

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 Gondwanan 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 crustal 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.

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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; Franěk 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; Ganne 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 to any previous or subsequent tectonic reworking process these domains may have experienced, resulting in complex ensembles when observed on Earth's surface (e.g., Harris et al., 2002). In the absence of clear structural markers, such as way-up criteria, homogeneous stratigraphic units, or age data, tracing of the large structures needs to rely on simpler criteria. Luckily, many high-grade domains in orogenic belts share a tectonostratigraphy composed of an alternation of metamorphic rocks with different protolith and rheological behavior (i.e., orthogneisses and paragneisses, schists, marbles, etc.). Mapping and structural analysis of these ensembles can help decipher the geometry and interference of exhumation structures, whereas

the analysis of their metamorphic evolution and regional setting may constrain the context of deformation (i.e., contractional 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 oroclinal bends, 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 (D_1), followed by (2) syn- and late-orogenic collapse of the orogenic belt synchronous with the development of intracontinental strike-slip systems and oroclinal bends (D_2 – D_4). In the autochthon, folds and thrusts produced initial crustal thickening (D_1 ; 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 (D_2 ; 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 (D_3 ; 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

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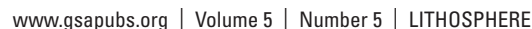


Figure 1. (A) Terranes and oroclines of the Variscan belt (Martínez Catalán, 2011). Arcs: BA—Bohemian; CIA—Central Iberian; IAA—Ibero-Armorican; MCA—Massif Central. Zones of the Iberian Massif: CIZ—Central Iberian; CZ—Cantabrian; GTMZ—Galicia-Tras-os-Montes; OMZ—Ossa-Morena; SPZ—South Portuguese; WALZ—West Asturian-Leonese. Shear zones and faults: BCSZ—Badajoz-Cordoba; JPSZ—Juzbado-Penalva; LPSZ—Los Pedroches; NPF—North Pyrenean; PTSZ—Porto-Tomar; SISZ—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; Martín 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- D_1 to syn- D_2 Barrovian metamorphic zonation formed by the chlorite, biotite, garnet, and staurolite zones; Rubio Pascual et al., 2012). The middle layer preserves some higher-grade D_1 Barrovian mineral associations (kyanite, staurolite, almandine, garnet) but is mainly overprinted by D_2 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; Hernaiz 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 D_2 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 culminations. There, transposition of D_1 fabrics by D_2 penetrative shearing is very intense and took place under high-grade metamorphic conditions (sillimanite zone and higher).

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 Repository¹) 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; Viallette et al., 1987; Navidad and Castiñeiras, 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 metatextite 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 concordant 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

D_1 : 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 (D_1) formed a planar anisotropy (S_1), which is preserved as mineral relicts within S_2 -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, relicts 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; Barbero and Villaseca, 2000) indicate that some of the lower structural sections of Central Iberia were buried to depths around 40 km. D_1 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, D_1 folds trend parallel to the orogenic belt and verge to the foreland (Macaya et al., 1991).

D_2 : Flat-Lying Isoclinal Folds

The regional foliation (S_2) formed during a second phase of deformation (D_2). The attitude of S_2 depicts a concentric pattern over the external parts of the dome, whereas in the migmatite domain, S_2 outlines second-

¹GSA Data Repository Item 2013279, color versions of photos presented in Figures 4, 5, 6, 10, and 11, is available at www.geosociety.org/pubs/ft2013.htm, or on request from editing@geosociety.org, Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, USA.

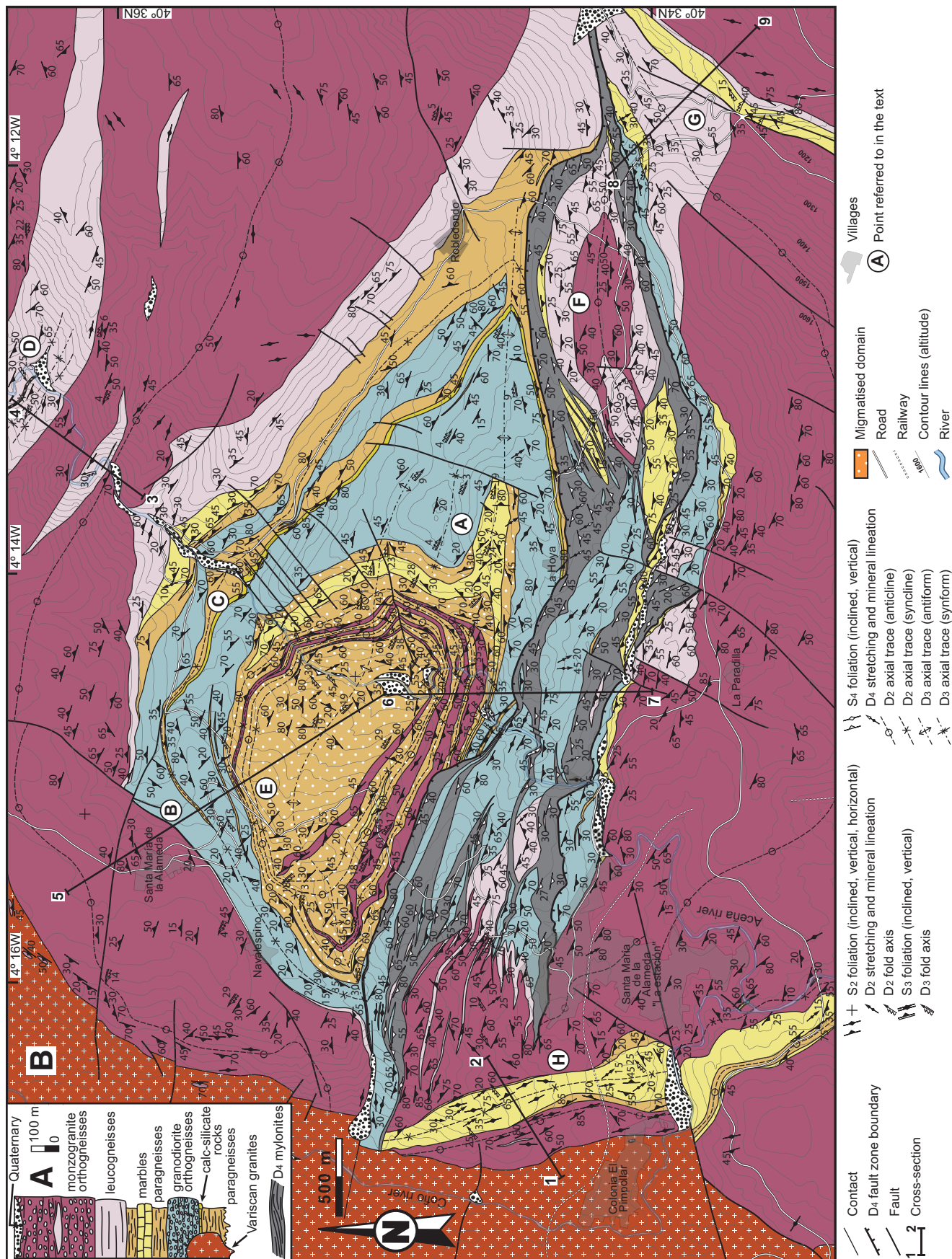


Figure 2. (A) Lithostratigraphic column and (B) geological map of the Santa Maria de la Alameda dome.

