

Tectonic evolution of a continental subduction-exhumation channel: Variscan structure of the basal allochthonous units in NW Spain

Rubén Díez Fernández,¹ José Ramón Martínez Catalán,¹ Ricardo Arenas Martín,² and Jacobo Abati Gómez²

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[1] A regional study starting from detailed geological mapping has been carried out in the Malpica-Tui Complex of Galicia in NW Spain. The complex is formed by two units representing pieces of the external edge of Gondwana, subducted and exhumed during the Variscan collision. The study shows that synsubduction and early synexhumation structures in continental subduction channels tends to be obscured and even erased once exhumation is complete. Detailed structural analysis, matched with the knowledge of the history, and available data for other Galician basal units have elucidated the major structures developed during the subduction-exhumation process. The results include evidence of the plate convergence causing early Variscan continental subduction of the Gondwana margin. Subduction was followed by exhumation driven by ductile thrusting within the subduction channel, which, in turn, provoked crustal duplication in the subducted slab and modified the initial tectonometamorphic architecture of the subduction wedge. The next step was accretion to the adjacent continental domains, placing the subduction wedge on top of unsubducted parts of the Gondwana margin via ductile thrusting. Thrusting was preceded by progressive propagation of a train of recumbent folds toward the foreland that affected the previous structural stack. Subsequent transference of oceanic (Rheic) and peri-Gondwanan terranes to the Gondwana margin took place by out-of-sequence thrusting followed by crustal extensional collapse and strike-slip tectonics.

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1. Introduction

[2] Subduction is one of the most important tectonic processes preceding continental collision in mountain building, affects both oceanic and continental crust, and is the first mechanism of plate convergence at the onset of major orogenies. Mountain belts provide most of the preserved records of Earth's subduction systems developed in the geological history. Relicts of ancient to relative young subduction zones (Archean to Alpine) in mountain belts are a rich database for exploring subduction mechanisms.

[3] Exposed subduction systems share the fact that they no longer occur at the place where they developed [Searle *et al.*, 1994; Rawling and Lister, 2002; Avigad *et al.*, 2003; Gray *et al.*, 2004; Le Bayon and Ballèvre, 2006; Ring *et al.*,

2007]. Their position does not match a model of simple accretion without subsequent tectonic transport, as evidenced by widespread penetrative foliation and retrogression of high-pressure (HP) assemblages occurred during exhumation [e.g., Platt and Lister, 1985; Le Bayon and Ballèvre, 2006; Hacker *et al.*, 2010]. These fabrics account for a deep distortion of the subduction record via ductile structures, as well as for the displacement of fossil subduction wedges to shallower levels in the lithosphere. These evidences set the basis to analyze exhumed subduction systems, which must be studied tackling the exhumation history, and considering its structural and metamorphic implications. Although this workflow has been followed in many HP belts, most of the efforts have been focused on the structural relationships between the HP domains and their relative autochthon, treating the first as relatively simple tectonic sheets. However, many HP belts, and especially those affecting continental, transitional or arc-derived crusts, depict a varied lithostratigraphy of metaigneous and metasedimentary rocks. We show here how this layered disposition may be used to unravel the mesostructures to macrostructures developed within HP belts during exhumation, and even to get some hints on the

¹Departamento de Geología, Universidad de Salamanca, Salamanca, Spain.

²Departamento de Petrología y Geoquímica and Instituto de Geología Económica, Universidad Complutense/Consejo Superior de Investigaciones Científicas, Madrid, Spain.

subduction architecture. Furthermore, the metamorphic evolution is used too, since it provides constraints on position along the subduction wedge and the reliability of exhumation patterns obtained from structural analysis. Our study is centered on the basal units of the allochthonous complexes in NW Iberia, which represent a subduction-exhumation channel developed at the early stages of Pangea's assembly.

2. Geological Setting

[4] The Variscan belt rose from the Devonian-Carboniferous collision between Gondwana and Laurussia [e.g., *Matte, 1991; Martínez Catalán et al., 2007*]. The structural pile of the NW Iberian Massif preserves the main Paleozoic paleogeographic domains of the northern Gondwanan margin involved in the Variscan collision, with the peripheral and outermost realms placed on top (Figure 1) [*Martínez Catalán et al., 2009*].

[5] The allochthonous complexes consist in a nappe stack cropping out as klippen in synformal structures [*Martínez Catalán et al., 2007*]. The Paleozoic geodynamic evolution of the northern peri-Gondwanan realm is preserved in NW Iberia as a collage of exotic terranes with ophiolites and HP rocks [*Arenas et al., 1986*] that delineate the Rheic suture in southern Europe [*Martínez Catalán et al., 1997*].

[6] Once formed, the suture was transferred to the Gondwana mainland during the Variscan collision [*Martínez Catalán et al., 2002*], being exposed in the allochthonous complexes of Cabo Ortegal, Órdenes, Morais, Bragança and Malpica-Tui. Three types of tectonometamorphic units can be recognized in these complexes (Figure 2), which are, from top to bottom: the upper, ophiolitic, and basal units.

[7] The upper units are pieces of a Cambro-Ordovician ensialic island arc detached from Gondwana [*Abati et al., 1999; Andonaegui et al., 2002; Santos et al., 2002*], while the middle units are ophiolitic, and represent remnants of the Rheic Ocean [*Díaz García et al., 1999; Pin et al., 2006; Arenas et al., 2007; Sánchez Martínez et al., 2009*], whose closure marked the beginning of Pangea's assembly [*Gómez Barreiro et al., 2010a*]. The basal units represent the most external part of the Gondwana margin [*Martínez Catalán et al., 1996; Díez Fernández et al., 2010*]. These units are the focus of this article and have been grouped in two tectonic sheets according to their lithostratigraphic affinities (Figures 2a and 2b) [*Díez Fernández et al., 2010*]. The basal units were subducted beneath Laurussia at the onset of the Variscan collision [*Arenas et al., 1995, 1997; Santos Zalduegui et al., 1995; Rodríguez et al., 2003*], and subsequently exhumed through crustal-scale thrusting accompanied by recumbent folding and extensional tectonics [*Martínez Catalán et al., 1996, 1997; Díez Fernández and Martínez Catalán, 2009; Díez Fernández, 2011*].

[8] The allochthonous pile was emplaced on top of the Gondwanan sedimentary cover, which was detached to form an imbricate thrust sheet known as parautochthon [*Ribeiro et al., 1990*] or Schistose Domain [*Marquinez García, 1984; Farias et al., 1987*]. It is composed of siliciclastic sedimentary and volcanic rocks, separating the allochthonous complexes from the Iberian autochthon, which occurs below the nappe stack. The autochthon consists of a thick metasedimentary sequence intruded by widespread Cambro-Ordovician plutons that, together with the Schistose Domain

and the basal units of the allochthonous complexes, was deposited along the northern margin of Gondwana. The autochthon and parautochthon will be treated as a single ensemble in this article, and will be referred to as the relative autochthon. Two broad lithological units have been distinguished: a lower layer of glandular orthogneisses, and an upper metasedimentary sequence of schists and migmatitic gneisses.

