From thickening to extension in the Variscan belt — kinematic evidence from Sardinia (Italy)

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ABSTRACT

The Variscan nappe stack of SE Sardinia originated as a result of several stages of nappe imbrication during the Lower Carboniferous phases of the Variscan orogeny. The crustal shortening caused regional SSW-and W-directed thrusting, greenschist facies metamorphism and open-to-isoclinal polyphase folding. The final stage of shortening produced large-scale antiforms and synforms.

Post-collisional deformation resulted in inversion of earlier thrusts as normal faults, development of low-angle normal faults, and refolding of earlier foliation and thrust planes by asymmetric folds with subhorizontal axial planes. Facing directions of these latest folds are directed horizontally outward from the hinge zones of main antiforms, suggesting that they cannot be regarded as parasitic folds of the latest thickening phase, but instead are the consequence of vertical shortening during gravitational collapse of dome-like km-scale antiforms, leading to denudation of antiformal culminations.

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Introduction

Post-collisional deformation all along the Variscan chain of southern Europe has been well-constrained, particularly where core complex-type structures developed involving exhumation of deep crustal levels (Echtler and Malavieille, 1990; Malavieille *et al*., 1990; Gibson, 1991; Vissers, 1992; Brun and Van Den Driessche, 1994; Doblas *et al*., 1994; Faure, 1995). In all these cases, petrological data strongly support an extensional tectonic setting following the orogenic thickening stage. In the Sardinian Variscides, evidence of late orogenic extension, thermal relaxation and formation of metamorphic core complexes is still poorly constrained. Late Variscan extensional features are described only from restricted areas, namely: (i) basement pendants (High Grade Metamorphic Complex in Fig. 1) resting on the large granitoid intrusion of northern Sardinia (Di Pisa and Oggiano, 1987; Oggiano and Di Pisa, 1988; Ricci, 1992); (ii) foliated granitoids in central Sardinia (Monte Grighini), outcropping in an isolated erosional window below the post-Variscan cover (Musumeci, 1992); and (iii) the southernmost promontory of Sardinia (Capo Spartivento), where basement rocks are largely intruded by granites (Mazzoli and Visonà, 1992; Carmignani *et al*., 1994; Carosi *et al*., 1995). In all of these areas, extensional features, such as normal faults and associated folds, can be traced only to a limited extent, and the complete evolution of basement rocks from collision to late orogenic collapse is still a matter of debate.

The most complete section of the Variscan basement of Sardinia crops out in the southeastern part of the island, east of the Tertiary Campidano graben (Fig. 1). Here nappe superposition is well documented, and a significant amount of structural and stratigraphic data are available from previous works (Carmignani *et al*., 1978, 1982, 1994; Naud, 1979; Arthaud and Sauniac, 1981). However, the late orogenic evolution has not been studied in detail. Major tectonic boundaries are usually regarded as thrusts or thrusts only locally reactivated as normal faults. The aim of this paper is to demonstrate: (i) the primary role of extensional tectonics during the late orogenic evolution in this sector of the Variscan orogen; (ii) the close relationships between normal faulting and folding during extension and exhumation; and (iii) the role of inherited syn-collisional features in the development of low-angle normal faults.

Geological setting

Deformation in the Variscan basement of Sardinia resulted from N–S collision, in present-day coordinates, between the northern Armorican and the southern Gondwana continents during the Carboniferous period (Matte, 1986). The suture zone with oceanic crust remnants is exposed in northern Sardinia along the Posada–Asinara line (Cappelli *et al*., 1992) (Fig. 1). Deformation and metamorphism increase from SW (low-grade metamorphic rocks) to NE Sardinia (gneisses, migmatites), approaching the Posada–Asinara line. During the final stages of the Variscan orogeny, emplacement of large granite bodies occurred, together with normal and strike-slip faulting, resulting in the formation of Upper Carboniferous–Lower Permian basins. The Mesozoic–Tertiary carbonate sequence overlies the Variscan basement along a horizontal nonconformity. This sequence shows only limited evidence of Alpine deformation, in the form of minor normal and strike-slip faulting.

