. Going northwards, D_2 folds gradually become more and more closed, finally turning into northwards recumbent tight isoclinal folds with a southwards dipping penetrative S_2 schistosity that almost completely obliterated older D_1 structural features. The third deformation phase D_3 generated chevron, box or kink folds, with N-S trending axes locally associated with strain-slip or fracture schistosity.

A slightly more complex framework was proposed for the Baronie region (Posada area) by Elter in Oggiano and Di Pisa (1992, 155-157). The main difference is the addition of a D_4 and D_5 deformation phase. The D_4 phase generated a mylonitic complex characterised by continuous, gradual transition from an S-C structure to an ultramylonitic one (Elter, 1987). The late D_5 phase, deforming all previous structures in Elter's reconstruction, produced, according to Helbing (2003), a kilometre-sized flexure with a sub-horizontal axis parallel to the orogenic trend.

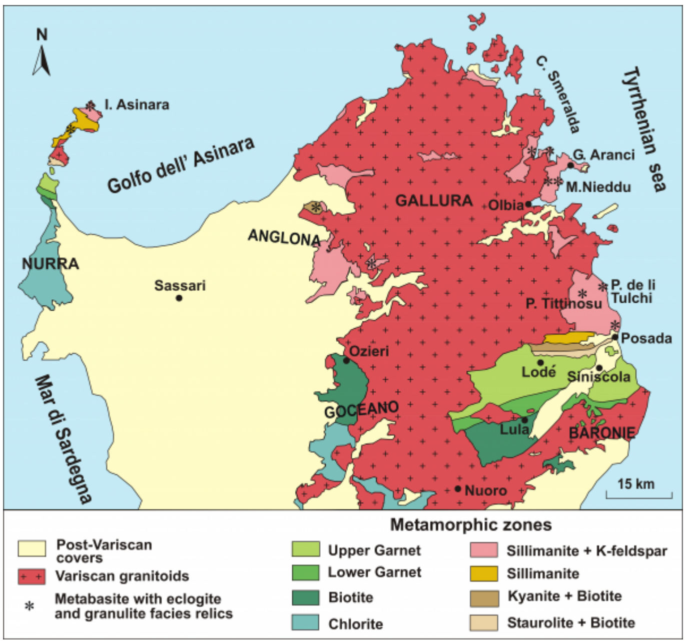
Several papers have been published on the Variscan deformations in Anglona, western Gallura, Nurra and Asinara: Elter et al. (1990); Franceschelli et al. (1990); Oggiano and Di Pisa (1992); Carmignani et al. (1992, 1994, 2001); Carosi and Oggiano (2002); Carosi et al. (2004).

Central and southern Sardinia

The tectonic setting of central and southern Sardinia has been discussed by Carmignani and Pertusati (1977), Dessau et al. (1983), Carosi et al. (1990), Carosi and Pertusati (1990), Carosi et al. (1992a,b), Musumeci (1992), Carmignani et al. (1994) and Conti et al. (1998, 1999).

Conti et al. (2001) have recently tried to reconstruct the tectonometamorphic evolution of the Sardinian Variscan segment as a whole, correlating deformation events in the central, southeastern and southwestern parts of Sardinia. Three deformation phases (D_1, D_2 and D_3) were identified by Conti et al. (2001, Figure 3). D_1 deformation can be subdivided into four different phases: Gerrei, Meana, Sarrabus and Flumendosa. The Gerrei phase is characterised by mylonitic deformation confined to narrow zones beneath the main thrust planes, kilometre-sized southward vergent isoclinal folds and metamorphic recrystallisation under greenschist facies conditions. The result is the generation of steeply-dipping S_a foliation parallel to S_0 stratigraphic contact. However, S_a/S_0 surfaces are cut by a flat-lying S_b crenulation cleavage parallel to the Meana Sardo thrust surface, more and more penetrating as it approaches the Meana Sardo thrust. S_b was considered to be related to the Meana Sardo phase leading to the S-directed emplacement of the Meana Sardo and Barbagia Units.

Figure 3. Metamorphic zonation



Metamorphic zonation in metapelite and metapsammite from the Variscan basement of northern Sardinia, (from Franceschelli et al., 1982b, modified). The locations of the main outcrops of metabasite with eclogite and granulite facies relics are also shown. See text for explanation and data sources.

The Sarrabus phase marks a drastic change in tectonic transport from S-directed to W-directed, identified by Conti et al. (2001) within the Arburese and Sarrabus Units, respectively west and east of the Campidano graben. This means, according to Conti et al. (2001), a 90° rotation of the tectonic trasport direction, taking place during the Early Carboniferous.

The final phase, named by Conti et al. (2001) the Flumendosa phase, brought a renewed N-S shortening phase, characterised by E-W to ESE-WNW trending upright folds, and generated the Flumendosa and Gennargentu antiforms and the Barbagia synform.

The D_2 deformation, named the Riu Gruppa phase, took place during exhumation and produced normal faulting, NW-SE trending folds and crenulation cleavage. The last D_3 deformation is revealed by NE-SW folds and crenulation cleavage.

As regards SW Sardinia, an initial tectonic framework was proposed by Arthaud (1963), who distinguished four deformation phases in the Cambro-Ordovician sequences of SW Sardinia:

1) Sardic phase producing E-W trending open folds; 2) first Variscan phase generating E-W trending folds; 3) second and main Variscan phase yielding N-S trending folds; 4) third Variscan phase producing weak deformation structures with variable orientation.

Conti et al. (2001) have recently questioned the pre-Variscan age of the E-W trending open folds on the basis of a careful field review of existing data. The authors note that contact between the Middle Ordovician conglomerate (Puddinga Auct.) and the underlying Iglesias syncline and Gonnesa anticline consists of a steep reverse fault surface of Variscan age and cannot be considered a Caledonian unconformity (Sardic phase) between pre-Variscan and Variscan sediments.

Shear zones

In the evolution of the Sardinian Variscan basement, a key role was played by a composite network of shear zones (Elter et al., 1986,1990, 1999; Elter and Ghezzo, 1995) displaying complex evolution from early HT/LP stages to late MT-LT/LP conditions associated with changes in the sense of movement. Two shearing events were distinguished: an early shear event (ESE) and a late shear event (LSE). Five shear zones (SZ) associated with synkinematic HT-LP metamorphism were active during the ESE: 1) Barrabisa; 2) Golfo Aranci; 3) Porto Ottiolu; 4) Siniscola-Mamone; 5) Posada Valley. The LSE gave rise to two subsystems of strike-slip faults, the first associated with synkinematic intrusions and the second with synkinematic retrograde metamorphism. LSE shear zones from south to north are: 1) Monte Grighini; 2) Ottana-Monte Senes; 3) Barrabisa II; 4) Posada Valley II; 5) Anglona; 6) Porto Ottiolu II; 7) Golfo Aranci II.

As regards metamorphic grade, the Golfo Aranci shear zone is the only one showing structural features related to an extensional phase with amphibolite facies conditions (Elter and Ghezzo, 1995; Elter et al., 1999), while other shear zones are characterised by cataclastic to mylonitic products, generated for the most part under greenschist facies conditions.

Variscan magmatism

Three main magmatic cycles have been distinguished in the Variscan basement of Sardinia by Memmi et al. (1983), Di Pisa et al. (1992) and Franceschelli et al. (2003). The first one led to the emplacement of basaltic to rhyolitic calc-alkaline volcanic rocks during the Middle Ordovician (pre-Caradoc cycle). Metamorphosed porphyritic acid rocks of this cycle are known as "porphyroids" and derived from original rhyolitic, rhyodacitic to dacitic lava flows and domes or ignimbrites, widespread in Sarrabus, Gerrei, Sarcidano, Goceano and Nurra. A radiometric age of 474 ± 13 Ma (Table 1) has ben obtained for the Lula porphyroids by Helbing and Tiepolo (2005).

These volcanic products are considered late orogenic calc-alkaline crustal rocks. In some regions (Sarcidano, Goceano), the rocks of the pre-Caradoc cycle show a clear andesitic (Serra Tonnai Fm) to sub-alkaline basaltic affinity. The first cycle lasted from the Arenig to Caradoc (Carosi et al., 1987). Ordovician acidic plutonic and basic volcanic rocks, at present orthogneiss or amphibolite, occur in the northern Sardinia.

A second cycle attributed to the Silurian by Memmi et al. (1983) and to the Caradoc-Ashgill by Di Pisa et al. (1992) gave rise to alkaline within-plate basalts (Beccaluva et al., 1981) emplaced as subvolcanic body sills and dykes, in Goceano, Gerrei, Iglesiente-Sulcis and Sarcidano.

The third cycle, considered by Di Pisa et al. (1992) and Garbarino et al. (2005) of Devonian (?) to Carboniferous age, produced within-plate alkali basalts now outcropping as metavolcanic rocks embedded in the Palaeozoic sequences of Sulcis, Sarrabus, and metagabbbros and metadolerites intruded into black metapelites at Nurra. Franceschelli et al. (2003) demonstrated that Nurra metagabbros emplaced during the extensional phase prior to Variscan continent-continent collision originated from an ocean island alkali basalt-like asthenospheric mantle enriched with incompatible elements but devoid of a crustal component.

A Middle Devonian episode (387 ± 2 Ma) with calc-alkaline acid products of orogenic affinity was reported by Garbarino et al. (2005) in southern Sulcis. However, further geochemical and geochronological data need to prove the existence of this episode.

During the Variscan tectono-metamorphic event, widespread plutonic activity (Di Vincenzo et al., 1994; Carmignani and Rossi, 1999) led to the formation of one of the largest batholiths in SW Europe. The intrusive sequence consists of: 1) an earlier syn-tectonic Mg-K calc-alkaline association (northwestern Corsica) emplaced at ~330-345 Ma; 2) a late- to post-tectonic high-K calc-alkaline association cropping out in Corsica and Sardinia (Rossi and Cocherie, 1991) emplaced from 310 to 280 Ma. 3) Peraluminous association. The high-K calc-alkaline late-tectonic intrusions range in composition from gabbro and diorite to leucogranite, whereas post-tectonic intrusions consist of leucogranites. According to Tommasini et al. (1995), the source of late orogenic Variscan gabbros is a subcontinental mantle contaminated by 5% of material subducted 450 Ma ago during a previous Ordovician cycle. The high-K calc-alkaline association also includes highly-peraluminous granitoids (Di Vincenzo et al., 1994 and references therein). The latter mainly consist of granodiorites and monzogranites, with minor tonalites and leucogranites. During the Upper Carboniferous-Permian, ignimbrites, lava flows and subvolcanic bodies were emplaced.

Variscan Metamorphism

NE Sardinia

The most impressive feature of metamorphic zoning in NE Sardinia (Figure 3) is the rapid increase in metamorphic grade in a very restricted area, from biotite to a sillimanite + K-feldspar zone in only 40 km. The metamorphic zoning of metapelitic and metapsammitic sequences outcropping in northern Sardinia has been studied by observing the regional distribution of AKFM minerals (Franceschelli et al., 1982b; Elter et al., 1986). Six zones were distinguished from south to north: 1) biotite; 2) garnet; 3) staurolite + biotite; 4) kyanite+biotite 5) sillimanite; 6) sillimanite + K-feldspar (Figure 2). The garnet zone has been further subdivided into lower and upper garnet zones (i.e. garnet + albite and garnet +albite-oligoclase zones of Franceschelli et al., 1982a). This zoning is apparently continuous and gradual from medium- to high-grade, in spite of the abrupt lithological change and the major tectonic line (Posada-Asinara Line of Cappelli et al., 1992) dividing high-grade rocks from medium-grade ones.

The biotite zone is defined by the first appearance of biotite. The mineralogy is quartz, albite, muscovite, biotite and chlorite, with minor K-feldspar, epidote, carbonate and ilmenite in variable combinations. The sedimentary bedding of the protolith is still recognisable a few kilometres south of the village of Lula (Franceschelli et al., 1982b; Ricci et al., 2004). Mica, quartz and graphite helicitic inclusions sometimes occur in albite porphyroblasts. White mica parallel to S_1 schistosity is a paragonite-poor, celadonite-rich muscovite. Chlorite is an Mg-Fe chlorite and biotite shows an X_{Mg} in the 0.4-0.5 range.

The garnet zone is defined by the first appearance of almandine-rich garnet and by the coexistence of garnet, biotite and chlorite. The main mineral assemblage is quartz, albite, oligoclase, garnet, muscovite, biotite, chlorite, ilmenite, sphene, and, locally, chloritoid and incipient staurolite in varying combinations.

Diachronous development characterised the main constituents; S_1 minerals, quartz, muscovite, biotite and chlorite, are recrystallised parallel to S_2, while the first garnet and chloritoid are post-D_1, pre-D_2; that same initial garnet can have a post-D_1, pre-D_2 core and a syn-D_2 rim. A second garnet generation represented by small subhedral to euhedral crystals appears clearly syn-D_2. Helbing and Tiepolo (2005) attributed all pre-D_2 minerals to a pre-Variscan metamorphic event.