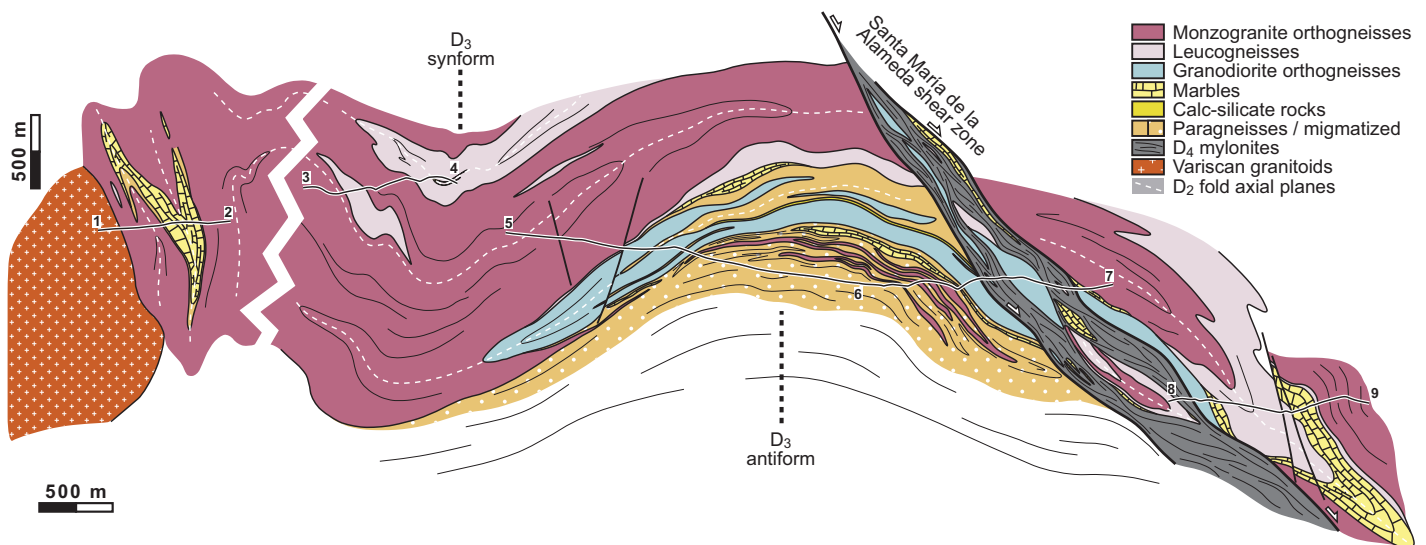


Figure 3. Composite geological cross section of the Santa María de la Alameda dome. See Figure 2B for location. Quaternary deposits are not included.

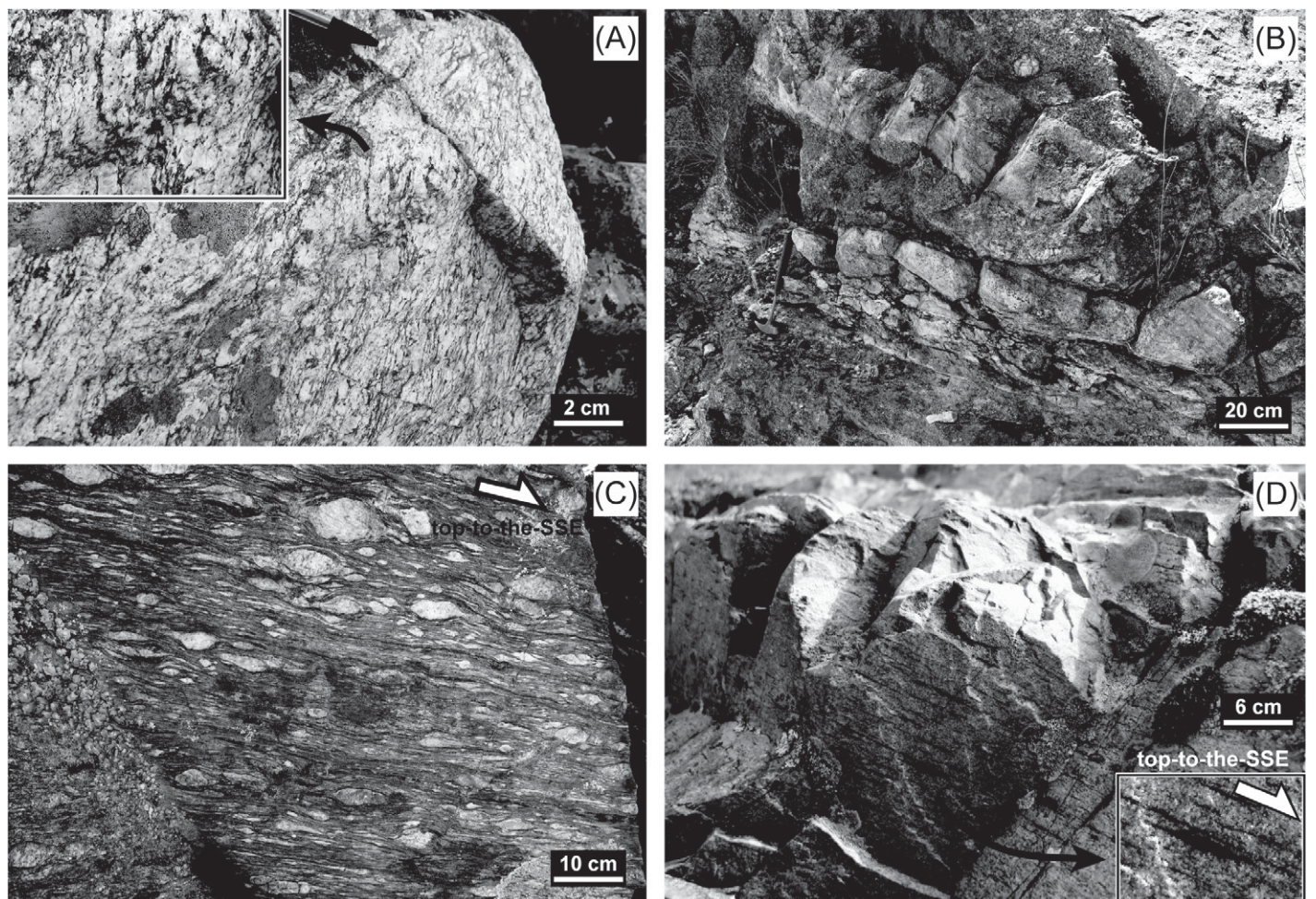


Figure 4. (A) Granodiorite/adamellite augen orthogneiss showing D_2 crenulation (inset). (B) Folded (D_2) aplite/pegmatite dike in the granodiorite orthogneisses. (C) Main foliation (S_2) in the monzogranite augen orthogneisses and (D) leucogneisses. Note the kinematics indicated by asymmetric feldspar porphyroclasts and melanocratic aggregates (sections parallel to finite stretching lineation).