[9] Crustal thickening was followed by thermal relaxation causing temperature increase and partial melting. It facilitated viscous flow and the gravitational extension of the whole orogenic wedge [*Martínez Catalán et al., 2002*]. The final result was the formation of high-T low-P rocks cropping out in migmatitic domes, with voluminous Variscan granitoids [*Gómez Barreiro, 2007; Gómez Barreiro et al., 2010b; Díez Fernández, 2011*]. Extension was followed by strike-slip tectonics [e.g., *Iglesias Ponce de León and Choukroune, 1980; Llana-Fúnez and Marcos, 2001*], which affected the whole Variscan belt, overprinting many of the previous tectonic contacts and fabrics.

3. Tectonostratigraphy of the Basal Allochthonous Units

[10] The sediments of the lower sheet are Late Neoproterozoic in age and represent a pile of immature sandstones intruded by acid and basic Cambro-Ordovician calc-alkaline and rift-related igneous rocks [*Floor, 1966; Ribeiro and Floor, 1987; Pin et al., 1992; Rodríguez et al., 2007; Abati et al., 2010*]. The sediments of the upper sheet represent a Cambro-Ordovician pelitic succession that alternates with basic rocks.

[11] During the Variscan cycle, the sediments of the lower sheet were transformed into albite-bearing schists and paragneisses, which alternate with minor layers of graphitic schists, calc-silicate rocks and quartzites. The granitic rocks include a calc-alkaline suite (tonalites, granodiorites and high-K granites) and an alkaline association with metaluminous alkali-feldspar quartz syenites and granites, peraluminous alkali-feldspar granites and peralkaline granites [*Rodríguez Aller, 2005*]. They were metamorphosed to biotitic, felsic, alkaline, and peralkaline orthogneisses. The basic rocks were transformed into eclogites and then, depending on their retrogression, into different types of amphibolites and greenschists. As for the upper sheet, the sediments and basic rocks were transformed into a monotonous pile of albite-bearing mica-schists alternating with minor lenses of amphibolites and greenschists, graphite schists, metacherts, calc-silicate lenses, limestones, greywackes and quartzites.

[12] The two tectonic sheets of the basal units maintain the same relative structural position. The present contacts between them are extensional detachments [*Gómez Barreiro et al., 2010b; Díez Fernández, 2011*], their relative position thus showing the original structure of the subduction wedge.

4. Microstructural Analysis of the Basal Units

[13] Five deformation phases can be recognized in the basal units. The first phase (D_1) is related to the subduction event. It is recorded all over the basal units as mineral relicts (S_i) enclosed within porphyroblasts grown during the exhumation (D_2), or as a high-pressure (HP) fabric (S_1), preserved either in nonretrogressed lenses of HP rocks or in partly

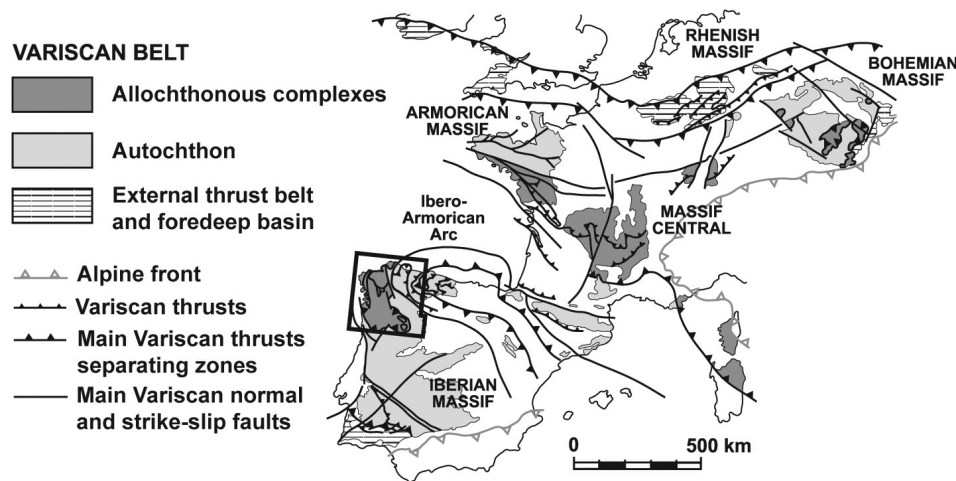


Figure 1. Location of the study area in the Variscan belt.

retrogressed schists, gneisses and amphibolites. D_1 fabrics are overprinted by a regional foliation (S_2) developed during the exhumation and emplacement of the basal units on top of their relative autochthon. D_2 deformation was heterogeneous and related to recumbent folding and thrusting. It was followed by the development of an out-of-sequence thrust system (D_3), and the gravity-driven extensional collapse of the orogen (D_4). During D_4 , the collisional wedge was affected by extensional detachments, and bent into regional, open antiforms and synforms, both related to the development of migmatitic domes. Later on, a complex dextral strike-slip system developed (D_5), producing subvertical, low-grade transcurrent shear zones and faults responsible for the flattening and reorientation of the previous folds and the development of new upright structures.

[14] The last three phases of deformation are not related to the evolution of the continental subduction-exhumation system and subsequent emplacement, and they will be briefly described here.

4.1. Continental Subduction (D_1)

[15] D_1 relicts can be observed all over the basal units, what implies that subduction transformed the two sequences identified into a tectonic pile whose sedimentary and igneous record was turned into a layered lithostratigraphy consisting of metasediments alternating with large, lens-shaped massifs of granitic orthogneisses, and with lenses of metabasites. However, given that undeformed and variably deformed igneous protoliths occur in the basal units, a heterogeneous character of deformation is assumed for this phase.

[16] The units representing the upper sheet (Ceán and Lamas de Abad; see Figure 2) are characterized by a HP mineral assemblage developed under blueschist facies conditions [Arenas *et al.*, 1995; Rodríguez *et al.*, 2003; Gómez Barreiro *et al.*, 2010b; López-Carmona *et al.*, 2010]. The HP parageneses are composed of glaucophane + chloritoid + garnet + phengite + chlorite + epidote + quartz + rutile in the metasediments (mica-schists) [López-Carmona *et al.*, 2010], and lawsonite + omphacite? + garnet ± quartz ± rutile ± glaucophane? ± aragonite? in the metabasites [Rodríguez *et al.*, 2003]. No relicts of such a HP fabric have been identified in its uppermost levels, where the temperature

and perhaps the pressure were the lowest registered in the basal units.