Carosi and Palmeri (2002) discovered that syn- to post-D_1 garnet crystals, included in albite porphyroblasts, show increasing MgO content from core to rim and lower CaO content with respect to the core of garnets growing along S_2 in the same thin section. This behaviour clearly indicates the attainment of peak temperature before the beginning of the D_2 phase, and subsequent decompression during this phase. The sequence peak temperature attainment - decompression is also suggested by the chemical zoning observed by Carosi and Palmeri (2002) in white mica flakes from the garnet zone: the core is still a celadonite-rich mica

(Si^{4+} = 3.3), while the rim is pure muscovite (Si^{4+} = 3.01). The same result, syn-D_1 celadonite-rich mica and stable syn-D_2 muscovite, was found by Di Vincenzo et al. (2004). The great decrease in celadonite content may be ascribed to a temperature increase and /or pressure decrease, probably to both simultaneously. Plagioclase, which is almost pure albite (An = 1-4%) in the albite-bearing zone, appears, in the oligoclase-bearing zone, as large albite (An= 1-4%) porphyroblasts, often surrounded by a thin oligoclase rim (An= 18-22%) or as small syn-D_2 oligoclase crystals (An= 16-18%). Abundant chlorite (X_{Mg} =0.4-0.5) and biotite (X_{Mg}= 0.4-0.5) occur as syn-D_2 flakes (parallel to S_2), wrapped around garnet and plagioclase porphyroblasts.

Garnet was generated by means of the following reactions (Franceschelli et al., 1982a): 1) chlorite + muscovite + quartz = garnet + biotite + H_2O; 2) carbonate + epidote + quartz = garnet + CO_2 + H_2O. The latter reaction is proposed to explain the remarkable grossularite content, particularly in garnet cores.

The staurolite + biotite zone is defined by the first appearance of the staurolite + biotite association observed on contact between metapelites and granodioritic orthogneisses and augen gneisses in the Lodé Antiform. Fractured staurolite and garnet porphyroblasts, flattened and rotated by D_2, are often embedded in a mylonitic matrix consisting of sheet silicates and quartz. Plagioclase, staurolite, garnet and biotite have syn-D_1 pre-D_2 cores and sometimes syn-D_2 rims (Franceschelli et al., 1982a, b; Carosi and Palmeri, 2002). Garnet porphyroblasts display bell-shaped Mn zoning and a gradual Mg, Fe increase and Ca decrease from core to rim. Staurolite has an X_{Mg} ratio ≈ 0.17, biotite X_{Mg} ratio= 0.50-0.60 and muscovite low celadonite content. The staurolite-producing reaction in the KFMASH system (Spear and Cheney, 1989) is: garnet + chlorite + muscovite = staurolite + biotite + quartz +H_2O.

The kyanite + biotite zone is characterised by the mineral assemblage kyanite, biotite, garnet, white mica and plagioclase. Staurolite relics are rimmed by new biotite. Garnet principally occurs as 0.1-0.2 mm sized idioblasts with cloudy cores rich in optically-undetectable inclusions. The average core and rim compositions are respectively Alm76, Prp11, Sps7, Grs6 and Alm75, Prp10, Sps9 and Grs6. Limpid garnet crystals contain microinclusions of idiomorphic anorthite, epidote and margarite and show the following average composition: Alm 62, Grs18, Prp7, Sps3, for the cores and Alm 67, Grs17, Prp10, Sps6 for the rims (Connolly et al., 1994). Staurolite, biotite, kyanite, garnet, white mica and plagioclase are all syn-D_1, pre-D_2 (Franceschelli et al., 1982b), at least in the cores, perhaps syn-D_2 in the rims (Carosi and Palmeri, 2002).

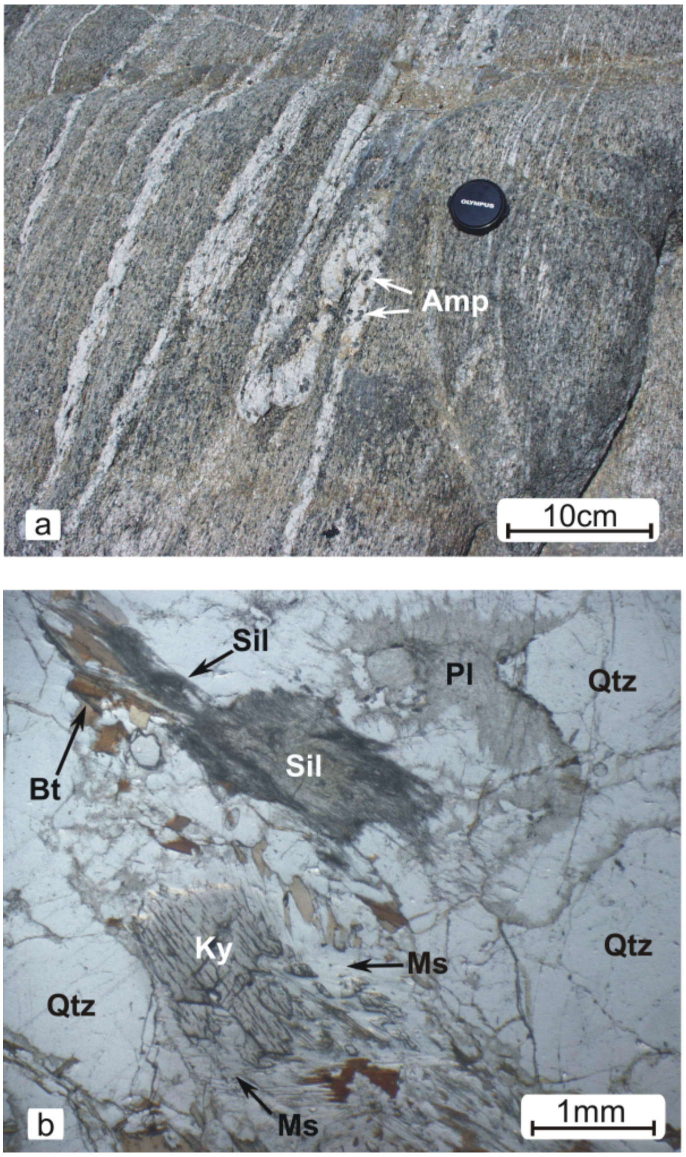
Plagioclase inclusions in limpid garnet are strongly calcic (An = 67-99%), while the same inclusions in cloudy garnet show An = 22-59%. The X_{Mg} of biotite is 0.4-0.5. White mica is a Na-and Fe-, Mg-poor muscovite. The following KFMASH discontinuous reaction (Spear and Cheney, 1989) staurolite + chlorite + muscovite = kyanite +biotite + quartz +H_2O, accounts for the appearance of kyanite + biotite.

The beginning of the sillimanite zone is marked by the entry of sillimanite and the disappearance of kyanite. Near the boundary with the kyanite + biotite zone, sillimanite-bearing rocks show mylonitic texture and, northwards, compositional layering due to the alternation of 0.1-0.3 mm thick mafic and quartz-feldspathic layers aligned along the main S_2 schistosity. The mineral assemblage includes quartz, plagioclase, garnet, sillimanite, muscovite, biotite and rare K-feldspar. Very fine-grained sillimanite needles, sometimes aggregated in nodules, are associated with small anhedral biotite flakes and/or included in quartz, plagioclase and muscovite crystals. Garnet occurs as fine-grained unzoned euhedral crystals with high spessartine content (up to 22%) and grossularite in the 3-7 % range, X_{Mg} ratio ≈ 0.16. Matrix biotite shows X_{Mg} in the 0.35-0.43 range and TiO_2 up to 2.6 %, muscovite with an Na/(Na+K) ratio of 0.06-0.09. Plagioclase has a uniform oligoclase-andesine composition.

The sillimanite + K-feldspar zone consists mainly of gneiss, nebulitic and layered migmatites. Mesosomes are medium-grained, with biotite flakes parallel to S_2 schistosity. They consist of quartz, plagioclase, garnet, fibrolitic sillimanite, biotite, K-feldspar and sporadic relics of kyanite in variable combinations. Retrograde overgrowth of muscovite on fibrolite has been observed locally. Leucosomes are medium- to coarse-grained poorly foliated rocks, with granitic to trondhjemitic composition. Plagioclase with An 25-32 sometimes shows myrmekitic intergrowth with K-feldspar. Garnet composition is Alm 59-67, Prp 0-11, Sps 10-25, Grs <4. Muscovite has low celadonite (0.06-0.09 a.p.f.u.) and paragonite content (0.06-0.10). K-feldspar is a frequently-zoned microcline with low anorthite content and a moderate Na/(Na+K) ratio (0.10-0.12). Biotite shows an X_{Mg} in the 0.30-0.50 range. Leucosomes may be regarded as having mainly derived from the muscovite dehydration melting reaction (Franceschelli et al., 1989; Cruciani et al., 2001).

Cruciani (2003) described two varieties of migmatite found at Punta Sirenella (north of Olbia): 1) amphibole-bearing migmatites; 2) kyanite + sillimanite bearing migmatites. Amphibole-bearing migmatites consist of close alternation of coarse- to fine/medium-grained leucosomes made up of quartz, plagioclase, ± K-feldspar, biotite and amphibole, and grey mesosomes (Figure 4 a). Kyanite + sillimanite migmatites have trondhjemitic leucosomes consisting of quartz, plagioclase, biotite, medium-grained kyanite crystals, sillimanite, garnet and retrograde muscovite (Figure 4 b).

Figure 4. NE Sardinia



A: Field photography of amphibole migmatites along the coast of NE Sardinia. Arrows indicate the coarse-grained amphibole (ferroan pargasite) on the D_2-folded trondhjemitic leucosome.

B: Photomicrograph of kyanite-sillimanite migmatites along the coast of NE Sardinia. One polar.

North-Central Sardinia

There are three major outcrops of low- to high-grade metamorphic rocks in north-central Sardinia: western Gallura, Anglona (Lago Coghinas) and Goceano (Figure 1 , 3). Studies on metamophic rocks from western Gallura and Anglona were carried out by Oggiano and Di Pisa (1992) and Ricci (1992).

In the western Gallura region, the L-MGMC is separated from the HGMC by an ENE-verging mylonitic belt. The L-MGMC is principally made up of micaschists with subordinate paragneiss bodies. The micaschists are strongly foliated rocks consisting mainly of quartz, plagioclase, biotite, muscovite, staurolite, kyanite and garnet. Along the shear zone, a phyllonitic belt a few hundred metres thick was formed from the original micaschists.

Migmatites of the HGMC outcrop northwards from the mylonitic belt. Although the migmatites show great textural and compositional variety, two main groups may be distinguished: metatexites and diatexites (Oggiano and Di Pisa, 1992). Metatexites are stromatic migmatites defined by the discontinuous alternation of leucosomes, melanosomes and mesosomes. Leucosomes, mainly trondhjemitic in composition, are granoblastic rocks characterised by vague foliation. These migmatites are very common in northern Sardinia, and their origin has been interpreted as the result of subsolidus differentiation (Ferrara et al., 1978; Oggiano and Di Pisa, 1992; Palmeri, 1992).

Diatexite structures vary greatly (agmatitic, nebulitic, schlieren, stromatic, etc.), revealing that these rocks were generated by in situ partial melting followed by melt segregation and mobilisation. Diatexites consist of plagioclase, quartz and K-feldspar in modal proportions similar to those of the minimum melting point. According to Del Moro et al. (1991), these rocks were probably generated by dehydration melting of muscovite and/or biotite deriving from crustal protoliths rich in arenaceous components. Diatexites show both trondhjemitic leucosomes, similar to those of metatexites, and granitic leucosomes. According to Oggiano and Di Pisa (1988), trondhjemitic leucosomes were generated before granitic ones. Migmatitic layering and trondhjemitic leucosomes are folded by D_2.

The metamorphic rocks in the Anglona region outcrop in the area of the Coghinas lake, between Limbara leucogranites to the north and eastern and western volcano-sedimentary successions of Oligo-Miocene age. Metamorphic rocks from Anglona are pelitic to arenaceous metasediments with minor calc-silicates, amphibolite lenses, quartzites and granitoid bodies of varying composition.

Three main deformation phases (D_1, D_2, D_3) have been recognised in this area (Oggiano and Di Pisa, 1992). In this region between Nurra and the Posada valley, the metamorphic basement shows LP/HT syn-late D_2 to pre-D_3 metamorphism that perhaps identifies a thermal dome with sillimanite in the central part, andalusite + sillimanite in adjacent NE and SW areas and only andalusite in the extreme SW corner of the region (Oggiano and Di Pisa, 1992; Ricci, 1992). Al_2SiO_5 polymorphs are associated to K-feldspar and cordierite. This LP/HT metamorphism is a late overprint produced by decompression on a previous Barrovian metamorphic event, testified by frequent relics of staurolite, garnet and plagioclase. According to Oggiano and Di Pisa (1992), similar evolution characterises Baronie and southern Gallura.