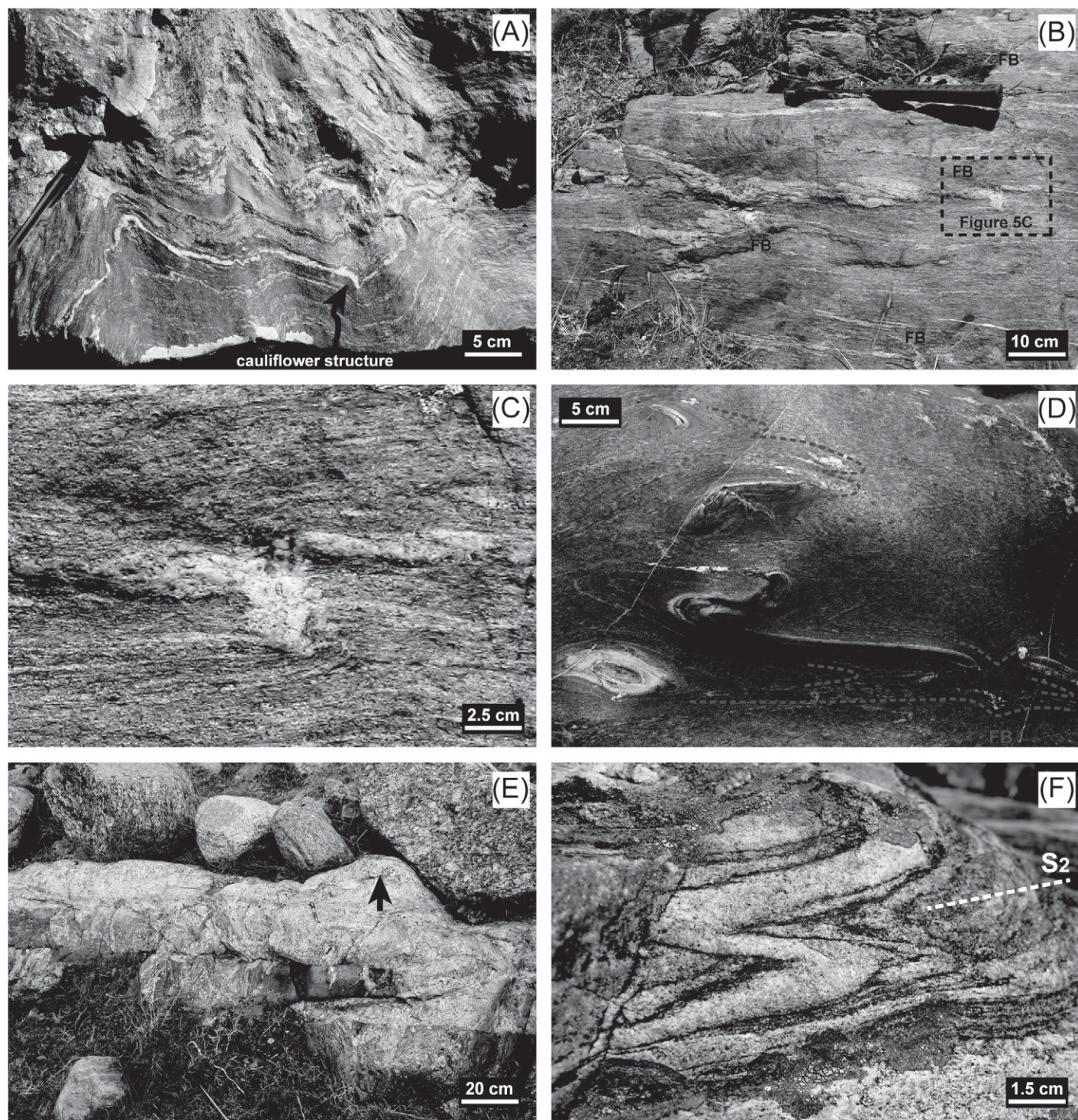


Figure 5. (A) D_3 subvertical crenulation of S_2 in the migmatized paragneisses. Note the development of a crenulation lineation in the upper half part and the way-up criteria (arrow) indicated by cauliflower structures. (B) Isoclinally folded (D_2) leucosomes in migmatized paragneisses. Melt-filled 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_2 isoclinal folds, and (3) late- D_2 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 María de la Alameda dome. Note the convex-up shape of the upper boundary and the flat shape of the lower one. (F) D_2 isoclinal folds in paragneisses.