[17] In the lower sheet, D_1 is characterized by eclogite facies conditions to the west (Malpica-Tui, Santiago and Agualada units), with a progressive transition to blueschist facies conditions to the east in the Lalín and Forcarei units [Gil Ibarguchi and Ortega Gironés, 1985; Gil Ibarguchi, 1995; Martínez Catalán *et al.*, 1996; Arenas *et al.*, 1997; Rubio Pascual *et al.*, 2002; Rodríguez *et al.*, 2003; Díez Fernández, 2011]. The HP assemblage in the metasediments includes garnet + phengite + rutile + quartz ± epidote ± chlorite, while in the metabasites (eclogites *sensu stricto*) it consists of garnet + omphacite + rutile + phengite + quartz ± zoisite ± kyanite. The evidences for HP metamorphism in the granitic, granodioritic, alkaline and peralkaline orthogneisses are less abundant, the most significant one being the development of coronitic garnet around igneous biotite in the metagranodiorites [Gil Ibarguchi, 1995]. However, the tonalitic granitoids were completely transformed into eclogitic gneisses consisting of garnet + omphacite + phengite + quartz + rutile + kyanite + zoisite [Rodríguez *et al.*, 2003], which defines the main foliation in these rocks.

[18] Rare glaucophane in the eclogites and scarce chlorite in the tonalitic gneisses are preserved within the eclogitic garnets, thus supporting a prograde character for the eclogitic assemblages. Moreover, the eclogitic assemblage in the basic rocks and the tonalitic gneisses is sometimes preserved within glaucophane porphyroblasts and Mg-katophorite blasts [Rodríguez Aller, 2005], respectively. This happens in the upper part of the Malpica-Tui Unit (MTU), constituting the first retrogression found in the HP assemblages, and suggesting that early exhumation was accompanied by cooling.

[19] According to the HP metamorphism, the MTU has been classically divided into two domains [Collée, 1964; van der Wegen, 1978]. The northern domain shows widespread evidence of eclogite facies metamorphism, whereas the southern domain almost lacks this feature. In the latter, the eclogite facies metamorphism is only traceable through scarce mineral relicts (garnet, clinopyroxene, rutile-illmenite), and decompressive reactions and textures in basic rocks (amphibole-plagioclase symplectites), as well as by the existence of equivalent S_1 fabrics in metasediments. Scarcity of

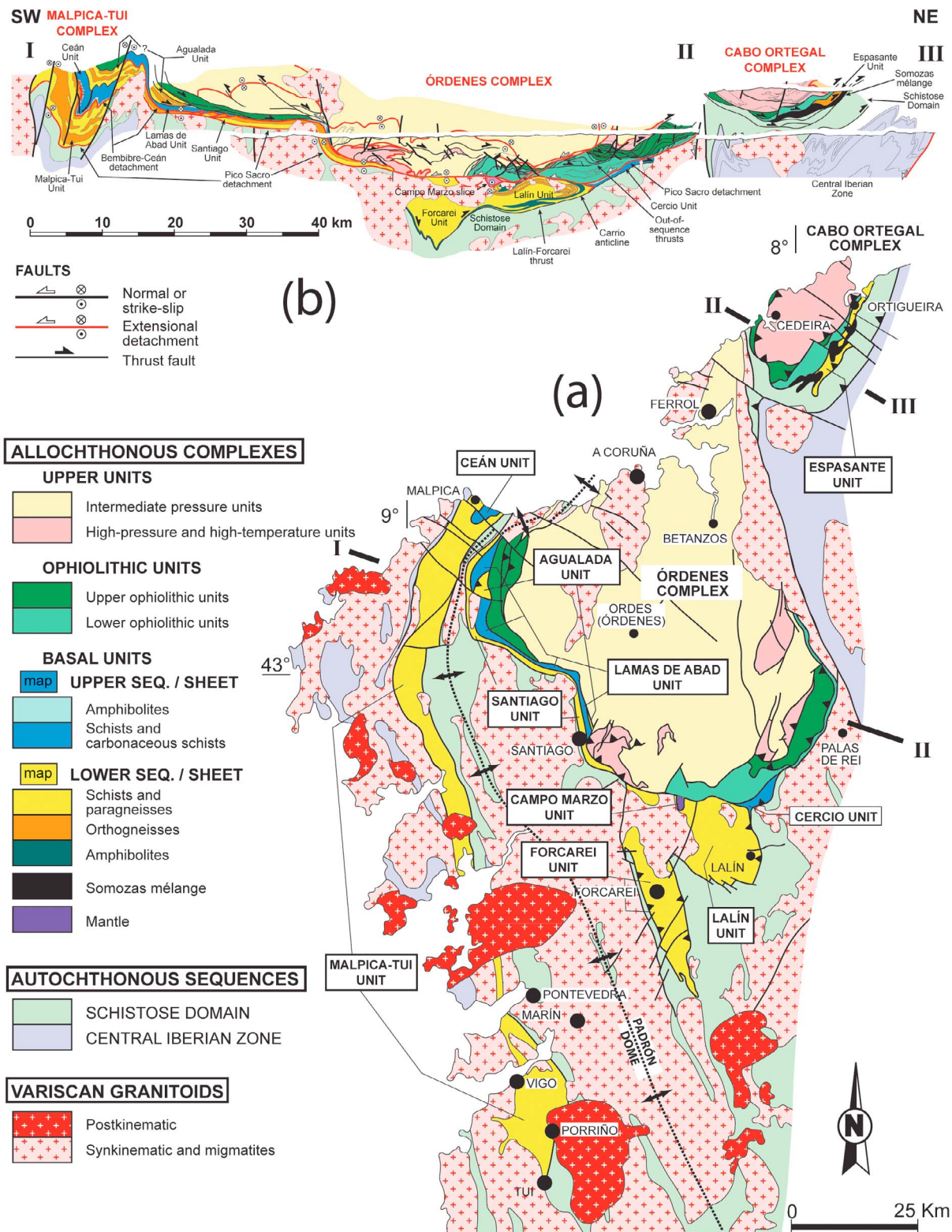


Figure 2. (a) Map showing the allochthonous complexes of Órdenes, Cabo Ortegal, and Malpica-Tui in NW Spain. (b) Composite cross section showing the general structure of the allochthonous complexes. For the basal units, the cross section includes further details than the map (see legend). Geographical distribution of the basal units, as well as their names and those of the main structures are included. The basal units have been grouped in two tectonic sheets. The upper sheet includes the Ceán, Lamas de Abad, and Cercio units, while the lower sheet includes the Malpica-Tui, Santiago, Agualada, Lalín, Forcarei, and Espasante units.