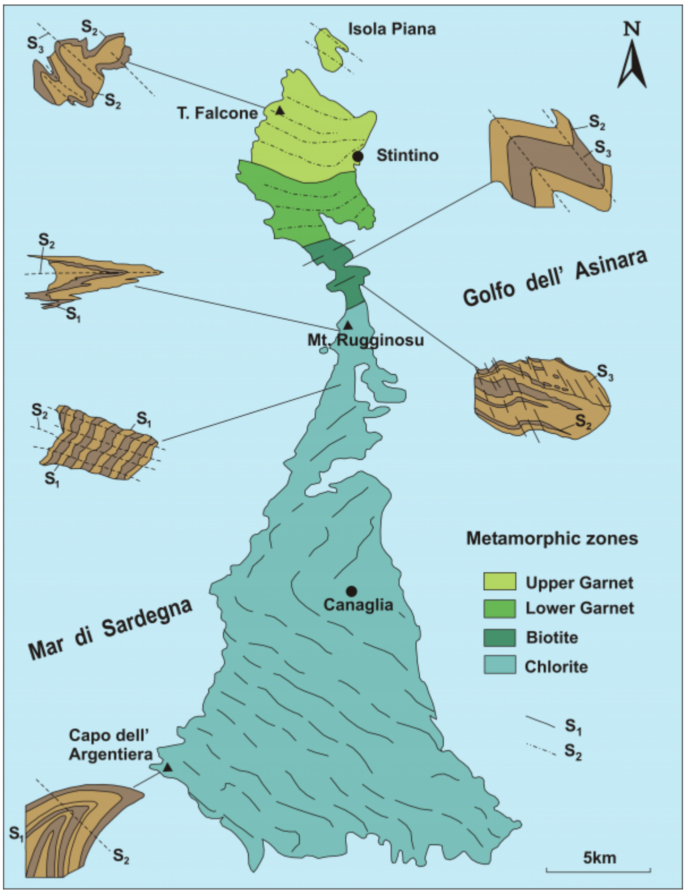
In the Anglona rocks, inclusion trails parallel to S_1 schistosity are often recognisable in plagioclase and staurolite porphyroblasts, suggesting post-D_1, pre- to syn-D_2 growth. The helicitic and sigmoidal texture of garnet indicates syn-D_2 growth and destral shear movement (Oggiano and Di Pisa, 1992, Figure 4). High-T low-P minerals developed during the late stages of D_2 and before D_3 as suggested by: a) fibrolitic sillimanite, parallel to D_2 and deformed by D_3; b) andalusite porphyroblasts, including poorly-crenulated S_2 schistosity, preceeding the highly-crenulated S_2 schistosity of the matrix; c) andalusite porphyroclasts deformed by D_3 and enveloped by the composite foliations of the Anglona shear zone; d) cordierite growth coeval with that of andalusite.

NW Sardinia

Nurra region

Detailed studies of metamorphic zoning in the Nurra region have been carried out by Carmignani et al. (1979, 1982b) and Franceschelli et al. (1990). The major difference with respect to NE Sardinia is represented by the existence of an extensive chlorite zone (Figure 2 , 5).

Figure 5. Nurra Paleozoic basement



Structural geological sketch map of the Nurra Paleozoic basement, showing metamorphic zonation and traces of S_1 and S_2 schistosity. Graphical representation of changing folding style is also shown (after Franceschelli et al., 1990, modified).

The chlorite zone shows close alternation of granoblastic quartzitic and micaceous layers, consisting of white mica (muscovite ± paragonite), albite, chlorite, and chloritoid in varying combinations. Accessory minerals are carbonate, epidote, Fe-Ti oxides and graphite. Slaty cleavage is replaced northwards by strain-slip S_1 schistosity. The increasing importance of S_2 schistosity, observed moving northwards, explains why sheet silicates are oriented parallel to S_2 in the northern part of the chlorite zone (Figure 5). Here, subrounded albite porphyroblasts still preserve inclusion trails parallel to relict S_1 schistosity, now oriented discordantly with respect to the enveloping S_2 one.

In the northernmost part of the chlorite zone, crenulations of S_2, mainly defined by opaque minerals, begin to appear on a regional scale. Albite porphyroblasts (An 1-4) occur as subrounded crystals of 0.1 mm. S_1 muscovite shows a broad range of Si a.p.f.u., with maximum values of 6.65-6.70 in metarhyolite samples and minimum values of 6.07-6.20 in Al-rich samples. Paragonite yielded MgO <0.20 wt%, FeO <0.40 wt% and a K/(K-Na) ratio ranging from 0.06 to 0.10. Chlorite shows the following composition: (a.p.f.u.) Al^{VI}= 2.76-3.14; Fe= 5.46-6.00 Mg=2.90-3.42.

The biotite zone - Carmignani et al. (1882b) discovered that the bulk composition of rocks may influence the first appearance of biotite. Early appearance of biotite was observed in metabasites at Monte Rugginosu, outcropping in the chlorite zone: low Al content anticipates while high Al content delays biotite entry into a prograde sequence.

Simpson (1998) observed that the D_2 phase produced open folds with interlimb angles of 40°-140° (mean 80°) in the chlorite zone and tight N-vergent isoclinal folds with interlimb angles of 25° at the onset of the biotite isograd. The abrupt change in interlimb angles defines the transition from dehydration-absent to dehydration-active metamorphic environments.

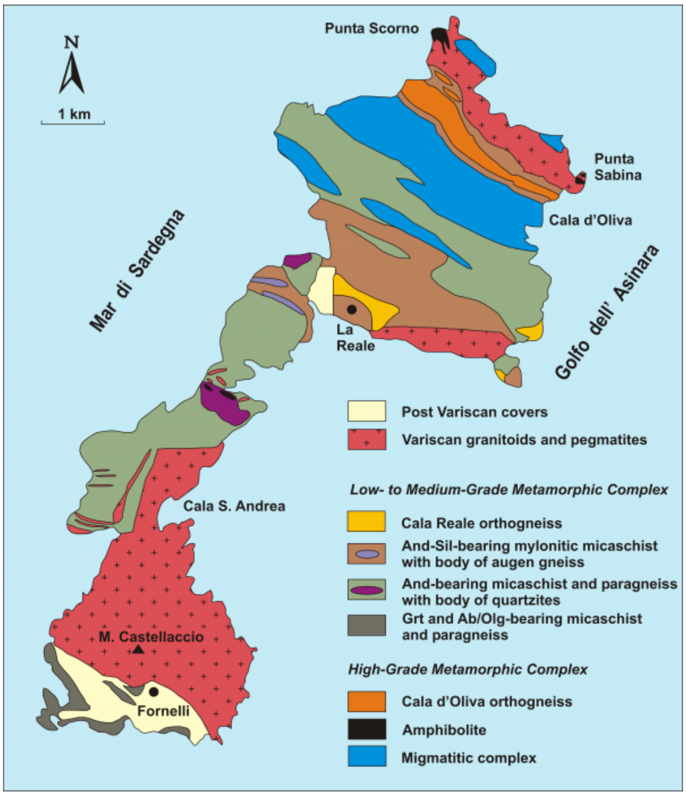
In the biotite zone, S_1 still survives in lens-shaped micaceous layers, where its presence is defined by mica blades sub-orthogonal to enveloping S_2 surfaces. Albite porphyroblasts show a xenoblastic core, including trails of opaque, quartz and phyllosilicate grains and a limpid rim devoid of inclusions. The trails, generally curved, denote growth, mostly subsequent to the beginning of the D_2 phase. The inclusion trails of biotite porphyroblasts depicted S_1 orientation and form wide angles with S_2 schistosity. Biotite (Al^{VI}=0.76 a.p.f.u.; Fe =2.65 a.p.f.u.; Mg =2.11 a.p.f.u.) is Al-rich annite (Carmignani et al., 1982b).

Garnet zone - This zone is defined by the first appearance of almandine- rich garnet in pelitic rocks. The analogy between NE Sardinia and Nurra as regards the presence of upper and lower garnet zones is worth noting. The garnet zone is characterised by an increase in mineral grain-size from 0.1-0.2 to 1-2 mm, and by the mineral assemblage albite, oligoclase, garnet, muscovite, biotite, chlorite, and minor chloritoid. S_2 is the main schistosity and S_3 becomes more and more evident towards the north. Inclusion trails in albite porphyroblasts may be rectilinear, gently or strongly folded and sometimes characterised by two orientations. These microstructures suggest a gradual deformation of S_1 schistosity during the D_2 phase. Garnet porphyroblasts of 0.1-0.2 mm occur in the matrix and as fresh idioblastic crystals within albite porphyroblasts. Garnet is Alm 59-82 with subordinate Grs 9-25, Sps 1-12 and Prp 5-9. Mn content and a Ca/(Ca+Mg+Fe) ratio decrease from core to rim. The opposite trend characterises Fe and Mg. A slight increase in Mg/(Mg+Fe) ratio was observed from core to rim. Biotite composition is as follows: Al^{VI}= 0.79-0.91 a.p.f.u.; X_{Mg} = 0.39-0.48. In the upper garnet zone, albite porphyroblasts are mantled by oligoclase rims (An20-24).

Asinara island

The first description of Asinara metamorphic rocks (Figure 6) can be found in Ricci (1972), who reported albite-oligoclase, garnet, K-feldspar, sillimanite and relict staurolite as characteristic metamorphic minerals. The occurrence of these minerals suggested that Asinara was affected, during the first Variscan phase, by the same Barrovian metamorphism well-known all over northern Sardinia. The "Barrovian" hypothesis was confirmed by the discovery of kyanite relics replaced by sillimanite within melanosomes of Punta Scorno migmatites (Oggiano and Di Pisa, 1998).

Figure 6. Asinara Island



Geological sketch map of Asinara Island (after Carosi et al., 2004, modified).

Two metamorphic complexes, juxtaposed along narrow belts of high-strain concentration, have been distinguished by Carosi et al. (2004) on Asinara Island: a low- to medium-grade metamorphic complex (L-MMC) and a high-grade metamorphic complex (HGMC). The L-MGMC consists mainly of (Figure 6) : 1) fine-grained and porphyroblastic paragneisses; 2) andalusite-and sillimanite-bearing paragneisses and micaschists; 3) massive amphibolites; 4) quartzites; 5) La Reale orthogneisses; 6) augen gneisses,7) mylonitic micaschists.

Fine-grained and porphyroblastic paragneisses consist of quartz, oligoclase, garnet, biotite and white mica; accessory minerals are epidote, monazite, zircon and oxides. Andalusite- and sillimanite-bearing paragneisses and micaschists show very high modal contents of andalusite, abundant relict garnet and staurolite replaced by an andalusite + biotite +oxide paragenesis. Accessory minerals are tourmaline, ilmenite, apatite, epidote and zircon. Massive amphibolites consist of hornblende and plagioclase, associated with minor amounts of biotite, chlorite and opaque minerals.

The HGMC consists of 1) migmatites, 2) Cala d'Olivo and Punta Scorno orthogneisses; 3) Punta Scorno banded amphibolites.

Diatexites and metatexites have been distinguished among the migmatites. Diatexites consist of centimetre-thick quartz-feldspar leucosomes, biotite schlieren and restitic polydeformed amphibolite bodies. Metatexites have mesosomes made up of biotite + sillimanite + quartz + cordierite ± K-feldspar and scattered neosome patches. Amphibolites occur as banded rocks made up of alternating centimetre- to decimetre-thick amphibolite and leptynite layers or as massive hornblende-plagioclase rocks. Relics of Ca-clinopyroxene and garnet testify to a previous granulite stage.

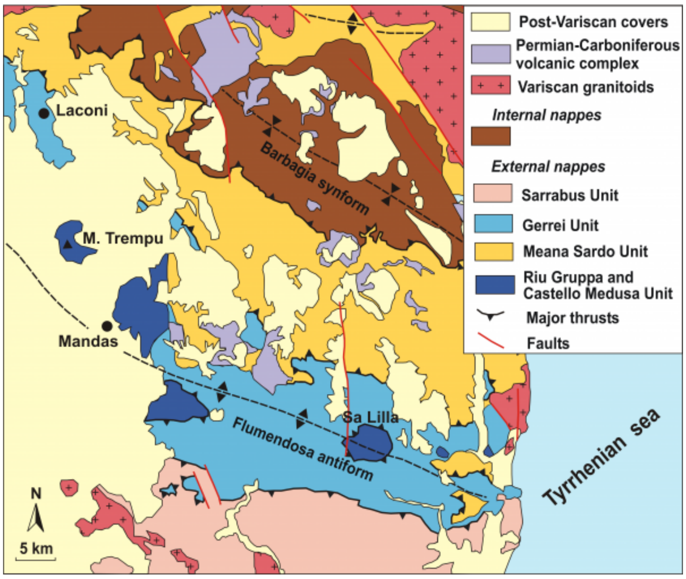
Asinara metamorphic rocks were affected by three deformation phases related to crustal thickening in a compressive and transpressive tectonic regime, followed by a later phase of extensional deformation (Carosi et al., 2004). In spite of a general HT/LP metamorphic overprint, Barrovian assemblages, pre- to syn-kinematic with respect to the D_2 deformation phase, are still recognizable.

These authors, mapping the distribution pattern of HT-LT mineral assemblages, hypothesized the existence of a regional Abukuma-type HT/LP metamorphism preceding the last deformation event.

Central Sardinia

Very few detailed studies have thus far been done on the tectono-metamorphic evolution of the Nappe and External zones (Figure 7) during Variscan orogeny. Metamorphism within the Nappe zone in central Sardinia took place during the complex event of Nappe emplacement and stacking and was prograde from low- to medium-grade P-T conditions. Two metamorphic events, M_1 and M_2 (D_1 and D_2 phases) mark the attainment of intermediate Barrovian P-T conditions (Carosi et al., 1992a). During the M_1 event, low-grade metamorphism was attained with blastesis of muscovite + chlorite + albite.

Figure 7. Paleozoic basement



Geological sketch map of the Paleozoic basement around the Flumendosa antiform, east-central Sardinia (after Carmignani et al. 2001, modified).

Further data on M_1 events were supplied by illite crystallinity (IC) measurements ($\Delta 2\theta$ values) on the six tectonic units of the Nappe pile. From bottom to top they are: Riu Gruppa, Gerrei, Meana Sardo, Arburese and the Barbagia metamorphic complex (Figure 2). Franceschelli et al. (1992) obtained the following IC values: Riu Gruppa, 0.18-0.26; Gerrei Unit, 0.24-0.32 at the front, 0.20-0.26 at the root; Arburese Unit, 0.20-0.26; Arburese unit, on a regional scale, 0.17-0.28 with clustering values in the 0.18-0.22 range (Eltrudis et al., 1995); Genn'Argiolas, 0.20-0.28; Meana Sardo, 0.20-0.32, with slightly decreasing values from front to roots; Barbagia Metamorphic Complex, front values of 0.24-0.34 and root ones of 0.18-0.22.

