Large-scale flat-lying isoclinal folding in extending lithosphere: Santa María de la Alameda dome (Central Iberian Massif, Spain)

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ABSTRACT

The exhumation mechanisms of deep-seated continental crust can be constrained by analyzing the structural and metamorphic imprints left in lithological ensembles. The Santa María de la Alameda dome formed during the collision of Gondwana and Laurussia in late Paleozoic time and is located in the Central Iberian Zone of the Iberian Massif (Spain). Rocks of the dome are part of the autochthonous Gondwana sections of the Variscan belt, and they occur in the Variscan hinterland. The lithostratigraphy of the dome consists of metasedimentary rocks alternating with orthogneiss massifs showing irregular and sinuous structure. The metamorphic record indicates peak pressures indicative of lower-crust depths followed by isothermal decompression to middle-upper-crust levels. Exhumation resulted in the exposure of different crystalline levels (represented by subsolidus vs. supersolidus mineral assemblages). The exhumation was accompanied by initial layer-parallel stretching and subsequent large-scale isoclinal folding developed in a heterogeneous, flat-lying shear zone with top-to-the-SE kinematics. SE-directed shearing and lateral extensional flow occurred in response to thermomechanical disequilibrium of previously thickened orogenic crust, probably assisted by coeval accretion of tectonic slices and lithospheric bending about a vertical axis. Positive feedback among partial melting, exhumation, and crustal attenuation resulted in the formation of a NE-SW–trending, migmatite-cored dome, and in refolding of early isoclinal folds and an associated axial surface regional foliation. The dome formed beneath a set of extensional detachments and was reshaped by WNW-ESE upright folds during later convergent deformation. The latter event brought in further instabilities throughout the belt, triggering in this region the development of a late extensional detachment under low-grade metamorphic conditions (top-to-the-S kinematics). The development of a regional train of flat-lying isoclinal folds is presented here as the macrostructural expression of the combination of vertical and lateral extensional flow, both of which are particularly common in orogens worldwide.

Introduction

Unraveling the tectonic evolution of the deep sections of collisional systems is fundamental to understanding the construction and dismantlement of orogens (e.g., McClelland and Gilotti, 2003; Franck et al., 2006; Schulmann et al., 2008; Jamieson and Beaumont, 2011) and to gaining knowledge about the rheology of the lower crust (Bürgmann and Dresen, 2008). Establishing the patterns and geometry of mass flow at a large scale requires proper identification and analysis of the regional structures and their interference (e.g., Alsop et al., 2001; Game et al., 2005). Deep-seated rocks follow a long deformation path and undergo severe metamorphic transformation as they are exhumed to the upper crust. That record adds the analysis of their metamorphic evolution and regional setting may constrain the context of deformation (i.e., contractual or extensional).

The Variscan belt formed from the collision between Gondwana and Laurussia in late Paleozoic time (Matte, 1991; Franke, 2000). It includes several crystalline massifs depicting a series of oroclines, like the Central Iberian and the Ibero-Armorican arcs in Iberia (Fig. 1A; Martínez Catalán, 2011). The Central Iberian Zone of the Iberian Massif represents an autochthonous section of the margin of Gondwana resting below a set of peri-Gondwanan allochthonous terranes with ophiolites and high-pressure rocks (Martínez Catalán et al., 2009). The present structure of the Central Iberian Zone is the result of (1) Paleozoic transference of peri-Gondwanan terranes onto the adjacent margin of Gondwana and associated crustal thickening (D1), followed by (2) syn- and late-orogenic collapse of the orogenic belt synchronous with the development of intracontinental strike-slip systems and orocline bends (D2–D4). In the autochthon, folds and thrusts produced initial crustal thickening (D1; Bellido et al., 1981; Capote et al., 1981; Díez Balda, 1986; Macaya et al., 1991; Rubio Pascual et al., 2012). The subsequent development of extensional shear zones and magmatic activity accompanied thermal and gravitational re-equilibration during crustal thinning (D3; Doblas, 1991; Doblas et al., 1994; Escuder Viruete et al., 1994; Barbero, 1995; Díez Balda et al., 1995; Díez Montes et al., 2010; Rubio Pascual et al., 2012). Extension was accompanied and followed by transcurrent shear zones and upright folds (D4; Martínez Catalán, 2011).

The Central System is an alpine mountain range that contains occurrences of the deep parts of the Variscan orogen. The Variscan collisional...
Figure 1. (A) Terranes and oroclines of the Variscan belt (Martínez Catalán, 2011). Arcs: BA—Bohemian; CIA—Central Iberian; IAA—Ibero-Armorican; MCA—Massif Central. Zones of the Iberian Massif: CIZ—Central Iberian; CZ—Cantabrian; GTMZ—Galicia-Tras-os-Montes; OMZ—Ossa-Morena; SPZ—South Portuguese; WALZ—West Asturian-Leonese. Shear zones and faults: BCSZ—Badajoz-Cordoba; JPSZ—Juzbado-Penalva; LPSZ—Los Pedroches; NPF—North Pyrenean; PTSZ—Porto-Tomar; SIISZ—Southern Iberian. (B) Location of the study area in the Spanish Central System. The map is a synthesis from our data and previous work (Hernández Samaniego et al., 1982; Bellido Mulas et al., 1990a, 1990b, 1990c, 1990d, 1990e; Portero García et al., 1990; Azor et al., 1991; Díaz de Neira et al., 2000; Díaz de Neira and López Olmedo, 2007; Martin Parra et al., 2008).
crust here displays a three-layer crustal section. The upper layer includes upper-crustal rocks deformed under low- to medium-grade conditions (syn-D1 to syn-D3, Barrovian metamorphic zonation formed by the chlorite, biotite, garnet, and staurolite zones; Rubio Pascual et al., 2012). The middle layer preserves some higher-grade D3 Barrovian mineral associations (kyanite, staurolite, almandine, garnet) but is mainly overprinted by D3 penetrative deformation under medium- to high-grade metamorphic conditions (staurolite and sillimanite zones). This layer hosts a set of extensional detachments (González Casado and Capote, 1987; Barbero, 1995; Hermaiz Huerta et al., 1996; Rubio Pascual et al., 2012). These low-angle faults were located at or near the brittle-ductile transition and were responsible for the progressive exhumation of high-grade rocks, and therefore overprinting of earlier formed structures (e.g., Martín Escorza, 1981, 1988; Capote et al., 2000; Tsige et al., 2002; Martín-González, 2007; Díaz-Alvarado et al., 2012). In the Guadarrama region (Fig. 1B), a major D3 ductile detachment zone of regional scale, generally referred to as the Berzosa shear zone (Capote et al., 1977; Arenas et al., 1982; Escuder Viruete et al., 1998), separates the upper-crustal (low-grade rocks) and the middle-crustal (high-grade rocks) layers. The lower layer is characterized by the presence of migmatites and is generally exposed in the cores of dome-shaped culmination. There, transposition of D1 fabrics by D2 and D3 is mainly caused by the presence of migmatites and is generally exposed in the cores of dome-shaped culmination. There, transposition of D1 fabrics by D2 and D3 is mainly caused by the presence of migmatites and is generally exposed in the cores of dome-shaped culmination.