The Palaeozoic lithostratigraphic succession in southeastern Sardinia is similar in all the tectonic units. It starts with Cambrian–Lower Ordovician metasandstones, phyllites and quartzites, overlain by metaconglomerates and metavolcanic rocks (rhyolites, andesites, tuffs, etc.) of Middle Ordovician age. The Upper Ordovician is characterized by meta-arkoses and metasiltstones which pass into Silurian–Lower Devonian black shales, phyllites and metalimestones. The Middle–Upper Devonian is represented by thickly bedded metalimestones and marbles, which are overlain by Lower Carboniferous syntectonic flysch deposits (metaconglomerates, metasandstones, phyllites and quartzites with large olistolith bodies).

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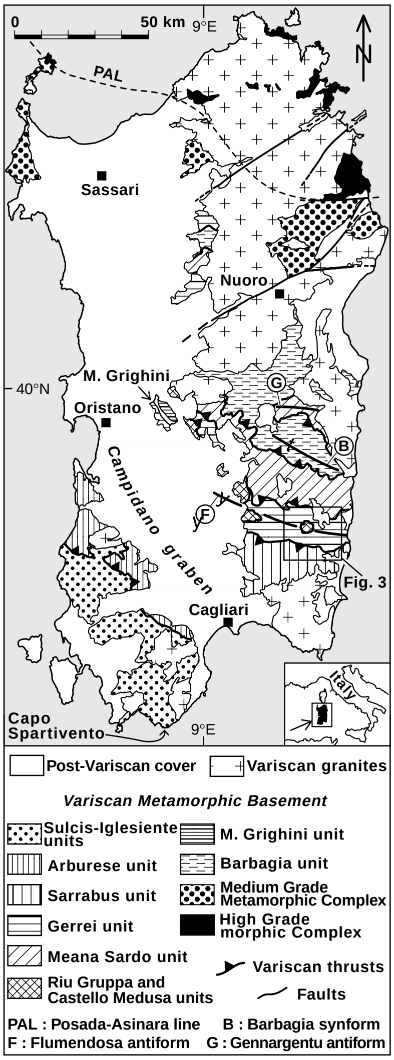

Fig. 1 Tectonic map of the Variscan basement of Sardinia.