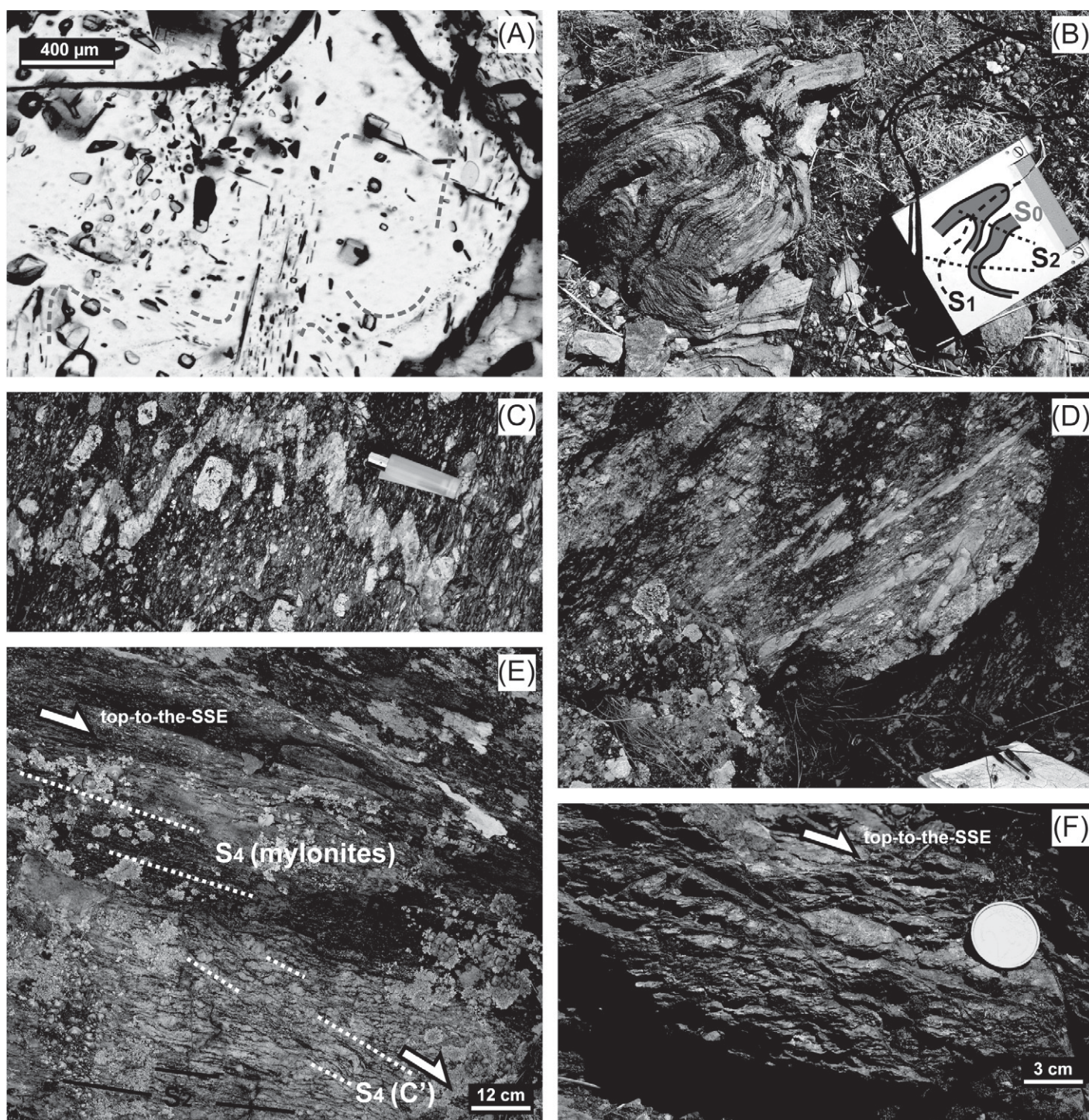


Figure 6. (A) Mineral fabrics included in a D_2 garnet porphyroblast grown in the paragneisses. The inclusions seem to represent a pre- S_2 crenulation cleavage(?). (B) Marble beds (S_0) depicting the interference between D_2 and previous (D_1), homoaxial folds. See sketch on compass. (C) S_2 axial plane foliation formed in a hinge zone of D_2 folds and affecting a pegmatitic dike intruded in monzogranite augen orthogneisses. (D) Asymmetrically folded (D_2) fine-grained dikes in the monzogranite augen orthogneisses. The picture shows a normal-to-the fold axes and S_2 view. Note that the microfractures of the feldspar porphyroclasts are orthogonal to the foliation and lineation. (E) D_4 mylonitic bands (dark colors) bounding a less-deformed block of orthogneiss (C' shear bands) in the Santa María 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_2 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_2), whereas in the orthogneisses, this fabric is a gneissic banding alternating quartz and feldspar domains, aluminosilicates, and ferromagnesian minerals (S_2). 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_2 -s).

In the orthogneisses, the development of microfractures normal to L_2 -s is common within feldspar porphyroclasts (Figs. 6C and 6D). L_2 -s lineations are normal to D_2 isoclinal fold axes (Figs. 6C, 6D, and 7), and, locally, a crenulation lineation can be observed (Fig. 4A). The orientation of L_2 -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_2 tectonic transport in this region, a SE-SSE fold vergence will be assumed for nomenclature purposes of the D_2 major structures described herein.

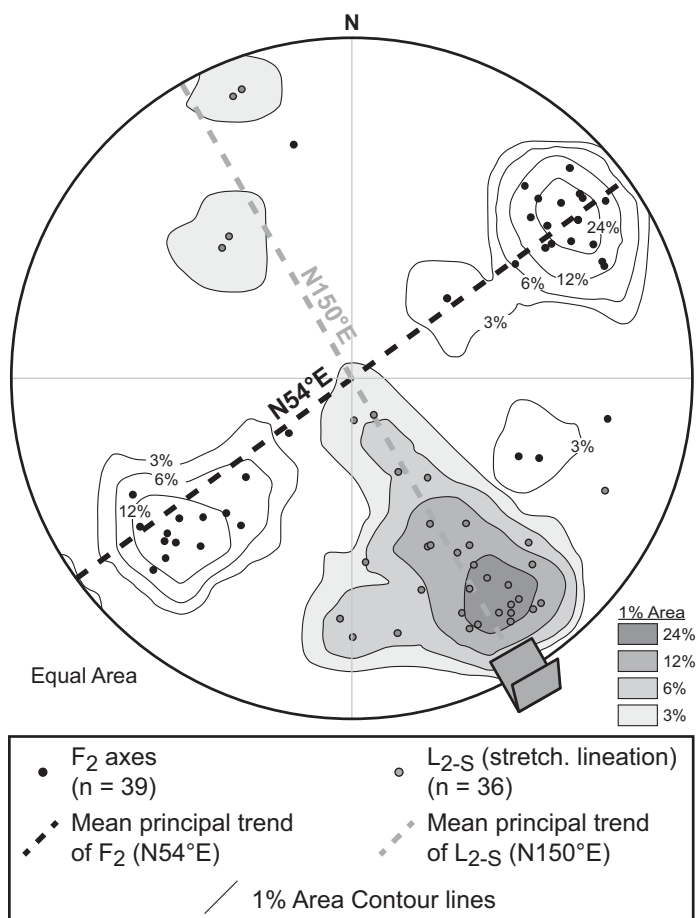


Figure 7. Equal-area stereoplot showing the NNW-SSE orientation of the D_2 finite stretching lineation (L_2 -s; top-to-the-SSE kinematics) and the broad perpendicular trend of the D_2 fold axes (F_2 ; black dots). 1% area contour lines of lineations (gray) and fold axes (white) are included.

Figure 3 shows a geological cross section of the study area, including a train of isoclinal folds (D_2) 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_2 crosscuts the hinge zone of the syncline and is parallel to its limbs and axial zone. Equivalent D_2 macrostructures were identified using these markers and confirmed by in situ observation of folded layers (Figs. 4B, 5B, 5D, 5F, 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_2 folds depicted by thin layers of monzogranite orthogneiss (Fig. 2B, zone E), and the resulting pattern provides map-scale evidence that D_2 fold axes trend NE-SW (F_2 ; Fig. 7). Dismembered D_2 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_2 anticline (Fig. 2B, zone G). The D_2 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_2 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_2 in the orthogneisses. $S > L$ S_2 fabrics are more common in the D_2 hinge zones, while $L > S$ S_2 fabrics are found both on D_2 fold limbs and hinge zones and exhibit larger strain. A qualitative deformation gradient is recognized at a regional scale: S_2 increases in intensity toward the lower structural sections (Fig. 4C), and folds tighten into mylonitic shear zones that parallel the S_2 fabric (Fig. 5B). The lower boundary of this high-strain domain is unknown.

D_2 fabrics in this part of the orogen are dated at 335–327 Ma (U-Pb in monazite—Valverde Vaquero et al., 1996; $^{40}\text{Ar}/^{39}\text{Ar}$ 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; $^{40}\text{Ar}/^{39}\text{Ar}$ 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_3 : Upright Folds

S_2 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_3 is revealed by the elongate character of the dome in a WNW-ESE direction, coinciding with the fold axis trend of open upright folds affecting S_2 at a minor scale (Fig. 5A). The plunge direction of D_3 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_3 folds at a larger scale (Fig. 2B, zone A). D_3 folds (F_3) trend WNW-ESE (Fig. 8) and can be accompanied by local, subvertical crenulation cleavage (S_3) 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 moder-

ately 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 (L_{4-s}) that can be marked by quartz, mica, amphibole, and/or feldspar. L_{4-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).