Throughout the middle- and lower-crustal layers, the medium- to high-grade penetrative foliation records information about the structural history of deep crustal sections before being cut by (discrete) extensional detachments. A concentric pattern of trajectories of this fabric depicts domes at a regional scale (e.g., Macaya et al., 1991; Rubio Pascual et al., 2012). However, the geometry of the large structures associated with that fabric has not previously been identified, and therefore the mechanisms and mechanical behavior of this part of the collisional crust during most of its exhumation process are not fully understood.

The Santa María de la Alameda dome is located in the central-eastern part of the Central Iberian Zone and constitutes the core of the El Escorial Massif (Fig. 1B). The dome contains the transition zone between the middle- and lower-crustal layers and occurs structurally below the Berzosa shear zone. The main tectonic fabric in this region is a flat-lying high-grade foliation that pitches south and southeast under the prolongation of the Berzosa shear zone to the south, and also pitches west under the low-grade domains of the La Cañada Massif (Fig. 1B). El Escorial Massif is separated from the rest of the high-grade metamorphic domains located to its west, north, and east by a number of Variscan granitoids, thus presenting the Santa María de la Alameda dome as an interesting section to study the processes involved in the thermal re-equilibration of the Variscan collisional crust. Here, we present the tectonometamorphic evolution of this dome with the aim of describing the exhumation mechanisms, geometry of crustal flow, and resulting macrostructures formed during the transient stage spanning the crustal thickening climax of the continental collision and the advanced stages of late crustal thinning in a section of the hinterland of the Variscan belt.

Santa María de la Alameda Dome

Lithostratigraphy

The Santa María de la Alameda dome contains a layered lithostratigraphy (Fig. 2) with contacts among rock bodies showing an irregular and sinuous appearance in map view due to their strongly flattened and folded nature (Fig. 3). The lithological series in the study area consists of paragneisses with a discontinuous, yet regionally traceable, horizon of marbles, calc-silicate rocks, and two main layers of orthogneiss (Fig. 2A). One of these granitoid layers is made of granodiorite/adamellite augen orthogneisses (Fig. 4S; color pictures are available in the Data Repository) and contains possible metasedimentary and meta-igneous (quartz-diorite) enclaves and aplite/pegmatite dikes (Fig. 4B). The other layer includes monzogranite augen orthogneisses (Fig. 4C) intercalated with leucogneisses (meta-leucogranite; Fig. 4D) and minor lenses of metasedimentary rocks (Peinado, 1973; Peinado and Alvaro, 1981). The crosscutting relationships between the orthogneisses (protolith ages of ca. 490 Ma; Vialete et al., 1987; Navidad and Castañeras, 2011) and the paragneisses places a Cambrian–Ordovician minimum depositional age on the metasedimentary rock sequence, which is considered late Neoproterozoic to early Paleozoic (Capote and Fernández Casals, 1975; Navidad and Peinado, 1977).

The Santa María de la Alameda dome is cored by metatexite migmatites in which syn-migmatitic layering parallels the shape of the dome and the virtual boundary of the transition between the migmatized and non-migmatized domains. Leucosomes are found in veins and pods concurrent with the regional foliation (Fig. 5A), affected by regular folds (Fig. 5B), and in boudin necks (Fig. 5C), pressure shadows, and in small-scale shear zones (Fig. 5D). Partial melting increases toward the lower structural levels and mostly affects the paragneisses. In the upper part of the migmatized domain, the granitic veins are thin (millimeter to centimeter scale), with limited lateral continuity, and usually parallel to the compositional layering of the host rock. The percentage of discordant granitic veins increases toward the lower structural levels, as the thickness and lateral continuity of leucosomes grow. Local melting is accompanied by an increasing number of sheet-like intrusions of felsic granitoids and pegmatites (Fig. 5E), which may cut and/or be subconcordant with the regional foliation. Syn-migmatization way-up criteria, such as cauliflower structures (Fig. 5A; Burg, 1991) and the convex-up shape of the sheet-like intrusions (Fig. 5E), indicate normal polarity across the migmatized domain.

Phases of Deformation and Interference

D1: Relicts of Early Deformation

Within the study area, preservation of early tectonic events is rare and fragmentary due to intense overprinting during later deformation (Fig. 5F). A first crustal thickening event (D1) formed a planar anisotropy (S1), which is preserved as mineral relicts within S1-porphyroblasts (Fig. 6A) and as crenulated quartz ribbons and tectonic banding (Fig. 4A). Although the associated mineral assemblages and kinematics could not be established in this work, relics of kyanite in the study area (Peinado, 1973; Peinado and Alvaro, 1981) along with the presence of retro-eclogites in adjacent domains (1.4 GPa; Babero and Villaseca, 2000) indicate that some of the lower structural sections of Central Iberia were buried to depths around 40 km. D1 also produced folds (Fig. 6B), which in this part of the belt trend NE-SW. The original geometry and extent of these folds are unknown for the study area, but in upper-crustal sections exposed next to it, D1 folds trend parallel to the orogenic belt and verge to the foreland (Macaya et al., 1991).