HP evidences in the south cannot be explained alone by post- D_1 retrogression, since they are rare even where retrogression is weak. On the contrary, widespread evidence of HP metamorphism can be found in the north, independently of the intensity of posteclogitic overprint.

[20] No evidence of HP metamorphism has been found in the relative autochthon, thus suggesting that both the parautochthon and the autochthon escaped the subduction zone.

4.2. The Initial Stages of the Exhumation (D_2)

[21] The initial exhumation was driven by heterogeneous ductile shearing. This phase is responsible for the main regional foliation (S_2) and stretching lineation in the basal units [Gil Ibarra and Ortega Gironés, 1985; Martínez Catalán et al., 1996; Llana-Fúnez and Marcos, 2002; Rodríguez et al., 2003; Díez Fernández and Martínez Catalán, 2009; Díez Fernández, 2011].

[22] The main foliation in the metasediments is a schistosity defined by statistically oriented quartz, mica, and albite porphyroblasts with internal foliation (S_1). S_2 defines a normal regional metamorphic zonation that ranges from amphibolite to greenschist facies conditions. It includes quartz + biotite + garnet + white mica + albite \pm staurolite \pm illmenite at the base of the MTU, and quartz + biotite + white mica + albite \pm garnet \pm illmenite in an intermediate position of the MTU, as well as the base of the Santiago Unit and the whole Lalín Unit. To the top of the Santiago Unit and in the Forcarei Unit, S_2 consists of quartz + white mica + albite \pm chlorite \pm biotite \pm illmenite. In the Ceán Unit, the S_2 assemblage consists of quartz + biotite + garnet + white mica + albite + chloritoid \pm epidote \pm tourmaline \pm opaque minerals at the base, progressively losing garnet, chloritoid, biotite and albite, and containing more chlorite and opaque minerals toward the top. The D_2 assemblages in the Lamas de Abad and Cercio units (Figure 2a) are similar to those in the Ceán Unit.

[23] The pre-Variscan granitic rocks show an S_2 gneissic foliation and a mineral lineation formed by an alternation and alignment of quartz-feldspathic bands and different ferromagnesian minerals depending on the composition and metamorphic evolution. The grains are oriented and aligned, individually or as aggregates, parallel to the main foliation, and defining a stretching lineation. S_2 in the orthogneisses includes planar, linear and planilinear varieties. For an extended petrographic description of the orthogneisses, see Floor [1966], Arps [1970], Rodríguez Aller [2005] and Díez Fernández [2011].

[24] The orientation of the D_2 stretching lineation ranges from orogen-parallel to orogen-perpendicular trends, showing a fan-like disposition respect to the D_5 strike-slip zones [Díez Fernández, 2011]. The D_5 shear zones are mostly orogen-parallel features in this section of the Variscan belt [Iglesias Ponce de León and Choukroune, 1980; Llana-Fúnez and Marcos, 2001]. Based on this fact, Díez Fernández [2011] suggested that the D_2 stretching lineations were reoriented by the late strike-slip systems to an orogen-parallel fashion, being the major D_2 stretching direction roughly normal to the orogen. Top-to-the-east and top-to-the-SE kinematics was determined by using offset mesoscopic and microscopic shape fabrics in D_2 structures (composite foliation S-C, C' fabrics, asymmetric boudins, σ clasts, and δ clasts). This kinematics is supported by tex-

tural analyses of the regional foliation (quartz-CPO fabrics [Llana-Fúnez, 2002; Gómez Barreiro et al., 2010b]), and has been used, together with the orogen-perpendicular trend of the main D_2 stretching direction, to set a roughly normal to the orogen D_2 tectonic flow (top-to-the-east in present coordinates), as previously stated by Martínez Catalán et al. [1996].

[25] Two main groups of metasediments and orthogneisses can be distinguished: those that record very strong (mylonitic) deformation and those with variably but weaker (nonmylonitic) deformation. It must be pointed out, however, that such a distinction is made in a region with a generally high degree of deformation, and where transitions between differently deformed facies of the same protolith are common.

[26] The mylonitic varieties developed only in the structurally upper parts of the MTU. These upper parts crop out only in the northern domain of the MTU, showing abundant evidences of high-pressure metamorphism. However, S_2 includes a long, complex record of ductile shearing. This fabric is the result of superimposing a regional axial planar foliation related to large recumbent folding (S_2 sensu stricto) to a tectonic, mylonitic layering developed during the early stages of D_2 (proto- S_2).

[27] In the nonmylonitic granitic and granodioritic gneisses, S_2 is composed of quartz-feldspathic bands and lenses formed by oriented K-feldspar (augen), quartz and plagioclase, alternating with aggregates of biotite, titanite, allanite, zircon, apatite, opaque minerals, amphibole, garnet and white mica. In their mylonitic varieties, grain-size reduction is intense, with the garnet-phengite-epidote association constituting most of the ferromagnesian fraction. The garnet-phengite-epidote trio may be trapped within plagioclase porphyroblasts but it is frequently retrogressed to a greenschist facies assemblage made up of chlorite, quartz, plagioclase and rare biotite when it occurs in the matrix. Such retrogression together with the alignment and recrystallization of the quartz-feldspathic fraction form the regional S_2 foliation, which is a crenulation cleavage of a previous planilinear fabric that consisted of garnet-phengite-epidote lenses (proto- S_2). The greenschist facies assemblage does not participate in the mylonitic layering, which is mostly dominated by the garnet-phengite-epidote trio, thus indicating that mylonitization developed at early D_2 stages.

[28] Rodríguez et al. [2003] and Rodríguez Aller [2005] identified a series of metamorphic reactions that match the transformation at HP conditions of igneous biotite in the granitic and granodioritic orthogneisses into garnet, phengite, epidote and rutile. The transformation of igneous plagioclase into phengite, epidote and metamorphic plagioclase is another HP reaction that would likely explain the abundance of phengite in the mylonitic lithologies [Gil Ibarra, 1995]. Moreover, the Si content in phengite together with the absence of omphacitic clinopyroxene in these orthogneisses point to a pressure of 12–15 kbar [Rodríguez et al., 2003]. Given the peak pressure calculated for the eclogites surrounded by these mylonitic varieties (26 kbar [Rodríguez et al., 2003]), the mylonitization clearly accounts for the initial exhumative evolution of the basal units.

[29] A hornblende + plagioclase assemblage is preserved all over the basic rocks within the MTU, Santiago, Agualada and Lalín units and in the base of the Ceán Unit, suggesting that they were exhumed under amphibolite facies

conditions during D_2 , once these units reached crustal levels ($P < 12$ kbar).