Eltrudis and Franceschelli (1995) supplied further IC data for the Mulargia Lake area. From bottom to top, IC values of 0.18-0.27 for the Mulargia Unit, 0.26-0.42 for the Gerrei Unit and 0.34-0.43 for the Meana Sardo Unit reveal metamorphic zonation from the deepest epizonal to the shallowest anchizonal units.

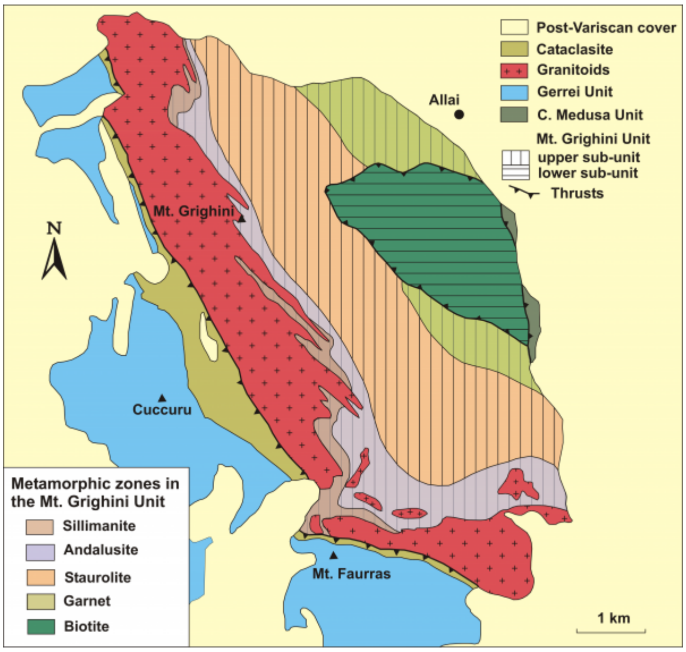
During the M_2 event, the low-grade metamorphic complex of Barbagia shows persistence of low-grade conditions, with syn-S_2 blastesis of chlorite + muscovite + albite (Dessau et al., 1983). Along the axis of the Flumendosa Antiform (Figure 7), prograde metamorphism was observed from SE (Riu Gruppa Unit) to NW (Castello Medusa Unit), applying the calcite-dolomite geothermometer to dolostone levels from Siluro-Devonian marbles (Carosi et al., 1990). The values obtained confirm prograde metamorphism from the Riu Gruppa Unit (average ≈ 300°C).

A higher metamorphic grade was attained only in the deepest units of Castello Medusa and Monte Grighini and in the metamorphic complexes of Sa Lilla, Mandas, Monte Trempu and Asuni.

The Monte Grighini (Figure 8) is a NW-SE trending metamorphic complex consisting of an upper Gerrei Unit and a lower Monte Grighini Unit (Musumeci, 1992). The Gerrei Unit, made up of Middle Ordovician to Siluro-Devonian sedimentary sequences, is characterised by low-grade metamorphism (chlorite zone). The Monte Grighini Unit, considered by Carosi et al. (1990) the deepest and most metamorphic unit in the whole Nappe zone, is divided by Musumeci (1992) into two subunits (Figure 8) : 1) one, consisting of low grade phyllites and acid metavolcanic rocks (biotite zone) and 2) the other, made up of paragneisses and micaschists reaching garnet to sillimanite zones. Along the NE corner of Monte Grighini, the Monte Grighini Unit is locally overlain by the Castello Medusa Unit consisting of low grade (biotite zone) phyllites, metavolcanic rocks and calc-schists. The Monte Grighini Unit is intruded by two NW-SE elongated granite plutons, one calc-alkaline monzogranitic to tonalitic, the other peraluminous leucogranitic. According to Musumeci (1992), P-T peak conditions at the end of the D_2 -M_2 event were T= 600° ± 50°C and P= 0.6 ± 0.1 GPa. The following M_3 event, linked to the intrusion of plutonic rocks, gave rise to the crystallisation of fibrolitic sillimanite, andalusite and cordierite and subsequent generation of biotite and garnet. Thermobarometric calibrations (Carosi et al., 1992a) yielded values of 470°C for the garnet +biotite zone and 560°C for the staurolite+ biotite zone, with pressure values of 0.5

± 0.1 GPa. The latter stage marks the initial exhumation of the Variscan Belt by means of extensional tectonics caused by the late gravitational collapse of the thickened Variscan crust.

Figure 8. Monte Grighini Unit



Metamorphic zonation of the Monte Grighini Unit, west-central Sardinia (after Carosi et al., 1992a, modified). See text for explanation.

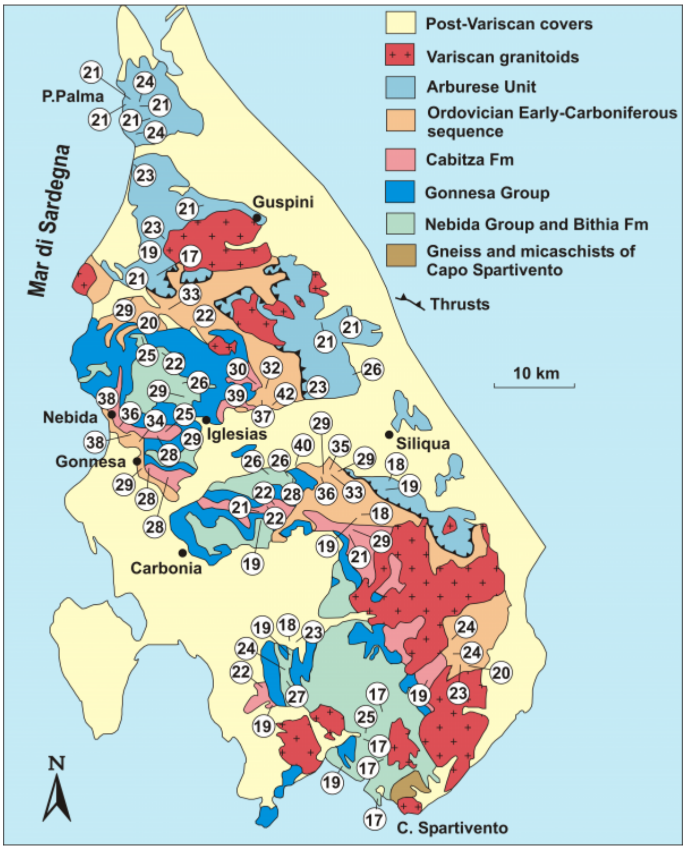
In the tectonic windows of Sa Lilla, Mandas, Monte Trempu, and Asuni medium- to high-grade rocks show rare S_1 relics, largely prevailing S_2 schistosity and the following prograde sequence of metamorphic zones: 1) chlorite; 2) biotite; 3) cordierite + andalusite; 4) cordierite + sillimanite. The coexistence of Al-silicates with cordierite occurs within the muscovite + cordierite + quartz stability field, yielding a nearly isobaric P-T path for higher-grade assemblages, corresponding to P=0.3 GPa and T= 500°-600°C, with a gradient of 60°C/km (Cappelli, 1991).

SW Sardinia

Illite crystallinity zonation

A detailed picture of illite crystallinity regional distribution (Figure 9) in the external zone of the Variscan chain in Sardinia is given by Eltrudis et al. (1995, Figure 1). For the whole sequence, from Puddinga to early Carboniferous metasediments, they found a IC range of 0.26-0.45, with a cluster of values from 0.32 to 0.38 (anchizone). Going northwards from the area around Gonnesa, the Cabitza Fm yielded IC values of 0.20-0.30, (average 0.26) with six values > 0.23 and only two < 0.23. Going southwards from the area near Gonnesa, the Cabitza Fm displays opposite behaviour, yielding IC values of 0.19-0.29 (average 0.22). The Nebida group is characterised by IC values in the 0.17-0.29 range, with a regional distribution similar to that described for the Cabitza Fm: moving northwards from the Gonnesa area, the Nebida Fm shows a IC range of 0.22-0.29, with an average of 0.26; going southwards from the Gonnesa area, results obtained for the Nebida group are: IC range of 0.17-0.23, average value of 0.20.

Figure 9. Average IC values



Distribution of average IC values (Δ°2θ x 100) in Paleozoic units from SW Sardinia; after Eltrudis et al. (1995), modified.

The following observations are suggested by the above-reported data: 1) There is a significant jump in metamorphic grade between the anchimetamorphism of Puddinga and the epimetamorphism of the underlying Cambrian to early Ordovician formations (see also Conti et al., 1978); 2) Metamorphic zonation of the Cabitza and Nebida formations, showing increasing metamoprhic grade from north (average IC values of 0.26) to south (average IC values of 0.20-0.22), is opposite to Variscan zonation, characterised by a northward increase in metamorphic grade.

Capo Spartivento metamorphic rocks

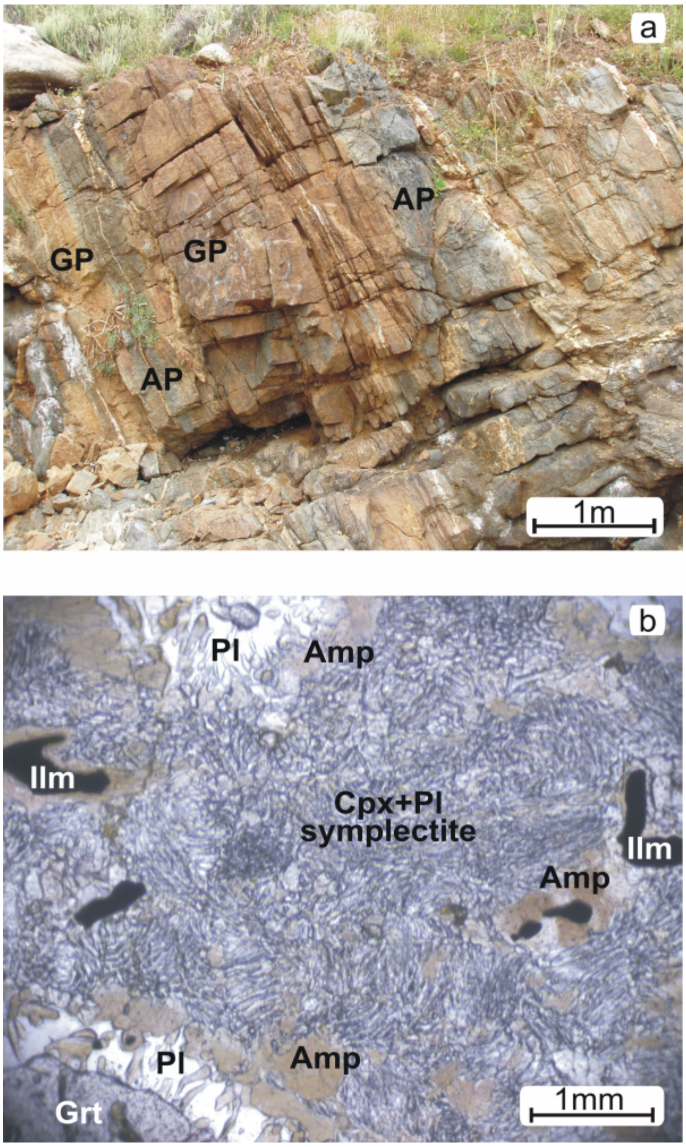
Capo Spartivento metamorphic rocks (Figure 9) were described by Cocozza et al. (1977) and Mazzoli and Visonà (1992), who considered the Monte Filau gneissic body an original granitic intrusion and Settiballas micaschists as its original country rock envelope. The medium- to coarse-grained Monte Filau orthogneisses consist of quartz, plagioclase, K-feldspar, biotite, muscovite and andalusite, sillimanite and garnet. Monte Settiballas micaschists are polymetamorphic, well-foliated rocks characterised by an alternation of granoblastic and lepidoblastic millimetric layers. According to Sassi and Visonà (1989), two main tectono-metamorphic events have been recognised in the micaschists: the first one is documented by relics of andalusite, cordierite, garnet and muscovite, suggesting a high metamorphic gradient. The second one led to the development of compositional layering and the growth of new biotite and muscovite flakes along foliation planes. The Bithia Fm consists of a sequence of phyllites, quartzites, quartzitic metasandstones and metapelites, with minor lenses of marbles and metabasites. The relationships between the Monte Filau orthogneiss and the Bithia Fm support the idea that D_1 deformation was pre-Variscan and contemporaneous to the emplacement of the orthogneiss protolith during the Ordovician (Carosi et al., 1995). The D_2 phase was probably related to a compressional phase, with maximum shortening in the WSW-ENE direction. This phase can be correlated with the main Variscan deformation of the Iglesiente-Sulcis region. Carmignani et al. (1992, 1994) interpreted the Capo Spartivento structure as a metamorphic core complex generated during the extensional collapse of the chain.

Metabasites with Eclogite Facies Relics

Metabasites with eclogite facies relics have been described at Punta de li Tulchi, Costa Smeralda, Anglona and Golfo Aranci (Figure 3). The origin and age of these rocks are still controversial and highly debated in literature. In particular, except for Cappelli et al. (1992), who suggest an age of 957 ± 93 Ma, U-Pb zircon data (Table 1) point to protolith ages of about 460 Ma (Cortesogno et al., 2004; Palmeri et al., 2004; Giacomini et al., 2005a).