D2: Flat-Lying Isoclinal Folds

The regional foliation (S2) formed during a second phase of deformation (D2). The attitude of S2 depicts a concentric pattern over the external parts of the dome, whereas in the migmatite domain, S2 outlines second-
Figure 2. (A) Lithostratigraphic column and (B) geological map of the Santa María de la Alameda dome.
Folding in extending lithosphere: Santa María de la Alameda dome (Iberia)

Figure 3. Composite geological cross section of the Santa María de la Alameda dome. See Figure 2B for location. Quaternary deposits are not included.

Figure 4. (A) Granodiorite/adamellite augen orthogneiss showing D2 crenulation (inset). (B) Folded (D3) aplite/pegmatite dike in the granodiorite orthogneisses. (C) Main foliation (S2) in the monzogranite augen orthogneisses and (D) leucogneisses. Note the kinematics indicated by asymmetric feldspar porphyroclasts and melanocratic aggregates (sections parallel to finite stretching lineation).
Figure 5. (A) $D_3$ subvertical crenulation of $S_2$ in the migmatized paragneisses. Note the development of a crenulation lineation in the upper half part and the way-up criteria (arrow) indicated by cauliflower structures. (B) Isoclinally folded ($D_3$) leucosomes in migmatized paragneisses. Melt-filled boudin necks that developed after folding are observed (inset in part C). FB—foliation boudinage. (C) Inset of a foliation boudinage of $S_2$ in the migmatized paragneisses (melt-filled neck). See location in B. (D) Marble lens within migmatized paragneisses. The lens is affected by (1) shearing and boudinage ($D_1$-$D_2$), (2) $D_3$ isoclinal folds, and (3) late-$D_3$ foliation boudinage. The first boudinage is not coeval with the isoclinal folding, as demonstrated by folded boudin necks (whitish granitic strips between lenses). (E) Variscan felsic granitoids and pegmatites cropping out in the core of the Santa Maria de la Alameda dome. Note the convex-up shape of the upper boundary and the flat shape of the lower one. (F) $D_3$ isoclinal folds in paragneisses.
Figure 6. (A) Mineral fabrics included in a D₂ garnet porphyroblast grown in the paragneisses. The inclusions seem to represent a pre-S₁ crenulation cleavage(?). (B) Marble beds (S₀) depicting the interference between D₂ and previous (D₁), homoaxial folds. See sketch on compass. (C) S₁ axial plane foliation formed in a hinge zone of D₂ folds and affecting a pegmatitic dike intruded in monzogranite augen orthogneisses. (D) Asymmetrically folded (D₂) fine-grained dikes in the monzogranite augen orthogneisses. The picture shows a normal-to-the-fold axes and S₂ view. Note that the microfractures of the feldspar porphyroclasts are orthogonal to the foliation and lineation. (E) D₄ mylonitic bands (dark colors) bounding a less-deformed block of orthogneiss (C‘ shear bands) in the Santa Maria de la Alameda shear zone. (F) Inset into the mylonitic bands showing C‘-S structures.
order domes and basins (Fig. 2). However, a continuity of structural fabrics from the adjacent mantling rocks into the core of the dome is observed. Additionally, S, is parallel to the axial surface of flat-lying isoclinal folds (Fig. 5F) and is also affected by foliation boudinage (Figs. 5C and 5D).

The main fabric in the paragneisses is a gneissic foliation formed by the shape-preferred orientation of quartz, feldspars, aluminosilicates, and mica (S), whereas in the orthogneisses, this fabric is a gneissic banding alternating quartz and feldspar domains, aluminosilicates, and ferromagnesian minerals (S). The mineral grains with shape-preferred orientation are aligned parallel to the main foliation either individually or as aggregates, and so are the long axes of deformed enclaves in the orthogneisses, altogether defining a mineral and a stretching lineation (L-s).

In the orthogneisses, the development of microfractures normal to L-s is common within feldspar porphyroclasts (Figs. 6C and 6D). L-s lineations are normal to D isoclinal fold axes (Figs. 6C, 6D, and 7), and, locally, a crenulation lineation can be observed (Fig. 4A). The orientation of L-s varies from NW-SE to N-S. The sense of shear inferred from mesoscopic and microscopic kinematic indicators (asymmetric boudins, σ-type porphyroclasts, and C′) is consistently to the SE-SSE (Figs. 4C, 4D, and 7). Considering the main direction of D rectilinear transport in this region, a SE-SSE fold vergence will be assumed for nomenclature purposes of the D major structures described herein.

Figure 3 shows a geological cross section of the study area, including a train of isoclinal folds (D) marked by orthogneiss layers with different thickness. The massif of granodiorite augen orthogneisses defines a periclinal closure of a large, flat-lying isoclinal syncline in the north part of the dome (Fig. 2B, zone B). Whereas the local lithostratigraphy is repeated in that area (Fig. 2B, zone C), S crosses the hinge zone of the syncline and is parallel to its limbs and axial zone. Equivalent D macrostructures were identified using these markers and confirmed by in situ observation of folded layers (Figs. 4B, 5B, 5D, 6F, and 6B). As an example, we found a folded bed of marble hosted in the core of a layer of leucogneiss (Fig. 2B, zone D). The alternation of leucogneisses and monzogranite augen orthogneisses could be an inheritance from a primary igneous architecture. However, given that the leucocratic varieties are typical border facies of the orthogneiss massifs from this region (Peinado and Alvaro, 1981), the symmetrical repetitions of these orthogneisses can be explained by flat-lying isoclinal folds. Local structural criteria, such as minor fold asymmetry (Figs. 6C and 6D), fully support this interpretation. At its core, the structural dome interferes with other D folds depicted by thin layers of monzogranite orthogneiss (Fig. 2B, zone E), and the resulting pattern provides map-scale evidence that D fold axes trend NE-SW (F; Fig. 7). Dismembered D fold limbs and the hinge zone of an anticline are preserved as tectonic blocks-in-mylonitic matrix within the Santa María de la Alameda shear zone (Fig. 2B, zone F). Rimming the uppermost part of its fault zone to the SE, discontinuous layers of marble, granodiorite augen orthogneisses, and leucogneisses depict the hinge zone of a D anticline (Fig. 2B, zone G). The D macrostructure exposed around the core of the dome is completed to the SW by a pair of flat-lying isoclinal folds marked by marbles and paragneisses and brought to an upright orientation (Fig. 2B, zone H). Altogether, the D folds depict a train of folds with no clear asymmetry at map scale (Fig. 3).