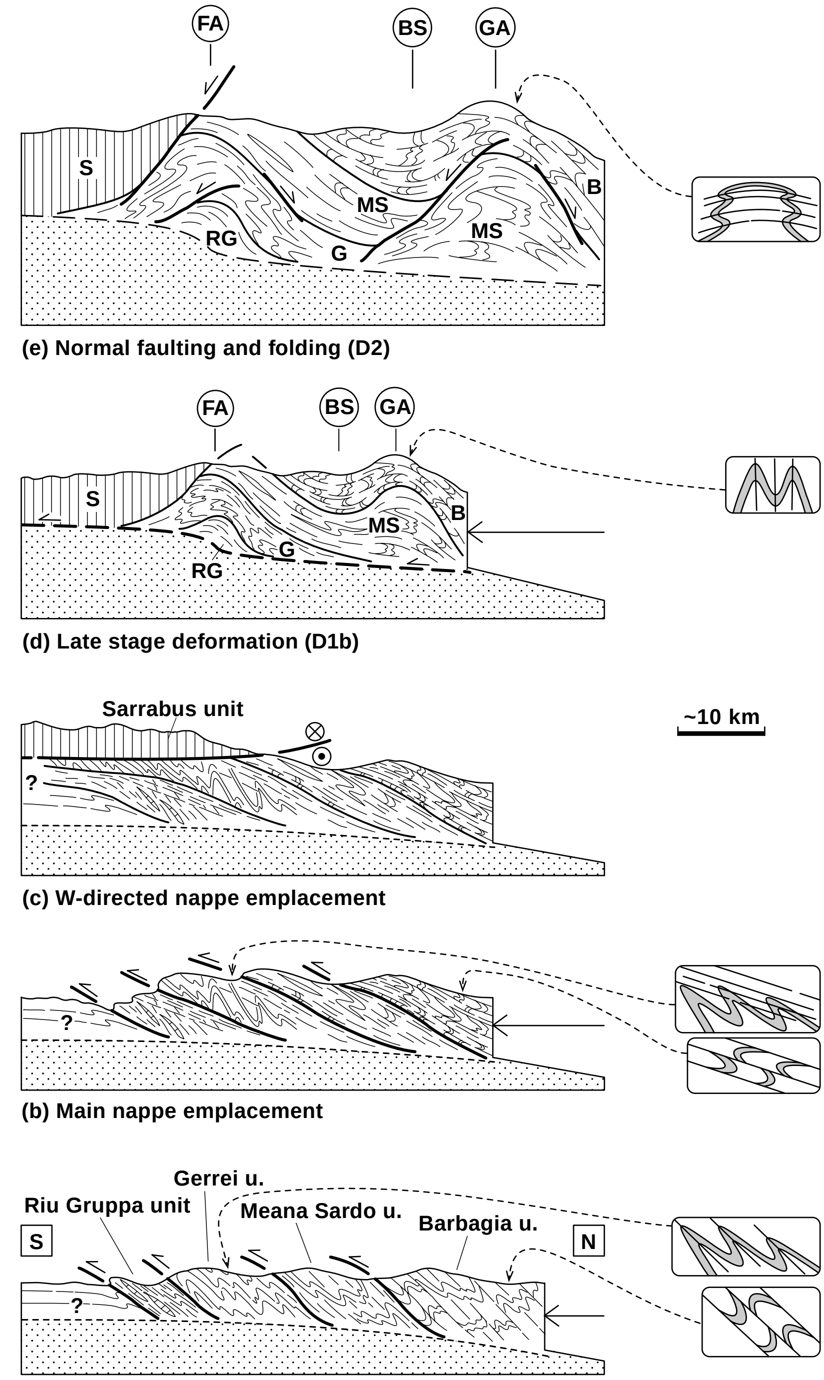

Fig. 2 Tectonic evolution of the Variscan basement of SE Sardinia during continental collision and crustal thickening (D1 phase) and subsequent extension (D2 phase), along a NNE-SSW profile. (a) S-directed early thrusting and folding phase. (b) S-directed main nappe emplacement phase. (c) W-directed nappe emplacement. (d) Late stage deformation. (e) Tectonic exhumation through normal faulting and folding (D2). S, Sarrabus unit; G, Gerrei unit; RG, Riu Gruppa unit; MS, Meana Sardo unit; B, Barbagia unit; FA, Flumendosa antiform; BS, Barbagia synform; GA, Gennargentu antiform.

Collisional evolution

The Variscan basement of central-southern Sardinia is characterized by regional thrusting, SSW-directed nappe emplacement, km-scale isoclinal folding and syntectonic regional greenschist facies metamorphism (Carmignani *et al.*, 1994 and references therein). The nappe stack is composed of (from bottom to top) Castello Medusa–Riu Gruppa unit, Gerrei unit and Meana Sardo unit (Figs 1 and 2a). In central Sardinia the Barbagia unit overrides the Meana Sardo unit, whereas in southern Sardinia the Sarrabus unit overlies on both the Gerrei and Meana Sardo units. During crustal thickening, nappe transport with top-to-the-SW sense of shear is well documented. The transport direction is well-constrained by SSW-facing isoclinal folds, NNE–SSW striking stretching lineations and top-to-the-S shear sense indicators