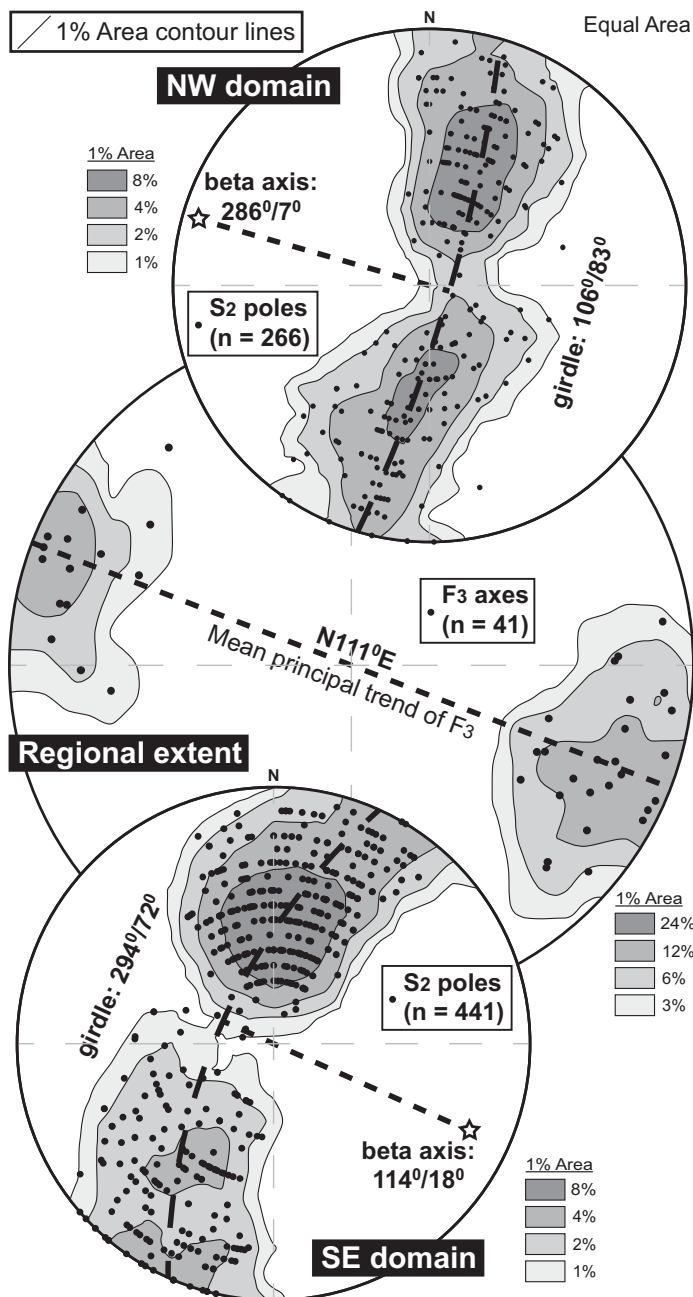


Figure 8. Equal-area stereoplots showing beta-axis calculation of the D₃ fold axis (F₃) obtained from S₂ poles in the NW domain (upper plot), and SE domain (lower plot). Actual measurements of D₃ folds over all of the study area are included in the central plot. 1% area contour lines of linear features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

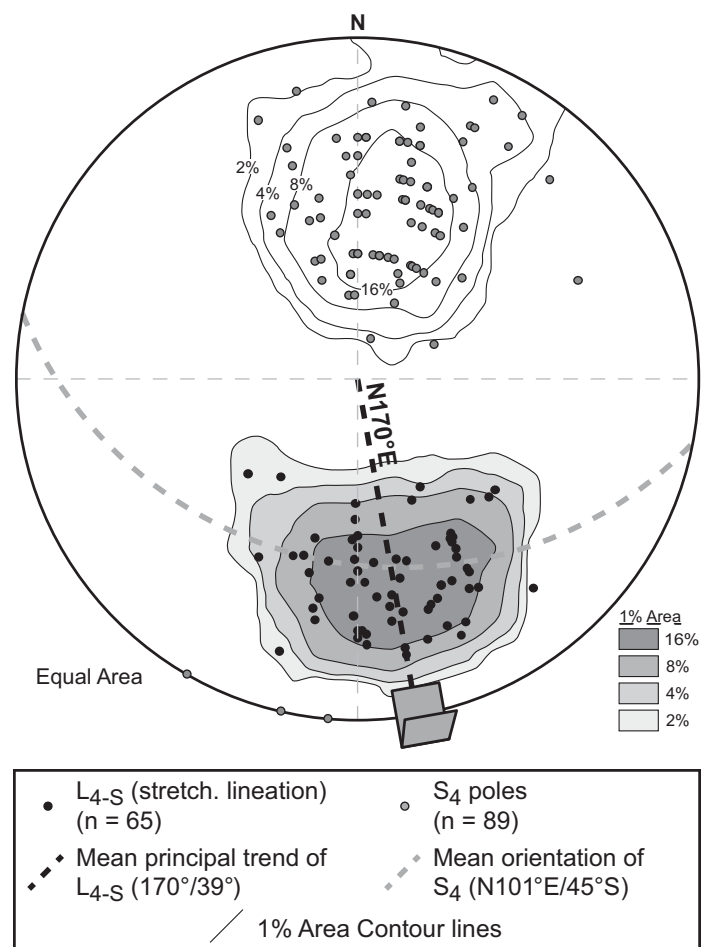


Figure 9. Equal-area stereoplots showing the mean orientation of S₄ (dashed gray line) and principal trend of its associated stretching lineation (L_{4-s}; top-to-the-SSE kinematics). Data extracted from Martín-González (2007) have been also incorporated. 1% area contour lines of lineations (gray) and poles to S₄ (white) are included.