There are L > S (Figs. 4A and 4B) and S > L (Fig. 4D) varieties of S in the orthogneisses. S > L S fabrics are more common in the D hinge zones, while L > S S fabrics are found both on D fold limbs and hinge zones and exhibit larger strain. A qualitative deformation gradient is recognized at a regional scale: S increases in intensity toward the lower structural sections (Fig. 4C), and folds tighten into mylonitic shear zones that parallel the S fabric (Fig. 5B). The lower boundary of this high-strain domain is unknown.

D fabrics in this part of the orogen are dated at 335–327 Ma (U-Pb in monazite—Valverde Vaquero et al., 1996; 40Ar/39Ar in white mica—Rubio Pascual et al., 2012), and equivalent fabrics in other zones are found to have a similar or slightly older ages (335–310 Ma; 40Ar/39Ar in multigrain concentrate—Dallmeyer et al., 1997; U-Pb in zircon from synkinematic granites—Fernández-Suárez et al., 2000; U-Pb in monazite—Díez Montes, 2007; U-Pb in zircon and monazite—Carosi et al., 2012, and references therein).

D: Upright Folds

S, and related structures are gently bent into a regional structural dome, which has rhomboid shape in map view. The northwestern and southeastern limbs of the dome are shorter in length and trend approximately SW-NE, whereas the southwestern and northeastern limbs are longer and trend NW-SE (Fig. 2B). D is revealed by the elongate character of the dome in a WNW-ESW direction, coinciding with the fold axis trend of open upright folds affecting S, at a minor scale (Fig. 5A). The plunge direction of D folds is variable, being to the WNW in the northwestern half of the dome and to the ESE in the southeastern half (Fig. 8). The sinuous character of the lithological contacts in map view over the eastern and western periclinal closures of the dome and their tighter nature compared to that of the northern and southern closures reveal the existence of D folds at a larger scale (Fig. 2B, zone A). D folds (F) trend WNW-ESW (Fig. 8) and can be accompanied by local, subvertical crenulation cleavage (S, and crenulation lineation (Fig. 5A).
D3 upright folds are associated across the Iberian Massif with strike-slip ductile shear zones that moved at 315–305 Ma (40Ar/39Ar in multigrain concentrate—Dallmeyer et al., 1997; 40Ar/39Ar in white mica—Rodríguez et al., 2003; U-Pb in zircon and monazite—Valle Aguado et al., 2005).

**D4: Santa María de la Alameda Shear Zone**

The southern limb of the dome is cut by the Santa María de la Alameda shear zone (Capote et al., 2000; Tsige et al., 2002). It is a low to moderately south-dipping fault consisting of a network of low-grade mylonitic shear bands (D4), which wrap around tectonic blocks ranging from several meters up to several hundred meters in size (Fig. 2B; Martín-González, 1999). The lithostratigraphy and internal structures of these blocks mirror those of the surrounding gneisses. The trace of the fault zone is parallel to the rhomboid shape of the southern half of the dome, narrows to the east and west, and reaches a maximum thickness of ~500 m (Fig. 3). The mylonitic foliation (S4) is generally parallel to the boundaries of the shear zone or the tectonic blocks (Fig. 2B) and dips to the south (Fig. 9). D4 shear planes carry a mineral and stretching lineation (L4-S) that can be marked by quartz, mica, amphibole, and/or feldspar. L4-S trends variably between N-S and NNW-SSE (Fig. 9). Kinematic criteria, such as composite foliation S-C, shape fabrics, and C′ shear bands (Figs. 6E and 6F), consistently indicate top-to-the-S/SE tectonic transport (Capote et al., 2000). Cataclasites and fault gouges representing the last and colder pulses of deformation are common in the upper structural levels and yet show similar kinematics (Fig. 10A).

The Santa María de la Alameda shear zone (D4) is cut to the west by a massif of Variscan granitoids dated at ca. 302 ± 4 Ma (Rb-Sr and Nd in whole rock—Casillas et al., 1991; Rb-Sr in whole rock—Villaseca et al., 1995).

**Figure 8.** Equal-area stereoplots showing beta-axis calculation of the D3 fold axis (F3) obtained from S2 poles in the NW domain (upper plot), and SE domain (lower plot). Actual measurements of D3 folds over all of the study area are included in the central plot. 1% area contour lines of linear features are shown.

**Figure 9.** Equal-area stereoplot showing the mean orientation of S4 (dashed gray line) and principal trend of its associated stretching lineation (L4-S; top-to-the-SSE kinematics). Data extracted from Martín-González (2007) have been also incorporated. 1% area contour lines of lineations (gray) and poles to S4 (white) are included.
Figure 10. (A) Fault gouges and cataclasites affecting the mylonitic bands in the upper part of the Santa María de la Alameda shear zone. Transition from a ductile to a ductile-brittle regime. (B) S2 tectonic banding in sillimanite + biotite paragneiss. (C) Late S2-cordierite (pinnitized) growing at the expense of S2-biotite and S2-sillimanite needles in paragneiss. (D) Shape-preferred orientation of mineral grains (S2) in calc-silicate rocks. Note the porphyroblastic texture of plagioclase. (E) Gneissic banding (S3) in monzogranite augen orthogneiss. Mineral abbreviations after Kretz (1983).
**D2 Subvertical Faults**

Two families of near-vertical faults with NE-SW and E-W to NW-SE orientation and limited offsets cut the entire structural record described here (D2), even the massif of granitoids to the west. These faults could be Variscan and/or Alpine in age (de Vicente et al., 2007).