[30] Greenschist facies conditions (biotite-chlorite zone) were reached during D_2 retrogression at the top of the MTU, the basal levels of the Santiago Unit [Gómez Barreiro *et al.*, 2010b], and the base of the Ceán Unit. In the rest of the Ceán Unit, D_2 exhumation was accompanied by weak heating followed by cooling [López-Carmona *et al.*, 2010]. Conversely, the intermediate and basal levels of the MTU experienced only D_2 isothermal decompression (amphibolite facies) followed by cooling [Martínez Catalán *et al.*, 1996]. However, the late D_2 metamorphic evolution in the Santiago, Lalin and Forcarei units, as well as in the top of the Lamas de Abad Unit was controlled by synexhumation (D_2 - D_3) thermal overprinting exerted by the emplacement of a high-temperature tectonic sheet over the basal units of the Órdenes Complex [Arenas *et al.*, 1995; Martínez Catalán *et al.*, 1996; Gómez Barreiro, 2007].

4.3. Out-of-Sequence Thrust System (D_3)

[31] The Agualada Unit [Díaz García, 1990] is a key element to understand the subsequent exhumation process (Figure 2). This unit shows high grade and contains tectonic intercalations of ultramafic rocks [Arenas *et al.*, 1997; Abati and Dunning, 2002]. Its lithostratigraphy shows affinities with the MTU [Gómez Barreiro *et al.*, 2010b]. The Agualada Unit is assumed to have occupied the uppermost structural position in the Malpica-Tui Complex before being removed by erosion (Figure 2b). It is considered the most internal relict of a high-temperature nappe emplaced onto the upper sheet of the basal units [Arenas *et al.*, 1995; Martínez Catalán *et al.*, 1996; Gómez Barreiro, 2007]. Its discontinuous appearance may be due to dismembering during its emplacement, or to subtraction by extensional structures [Gómez Barreiro *et al.*, 2010b]. Several of its blocks are involved in a tectonic mélange located at the front of the allochthonous complexes [Arenas *et al.*, 2009].

[32] We interpret the Agualada Unit as a sliver of the lower sheet emplaced onto the upper sheet (Figure 2b). Its regional distribution fits the trace of the east directed out-of-sequence thrust system responsible for the final obduction of the ophiolitic units and emplacement of the upper units of the allochthonous complexes onto the basal units, crosscutting previous thrust faults [Martínez Catalán *et al.*, 2002]. For this reason, we assume the emplacement of this high-grade nappe to be linked with this out-of-sequence thrust system.

4.4. Extensional Collapse (D_4)

[33] During this phase, retrogression of the D_2 fabrics, ductile extensional deformation and open upright folding affected the basal units. In the Malpica-Tui Complex and the western half of the Órdenes Complex, the metasediments at the basal levels of the lower sheet developed low- P high- T schistosity (S_4), whereas a widespread low- P low- to medium- T metamorphism retrogressed the D_2 fabrics all over the upper sheet [Gómez Barreiro *et al.*, 2010b; Díez Fernández, 2011].

[34] Ductile and brittle structures developed during this event, attributed to a gravitational collapse. Ductile structures, represented here by S_4 , developed in relation to a set of extensional detachments that mark the present contacts

between the two tectonic sheets of the basal units, as well as between them and their relative autochthon around the Padrón dome (Figure 2b). They are associated to partial melting and the development of migmatitic domes, with coeval D_4 open upright folding related to regional doming [Martínez Catalán *et al.*, 2002; Gómez Barreiro, 2007; Gómez Barreiro *et al.*, 2010b; Díez Fernández, 2011].

4.5. Strike-Slip Tectonics (D_5)

[35] All of the previous structures were affected by a set of orogen-parallel low-grade transcurrent shear bands and subvertical faulting that caused the reorientation of previous planar and linear features, flattening of the previous folds and creation of new upright folds and development of axial planar fabrics (S_5).

5. Macrostructural Analysis of the Basal Units

[36] Most of the present internal structure of the basal units is related to a regional-scale train of D_2 recumbent folds identified by detailed mapping and corroborated by normal and reverse way-up criteria in the less deformed metasediments, S_0 - S_2 and S_1 - S_2 intersections, as well as in situ observation of minor recumbent folds affecting the layering in metasediments and metagranitoids, S_1 relicts, and proto- S_2 tectonic banding.

[37] The first hint of the possible existence of mayor recumbent folds in the MTU comes from the map pattern of the peralkaline orthogneisses in its southern part (Figure 3). The roughly tabular body of peralkaline orthogneisses shows considerably widening and depicts an irregular shape reminiscent of either large-scale *boudinage* or folding. The fact that the main foliation (S_2) changes from parallel to oblique and even perpendicular to the external contact of the orthogneiss suggests D_2 folding, with the zones of higher obliquity corresponding to hinge zones. Upright folds also occur all over the Malpica-Tui Complex, given the way to a fold interference pattern.

[38] The hinge zones of the upright folds are defined by those orthogneiss inflexions parallel to the strike of S_2 and coincident with a change in S_2 dipping direction. On the contrary, those inflexions where S_2 is strongly oblique or normal to the orthogneiss contact can be interpreted either as the hinge zone of a D_2 recumbent fold or as the termination of a deformed granitic massif. These criteria demonstrated the existence of kilometer-scale recumbent folds in the basal units [Díez Fernández and Martínez Catalán, 2009], and have been used to unravel the whole D_2 macrostructure in the MTU.

[39] There is an important variation in the regional plunge of the upright folds in the MTU. They plunge to the south, defining a structural basin in the southern half of the MTU [Díez Fernández and Martínez Catalán, 2009], and statistically to the north, although defining minor domes and structural basins in its northern half. This implies that the uppermost levels of the D_2 structure preserved in the MTU occur to the north, while its southern and central sections share an equivalent lower D_2 structural position.