The metabasites with eclogite relics at Punta de li Tulchi consist of brownish to greenish layers, from a few cm up to 50 cm thick. Two main lithotypes have been distinguished by Franceschelli et al. (1998): garnet-pyroxene-rich and amphibole-plagioclase-rich layers (Figure 10 a). The complex evolution of the eclogite-bearing rocks may be summarised as follows:

Figure 10. Field photography



A: Alternation of garnet-pyroxene (GP) layers and amphibole-plagioclase (AP) layers in the eclogite outcrop at Punta de li Tulchi (NE Sardinia).

B: Diopsidic clinopyroxene and Na-plagioclase symplectite resulting from destabilisation of omphacites; in the upper and lower left corners, note the kelyphytic structure around garnet. One polar.

Pre-eclogite stage: The occurrence of euhedral tschermakite and zoisite in the garnet core might be interpreted as a pre-eclogite amphibolite-epidote stage, as suggested by Chang Whan Oh and Liou (1990). Amphibole composition plotting in the field of medium pressure amphiboles supports this hypothesis.

Eclogite stage: Maximum jadeite content in omphacitic pyroxene is about 30%; reasonable assumptions led Franceschelli et al. (1998) to hypothesize an original content of 50% jadeite. Syn-eclogite garnet was probably almandine-rich (55-60%), with 15-20% pyrope and 20-25% grossularite content. The mineral assemblage of the eclogite stage was probably garnet, omphacite, rutile, zoisite and quartz.

Granulite stage: The destabilisation of eclogite stage minerals is documented by the formation of sodic plagioclase-clinopyroxene symplectites (Figure 10 b), and the crystallisation of orthopyroxene, calcic plagioclase and new garnet. Probable mineral paragenesis at the granulite stage includes garnet, Na- and Ca-rich plagioclases, clinopyroxene and orthopyroxene as major minerals.

Amphibolite stage: The amphibolitisation of the granulite assemblage during exhumation of the chain led to the replacement of orthopyroxene by cummingtonite and the formation of coronitic rims of amphibole and plagioclase around garnet, plagioclase and clinopyroxene-plagioclase symplectites.

Greenschist stage: The mineral assemblage is: actinolite, chlorite, epidote ss., REE epidote and titanite, replacing garnet, clinopyroxene, hornblende and biotite.

In a recent paper, Cortesogno et al. (2004) have distinguished the eclogites enclosed in the Migmatite Complex, eclogites A, from those embedded in a medium-grade, mostly metapelitic Gneiss Complex, eclogites B. Metabasites with relics of eclogites A appear in Costa Smeralda (NE Sardinia) as rotated boudins within andalusite-cordierite-bearing metatexites or within diatexites derived from ortho- and para-gneisses showing a sillimanite +K-feldspar-bearing assemblage. Hectometric metabasite slices with eclogite B relics occur within staurolite + garnet + kyanite-bearing micaschists and paragneisses in the Anglona region, within the Posada-Asinara shear zone. Omphacite is preserved only as armoured relics in garnet; its former presence is suggested by symplectite pseudomorphs that have completely replaced the originally subhedral isoriented omphacite crystals. The decompression event caused the breakdown of omphacite into symplectite aggregates, the growth of kelyphytic rims and finally the attainment of retrograde amphibolite facies P-T conditions, with patchy growth of fine-grained to acicular amphibole. A slightly different framework is proposed by Giacomini et al. (2005a) for the banded amphibolites and amphibolitised eclogites embedded within the Golfo Aranci metamorphic basement (NE Sardinia, north of Olbia). The banded amphibolites are foliated and layered rocks with variable proportions of amphibole and plagioclase. The amphibolitised eclogites consist of clinopyroxene-garnet-amphibole rocks with compositional layering defined by alternating garnet-rich and garnet-poor bands. Eclogite relics are more frequent in garnet-rich layers and mainly represented by omphacite relics surrounded by clinopyroxene-plagioclase symplectite and, towards the garnet grains, by a rim of bluish-green amphibole. The most important novelty with respect to Punta de li Tulchi eclogites is the appearance of an inclusion-free kyanite displaying composite coronitic texture. The latter texture consists of outer coarse-grained plagioclase + clinopyroxene + orthopyroxene symplectite, an intermediate polycrystalline plagioclase domain and an inner symplectitic rim made up of sapphirine + anorthite ± corundum ± spinel passing to spinel + anorthite ± corundum in the innermost part, in contact with relict kyanite. Garnet-poor layers rarely show eclogitic relics and consist mainly of clinopyroxene-plagioclase symplectites, quartz and scattered garnet porphyroblasts. In extreme cases, the amphibolitisation process completely transformed the original rock into an aggregate of amphibole, plagioclase and quartz ± ilmenite, with few surviving relics of clinopyroxene-plagioclase symplectites. Sapphirine is a highly-peraluminous variety similar to that found by Godard and Mabit (1998) in the kyanite-bearing eclogites at Pays de Léon, France. For the final amphibolitisation stage, Giacomini et al. (2005a) hypothesize that zoisite breakdown provided a limited amount of water, but that major amounts of fluids were provided by metapelite dehydration processes.

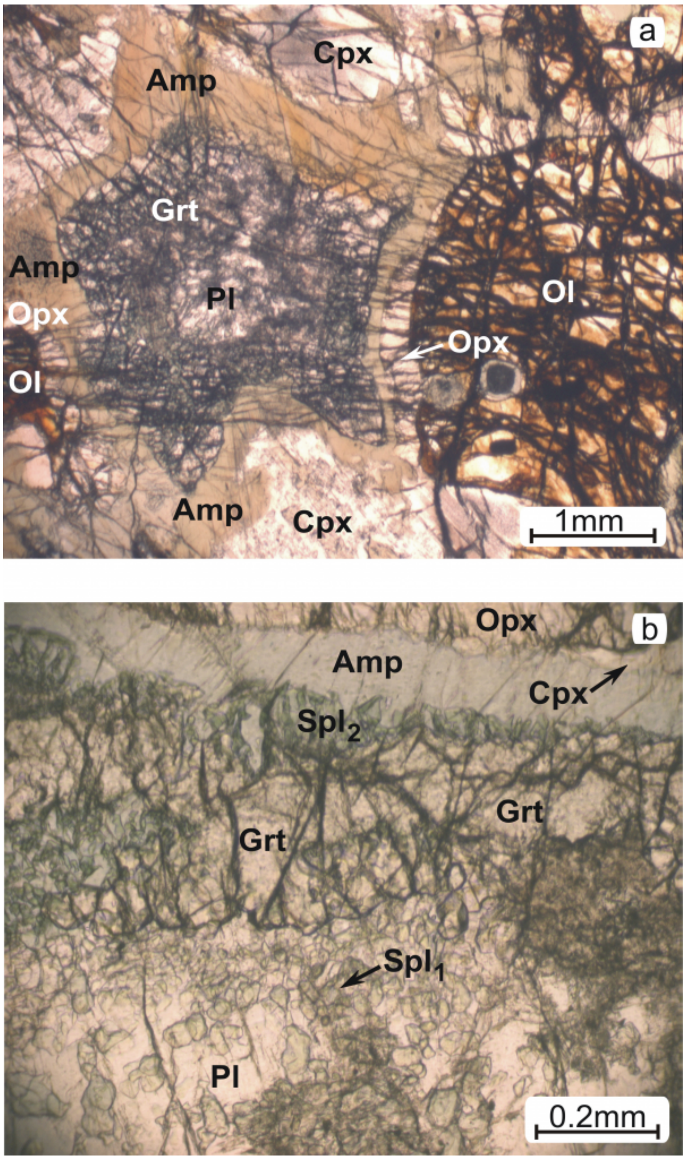
Metabasite with Granulite Facies Relics

Metabasites with granulite facies relics occur as a lenticular body (2 km long and 100-150 m thick) embedded in the migmatites at Montiggiu Nieddu (Figure 3) (Ghezzo et al., 1979; Franceschelli et al., 2002). Two lithotypes have been distinguished: ultramafic amphibolites and plagioclase banded amphibolites. Ultramafic amphibolites appear as a dark-green to black, massive to slightly schistose body divided into three compositional layers: A: chlorite-rich amphibolite; B: plagioclase-rich greenish amphibolite; C: dark green garnet-rich amphibolite. These layers include 20-30 cm thick intercalations consisting almost totally (up to 90-95%) of coarse-grained amphibole. Garnet-rich nodules and veins are also frequently observed.

Plagioclase banded amphibolites consist of alternating dark green (amphibole-rich) and white (plagioclase-rich) decimetre- to decametre-thick layers. The dark green layers are made up of amphibole, plagioclase, garnet, relics of orthopyroxene and clinopyroxene, chlorite, biotite, epidote and sphene. The white layers consist of plagioclase, amphibole, garnet and rare clinopyroxene. In ultramafic amphibolites, the igneous minerals are plagioclase, olivine, clinopyroxene and orthopyroxene.

The post-igneous metamorphic evolution of Montiggiu Nieddu metabasites may be divided into three stages: a first stage (corona stage), with the appareance of clinopyroxene, orthopyroxene, spinel and garnet (Figure 11 a,b); a second stage, mainly characterised by the widespread appearance of clinoamphibole, chlorite, spinel and orthoamphibole; third stage minerals, mostly replacing mafic minerals, consist of tremolite, chlorite, fayalite, talc, epidote, albite, calcite, dolomite, serpentine and probably corundum.

Figure 11. Montiggiu Nieddu



Photomicrographs showing steps in the metamorphic evolution of granulite rocks from Montiggiu Nieddu (NE Sardinia). A: corona microstructure developed around igneous olivine and plagioclase; on the lower side, igneous clinopyroxene is replaced by brown amphibole. One polar. B: igneous plagioclase (anorthite) replaced by garnet. Note the spinel growth on plagioclase (Spl_1) and at the boundary between garnet and amphibole (Spl_2). One polar.

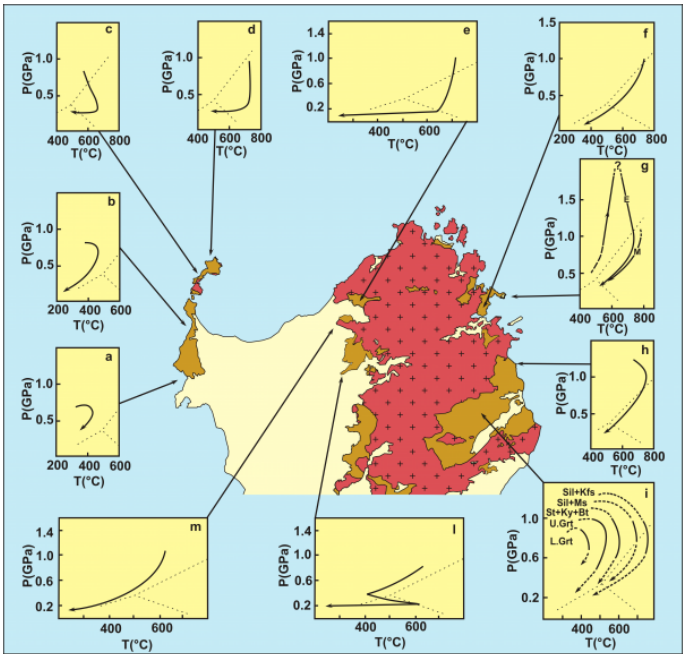
Tectono-metamorphic Evolution of Variscan Sardinia

P-T-t path

The main contributions towards a better reconstruction of P-T paths for the whole northern part of Sardinia are graphically reported in Figure 12 . Reference papers are those by Franceschelli et al. (1989, 1990, 1998, 2002), Ricci (1992), Carosi et al. (2004) and Giacomini et al. (2005a).

Ricci (1992, Figure 3) has also shown that P-T loops, varying in shape and time evolution, characterise the different portions of the metamorphic basement in northern Sardinia, probably owing to diachronous metamorphic evolution of the chain in different areas. However, the general P-T-t path for Variscan metamorphism in Sardinia is that of collisional belts, showing the typical clockwise trend attributed by England and Thompson (1984) to late thermal relaxation following the initial thickening stage. The P-T-t path in Nurra (NW Sardinia) is described by Franceschelli et al. (1990). In southern Nurra, the P-T path (Figure 12 a) is poorly constrained but seems to indicate P=0.7-0.8 GPa for peak pressure at 350°C (end of D_1) and T = 420°C for peak temperature at P=0.6 GPa (end of the early D_2 phase). For northern Nurra (Figure 12 b), the end of the D_1 phase is marked by the attainment of peak pressure at 0.8 GPa at a temperature of 400°C, while during early decompression of the D_2 phase, northern Nurra basement rocks reached peak temperatures of 450°- 500°C for pressures of 0.6-0.7 GPa. The late stage of the D_2 phase is characterised by a decrease in both pressure (from 0.6 to 0.4 GPa) and temperature (from 480°C to 420°C). During D_3, the P-T decrease was: T from 420°C to 300°C, P from 0.4 to 0.2 GPa.

Figure 12. Pressure and temperature evolution



Pressure and temperature evolution in the northern Sardinia basement: shapes of P-T paths are redrawn from: a, b (Nurra) Franceschelli et al. (1990); c, d (Asinara Island) Carosi et al. (2004); e, l, m (Western Gallura, Anglona) Ricci (1992); f (granulitic rocks from Montiggiu Nieddu) Franceschelli et al. (2002); g (E: retrogressed eclogites and M: migmatite near Golfo Aranci) Giacomini et al. (2005a); h (eclogitic rocks from P.ta de li Tulchi) Franceschelli et al. (1998); i (metapelitic rocks from garnet to sillimanite+ K-feldspar zones, NE Sardinia) Franceschelli et al. (1989), Ricci et al. (2004). U.Grt = upper garnet zone; L.Grt = lower garnet zone. Dashed line: Al_2SiO_5 triple point after Holdaway (1971).