**Petrography**

The paragneisses have granulepidoblastic texture and consist of quartz, plagioclase, K-feldspar (microcline), sillimanite (fibrolite), garnet (almandine; López-Ruiz et al., 1975), cordierite, biotite, kyanite, andalusite, chlorite, muscovite, apatite, tourmaline, and opaque minerals (Fig. 10B). These rocks may contain leucosomes concordant with the main foliation, but leucosomes may also occur at high angles relative to the planar fabrics and defining complex networks. The main foliation (S1) preserves a stable mineral assemblage of qtz + Kfs + sll + grt + bt + pl (mineral abbreviations after Kretz, 1983). S1-garnet may contain mineral trails (Fig. 6A; mineral phases have not been identified). S1-sillimanite usually occurs as fibrolite aggregates parallel to the foliation (Fig. 10B). Sillimanite also appears as pseudomorphs after kyanite (S1?) or grown at the expense of biotite (data published by Peinado and Alvaro, 1981; we have not found these rare augen porphyroclasts (S2; Fig. 11B). S2-biotite growth at the expense of (igneous?) biotite aggregates parallel to the main fabric (S1, and L2-s) at the expense of biotite, garnet, and sillimanite (Fig. 10C).

The marbles consist of a granoblastic mosaic of calcite/dolomite together with diopside, forsterite, clinohumite, spinel, tremolite, and phlogopite. Calc-silicate rocks show a similar texture and include diopside, plagioclase, microcline, phlogopite, amphibole, garnet (andradite-grossular), and wollastite (Fig. 10D). Calc-silicates exhibit shape-preferred orientation defining the main foliation (S1). These minerals also occur in calc-silicate-rich bands forming a compositional layering.

The granodioritic orthogneisses contain quartz, K-feldspar (microcline and orthoclase), plagioclase (oligoclase-andesine), biotite, muscovite, sillimanite (fibrolite), cordierite, garnet, zircon, apatite, allanite, tourmaline, chlorite, and opaque minerals (Fig. 10E). The mineralogy of the monzonitic orthogneisses is similar to that of the granodioritic orthogneisses, but they also contain andalusite and are generally richer in aluminosilicates. Growth of garnet at the expense of (igneous?) biotite aggregates is observed in the less-deformed facies. The matrix in these two types of orthogneiss is granulepidoblastic in texture, and small igneous clasts and newly formed S2 metamorphic minerals wrap around K-feldspar porphyroclasts ranging in size from 3–4 cm in the granodioritic gneisses up to 10 cm long in the monzogranitic varieties. Both the deformation of quartz and feldspar grains and the modal abundance in phyllosilicates and aluminosilicates result in grain-size reduction within deformation bands, particularly in the K-feldspar porphyroclasts, which have perthites, may preserve orthoclase surrounded by newly formed microcline, and developed symmetric and asymmetric pressure shadows consisting of feldspar, quartz, mica, and aluminosilicates (S2; Fig. 11A). Plagioclase with myrmekitic texture also appears as smaller porphyroclasts in the monzogranitic gneisses. The leucogneisses on the other hand consist of K-feldspar (with perthitic texture), plagioclase, and quartz, with small amounts of biotite, cordierite, sillimanite (fibrolite), garnet, muscovite, andalusite, chlorite, tourmaline, apatite, and zircon. The mineral grains define a granoblastic matrix with rare augen porphyroclasts (S2; Fig. 11B). S2-biotite growth at the expense of aggregates of igneous muscovite can be observed in the less-deformed facies of these gneisses. Index metamorphic reactions common in all of the orthogneisses include growth of S2-sillimanite at the expense of biotite, and late S2-cordierite at the expense of biotite, garnet, and sillimanite (Fig. 11C). S2-cordierite occurs as poikiloblasts parallel to the main foliation and can include the minerals it grows at the expense of. S2-cordierite can also include any other mineral defining the main foliation (Fig. 11D).

Other reactions observed during late metamorphism (D3–D4) are the breakdown of biotite rims to form muscovite and opaque minerals, growth of muscovite blasts, garnet and biotite retrogression into chlorite, feldspar sericitization, and pinnitization of cordierite, sillimanite, and andalusite. These low-grade processes, along with late pervasive, ductile-brittle and brittle deformation, are extensively found in mylonites cutting the main foliation (detailed description provided by Martín-González, 2007).

**Discussion**

**Qualitative Pressure-Temperature-Time (P-T-t) Evolution**

The Santa María de la Alameda dome can be divided in two broad metamorphic domains: migmatized (Figs. 5A and 5B) and non-migmatized. The degree of partial melting increases toward the core of the dome. There are no relevant lithological differences between the original series that characterized these two domains, such as a relative absence or a larger number of fertile horizons, so the percentage of melt can be considered as an indicator of a higher metamorphic peak toward lower structural levels. Migmatization preceded D2 folding (Fig. 5A), and textural relationships indicate that the generation of melt was coeval with D2. Melt-present deformation structures spanned a range including (1) early D2 deformation (Figs. 5B and 5D), (2) the configuration of a widespread penetrative foliation (S2, Fig. 5A), and (3) a late-D2 foliation boudinage (Fig. 5C). Yet, the migmatized domain is not bent into the D3 isochron fold macrostructure, so the partial melting climax in this zone must have occurred after the major isoclinal folds were formed and acquired a recumbent or flat-lying geometry.

Qualitative estimations of the P-T evolution are based on a petrogenetic grid for partial melting of pelitic rocks in the NaKFMASH system (Spear et al., 1999), and constrained by metamorphic reactions observed in pelitic paragneisses of the study area. The two P-T-t paths proposed in Figure 12 represent supersolidus and subsolidus trajectories for the migmatized (A) and non-migmatized (B) domains, respectively.