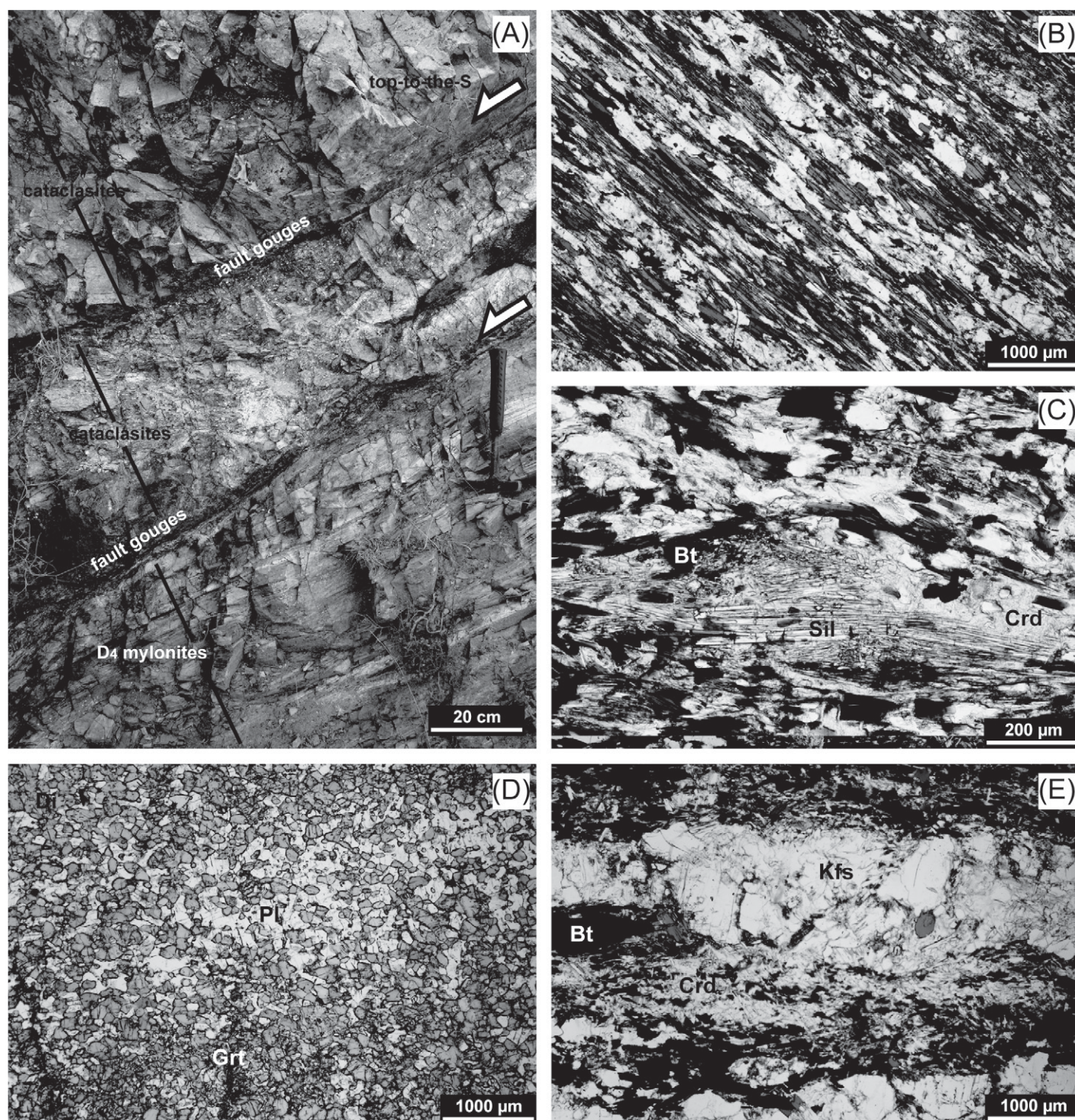


Figure 10. (A) Fault gouges and cataclasites affecting the mylonitic bands in the upper part of the Santa María de la Alameda shear zone. Transition from a ductile to a ductile-brittle regime. (B) S_2 tectonic banding in sillimanite + biotite paragneiss. (C) Late S_2 -cordierite (pinnitized) growing at the expense of S_2 -biotite and S_2 -sillimanite needles in paragneiss. (D) Shape-preferred orientation of mineral grains (S_2) in calc-silicate rocks. Note the porphyroblastic texture of plagioclase. (E) Gneissic banding (S_2) in monzogranite augen orthogneiss. Mineral abbreviations after Kretz (1983).

***D*₅: 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 (*D*₅), 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 granolepidoblastic 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 (*S*₂) preserves a stable mineral assemblage of qtz + Kfs + sil + grt + bt + pl (mineral abbreviations after Kretz, 1983). *S*₂-garnet may contain mineral trails (Fig. 6A; mineral phases have not been identified). *S*₂-sillimanite usually occurs as fibrolite aggregates parallel to the foliation (Fig. 10B). Sillimanite also appears as pseudomorphs after kyanite (*S*₁?) or grown at the expense of biotite (data published by Peinado and Alvaro, 1981; we have not found these textural relationships). *S*₂-sillimanite in some instances is overgrown by later *S*₂-cordierite porphyroblasts, which grew parallel to the main fabric (*S*₂ and *L*₂-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 wollastonite (Fig. 10D). Calc-silicates exhibit shape-preferred orientation defining the main foliation (*S*₂). 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 monzogranitic orthogneisses is similar to that of the granodioritic orthogneiss, 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 granolepidoblastic in texture, and small igneous clasts and newly formed *S*₂ 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 (*S*₂; 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 (*S*₂; Fig. 11B). *S*₂-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 *S*₂-sillimanite at the expense of biotite, and late *S*₂-cordierite at the expense of biotite, garnet, and sillimanite (Fig. 11C). *S*₂-cordierite occurs as poikiloblasts parallel to the main foliation

and can include the minerals it grows at the expense of. *S*₂-cordierite can also include any other mineral defining the main foliation (Fig. 11D).

Other reactions observed during late metamorphism (*D*₃–*D*₅) 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 configured 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 *D*₃ folding (Fig. 5A), and textural relationships indicate that the generation of melt was coeval with *D*₂. Melt-present deformation structures spanned a range including (1) early *D*₂ deformation (Figs. 5B and 5D), (2) the configuration of a widespread penetrative foliation (*S*₂; Fig. 5A), and (3) a late-*D*₂ foliation boudinage (Fig. 5C). Yet, the migmatized domain is not bent into the *D*₂ isoclinal 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 *S*₂-sillimanite can be found in both the upper and lower structural sections (Peinado, 1973; Peinado and Alvaro, 1981), suggesting a decompression trajectory for *D*₂ (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 *D*₂. 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 (Vanderhaeghe, 2001) and high-temperature conditions (>650 °C) at advanced stages of *S*₂ development. In both migmatized and non-migmatized domains, the breakdown of *S*₂-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 *D*₂ (Fig. 12, A3 and B3). Thermal conditions during *D*₂ 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 *D*₂ and the existence of kyanite relicts constrain the peak pressures to minimum values ~0.4–0.5 GPa in the upper domains and ~0.8 GPa in the lower ones (*D*₁). These values