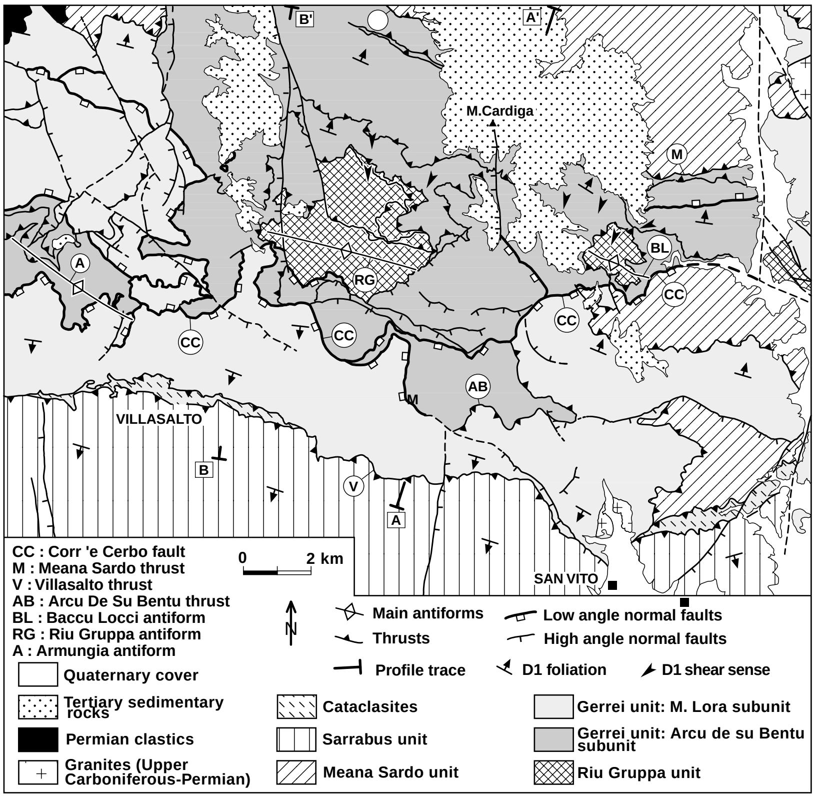

Fig. 3 Tectonic sketch map of the Variscan basement of SE Sardinia. See map location in Fig. 1.

along thrusts. The axial planar foliation of the isoclinal folds is the ubiquitous regional main foliation.

Microstructural studies and outcrop-scale observations point to a complex polyphase evolution during crustal thickening (Carmignani and Pertusati, 1977; Carmignani *et al.*, 1982; Dessau *et al.*, 1982; Carosi and Pertusati, 1990; Conti and Patta, 1998; Conti *et al.*, 1998), here collectively indicated as D1 phase, that can be summarized as follows:

(a) a SSW-directed *early thrusting and folding stage* that led to folding and mylonitization in the Riu Gruppa, Meana Sardo and Barbagia units, contemporaneous with isoclinal folding in the Gerrei units (Fig. 2a);

(b) a SSW-directed *main nappe emplacement stage*, with mylonites development in the Meana Sardo and Barbagia units and beneath the main thrusts (Fig. 2b);

(c) a *west-directed nappe emplacement stage*, documented only in the Sarrabus unit. During this stage the Sarrabus unit moved westward, overriding both the Gerrei and the Meana Sardo unit (Fig. 2c);

(d) a *late deformation stage* (D1b), again resulting from NNE–SSW-shortening, with development of large-scale upright antiforms and synforms (Flumendosa antiform, Barbagia synform, Gennargentu antiform, Fig. 1), with subhorizontal WNW–ESE-trending axes, which refolded earlier foliation and thrust planes (Fig. 2d). Small-scale folds related to this event have steep axial planes WNW–ESE-trending axes and upward facing in the normal limb of D1 folds. These folds are mostly developed in the hinge zone of the antiforms and are associated with a spaced crenulation cleavage affecting the earlier regional and mylonitic foliation. The Flumendosa antiform is the most prominent antiform developed during this stage, and we investigated it in more detail. It comprises minor axial culminations dislocated by later faulting: the Armungia antiform, the Riu Gruppa antiform and the Baccu Locci antiform (Fig. 3).

Post-thickening evolution

After crustal thickening, all the tectonic units suffered postcollisional deformation, late-orogenic extension and granite intrusion. Tectonic exhumation occurred through km-scale normal faulting and vertical shortening (D2 phase) that led to asymmetric overturned and recumbent folds. We describe below the tectonic evolution during extension of a selected area of the Variscan basement, the eastern part of the Flumendosa antiform (Fig. 3, Fig. 4), where the most complete exposure of the nappe stack crops out.