Very few, fragmentary data are available in literature for a precise reconstruction of the P-T path followed by the medium- and high-grade rocks on Asinara Island (Figure 12 c,d), perhaps owing to the strong thermal overprint of late HT/LP metamorphism, able to obliterate any evidence of the D_1 phase (early part of the P-T-t path). The highest P-T values calculated with thermobarometers for the D_2 phase in L-MGMC rocks are P= 0.8-1.0 GPa and T=560°C (Carosi et al., 2004, Figure 4) and T=600°-650°C (Ricci, 1992).

The first part of the P-T path shows an increase in temperature of about 100°C and a decrease in pressure of up to P=0.2-0.3 GPa. The final part of this path shows isobaric cooling, with a T decrease down to T=400°-500°C. For migmatite and high-grade rocks, Di Pisa et al. (1993) calculated a pressure decrease from 0.7-0.8 GPa at temperatures of 720°-740°C (migmatite stage) to 0.3-0.4 GPa at temperatures of 500°-600°C (amphibolite stage). However, more recently, in melanosomes from migmatites at Punta Scorno, Oggiano and Di Pisa (1998) have found kyanite relics containing sillimanite prisms evolving into acicular aggregates. The attainment of kyanite isograd force pressure at values higher than 0.8-0.9 GPa for high-grade rocks on Asinara.

According to Ricci (1992), and similar to the Asinara path, the Anglona and western Gallura areas (Figure 12 e,l,m) (northern coast facing Corsica) show a P-T-t path crosscutting the maximally relaxed geotherm due to a post-thickening heat supply (Thompson and England, 1984). Ricci (1992) attributed this behaviour to reheating of a previously-cooled and uplifted block by late intrusion (Anglona) or to extensional tectonics caused by shallow emplacement of syntectonic, peraluminous granitoids (Palau-Santa Teresa di Gallura).

Giacomimi et al. (2005a) calculated T=680-750°C P=0.8 GPa up to maximum values of T= 788°C and 1.02 GPa for the migmatite formation at Golfo Aranci (Figure 12 g, path M). For the amphibolite stage, the authors obtained temperatures of 582°C and 663°C and pressures of 0.6-0.7 GPa.

P-T-t paths of metabasite with granulite (Figure 12 f) and eclogite (Figure 12 g,h) facies relics in NE Sardinia have been defined by Franceschelli et al. (1998) and Giacomini et al. (2005a).

At Punta de li Tulchi, after the first papers by Miller et al. (1976) and Ghezzo et al. (1982), Franceschelli et al. (1998) reconstructed the P-T-t path (Figure 12 h) of HP metamorphism by calibrating different thermobarometers applied to relict eclogite and granulite assemblages still preserved within amphibolitic metabasites.

The following P-T conditions were obtained. For the eclogite stage, a temperature range of 564°-669°C (Grt-Cpx geothermometer using calibration by Ellis and Green, 1979) and a minimum pressure of 1.3 GPa at 700°C, based on jadeite content of relict omphacite, were calculated for a sample crystallised at the beginning of the eclogite-granulite transition. For the granulite stage, two temperature ranges of 765°-899°C and 789°-829°C were calculated with a Cpx-Opx thermometer; pressure estimates yielded 1.0-1.2 GPa. For the amphibolite stage, calculations provided temperatures of 550°-650°C and pressures of 0.3-0.7 GPa; temperatures of 300°-400°C and pressures of 0.2-0.3 GPa were calculated for the greenschist stage.

A similar reconstruction of the P-T path (Figure 12 g, path E) is now proposed for the metabasites with HP metamorphic relics at Golfo Aranci by Giacomini et al. (2005a, see Figure 12). The occurrence of the edenite-andesine pair as a small inclusion within a kyanite porphyroblast is interpreted by the authors as testifying to an early prograde amphibolite stage with T=580°-605°C and P=0.8-1.0 GPa. For the subsequent eclogite stage at temperatures of 550°-700°C, maximum pressures of 1.32-1.63 GPa and of 1.40-1.72 GPa were calculated on the basis of jadeite contents of 36% and 44% respectively, yielded by omphacite microprobe analyses. The clinopyroxene-garnet thermometer indicates a temperature range of 580°-690°C and a pressure range of 1.3-1.8 GPa. The decompression part of the P-T path is characterised by omphacite destabilisation and the appearance of clinopyroxene-plagioclase symplectites, typical of the granulite stage, for which geothermobarometry yielded T=650°-790°C and P=0.9-1.2 GPa. A further pressure decrease led to the almost complete transformation of symplectite-rich granulites into garnet amphibolites (HT amphibolites), for which temperatures of 650°-790°C and pressures of 0.9-1.2 GPa have been calculated. The final, complete amphibolitisation of the original granulitic rock generated amphibolites with lobate grain boundaries yielding equilibration temperatures of 550°-650°C at 0.5-0.7 GPa.

For the whole evolution of Montiggiu Nieddu metabasites, P-T conditions (Figure 12 f) were estimated as follows. Granulite stage equilibrium conditions were achieved only on a domain scale (Bethume and Davidson, 1997, with references). For two corona microdomains, the garnet-clinopyroxene thermometer, using calibrations by Ellis and Green (1979) and Powell (1985), yielded temperatures in the 700°-740°C range, while the garnet-orthopyroxene thermometer gave a range of 650°-700°C. Using the composition of garnet cores and orthopyroxene and plagioclase enclosed in porphyroblastic garnet, Franceschelli et al. (2002) obtained, with various calibrations, T= 680°-750°C and P=0.8-0.9 GPa. Comparable values of T =700°-750°C and P= 0.8-1.0 GPa were calculated for the coronas of ultramafic amphibolites by Franceschelli et al. (2002), substantially agreeing with the previous estimates of 750°C and 1.0 GPa proposed by Ghezzo et al. (1979). For the amphibolite stage, the garnet +amphibole + plagioclase +quartz assemblage suggests temperatures in the 540°-640°C range and the garnet-amphibolite-plagioclase-quartz barometer gives pressures of 0.4-0.6 GPa. The late appearance of chlorite, tremolite and epidote indicates re-equilibration at temperatures of 330°-400°C and pressures of 0.2-0.3 GPa in the final part of the P-T path.

For NE Sardinia, from Siniscola to Punta de li Tulchi, Franceschelli et al. (1989) and Ricci (1992) proposed the following P-T-t path (Figure 12 i). The ascending limb of the path corresponds to maximum thickening of the crust, with attainment of a peak pressure of 1.2 GPa at temperatures of 550°-600°C: this limb is characterised by prograde metamorphism and static post-D_1 crystallisation of key minerals from biotite up to the sillimanite+ K-feldspar zone. The highest pressure part of this path is calculated precisely only in the garnet zone, still maintaining minerals crystallised during the M_1 event (D_1 phase). From the staurolite + biotite zone to the sillimanite + K-feldspar one, rocks recorded only decompression and temperature decrease after the thermal climax, as strong D_2 deformation and peak temperatures had almost completely erased any evidence of the previous D_1 phase. This explains why, for medium- and high-grade zones, the peak pressure part of the path is hypothetical and backward-extrapolated from the peak temperature of the downgoing limb. Backward extrapolation from peak temperature of high-grade zones is partially constrained by the finding of relict minerals such as kyanite.

The folllowing estimates of P-T conditions have been published: Di Vincenzo et al. (2004), using various geothermobarometers, calculated T = 497°-544°C, P=1.0-1.1GPa for the D_1 phase and T= 521°-560°C, P=0.7-0.9 GPa for the D_2 phase in minerals of the garnet zone.

For the staurolite + biotite zone, Franceschelli et al. (1989) calculated T =610°C, P= 0.8-1.0 GPa; Di Vincenzo et al. (2004), T=588°-624°C, P= 0.6- 0.9 GPa. As regards the kyanite + biotite zone, Carosi and Palmeri (2002) estimated T < 650°C and P>0.9 GPa for the D_1 phase and T= 650-700°C, P=0.7-0.9 for the D_2 phase. For the sillimanite + K-feldspar zone (near Punta de Li Tulchi), Cruciani et al. (2001) found evidence that melting started in the kyanite field, migmatite later attained peak conditions in the sillimanite field at T= 700°-720°C, P= 0.6-0.8 GPa. The downgoing part of the P-T-t path ended at T 630°-670°C, P=0.4-0.6 GPa.

Variscan metamorphic events and geochronological data

Summarising all previously-reported data for the P-T-t path of Variscan metamorphic rocks from Sardinia and taking many recent radiometric data into account (Table 2), the following sequence of events may be proposed. An early phase of HP metamorphism characterised the beginning of the Variscan orogenic cycle in Sardinia. Evidence of this event has thus far been found only in some amphibolitic metabasites containing relics of eclogitic mineral assemblages enclosed in the HGMC or L-MGMC. The age of protoliths seems to be Middle Ordovician, as suggested by three U-Pb zircon ages of 453 ± 14 Ma, 457 ± 2 Ma and 460 ± 5 Ma, (Table 1) respectively found for metabasites with eclogite relics by Palmeri et al. (2004), Cortesogno et al. (2004) and Giacomini et al. (2005a).

Moreover, Palmeri et al. (2004), on the basis of SHRIMP zircon U-Pb data, put forward the hypothesis that 400 ± 10 Ma might be the age of the eclogite formation in NE Sardinia. For eclogites (eclogite A) included in the HGMC, Cortesogno et al. (2004) also obtained a zircon age of 403 ± 4 Ma, interpreted as dating the high-grade event.

According to Cortesogno et al. (2004), eclogites B and the hosted Gneiss Complex must be attributed to a Neoproterozoic (?) Cambro-Ordovician rifting event, giving rise to bimodal tholeiitic-rhyolitic volcanic rocks interbedded within a pelitic-psammitic sedimentary sequence, later subducted and intruded by granitic melts during the Early-Middle Ordovician.

For eclogites embedded within the Sardinian HGMC and Savona Crystalline Massif, Giacomini et al. (2005b) propose a Middle Ordovician protolith age, an Early Visean age for eclogite facies metamorphism and a Late Visean age for amphibolite facies metamorphism, i.e. respective ages of 460-450 Ma, 345 Ma and 320 Ma.

A question now arises: did migmatite in fact experience the HP event revealed by enclosed eclogite lenses? Except for kyanite relics found in plagioclase, no UHP mineral such as coesite or diamond has thus far been found in migmatite.

The following chronological sequence may be proposed for Variscan metamorphic events. According to Di Vincenzo et al. (2004), the early thickening stage of the D_1 - M_1 event did not start before ~360 Ma and probably took place at 345-340 Ma, in agreement with the Tournaisian age (355-345 Ma) of deformed metamorphic rocks from SE Sardinia, and with an ^{40}Ar-^{39}Ar age of 345 ± 4 Ma on actinolite from a metagabbro (Del Moro et al., 1991). Moreover, the time span from 360 to 345 Ma is consistent with the minimum theoretical time lapse of 20 Ma indispensable for the change from a passive to active collisional margin and for the beginning of exhumation (Thompson et al., 1997). More specifically, Di Vincenzo et al. (2004) found apparent ^{40}Ar-^{39}Ar ages of 340-315 Ma for muscovites in the garnet zone, with the oldest ages (340-335 Ma) for syn-D_1 white mica, and age clustering at 320-315 Ma for most syn-D_2 white mica.

Giacomini et al. (2005a), taking into account the petrological features of a 344 ± 7 Ma-old leucosome from a NE Sardinia migmatite (Ferrara et al., 1978), proposed an age of 345 Ma for the muscovite dehydration melting event that took place in Golfo Aranci gneisses.

Pervasive fluid infiltration into metabasite lenses from enveloping migmatites transformed anhydrous granulite facies assemblages into hydrated upper amphibolite facies assemblages at 352 ± 3 Ma (zircon resetting age by Giacomini et al., 2005a). The age values of 344 and 352 Ma seem to indicate that migmatisation, partial melting and related metasomatic processes took place soon after the end of the D_1-M_1 phase, and that peak temperatures in the basement were attained very close to the D_1-D_2 boundary. Similarly, according to Elter et al. (1999), age values around 350 Ma, 345 ± 4 (Del Moro et al., 1991) and 344 ± 7Ma (Ferrara et al., 1978) mark the boundary in the axial zone between the end of collisional tectonics, when peak pressure is reached (D_1 phase), and the beginning of the extensional tectonics responsible for exhumation and uplift (D_2 phase). Ricci et al. (2004) also stated that "the 345 Ma age could therefore be close to the collisonal stage or represent the beginning of the exhumation".

In Anglona and on Asinara Island, Barrovian metamorphic mineral assemblages were overprinted by late Variscan (Rb-Sr age of 303 ± 6 Ma on muscovite, Del Moro et al., 1991) HT/LP parageneses linked to gravitative collapse, exhumation of the chain and shallow emplacement of intrusive granitoids (high-K calc-alkaline type, 310-290 Ma, Di Vincenzo et al., 2004).

Ar-Ar and Rb-Sr ages from 305 to 298 Ma were obtained by Laurenzi et al. (1991) on muscovites from the synkinematic leucogranites of the Monte Grighini SZ, deformed by shear movements. An age of 300 Ma could reasonably be attributed to the Sardinian LSZ.