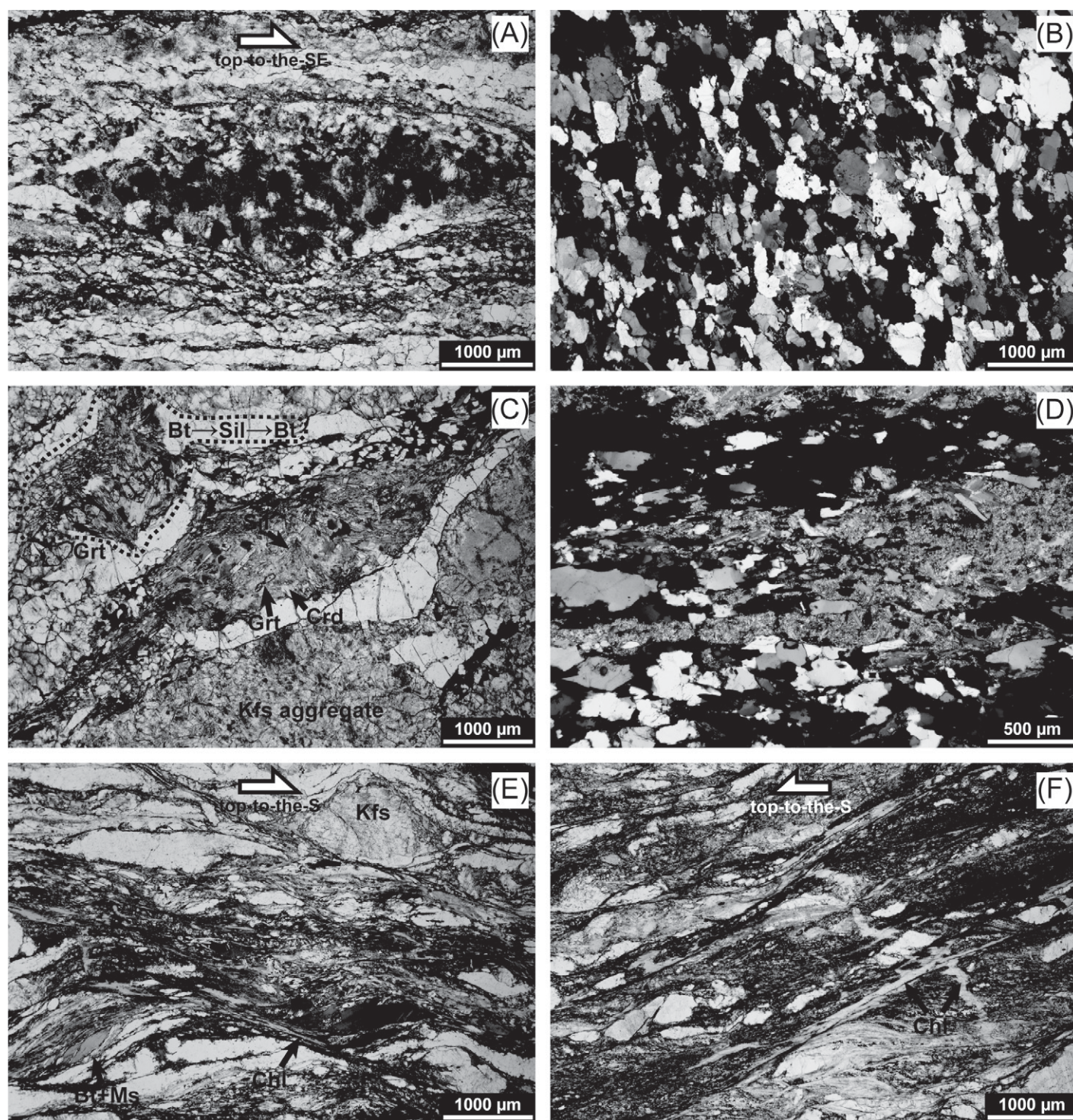


Figure 11. (A) Asymmetric K-feldspar porphyroblast (monzogranitic orthogneiss) transformed into an aggregate of microcline-orthoclase and surrounded by quartz ribbons and layers of aluminosilicate-mica-quartz-feldspar (S_2). (B) Main foliation (S_2) in the leucogneisses. (C) Transformation of igneous? biotite into S_2 -sillimanite (also retrograde to biotite), and growth of late S_2 -cordierite at the expense of biotite, garnet, and S_2 -sillimanite in monzogranitic orthogneiss. (D) Cordierite porphyroblasts, trapping mica and quartz, and parallel to the main foliation (S_2) 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).

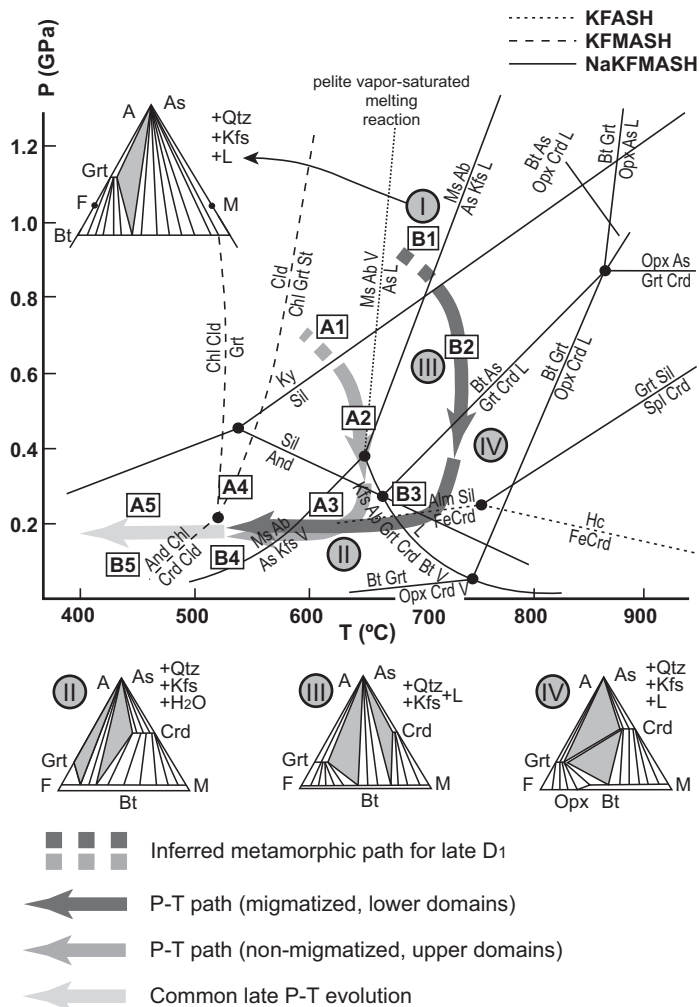


Figure 12. Qualitative pressure-temperature (P-T) evolution of the study area. Petrogenetic grid for KFMASH, NaKFMASH, and NaKFMASH systems with AFM (Alkali, Fe, and Mg oxides) diagrams for principal divariant fields is used (Powell and Holland, 1990; Spear et al., 1999). Non-migmatized domains were 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-ritization 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 (D₄) 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 contractional 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, D₂ and its regional equivalents elsewhere in the Variscan belt have been generally interpreted as developing in a syn-convergent 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 D₂ folds (Fig. 7). However, the stretching lineation associated with D₂ (L₂-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 L₂-s and the D₃ 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 D₂. Martínez Catalán (2012) bracketed the final closure of the Central Iberian arc loosely inside the interval of D₃ (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 oroclinal bend (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, oroclinal 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 exhumate 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 a 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 (S₂) pitches toward the foreland under the upper-crustal sections located east and southeast (Fig. 1B). This fea-

ture and the kinematics governing D_2 tectonic flow together resemble the geometry of a normal fault, and therefore, in the Santa María de la Alameda region, D_2 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 D_2 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 understacking 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 overthickened 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 synchronically (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 zonation all across the region during D_2 (Peinado, 1973; López-Ruiz et al., 1975; Bellido et al., 1981; Rubio Pascual et al., 2012). 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 migmatite 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 D_2 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 sheath 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 D_2 stretching direction is nor-

mal to D_2 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 D_2 (S_2 ; 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, D_3 crenulation accounts for a NW-SE antiform. Conversely, there is no evidence of subvertical crenulation developed after D_3 folding to explain the SW-NE limbs of the dome. Alternatively, a former open SW-NE antiform nucleated during late D_2 and/or its transition to D_3 could explain the interference as well as the variably plunging character of D_3 crenulation. The late D_2 character of the SW-NE antiform is evident from its interference with D_2 flat-lying structures (D_2 macrofolds, S_2 , and L_2 -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 D_2 . 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 D_2 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 D_2 strain (D_2 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 D_2 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- D_2 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_2), 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 Diot, 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 (Figs. 5A and 5E) point out the participation of vertical flow related to gravity inversion 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 (Kruckenberg et al., 2008) is compatible with a limited role of diapiric flow, which, however, must be added to the exhumation produced by isostatic 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 .

Coupling of Crustal Thinning and Large-Scale Folding

All of the D_2 structures define a section of a broad, heterogeneous ductile shear zone of unknown thickness, dipping toward the foreland of the belt, and with general noncoaxial tectonic flow dominated by simple shear. This is typified by (1) the development of a penetrative axial plane foliation and the amplification of isoclinal folds (S_2), (2) a stronger penetrative strain at deeper levels of the D_2 macrostructure and over fold limbs, and (3) the development of asymmetric structures, such as folds (Fig. 6D), σ -type porphyroclasts (Fig. 4C) and aggregates (Fig. 11A), and C' -S systems, plus (4) the orthogonal character of the principal finite stretching direction relative to the indicators of local vorticity axes seen in the best sections of the aforementioned structures (Passchier and Simpson, 1986; Robin and Cruden, 1994; Iacopini et al., 2008), also including boudin necks (Fig. 5C).