Low-angle normal faulting

The *late deformation stage* of crustal thickening produced km-scale upright antiforms (Fig. 2d). The orientation of pre-existing structures and weak layers played a dominant role in localizing D2 deformation. Low-angle normal faults developed on both limbs of antiforms, leading to tectonic unroofing of the antiformal hinge zones (Fig. 2e). Normal motion along faults is indicated by displacement of marker horizons, striations and shear sense criteria (recrystallized porphyroclasts, shear bands and S–C fabrics) in associated cataclasites and mylonites. Normal faults are WNW–ESE-striking on both sides of the antiform, parallel to antiformal axes, and dip SSW and NNE (Fig. 5). Some faults can be traced in the field for more than 15 km along-strike (e.g. Corr 'e Cerbo fault, Fig. 3). Faults show slip in the down-dip direction, with no strike-slip component. Later high-angle normal faults cut early low-angle normal faults. Special attention was paid to distinguishing S-dipping D2 faults from D1 thrusts rotated to a S-dipping orientation on the S limbs of the D2 antiform. From outcrop to thin-section-scale observation it is evident that D2 mylonites overprint D1 foliation, and that thrust planes and earlier foliations are re-orientated within D2 fault zones. The interpretation of S-dipping contacts as normal faults rather than overturned thrusts is based on the evidence that along normal faults emplacement of 'younger on older' rocks occurs, with elision of rocks in between; along thrusts instead 'older on younger' geometry occurs.

Along the southern edge of the Riu Gruppa unit (Figs 3, 4), a S-dipping low-angle normal fault is remarkably

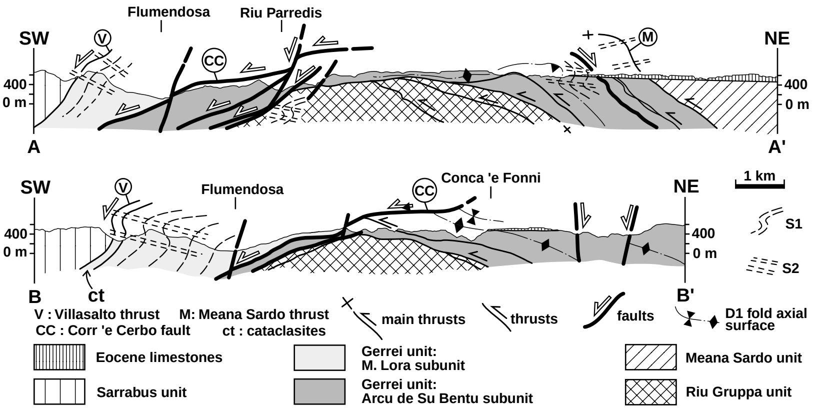

Fig. 4 Geological profiles across the Flumendosa antiform. See Fig. 3 for profile traces.

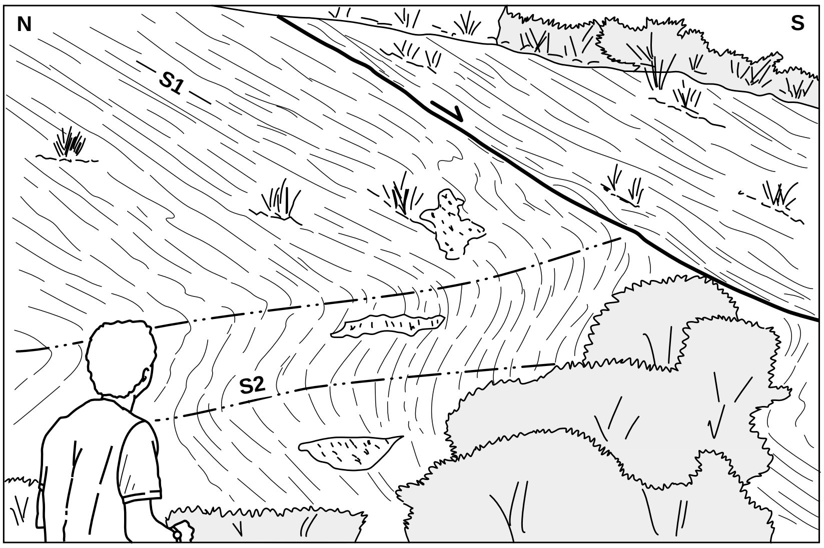

Fig. 5 D2 normal fault in central Sardinia (Barbagia unit near Aritzo). Normal faulting produced refolding of S1 foliation and development of recumbent folds, with crenulation cleavage axial plane foliation (S2).