Relicts of kyanite transformed into S2-sillimanite can be found in both the upper and lower structural sections (Peinado, 1973; Peinado and Alvaro, 1981), suggesting a decompression trajectory for D3 (Fig. 12, A1 and B1). In the upper, non-migmatized domains, sillimanite growth in the paragneisses suggests initial exhumation within a temperature range of 550–650 °C (Fig. 12, A2). In the migmatized domain, melt-present deformation structures suggest temperatures in excess of 650 °C for early D2. There, the absence of muscovite in the main fabric and a limited growth of K-feldspar point to peak temperatures in the transition to the second-sillimanite zone. The absence of orthopyroxene constrains peak temperatures below 800–850 °C (Fig. 12, B2).

In the migmatized domain, dilatant structural sites in the paragneisses (foliation boudin necks) were filled with melt, thus indicating melt flowing into low-pressure regions (Vandervaerghae, 2001) and high-temperature conditions (>650 °C) at advanced stages of S2 development. In both migmatized and non-migmatized domains, the breakdown of S2-sillimanite and biotite to form cordierite (Fig. 10C) plus the growth of andalusite porphyroblasts along with a still remarkable scarcity of muscovite after temperature peak indicate maximum pressures of ~0.4 GPa, and temperatures in excess of 550 °C for the latest D3 (Fig. 12, A3 and B3). Thermal conditions during D3 remained within a temperature range optimum for full ductile behavior of any quartzofeldspathic rock (>500 °C; Tullis and Yund, 1985).

The qualitative P-T paths inferred for D3 and the existence of kyanite relics constrain the peak pressures to minimum values ~0.4–0.5 GPa in the upper domains and ~0.8 GPa in the lower ones (D4). These values
Figure 11. (A) Asymmetric K-feldspar porphyroclast (monzogranitic orthogneiss) transformed into an aggregate of microcline-orthoclase and surrounded by quartz ribbons and layers of aluminosilicate-mica-quartz-feldspar (S₂). (B) Main foliation (S₁) in the leucogneisses. (C) Transformation of igneous? biotite into S₂-sillimanite (also retrograde to biotite), and growth of late S₂-cordierite at the expense of biotite, garnet, and S₂-sillimanite in monzogranitic orthogneiss. (D) Cordierite porphyroblasts, trapping mica and quartz, and parallel to the main foliation (S₁) in monzogranitic orthogneiss. (E) Medium-(Bt + Ms) to low-grade (Chl) C′-S mylonites of the Santa María de la Alameda shear zone. Chlorite occurs in C′ shear bands. (F) Very low-grade (Chl) C′-S mylonites of the Santa María de la Alameda shear zone. Mineral abbreviations after Kretz (1983).

Inferred metamorphic path for late D1

P-T path (migmatized, lower domains)

P-T path (non-migmatized, upper domains)

Common late P-T evolution

Figure 12. Qualitative pressure-temperature (P-T) evolution of the study area. Petrogenetic grid for KFASH, KFMASH, and NaKFMASH systems with AFM (Alkali, Fe, and Mg oxides) diagrams for principal divariant fields is used (Powell and Holland, 1990; Spear et al., 1999). Non-migmatized domains were eventually exhumed through divariant field II, but experienced former decompression out of the melting field (trajectory A1–A5). Migmatized domains evolved through divariant fields I, III, IV, and II (trajectory B1–B5). Mineral abbreviations after Kretz (1983).

agree with previous estimations (Peinado and Alvaro, 1981), although they may be too conservative if the presence of retro-eclogites in nearby areas is considered (Barbero and Villaseca, 2000).

D3 developed during retrograde metamorphism, as typified by pin nitization of cordierite and sillimanite, local growth of muscovite, and breakdown of garnet and biotite to form chlorite (Fig. 12, A4 and B4). Subsequent deformation within the Santa María de la Alameda shear zone (D3) took place at lower metamorphic grades (Fig. 12, A5 and B5), ranging from maximum temperatures of ~450–500 °C (ms ± bt; Fig. 11E) to ~250–350 °C (chl; Fig. 11F) during the latest tectonic pulses (Tsige et al., 2002; Martín-González, 2007).

Orogen-SCALE Perspective for Lateral Tectonic Flow during D2

D2 took place between two contractual phases of deformation (D1 and D3). This record dovetails with an equivalent structural sequence that alternates crustal thickening and thinning in other sections of the Central Iberian hinterland (Escuder Viruete et al., 1994; Díez Balda et al., 1995; Arenas and Martínez Catalán, 2003; Díez Montes, 2007; Díez Fernández et al., 2013). For this reason, D3 and its regional equivalents elsewhere in the Variscan belt have been generally interpreted as developing in a syncollisional tectonic setting (Druguet, 2001; Giacomini et al., 2008; Corsini and Rolland, 2009; Martínez Catalán et al., 2009; Carosi et al., 2012).

The Santa María de la Alameda dome is located in the hinge zone of the Central Iberian arc (Fig. 1), which bends several zones of the Iberian Massif and the D1 folds (Martínez Catalán, 2012). In the study area, this arc rotated the D1 folds to a NNE-SSW to NE-SW trend (Fig. 6B), parallel to the local trend of D1 folds (Fig. 7). However, the stretching lineation associated with D1 (L1-s) trends consistently NW-SE throughout this part of the Central Iberian Zone (Rubio Pascual et al., 2012), and all through the hinterland of the NW Iberian Massif in general (Díez Fernández et al., 2012, and references therein). Both L1-s and the D2 folds are parallel to the axial zone of this arc at a regional scale, giving no clue that any of them is notably bent about a vertical axis. This leads us to the idea that the formation of the arc was at least coeval with D1. Martínez Catalán (2012) bracketed the final closure of the Central Iberian arc loosely inside the interval of D1 (315–305 Ma), yet predating the Ibero-Armorican arc (305–295 Ma; Weil et al., 2013).