Geodynamic Framework

Previous models

The metamorphic basement of the Corsica-Sardinia microplate is a fragment of the southern European branch of the Variscan belt, considered by most authors a collisional belt generated by Siluro-Devonian deep subduction of the oceanic crust and by Upper Devonian-Carboniferous continental collision (Burg and Matte, 1978; Autran and Cogné, 1980; Bard et al., 1980; Matte and Burg, 1981; Matte, 1983; Burg et al., 1984; Matte, 1986; Pin and Peucat, 1986; Costa and Maluski, 1988; Franke, 1989; Pin, 1990), with ubiquitous emplacement and stacking of plurikilometric nappes made up of Paleozoic cover formations in the External zone (Arthaud and Matte, 1966; Julivert, 1971; Carmignani and Pertusati, 1977) and of the crystalline Precambrian basement in the Internal one (Ries and Shackleton, 1971; Mattauer and Etchecopar, 1976; Burg and Matte, 1978; Matte,1983; Behr et al.,1984).

For the Variscan Belt in Sardinia, Carmignani et al. (1992, 1994, 2001) proposed the following evolution:

1. Sea-floor spreading between the passive continental margins of Gondwana and Armorica from the Precambrian to Lower Ordovician;
2. Middle Ordovician convergence between Gondwana and Armorica, with subduction below the Gondwana margin indicated by volcanic arc products on the continental crust (Andean type);
3. Silurian subduction of the ocean crust below Armorica, accompanied during the entire Devonian by restored passive behaviour of the Gondwana margin;
4. Lower Carboniferous continental collision between the Andean margin of Gondwana and the Armorica crust, with crustal stacking of several tectonic units;
5. Gravitative collapse of the thickened orogenic wedge, with ascent of the deepest metamorphic cores;
6. Crustal extension allowing the emplacements of calc-alkaline granitoids coeval with late Paleozoic volcanism and the opening of continental molasse basins.

This model is based on the existence of an ocean area between the Armorica plate and Gondwana (Massif Central Ocean: Matte, 1986; South Armorican Ocean: Paris and Robardet, 1990).

The suture of this ocean has been identified in the Massif Central, France, in the external Alpine Massifs (Aiguilles Rouges, Pelvoux, Belledonne, Argentera-Mercantour) and inferred in the Maures Massif (southern Hercynian suture by Bodinier et al., 1986; Matte, 1986). According to Cappelli et al. (1992) and Carmignani et al. (1992), the continuation of this suture might be represented in Sardinia by the Posada-Asinara Line, a major shear zone (cataclasites up to phyllonites for a thickness of 10-15 km), with a NW-SE direction from Asinara Island to the Anglona region and a nearly E-W direction along the Posada Valley.

The proposal by Cappelli et al. (1992), who consider the Posada-Asinara Line as a southern continuation of the suture between Armorica and Gondwana, is based principally on the regional importance of this major tectonic line and the occurrence within mylonites of several metabasite bodies interpreted as fragments of the ocean crust. A different hypothesis was put forward by Elter et al. (1999), who defined the Posada-Asinara Line as a major shear zone, probably once connected to the Ramatuelle-Plan de La Tour shear zone described in the Maures Massif by Vaucher and Buffalo (1998). The same interpretation of the Posada-Asinara Line as a major late Variscan shear zone, in contrast to the suture hypothesis, is proposed by Helbing (2003) on the basis of several analogies between the two facing sides along the line, showing that both sides follow the same evolution.

Moreover, Helbing and Tiepolo (2005), taking into account that the coeval orthogneisses of Tanaunella and Lodé are placed on opposite sides of the Posada-Asinara shear zone, emphasized that "the displacement along the Variscan Posada fault did not exceed the dimensions" of the "Mid-Ordovician magmatic Belt" of NE Sardinia and cannot have played a major role as a suture between Gondwana and Armorica, a hypothesis also rejected by Giacomini et al. (2005a).

Controversial tectonic evolution of the Variscan Belt in Sardinia

According to Carmignani et al. (1992, 1994) and Carosi and Palmeri (2002), during the first collisional phase D_1, the high grade metamorphic complex (HGMC) on the NE side of the Posada- Asinara Line overthrust the L-MGMC on the SW side of the same tectonic line. This overthrusting produced early inversion of metamorphic zones in underthrust units (Carosi and Palmeri, 2002).

Similarly juxtaposition of midcrustal rocks with supracrustal sequences was reported by Schaltegger (2000) for the Schwarzwald basement. Fluck (1980) and Rey et al. (1992) described an overthrust of a hot granulitic lower crust over an amphibolitic medium one. According to von Raumer (1998, p.419 and references therein), uplift and exhumation of deeper crustal rocks appearing as thrust sheets above less metamorphic levels could explain the ubiquitous occurrence of lower crust remnants within Variscan migmatite outcrops in the Monts du Lyonnais (Massif Central), Schwarzwald, Vosges, Eastern Austroalpine basement, External Alpine and Maures Massifs.

In Sardinia, the D_2 phase is considered purely extensional by Carmignani et al. (1992, 1994, 2001), but transpressional by Carosi and Palmeri (2002). For all these authors, the D_2 phase was responsible for extensional tectonics with exhumation of deep crustal levels and, according to Carmignani et al. (1992, 1994, 2001), for the genesis of metamorphic core complexes such as the NE part of Sardinia, the Capo Spartivento dome and the Flumendosa Antiform. During this stage, thrust surfaces were reactivated as low-angle listric faults, enhancing the detachment of external Nappes occurring on the peripheral molasse and Culm basins. At the end of the Upper Carboniferous and in the Lower Permian, the sialic root of the collisional orogenic belt was completely destroyed by extensional tectonics and by the coeval intrusion of huge masses of calc-alkaline granitoids.

On the contrary, according to Carosi and Palmeri (2002), transpressional tectonics, due to the indentation of the Cantabrian block within the narrowing Ibero-Armorican arc, lasted during the entire D_2 and D_3 phases up to the uppermost Carboniferous, and caused not only a slowing of the exhumation rate but also both reversal and west-to-northwestward displacement of the L-MGMC, bringing internal crystalline Nappes onto the migmatites of the HGMC. This means that Nappe transport changed from orthogonal to the belt in D_1 to a parallel trend in D_2.

Finally, besides the Posada-Asinara line, another discontinuity in southern Sardinia is claimed by Loi and Dabard (1997). They found the Caradoc-Ashgill sediments of Sulcis-Iglesiente (SW Sardinia, External zone) radically differ from coeval sediments of Sarrabus (SE Sardinia, External Nappe zone) as regards source of detrital materials and faunas. Some baltic Taxa are present in Sarrabus, but lacking in Sulcis-Iglesiente. The authors think that Sulcis-Iglesiente was part of north Gondwana margin, Sarrabus was distant and intermediate between Gondwana and Baltica.

Pre-Variscan metamorphic events in Sardinia

The occurrence of pre-Variscan relics in Sardinia has long been debated. Naud (1979 and references therein) considered the eclogite and granulite relics in NE Sardinia, the augen gneisses at Capo Spartivento and the phyllite cover to be of Precambrian age. Carosi et al. (1995) described a pre-Variscan deformation in the Bithia Fm and Conti et al. (1978) proposed a pre-Variscan metamorphism for the Cabitza Fm. Carmignani et al. (1982a) attributed the polymetamorphic complex at Monte Grighini (central Sardinia) to a pre-Variscan basement. Carmignani et al. (1992) proposed a Precambrian (?) age for the protolith of sedimentary-derived migmatites in NE Sardinia, and Franceschelli et al. (2005) have recently hypothesised that the banded amphibolites at Monte Plebi, north of Olbia, were remnants of a pre-Variscan mafic-silicic layered intrusion.

Helbing (2003) and Helbing and Tiepolo (2005) attributed all pre-D_2 minerals crystallised during the M_1 event to a pre-Variscan basement in NE Sardinia intruded by late orogenic Ordovician granitoids (orthogneisses) and overlain by a cover with interlayered Middle-Ordovician acidic igneous rocks (porphyroids). D_1 structural features would be the expression of Cadomian Pan-African tectonics. On the contrary, Franceschelli et al. (1982a,b), Elter et al. (1986) and Carosi and Palmeri (2002) attributed D_1 deformations of the low-to medium-grade metamorphic rocks of NE Sardinia to Variscan orogeny.

According to Helbing and Tiepolo (2005), the garnet in the lower garnet zone (i.e. garnet +albite zone of Franceschelli et al., 1982a) is a pre-Variscan relict, whereas albite derives from Variscan metamorphism. Pre-Variscan and Variscan events reflect different metamorphic grades. Helbing and Tiepolo (2005) suggest a metamorphic jump (Figure 8, p. 693) from biotite to the garnet zone marking cover/basement contact, but field and petrological evidence of this jump is not provided.

Cortesogno et al. (2004) suggest that some eclogites associated with gneisses underwent lower Ordovician metamorphism in northern Sardinia. However, in spite of the several attributions of lithotypes and deformation phases to pre-Variscan orogenic events unconfirmed by radiometric or paleontological data, the only clear evidence of pre-Variscan deformation may be observed in the External zone (Iglesiente, Sulcis in SW Sardinia), where the weaker Variscan overprint still allows recognition of angular unconformity between Cambrian-Early Ordovician and Upper Ordovician sediments. This event has been present in the literature, and called the Sardic phase, for considerable time. However, Helbing and Tiepolo (2005) refuse to consider the Sardic phase a distinct orogenic event and attribute the Sardic unconformity to "the final deformational overprint" caused by the detachment of Avalonia from Gondwana. They assert: "the Sardic phase at its type locality of SW Sardinia is not, therefore, an orogenic event, but relates to rifting and evolution of the Rheic Ocean" (see following paragraph).

The existence of pre-Variscan deformation and metamorphism in northern Sardinia has not been proven to date with exhaustive, convincing argumentation. However, further field and petrographic work, such as new geochronological data, could help to finally solve this long-debated problem.

Hun Superterrane Framework: The role of Sardinia

In the last ten years, a new, more complex geodynamic model has been proposed for the history of the Variscan Belt. It is based on the hypothesis that the "Hun Superterrane (HS)" (Stampfli, 1996, 2000), a ribbon-like assemblage of basement blocks, separated in Siluro-Devonian times from the northern Gondwana margin and collided in the Late Devonian-Carboniferous with Laurussia or Laurussia derived-fragments.

Major review papers on the HS model are those by von Raumer (1998), Stampfli et al. (2002) and von Raumer et al. (2002, 2003), to which the reader is referred for more details and bibliography.

The HS model proposes that Variscan orogeny took place between the Early Silurian to Early Permian, overprinting two previous orogenic cycles. The first cycle is that of Cadomian- Pan African orogeny (0.54-0.60 Ga), linked to the opening of the Iapetus Ocean, recognizable only in the Erzgebirge Massif (Saxo-Thuringian zone) and in the adjacent Lusatia plutons (Tichomirowa et al., 2001), in Miranda do Duro (Spain) (Lancelot et al., 1985) and Bormes orthogneiss, Maures (Maluski and Gueirard, 1978).

The Cadomian event was followed by a Neoproterozoic to Cambrian cycle that shows the persistence of an active continental margin setting, still preserved in the Alpine External Massifs and in Penninic and Austroalpine Nappes.

The surprising evolutionary parallelism between pre-Variscan orogenic cycles and the persistence of a volcanic arc parallel to the future Variscan belt in front of the Gondwana margin has led to the hypothesis that starting in Neoproterozoic times, an extremely long, ribbon-like alignment of microplates repeatedly detached from and collided with Gondwana.

At the Cambrian/Ordovician boundary, the appearance of bimodal volcanism, alkaline granites and rhyolites and within-plate basalts testify to the initial rifting stages of a thickened crust leading to the opening of the Rheic Ocean. This rifting event is recognizable in Vanoise (Briançonnais), in the Penninic nappes (Guillot et al., 2002) and in the Maures Massif (Seyler, 1983).

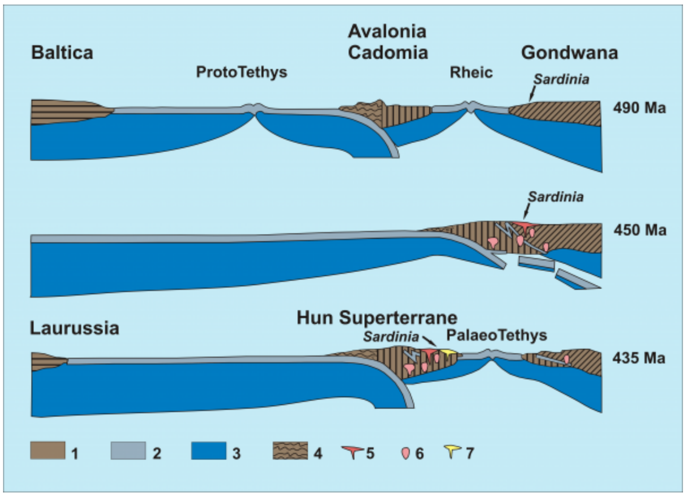
The Late Cambrian-Early Ordovician opening of the Rheic Ocean is revealed by several ophiolite associations occurring in the Alpine External Massifs and in some Austroalpine nappes (Schaltegger et al., 2002 and references therein).

Acidic magmatism from the same tectonic setting was active during the Middle to Late Ordovician. Metavolcanics and orthogneisses, once S granites and rhyolites, occur in Sapey, Ruitor gneisses (Briançonnais) (Guillot et al., 2002) and in the Argentera Massif (Rubatto et al., 2001).