well exposed. The fault cuts marbles and phyllites of the Riu Gruppa unit, and mylonites developed along it. Intracrystalline plastic deformation along the fault is restricted to dislocation creep in calcite; quartz only shows evidence of low-temperature plastic deformation (dislocation glide). The final product of deformation in the footwall of the fault was calcite mylonites, in which strain was accommodated by viscous flow in interconnected calcite layers, that enclose less-deformed pods of quartz-rich rock. The deformation mechanisms operating in mylonite, i.e. dislocation creep in calcite and dislocation glide in quartz, permit a rough temperature estimation of about 300°C during fault activity.

At shallow crustal levels the main D1 thrusts were also reactivated as normal faults, such as the Villasalto thrust east of Villasalto and the Arcu de su Bentu thrust. This can be demonstrated because these tectonic contacts now cut at a high-angle D1 foliation in the footwall. The thermobaric conditions of this reactivation can be constrained by the occurrence of epithermal synkinematic deposits of stibnite along the extension-related cataclasites at Villasalto and Ballao. According to Munoz *et al.* (1992), stibnite occurrence in the Variscan chain can be regarded as a metallogenic marker of brittle extension in the thermal range of 150–270°C at 0.1 kbar.

Geometry and kinematic interpretation of D2 folds

In the limbs of D1b upright antiforms there are well-developed asymmetric recumbent folds with WNW–ESE-trending horizontal axes (Fig. 6), which refold the main D1 regional. These folds have flat-lying axial planes and facing directions in the normal limb of D1 isoclinal folds that point sideways, outwards with respect to the hinge zone of the large-scale antiform (Fig. 7b). Whereas upright folds depicted in Fig. 7(a) can be regarded as parasitic folds formed during late D1 horizontal shortening and antiform development (Fig. 2d), this is not the case for the recumbent folds of Fig. 7(b). Folds with subhorizontal axial planes and 'outward' facing are interpreted to originate from vertical shortening of steeply inclined bedding and earlier foliation after antiform formation and are named hereafter 'D2 folds'. These folds cannot be regarded as parasitic folds of early D1 isoclinal folds because they refold both limbs of D1 folds. In addition, in the southern limb of the large antiform, D2 facing is SSW-orientated in the normal limb of D1 isoclinal folds (black arrow in Fig. 7b) and NNE-orientated in the overturned limb of D1 isoclinal folds (white arrow in Fig. 7b). An origin of these folds related to progressive D1 folding is unlikely because they show opposite asymmetry on the opposite limbs of the antiforms. No D2 stretching lineations developed during D2 folding, and a spaced axial planar crenulation cleavage developed only in phyllites.

Deformation was not homogeneous during vertical shortening. Metapelite layers and D1 thrust planes are sites where D2 low-angle normal faults preferentially developed. D2 folds mostly developed in the footwall of D2 normal faults and are confined to m-to-km-scale shear zones consistent with a normal component of shear. D2 folds are therefore coeval with southward normal shear in the southern limb of the Flumendosa antiform, and northward normal shear in the northern limb of the Flumendosa antiform. Vertical shortening is hence partitioned into zones of noncoaxial deformation with opposite shear sense. Low-angle normal faults developed also in the hinge zone of antiforms yet pure shear deformation is not observed in that area. Concluding,