Shaw et al. (2012) proposed margin-parallel translation resulting in buckling of this portion of the orogen to explain the Iberian arcs. Alternatively, Martínez Catalán (2012) linked the formation of the Central Iberian arc to lateral spreading of an extruding wedge during the extensional collapse of the orogen in a strike-slip tectonic setting. Regardless of the mechanism by which the Central Iberian arc was formed, variable lithospheric thickening in the inner parts of the orocline (e.g., Gutiérrez-Alonso et al., 2004) could have accompanied the emplacement of Variscan nappes (D1). The Santa María de la Alameda dome was probably located in the inner arc, so it would have occupied an overthickened section of the belt prone to large thermomechanical re-equilibration as the Central Iberian arc was being formed. Alternatively, orocline buckling may have developed by tangential longitudinal strain, thus producing no significant change in crustal thickness.

The top-to-the-SE kinematics governing D2 indicates a vector of lateral tectonic flow channeled away from the core of the Central Iberian arc. However, the flow is distributed tangentially along the axial plane of the arc and not radially toward its core, so complementary factors other than pure gravitational gradients must have shaped the D2 tectonic flow to that of an orogen-parallel extruding wedge, such as ongoing plate convergence (e.g., Díez Fernández et al., 2012), and/or the subhorizontal main stretching direction governing the core of the arc if developed by tangential longitudinal strain.

The overall tectonic interpretation of SE-directed lateral flow is different whether linked to reverse or to normal fault movement. Central and North Iberia were dominated by hinterland-dipping crustal accretion during the Variscan collision (Martínez Catalán et al., 2009). Accordingly, movement to the SE (foreland-directed in present-day geological coordinates) is the most likely vector toward which an early Variscan thrust (D1–D2) would have had to exhume previously accreted material following the shortest path to the upper crust over a former accretionary fault. However, no thermal response to thrusting nor a compatible thermal architecture at the regional scale (i.e., metamorphic inversion) is observed in the D2 metamorphic record. Unfolding of S2 and related structures in the study area results in a flat-lying geometry for D2, with no clear indication of any particular dipping direction of its shear planes before D3. However, at a larger scale, the regional foliation (S1) pitches toward the foreland under the upper-crustal sections located east and southeast (Fig. 1B). This fea-
ture and the kinematics governing D2 tectonic flow together resemble the geometry of a normal fault, and therefore, in the Santa María de la Alameda region, D2 corresponds to an extensional event.

Framing the Tectonothermal Model and Sources of Vertical Flow

Gravity forces can trigger lateral spreading and extension within orogenic crust (Dewey, 1988; Rey et al., 2001), whereas underthrusting of thicker and more buoyant continental slabs can force previously accreted thrust nappes to rotate and acquire a more horizontal orientation. Any of these two mechanisms alone would induce bulk decompression at a large scale and therefore can favor local melting of fertile horizons (early D2 migmatites; Fig. 5B).

The granulitic lower crust of the Central Iberian Zone lacks Variscan zircons older than 312 Ma (Fernández-Suárez et al., 2006), suggesting under-stacking of continental slabs not previously affected by medium- to high-temperature deformation in Variscan times (Rubio Pascual et al., 2012).

By channeling lateral and vertical mass flow, migmatite-cored domes contribute to the re-equilibration of both thermal and gravitational instabilities in thickened lithospheres (Whitney et al., 2004). The result is a mechanically weak layer of molten lithosphere capable of accommodating strain through the lower-middle crust, while low-angle faults (detachments) accomplish extension over the middle-upper crust synchronously (Rey et al., 2001; Vanderhaeghe, 2009). Exposures of a molten layer, along with low-angle faults assisting its exhumation, can be found in structural domes throughout the Variscan hinterland of Central Iberia (e.g., Doblas, 1991; Escuder Viruete et al., 1994; Barbero, 1995; Díez Balda et al., 1995; Hernaiz Huerta et al., 1996; Díaz-Alvarado et al., 2012). Herein, the migmatized domain of the Santa María de la Alameda dome features an additional occurrence of that particular layer, the overlying high-strain domain being the structural expression of its rheological weakness and a potential decoupling level between two mechanically different crustal layers (Vanderhaeghe and Teyssier, 2001).

The local metamorphic data and the regional data set compiled in previous works indicate no metamorphic inversion but a consistent normal regional isotherms along detachment faults (Rubio Pascual et al., 2012), illustrating that the exhumation of deep-seated crust to middle-crust levels is characterized by isothermal decompression (Fig. 12). This feature is common to high-grade rocks exposed in cores of migmatitic domes, in contrast to the mantling metamorphic series, which usually exhibit P-T paths involving cooling. The decompression paths in the Santa María de la Alameda dome (Fig. 12) indicate that D2 accommodated crustal thinning driven by tectonics and/or enhanced by erosion (i.e., Teyssier and Whitney, 2002).

Possible dome-forming mechanisms in orogeny were extensively discussed by Yin (2004). Typical cases in granite-migmatite terranes include superposition of several folding events (van Staal and Williams, 1983), buckling caused by compression perpendicular to extension direction (Yin, 1991), constrictional deformation with different amounts of shortening in two horizontal directions (Yang et al., 2011), en echelon folding (Platt, 1980), large-scale shear folding (Goscombe, 1991), footwall isostatic rebound in detachment systems (Spencer, 1984), extrusion of lower crust (Beaumont et al., 2001), and diapirism (Ramberg, 1980). Each model has predictable strain patterns and resulting geometry. However, natural migmatite domes may result from the combination of various mechanisms, either acting at different times, simultaneously, and/or in a progressive basis.

Sheath folding can be discarded as a dome-forming mechanism in the study area because of a complete lack of highly curved hinge lines and minor sheath folds, and since the principal D2 stretching direction is normal to D1 isoclinal fold axes at both minor (Figs. 6C and 6D) and regional scale (Fig. 7).

A model of coeval extension-perpendicular contraction is not supported by the structural record either. Although the NE and SW limbs of the dome constitute a macrofold with an axis parallel to the regional direction of extensional flow (NW-SE vector), this fold bends and crenulates the tectonic fabric formed during D2 (S2; Fig. 5A) and all of its related major structures. Moreover, the development of conjugate strike-slip shear zones is expected within extensional systems formed this way (Diamond and Ingersoll, 2002), and also in the domes resulting from en echelon folding. This type of shear zone has not been identified in the study area.