5.1. The Southern Section of the Malpica-Tui Complex

[40] Figure 3 shows an extended version of the macrostructural analysis carried out by Díez Fernández and

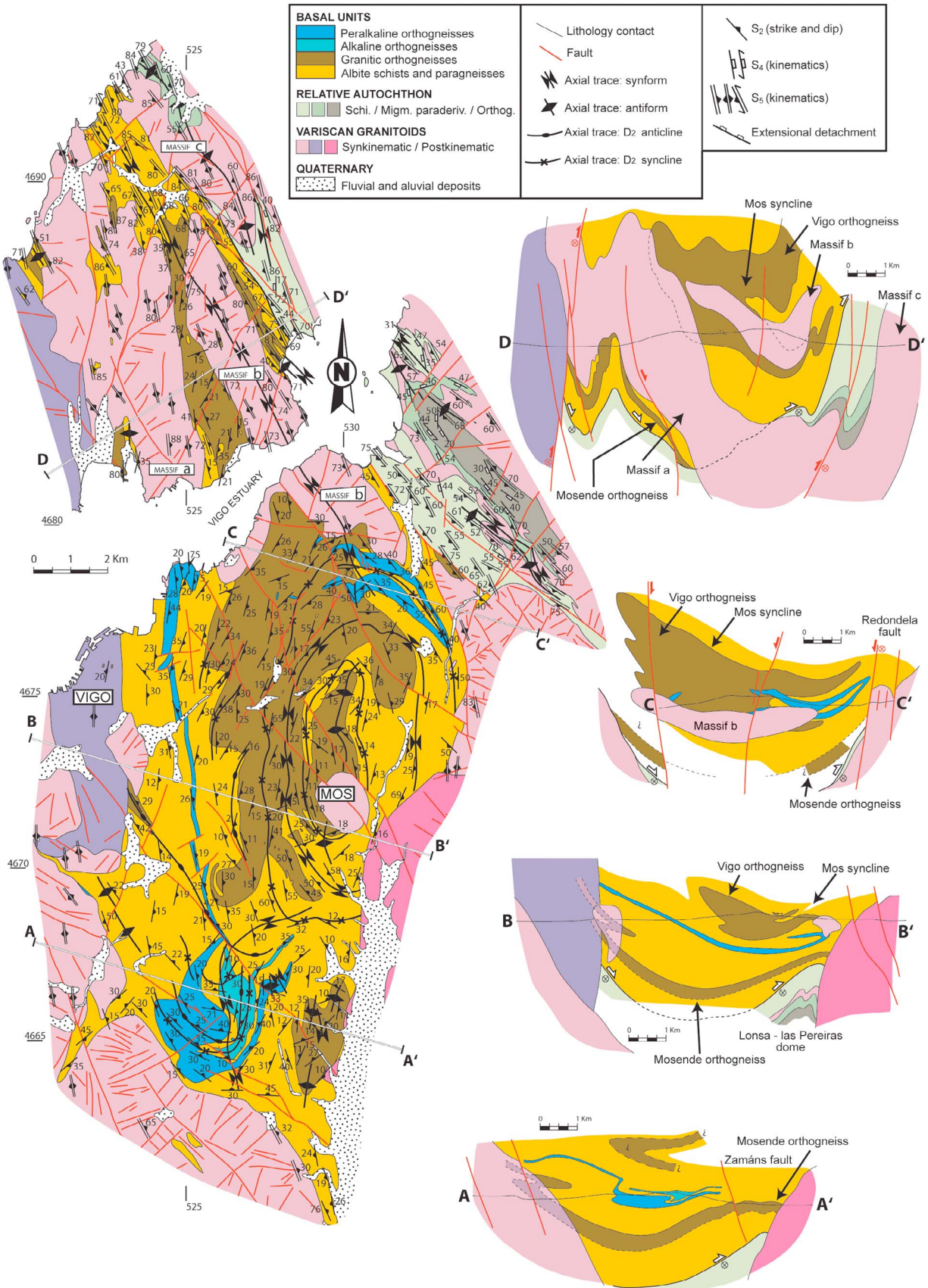


Figure 3. Structural map and cross sections of the southern part of the Malpica-Tui Complex. UTM coordinates. Extended version of previous results published by Díez Fernández and Martínez Catalán [2009].

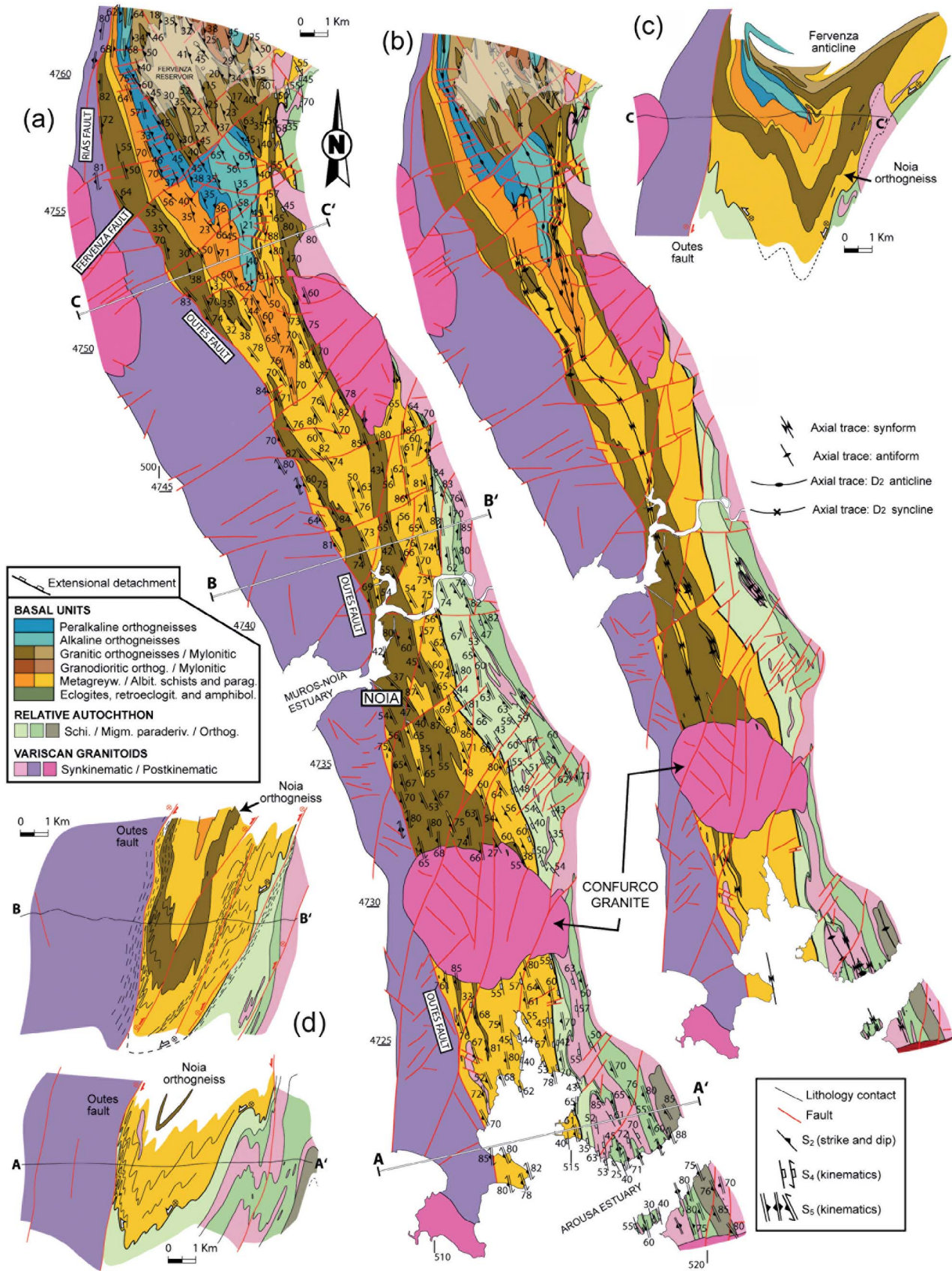


Figure 4. (a) Geological map of the central part of the Malpica-Tui Complex, including the attitude of the main foliations. UTM coordinates. (b) Axial traces of the two main fold systems. (c and d) Cross sections.