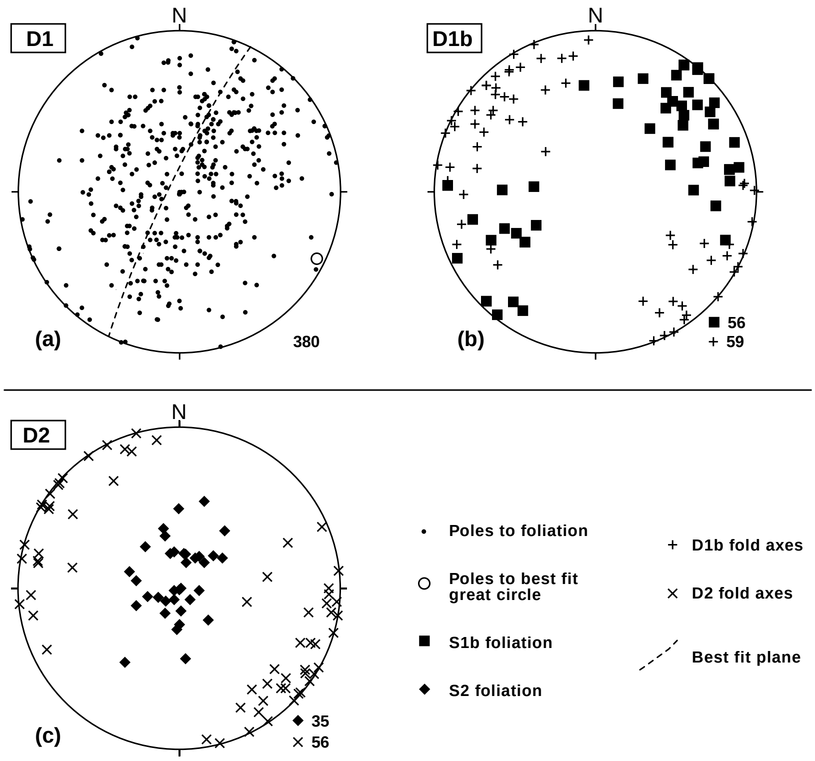

Fig. 6 Stereographic projections (equal area, lower hemisphere) of structural data from Fig. 3. The number of data plotted is shown at the lower right side of each projection. (a) Attitude of S1 foliation, refolded by D1b and D2 NW–SE orientated fold axes. (b) Attitude of S1b foliation and D1b fold axes. D1b foliation is steeply inclined in the area. (c) Attitude of S2 foliation and D2 fold axes. D2 foliation is steeply inclined in the area.

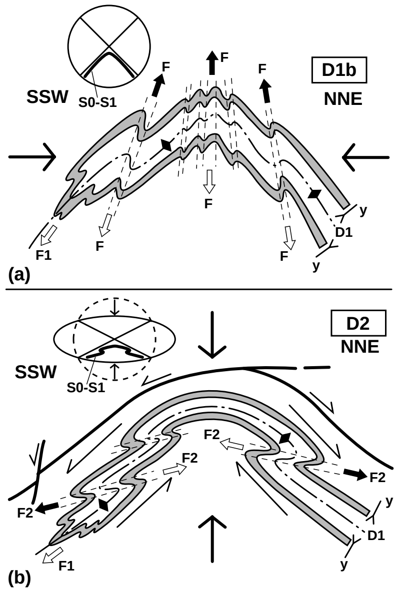

Fig. 7 Sketch showing relationships between D1 structures, major antiform development and facing directions of minor folds. D1, axial surface of isoclinal D1 folds; F1, facing direction of isoclinal D1 folds; F, Facing direction of later (postisoclinal) folds; F2, facing direction of D2 folds. Facing directions from normal limb of D1 isoclinal folds are marked by black arrows, facing directions from D1 overturned limb are marked by white arrows. (a) Facing directions of small-scale folds in the normal limb of D1 isoclinal folds are directed upward. Folds are parasitic folds formed during horizontal shortening, contemporaneously with antiform development. (b) Facing directions of small scale folds in the normal limb of D1 folds are directed sideways outward with respect to the hinge zone of the antiform. Small folds are not parasitic folds formed during antiform development, but must be younger, linked with vertical shortening.

hinge zones of the antiforms developed at the end of the thickening stage (Fig. 2d) and were unroofed during D2 normal faulting and folding (Fig. 2e).