The Santa María de la Alameda dome results from interference of two antiforms. On one hand, D2 crenulation accounts for a NW-SE antiform. Conversely, there is no evidence of subvertical crenulation developed after D2 folding to explain the SW-NE limbs of the dome. Alternatively, a former open SW-NE antiform nucleated during late D2 and/or its transition to D3 could explain the interference as well as the variably plunging character of D2 crenulation. The late D2 character of the SW-NE antiform is evident from its interference with D3 flat-lying structures (D3 macrofolds, S2, and L2-s). Although sequential or coeval formation of the two upright fold systems cannot be neatly distinguished based on available data, we ascribe the nucleation of the dome to the development of a SW-NE–trending antiform during late D2. Extension along a detachment fault would have produced crustal thinning and isostatic rebound of its footwall in response to tectonic denudation, which would then form folds with axes perpendicular to the transport direction (Spencer, 1984). The Santa María de la Alameda dome occurs in an intermediate position between two low-grade upper-crustal sections, located NW and SE (presently covered), and separated by extensional detachment systems (Fig. 1B). The extensional systems formed in Central Iberia are considered suitable for having produced significant tectonic unloading and for having induced upwarping of its footwall, as shown by exhumation of eclogitic rocks (Barbero and Villaseca, 2000) plus dramatic condensation of regional isotherms along detachment faults (Rubio Pascual et al., 2012), and by medium- to high-temperature conditions of D3 deformation (ductile behavior). The absence of geochronological data in the study area does not allow us to establish a direct link between extension carried out by detachment systems acting along the boundaries of upper- and middle-crust sections (i.e., Berzosa shear zone) and nucleation of the NE-SW antiform in the study area. However, the migmatized domain exposed in the Santa María de la Alameda dome occurs only a few meters below the structural levels showing more intense D2 strain (D2 mylonites). This zone can be considered equivalent to an extensional shear zone rooted in the middle-lower crust and could have controlled the initiation of the dome-forming process. Yet, the D2 shear zone does not juxtapose domains with highly contrasted thermal records (Fig. 12), so it may have produced no significant tectonic unroofing and thus limited isostatic uplift. This interpretation must be considered carefully because no quantitative data are available to evaluate the crustal attenuation associated with this high-strain zone.

Protracted ductile shearing with a significant pure shear component may eventually result in large-scale boudinage (e.g., Labrousse et al., 2002), which also agrees with the late-D2 record of the study area (Fig. 5C). SE-directed shearing would create perpendicular, NE-SW boudin necks. Regardless of scale, boudin necks represent domains of larger mechanical attenuation and relatively lower pressure. In this way, boudins may facilitate circulation/concentration of fluids and therefore can produce/host larger amounts of melt, which results in small-scale gravity inversions, from which domes are formed (or their growth is assisted).
The internal dynamics of the core of the Santa María de la Alameda dome are not independent from the structural evolution of its mantle, as indicated by presence of a continuous flat-lying regional foliation with consistent kinematics in both domains (S\(_0\)), the development of similar D\(_2\) structures across different sections of the dome (D\(_2\) isoclinal folds), and the absence of crosscutting relationships suggesting intrusion of the metatexite/diatexite domain into the overlying units (the extrusion model can be also discarded). In spite of that, second-order domes and basins in the core of this dome may account for multiple granitoid intrusions or convection within magmatic bodies (Bouchez and Diet, 1990; Weinberg, 1997), thus suggesting contribution of diapirs (currently unexposed) in the development of the dome (e.g., Gervais et al., 2004). Likewise, local evidences of incipient diapirism in the migmatized domain (Kruckenberge et al., 2008) is compatible with a limited role of diapirc flow in the exhumation of the core of the Santa María de la Alameda dome to some extent (e.g., Vanderhaeghe, 2004). The absence of cascading folds and shear zones formed during the upward motion of the migmatite core (Kruckenberge et al., 2008) is consistent with a limited role of diapirc flow, which, however, must be added to the exhumation produced by isotropic rebound. The absence of radial extensional flow suggests that the contribution of diapiric flow was very limited compared to other processes, but it gained importance during advanced stages of D\(_2\).

**CONCLUSIONS**

Regional mapping and tectonometamorphic analysis of deformed lithological ensembles provide knowledge about the mechanical behavior of the crust at a large scale. These approaches notably constrain the processes responsible for the wide physical distortion that the continental crust exhibits after deep burial.

Observations in the Santa María de la Alameda dome indicate that most of the exhumation of deep-seated Gondwanan crust in the autochthonous sections of the hinterland of the Variscan belt was driven by coupling of lateral extensional flow and vertical forces, probably related to footwall isostatic rebound and ascent of partially molten lithosphere. Progressive migmatization resulted in a rheologically weakened crust and the development of a broad ductile shear zone with top-to-the-SE kinematics. The combination of vertical gravity gradients and subhorizontal shearing facilitated the generation of flat-lying isoclinal folds, which preceded localized tectonic uplift and local doming as melting progressed. Regional data support a model of exhumation for the Santa María de la Alameda domain and related areas being coaxial with the development of the Central Iberian arc and also with ongoing crustal accretion. Due to renewed subhorizontal contraction, the dome was affected by steep folds, leading to further extensional activity localized in moderate-dipping normal faults.

These results corroborate field evidence from other migmatite-cored domes of the Central Iberian Zone and worldwide showing the fundamental linkages between extensional structures and gravitational and thermal re-equilibration of previously thickened orogenic crust. We also hypothesize that a comparable coupling between vertical and extensional flow could have existed in other domes of the Iberian Massif and in other orogens, and that the macrostructural expression of such a combination (large isoclinal flat-lying folds) is yet to be described in basin regions showing a layered nature. It is also important to remark that a significant part of the re-equilibrated Variscan basement along Europe is not currently exposed, and therefore there is limited access to its internal architecture. This work provides a feasible structural style for lower-crustal domains.
where vertical and lateral flows were coupled, including orogenic belts other than the Variscan.

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