Martínez Catalán [2009] in the southern part of the MTU. It includes the recumbent folds affecting the alkaline and peralkaline orthogneisses at the normal limb of the Mos syncline, which, in turn, folds the Vigo orthogneiss. A structurally lower massif, the Mosende orthogneiss, crops out in several places around the Vigo orthogneiss, often intruded by Variscan synkinematic granitoids. Both the recumbent folds and the Variscan granites were affected by upright folding, as shown in the cross sections.

[41] D_5 strike-slip shear zones developed at the borders of the Malpica-Tui Complex. They control the structural patterns toward the north, where the associated flattening is more intense. Moreover, widespread development of D_5 crenulations occurs all over this sector, accompanied by steeping and blurring of S_2 and S_4 foliations.

5.2. The Central Section of the Malpica-Tui Complex

[42] Figure 4 shows geological/structural maps and cross sections of this part of the complex. The width of the MTU here suggests intense D_5 strike-slip flattening, responsible for S_2 steeping and widespread development of D_5 crenulations.

[43] The north plunging regional upright synform in which the MTU is preserved here is marked by the two branches of granitic orthogneisses to the north of Noia. The two branches merge in the hinge zone, which broadens to the south due to plunge oscillation of the fold axis. Two thin bands of the Noia orthogneiss mark both limbs of the synform to the south of the Confurco granite (Figures 4c and 4d).

[44] The western branch of the Noia orthogneiss continues to the north folded by a D_5 antiform-synform pair and bounded by the Outes fault. The eastern branch runs along the eastern limb of the synform until the Fervenza normal fault. The granitic, alkaline and peralkaline orthogneisses, as well as the metagreywackes depict a hook-type interference pattern to the south of this fault (cross section CC' in Figure 4c). A D_2 recumbent fold, namely the Fervenza anticline, interferes with the major synform. The alkaline and peralkaline orthogneisses do not depict the hook-type interference pattern because their D_2 hinge zones occur just to the west of the axial surface of the synform.

[45] The contact between the granitic and the alkaline-peralkaline orthogneisses is strongly oblique toward the SE of the Fervenza reservoir. Such obliquity suggests a strong initial obliquity, thus supporting the interpretation of the alkaline suite as dikes [*Díez Fernández and Martínez Catalán*, 2009].

5.3. The Northern Section of the Malpica-Tui Complex

[46] The alkaline orthogneisses provide in this sector an excellent marker that delineates a train of recumbent folds developed in the normal limb of the Fervenza anticline (Figure 5a). Interference with the major synform causes the outcropping of the same D_2 hinge zones in both limbs of the synform and allows setting D_2 fold axis orientation (Figure 6). The results fit in situ measurements of D_2 fold axes. The fold interference pattern is further complicated due to the limited continuity of the alkaline orthogneisses, and the D_4 - D_5 structural basin bounded by the Riás and Fervenza faults (Figures 5b and 5c).

[47] The cross sections in Figure 7 show the fold interference pattern in this northern sector of the MTU. Pre- D_2 *boudinage* or original igneous irregularities can account for

the noncylindrical character of the folds, although thickening and stretching related to folding can also be envisaged. Furthermore, *boudinage* of the Fervenza anticline can be seen SE of Zas, where the fold is marked by an isolated band of alkaline orthogneisses that ends to the south and in depth close to mylonitic orthogneisses.

[48] Bands of mylonitic granitic orthogneisses, metagreywackes and more deformed metasediments occur south of Zas, representing the continuation of the recumbent folds identified around the Fervenza reservoir, but in the eastern limb of the main synform. NE of the reservoir, several bodies of metabasites and granodioritic orthogneisses help to trace the recumbent folds, while tonalitic orthogneisses occupy the normal limb of a recumbent fold at the core of the main synform (Figures 5a and 7).

[49] Gneissic bodies and metasediments show the same but imperfect D_2 - D_{4-5} fold interference pattern north of the Riás fault. The contacts of the granitic and granodioritic orthogneisses here depict more D_2 recumbent folds (Figure 8), which prolong to the north the normal limb of the Fervenza anticline (Figure 9). A recumbent anticline interfering with the main synform yields a hook-type pattern to the south, whereas a culmination exposes its hinge zone again toward the north (Figure 8).

[50] No cartographic linkage or structural correlation has been found between the two northern granodioritic bodies, which have been considered different plutons. Conversely, the granodioritic massifs occurring in both sides of the Riás fault have been correlated using the alkaline orthogneisses and the fault kinematics, which is oblique with dextral slip and down throwing of the northern block.

[51] The orthogneissic massif extending from the Fervenza reservoir to the northern coastal sections of the MTU will be named the Fervenza massif, whereas another large granitic and granodioritic massif occurring to the west will be referred to as the Borneiro massif. A pair of recumbent folds mapped in the latter indicates that they occur in the long (normal) limb of a major D_2 recumbent fold. The same can be said for the Fervenza massif (Figures 8 and 9). The normal limb is in both cases that of the Fervenza anticline. The granodioritic orthogneisses in the northern part of the Borneiro massif are separated from their granitic counterpart by a narrow mylonitic shear band.

[52] The Riás fault cuts a previous ductile shear band related to the D_5 strike-slip system developed at the west margin of the Malpica-Tui Complex [*Llana-Fúnez and Marcos*, 2001]. This fault is considered as a regional-scale C' structure that accounts for the dextral component of the strike-slip system, inducing the deflection, stretching and flattening of all of the gneissic bands, together with the development of an S_5 foliation [*Díez Fernández*, 2011]. The Riás fault hides the relationships between both limbs of the Padrón antiform and consequently, the correlation of the orthogneisses in the MTU with those in the Órdenes Complex.