In ductile shear zones, newly formed folds can result either from instability within the shear flow or because of favorable orientation of preshearing/folding layering (Carreras et al., 2005). The possibility that the major D_2 folds derive from amplified D_1 structures cannot be ruled out (e.g., Díez Fernández et al., 2013), although field observations suggest that early folds were bent into new flat-lying folds (Fig. 6B). D_2 deformed an ensemble that inherited a planar anisotropy (S_1 [Fig. 4A] and D_1 [Fig. 13A]). Indeed, subhorizontal shearing during the D_1 - D_2 transition could have also enhanced existing anisotropies and eased subsequent folding (Fig. 13B). The geometrical relationship between the initial D_2 shear planes in the study area and the former planar anisotropies ($S_0 + S_1$) favored layer-parallel extension (pre- D_2 folding boudinage; Fig. 5D). Such a relationship is compatible with S_1 dipping to the hinterland and west-dipping underthrusting of continental crust (Fig. 13A). Early D_2 shear zones evolved according to the local and regional gradients in pressure and probably linked to form larger extensional structures.

As tectonic uplift related to underthrusting progressed, it added to isostatic rebound triggered by ongoing lateral extension (Fig. 13C). Both of them gathered with an increasingly important vertical component of diapiric flow, which, through positive feedback between exhumation (decompression) and partial melting, led to progressive tilting of the infrastructure. Such a combination steepened the shear planes of the lateral components of tectonic flow and, eventually, caused the existing planar

anisotropies ($S_0 + S_1 + \text{proto-}S_2$) to lie in the shortening field, triggering the formation of the D_2 folds (Fig. 13D; e.g., Froitzheim, 1992), as well as their subsequent flattening and amplification during incremental shear strain (Fig. 13E; e.g., Ramsay et al., 1983). Individualized domes (e.g., Santa María de la Alameda dome) emerged during the climax of isostatic rebound (upwarping) and melt production (diapirs) and the waning stages of lateral D_2 flow (large-scale boudinage?) (Figs. 5C and 13F).

Horizontal compression gained importance across the orogen over the collision episode and gave rise to vertical structures in the study area, such as WNW-ESE steep folds (D_3 folds; Fig. 13G). At this stage, the Santa María de la Alameda dome would have acquired its present geometry. Thickening associated with that event brought the orogenic crust back to a disequilibrium state, fostering late extensional tectonics with dominant S-directed tectonic transport to release gravitational potential (D_4 ; Fig. 13H). The Central Iberian arc would also have a tight geometry at this point. The neutral surface (see Ramsay and Huber, 1987, p. 458) of this orocline separated domains with layer-parallel extension around the outer arc from domains with tangential, layer-parallel contraction around the inner arc. The inward migration of this surface toward the core of the arc as it tightened (Frehner, 2011) implies that domains subjected to horizontal contraction during initial orocline development may be eventually affected by horizontal extension in a similar direction. Interestingly, D_3 folds in the study area fit with compression perpendicular to the axial plane of the Central Iberian arc, while D_4 represents a shift to layer-parallel extension relative to the trace of the arc in the Santa María de la Alameda region.

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 coeval 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 basement 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

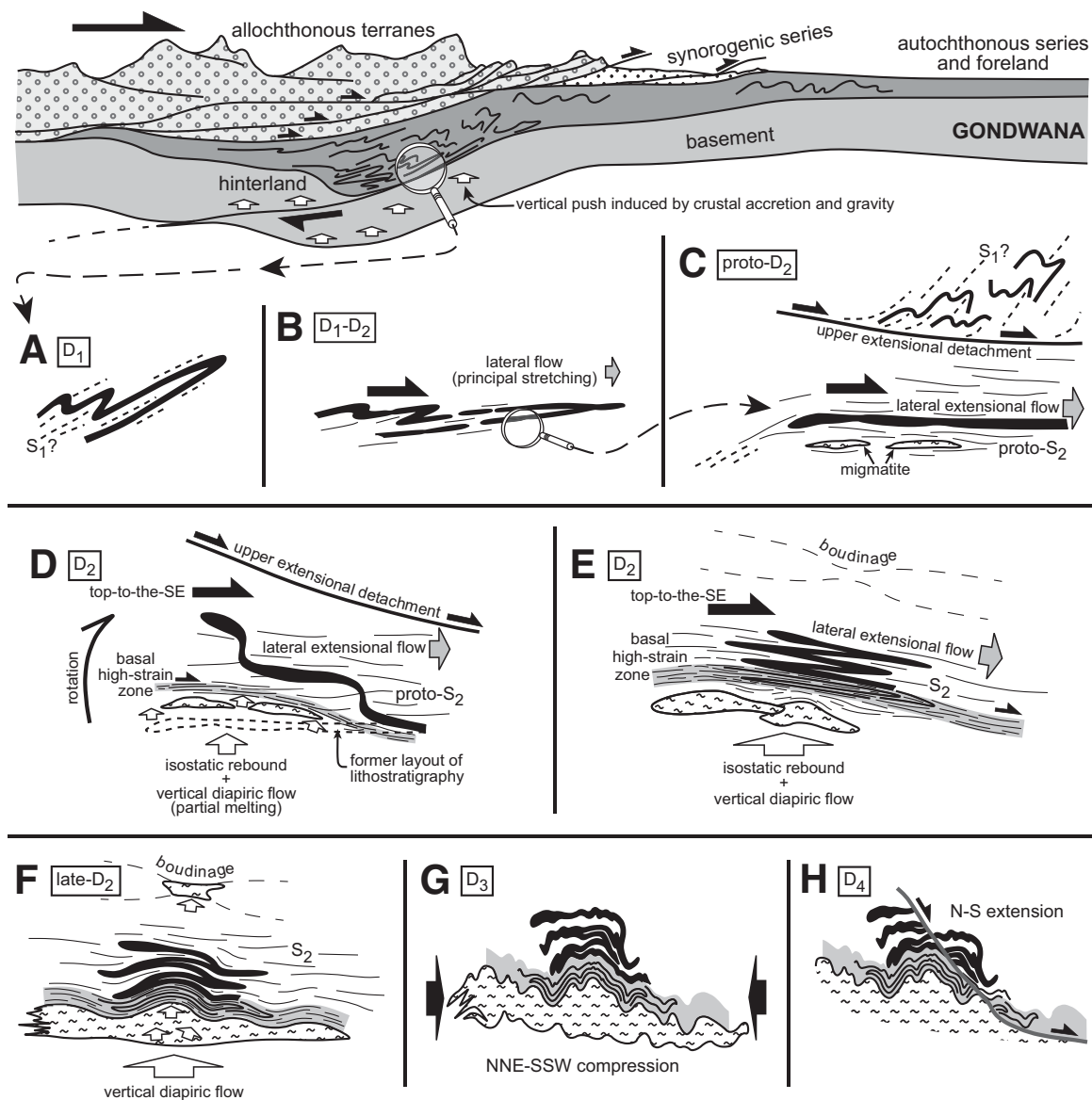


Figure 13. Structural model for Variscan deformation in the Santa María de la Alameda dome. Starting from a hypothetical collision architecture on top, the magnifying glasses zoom into the study case progressively. See explanations in the text.

where vertical and lateral flows were coupled, including orogenic belts other than the Variscan.

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