Lower to Middle Ordovician (500-460 Ma) mafic-ultramafic rocks reveal ocean crust formation during the opening of the Rheic Ocean that separated Gondwana from the Hun Superterrane. The slight age difference between early mafic-ultramafic associations (496-459 Ma, Rubatto et al., 2001, Schaltegger et al., 2002 and references therein) and late to post-orogenic acidic magmatism (471- 443 Ma, Rubatto et al., 2001; Guillot et al., 2002) emphasizes the very short duration of the Rheic Ocean rifting episode and of the related Ordovician orogenic cycle.

A comparison of the Corsica-Sardinia microplate with previously-described HS microplates (Figure 13) reveals that, up to now, Corsica-Sardinia shows no clear evidence of a Cadomian-Pan African orogenic event (580-540 Ma).

Figure 13. Baltica and Gondwana



Geodynamic evolution between Baltica and Gondwana during the Early Paleozoic (modified from von Raumer et al., 2002). The paleogeographic position of Sardinia at the Gondwana margin and subsequently in the Hun Superterrane can explain the magmatic evolution from Early Middle-Ordovician calcalkaline magmatism to Late Ordovician-Silurian alkaline volcanism. 1: continental crust; 2: oceanic crust; 3: lithospheric mantle; 4: accretionary prism; 5: calc-alkaline volcanics; 6: calc-alkaline and tholeiitic intrusives; 7: alkaline volcanics.

In Sardinia, metabasite intercalations of Late Precambrian to Early Cambrian age are known in the Bithia Fm, consisting of a regressive terrigenous sequence which evolved on a continental margin (Carmignani et al., 2001 and references therein). According to Tucci (1983), these metabasites are metandesites originated during a Precambrian continental rifting event (Carmignani et al., 2001).

Cortesogno et al. (2005) reveal the existence within the External nappe zone (Gerrei, Goceano) of andesitic-dacitic lava flows having calc-alkaline affinity, associated with rhyolites. The latter, on a geochemical basis, are considered by the Authors as comparable to the Ligurian Gneiss amphibolite and Sardinia Gneiss Complexes of the inner zone (as proposed by Cortesogno et al., 2004). All these rocks are attributed by Cortesogno et al. (2005) to a "Neoproterozoic(?) Cambro-Ordovician rifting and ocean spreading event emphasized by "bimodal tholeiitic basalt and rhyolite volcanism". Actually, discarding the unlikely Neoproterozoic to Cambrian age of this event, Cortesogno et al. (2005) claim, at least, the existence of an Early Ordovician rifting event in pre-Variscan Sardinia.

Comparable but slightly older bimodal volcanism (the Leptyno-amphibolitic complex by French authors) characterises the pre-Variscan basement of the Maures Massif, Provence, France (Seyler and Boucarut, 1979; Seyler and Crevola, 1982; Seyler, 1983, 1986), a Variscan microplate with which most geodynamic reconstructions associate the Corsica-Sardinia microplate.

However, most metabasites and orthogneisses from Sardinia yielded protolith ages ranging from 470-460 Ma to 450-440 Ma (Cortesogno et al., 2004; Palmeri et al., 2004; Giacomini et al., 2005a; Helbing and Tiepolo, 2005). The Middle to Late Ordovician orthogneisses of northern Sardinia and the coeval metavolcanites known as "Porfiroidi" in southern Sardinia are generally considered, on geochemical and isotopic grounds, as calc-alkaline rocks, defined by Carmignani et al. (1994) as "Andean type" rocks.

The bimodal nature of some Ordovician magmatism has attracted the attention of Giacomini et al. (2005b), who attributed the "nearly coeval felsic and mafic magmatic rocks" to "an arc-back-arc setting of the Sardinia Corsica microplate".

A new Late Ordovician-Early Silurian detachment of the HS from the Gondwana margin caused the opening of the PaleoTethys Ocean and, northwards, the diachronous subduction of the Rheic Ocean crust below the various HS microplates The leading edge of the HS was represented by NW Iberia, the southern part of Armorica, the Massif Central, Vosges, Schwarzwald, Alpine External Massifs, Penninic and Austroalpine Nappes, Bohemian Massif and Erzgebirge. The southern passive margin (Stampfli et al., 2002; von Raumer et al., 2003) of the European Hunic Terranes includes the central-eastern Pyrénées, Montagne Noire, Alboran plate, Corsica-Sardinia, Maures, Adria plate, Carnic and Julian Alps, Dinarides and Hellenides.

From the Late Ordovician to the end of the Devonian, while the northern HS leading edge was affected by the subduction of ocean crust-producing eclogites (440-360 Ma), the southern passive edge underwent only extensional tectonics and rifting processes that supplied subvolcanic bodies of within-plate alkali basalt affinity (Ricci and Sabatini, 1978; Memmi et al., 1983; Di Pisa et al., 1992; Franceschelli et al., 2003). At the Devonian/Carboniferous boundary, the southern passive margin became active owing to northwards subduction of the Paleo-Tethys Ocean crust below the southern margin (see Stampfli et al., 2002, Figure 4).

The great uncertainty regarding the age of eclogite genesis in NE Sardinia is linked to a lack of reliable data on the metamorphic zircons involved in the HP event. However, the position of Sardinia within the group of microplates belonging to the southern passive margin of the HS during its early northwards migration towards Laurussia suggests that the HP event took place in Sardinia later than in central-northern Variscan microplates and, probably, when the southern margin changed into an active one. For this reason, the Early Visean age proposed by Giacomini et al. (2005a) seems to be plausible.

After the early HP metamorphic stage, a first Variscan phase of continental collision between the northern HS margin and Laurussia or Laurussia-derived microplates took place between 360-320 Ma, with peak metamorphism in the 340-320 Ma range (Bussy et al., 1996; Dobmeier, 1998; Giorgis et al., 1999; Bussy et al., 2000; Rubatto et al., 2001) at the end of the first Variscan deformation phase.

From the previous paragraph on Variscan metamorphic events and geochronological data in Variscan Sardinia (southern margin), a likely age of 355-335 Ma may be attributed to the first D_1-M_1 Variscan phase, with the attainment of the thermal peak at 340-335 Ma. These values are comparable to those obtained for the first Variscan phase in northern HS microplates. The complex evolution of Variscan magmatism may be described as follows.

The oldest evidence of continental collision has been found in southern Brittany and the Massif Central, where metamorphism and anatexis with granitoid emplacement is dated at 380 Ma ago (Matte, 1986). After this early granite magmatism, the first important magmatic cycle took place approximately between 350 and 330 Ma; it was characterised by high K-Mg contents interpreted as the signature of a wet subcontinental mantle contaminated by crustal and older subducted material. This magmatism is well known from the Bohemian Massif, through Schwarzwald, Vosges, the Massif Central, Tauern Window, Alpine External Massifs and Corsica (see von Raumer, 1998 and references therein), but not in Sardinia.

After the attainment of peak metamorphism at the end of the first deformation phase, subsequent rapid exhumation accompanied by the upwelling of a more primitive asthenospheric mantle and by the thinning of the crust and lithosphere gave rise to a second deformation phase and to HT-LP metamorphism in all Variscan basement blocks. This second HT-LP phase took place between 330 Ma and

300 Ma in the basement blocks of the northern leading edge of the Hun Superterrane.

In Sardinia, the second D_2-M_2 metamorphic event may be dated at 335-320 Ma (Di Vincenzo et al., 2004), comparable to the 330-300 Ma range of the event in northern HS microplates.

As regards the tectonic evolution of Variscan Sardinia, Elter et al. (1999) proposed the following reconstruction. After 350-344 Ma, the age of Barrovian metamorphism, marking the end of collision, extensional tectonics driven by mantle underplating, started in the axial zone, while compression was still active in the external one. Extensional tectonics migrated southwards in the Nappe and External zones, producing detachment along low-angle shear faults, with consequent exhumation of midcrustal rocks.

Late high-K calc-alkaline magmatism producing granites, granodiorites, diorites and rhyolites took place in the Alpine External Massifs between 307 ± 1 Ma (Mont Blanc, Bussy and Hernandez, 1997) and 292 ± 11 Ma (Gotthard, Oberli et al., 1981).

The roll-back of the Benioff plane in the late stages of Variscan orogeny strengthened the post-orogenic extensional tectonics favourable to the emplacement of late- to post-orogenic calc-alkaline to andesitic ignimbrites and lava flows in a "Central European Lower Permian Depression" (Benek et al., 1996).

This late to post-orogenic magmatism took place from the Late Permian (308 Ma, Aiguilles, Rouges, Capuzzo and Bussy, 1998) to Middle Permian (268 Ma, Dora Maira, Bussy and Cadoppi, 1996) in a wide region including Harz and the Alpine External Massifs, the Ligurian Briançonnais, Corsica and the Monte Rosa Massif.

The late- to post-orogenic high-K calc-alkaline magmatism of Sardinia and the post-orogenic one yielded respective age values in the range of 310-280 Ma and 290-280 Ma (Di Vincenzo et al., 2004), very similar to those obtained for the same type of magmatism in the Alpine External Massifs and in the Central European Lower Permian Depression (see above).

It must be emphasized that the Corsica-Sardinia batholith is one of the largest among those in the Variscan Belt (≈ 12000 Km^2 and probably tens of thousands of cubic kilometres). The time interval of its genesis from early melting (migmatites of 344 ± 7 Ma age; Ferrara et al., 1978) to the emplacement of post-orogenic Early Permian igneous rocks (290-280 Ma) covers a range of 55-65 Ma, by far the longest life for a Variscan batholith.

The remarkable spatial and temporal dimensions that characterise the genesis of the Corsica-Sardinia batholith may be explained as follows. The southern edge of the Hun Superterrane, including Sardinia, was colliding with the huge Gondwana Supercontinent, far larger than the microplates (Lizard, south Portuguese, Moravo-Silesian, Harz, Giessen; Stampfli et al., 2002) which, after detachment from Laurussia, were moving against the northern HS leading edge. This difference in the dimensions of colliding plates on opposite sides of the Hun Superterrane could explain why, as compared with the batholiths generated within the northern HS edge, the Corsica-Sardinia one is characterised by remarkably greater volumes and longer gestation times.

According to Arthaud and Matte (1975), the latest orogenic tectonic phase (305-270 Ma) produced a conjugate system of strike-slip faults consisting of NW-SE dextral faults with displacement up to 50 km and of NE-SW sinistral faults. The first dextral movement of about 150-300 km brought Corsica-Sardinia from contiguity with Provence to a position adjacent to NE Spain, as proven by the parallelism between the Early Permian paleomagnetic declination of NE Spain and Sardinia. The Cretaceous opening of the Bay of Biscay by means of a 20° anticlockwise rotation of NE Spain and Sardinia along the north Pyrenean sinistral shear zone brought Sardinia near to the Maures Massif. Finally, an Oligo-Miocene anticlockwise rotation of 30° brought Sardinia to its present-day position.

It is our hope that this paper has provided a useful synthesis regarding Variscan Sardinia.

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A. Tables

Table A.1. Selected radiometric data on emplacement age of acidic and basic metamorphic rocks from the Variscan Belt of Sardinia

Rock-type/Locality	Mineral/whole rock	Method	Age (Ma)	Reference
Orthogneiss Capo Spartivento	whole rock	Rb/Sr isochron	427±33*	Cocozza et al. (1977)
	zircon	U-Pb	478±16	Delaperrière and Lancelot (1989)
Augen gneiss Lodé	whole rock	Rb/Sr	441±33	Ferrara et al. (1978)
Orthogneiss Lodé	whole rock	Rb/Sr isochron	458±31*	Di Simplicio et al. (1974)
	zircon	U-Pb LA-ICPMS	456±14	Helbing and Tiepolo (2005)
Orthogneiss Tanaunella	zircon	U-Pb LA-ICPMS	458±7	Helbing and Tiepolo (2005)
Porphyroid Lula	zircon	U-Pb LA-ICPMS	474±13	Helbing and Tiepolo (2005)
Metadacites Sulcis	zircon	U-Pb	475±10	Garbarino et al. (2005)
Metarhyodacites Sulcis	zircon	U-Pb	387±2	Garbarino et al. (2005)
Amphibolite Posada -Asinara	whole rock	Sm/Nd	957±93	Cappelli et al. (1992)
Retrogressed eclogite Punta de li	zircon core	SHRIMP U-Pb	453±14	Palmeri et al. (2004)

Table A.2. Selected metamorphic age, based on radiometric data, from acidic, basic, and pelitic rocks in the Variscan Belt of Sardinia

Rock-type/Locality	Mineral/whole rock	Method	Age (Ma)	Interpretation	Reference
Orthogneiss Lodé	Pl, Kfs, Bt	Rb/Sr isochron	294	cooling age	Di Simplicio et al. (1974)
	Pl, Kfs, Bt	Rb/Sr isochron	298	cooling age	Di Simplicio et al. (1974)
	Bt	K/Ar	289±9	cooling age	Ferrara et al. (1978)
	Bt	K/Ar	319±10	cooling age	Ferrara et al. (1978)
Augen gneiss Lodé	Kfs, Ms, Bt	Rb/Sr	306±10	cooling age	Ferrara et al. (1978)
	Ms	K/Ar	315±10	cooling age	Ferrara et al. (1978)
	Ms	K/Ar	306±9	cooling age	Ferrara et al. (1978)
	Bt	K/Ar	304±9	cooling age	Ferrara et al. (1978)
	Bt	K/Ar	284±9	cooling age	Ferrara et al. (1978)
Banded migmatite NW Torpè	whole rock	Rb/Sr	344±7	climax of metamorphism	Ferrara et al. (1978)
	Bt, whole rock	Rb/Sr	298±2	banded migmatite passed above 300° C	Ferrara et al. (1978)
Micaschist	Ms	Rb/Sr	350±16	amphibolite fa-	Del Moro et al.