Discussion

Metamorphic core complexes have been reported from many areas in the Variscan basement of southern Europe, where high-grade metamorphic rocks are exposed and juxtaposed against low-grade metamorphic rocks. In these areas, late orogenic extension affected wide continental domains, together with crustal thinning during the late stages of orogeny. Tectonic exhumation occurred along extensional shear zones and low-angle normal faults.

In southeastern Sardinia, late orogenic extension was not severe enough to develop a significant jump in metamorphic grade across normal faults and to exhume high-grade metamorphic rocks to the surface. From displacement of marker horizons we obtain few kilometres of displacement along the Corr ‘e Cerbo low-angle normal fault (Figs 3, 4), but the important penetrative deformation related to vertical shortening (D2 folding) could mark a more intense stretching of the crust during extension. All the other faults show minor displacement during fault activity. Inferences from deformation mechanisms along D2 low-angle normal faults point to low-temperature deformation (lower greenschists facies) during fault activity, with little change in metamorphic grade across normal faults and between upper and lower tectonic units. Although N-dipping faults are present, most of the D2 normal faults are dip southwards (Figs 3, 4). Extension is therefore asymmetric, and the Corr ‘e Cerbo fault can be regarded as the main normal fault along which most of the exhumation occurs.

In the study area, major low-angle normal faults and folds are only found close to the hinge zone of main antiforms, and both faults and antiforms show the same WNW–ESE orientation. This correspondence suggests that the large-scale upright antiforms strongly influenced the development of the normal faults and folds. Inheritence of extensional faults and folds from collisional (D1) features has been reported by many authors. In some cases normal faults and folds indicate an overall external-side down displacement with respect to the core of the antiform, and collapse occurred on earlier dome-like structures (Carmignani and Kligfield, 1990; Aerden, 1994; de Frizon Lamotte *et al.*, 1995; Hetzel *et al.*, 1995; Burg *et al.*, 1997; Dirks *et al.*, 1997; Marshak *et al.*, 1997). We consider this mode of collapse to have been a major process during vertical contraction, if earlier collisional tectonics lead to large-scale antiforms of which limbs can be vertically and asymmetrically shortened later. Further studies are needed to investigate the mechanism that led to late orogenic extension in Sardinia during the Lower Carboniferous. At the moment it is difficult to affirm whether the cause is gravitational adjustment of an unstable orogenic wedge (Platt, 1986), slab breakoff (Davies and von Blanckenburg, 1995) or convective removal of lithosphere (Platt and England, 1994). All the above models maintain high surface elevation during extension, and thick molasse basins are expected. These basins are lacking in the Variscan basement of Sardinia. Presently only a model that produces low topographic elevation, such as the subduction rollback model (Royden, 1993), can be applied. The migration in age of extension from north to south across Sardinia, as predicted by this model, however, has not yet been demonstrated. However, continental extension did not affect the Variscan basement of Sardinia for a long time, because movement along low-angle normal faults failed to produce classic metamorphic core complexes. Possibly late orogenic extension and adjustment of the unstable orogenic wedge did not go to completion, because of the onset of the well-documented strike-slip tectonics that affected the Variscan orogen during the Carboniferous and Permian (Arthaud and Matte, 1977; Badham, 1982; Ziegler, 1988).

Conclusions

The deepest tectonic units in the Variscan basement of Sardinia crop out in the core of the largest antiformal structure, the Flumendosa antiform. This WNW–ESE-trending antiform, together with other km-scale structures (Barbagia synform, Gennargentu antiform), did not originate during late orogenic extension, but resulted from NNE–SSW shortening during the latest stage of D1 crustal thickening.

During the beginning of late orogenic extension and exhumation, the limbs of the large-scale antiforms became preferred sites for normal faulting and recumbent folding. Most of faults are low-angle normal faults. Extension and tectonic exhumation were accompanied by vertical shortening, documented by folds developed in shear zones below low-angle normal faults and with subhorizontal axial planes, and opposite asymmetry and facing directions pointing outward with respect to the hinge zone of the antiform. These relationships suggest gravitational collapse of the thickened orogen as a possible cause for extension in this sector of the Variscan orogen.

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