5.4. Integration of the Regional Recumbent Fold Structure

[53] In the Órdenes Complex, the Santiago orthogneiss occupies the long normal limb of the Carrio recumbent anticline [*Martínez Catalán et al.*, 1996, 2002]. In the northern Malpica-Tui Complex, the orthogneisses occurring in the normal limb of the Fervenza anticline must continue

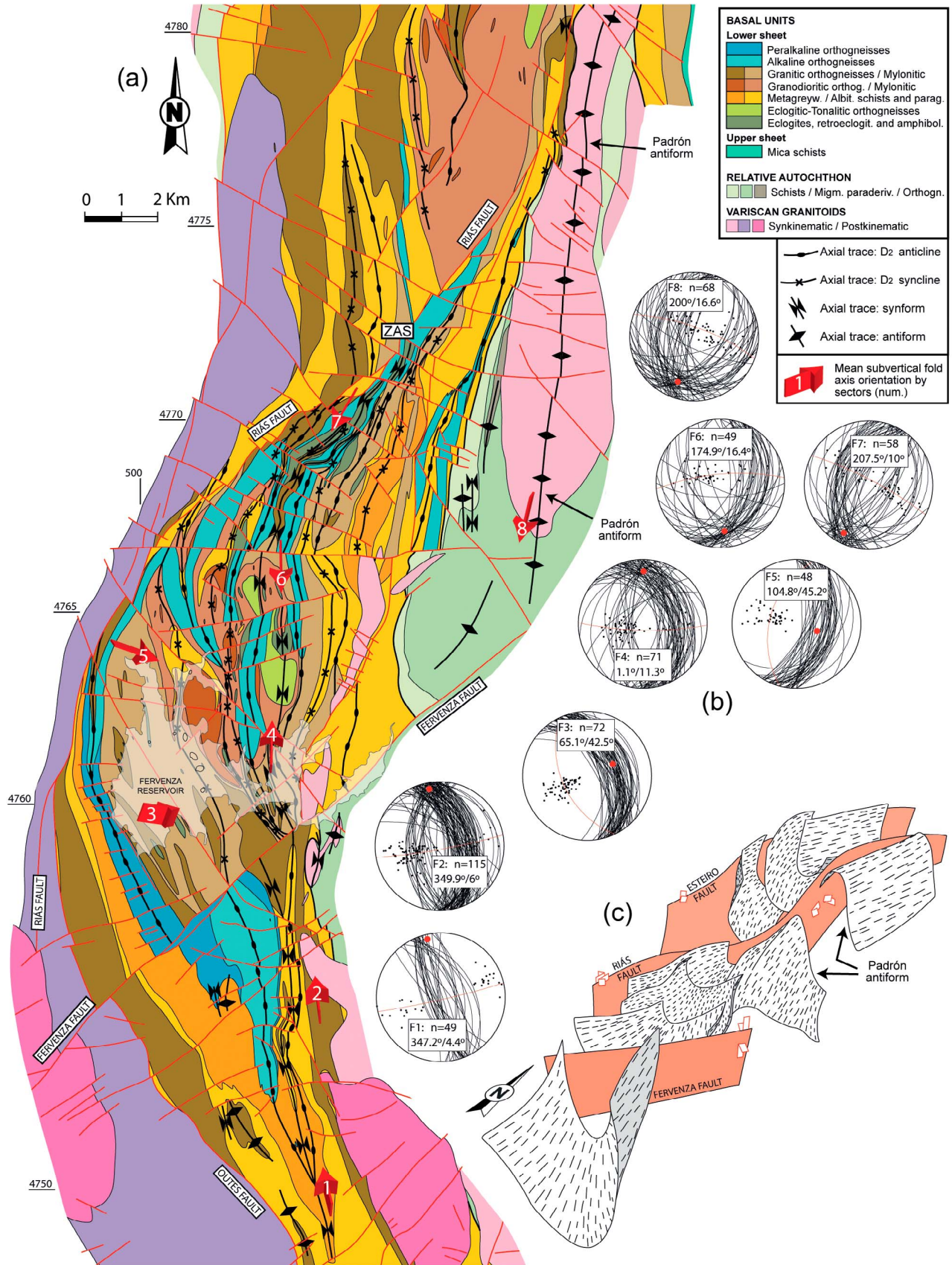


Figure 5

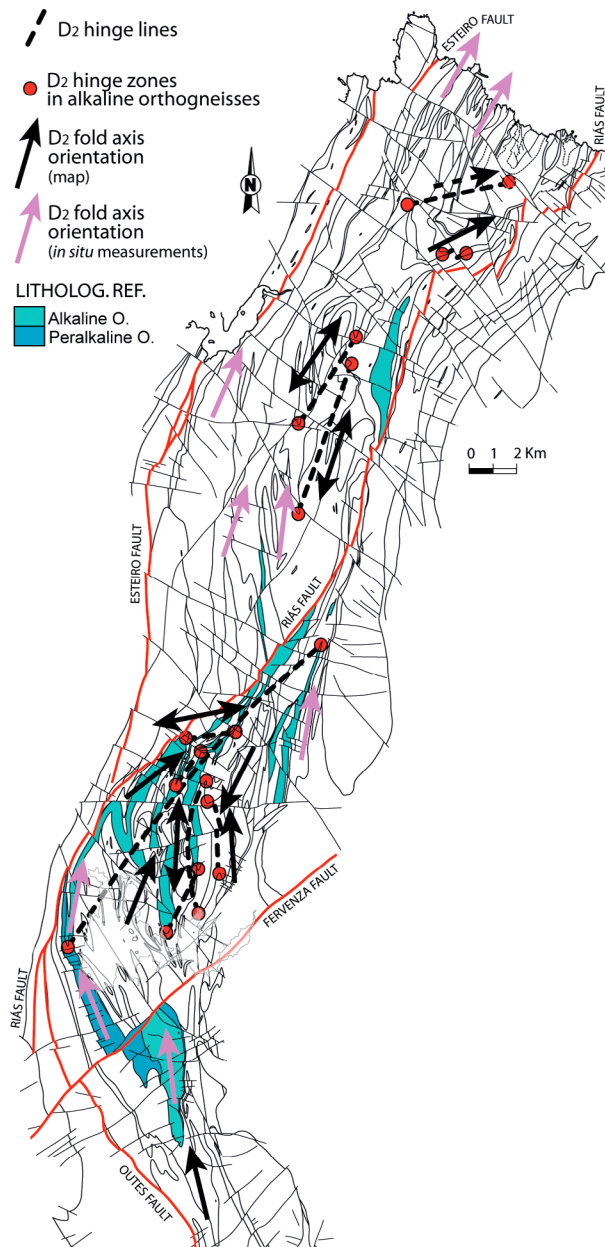


Figure 6. Correlation of D_2 hinge zones (circles) through the limbs of the main, late synform showing the attitude of regional recumbent fold axes (thick dashed lines and arrows), as well as the agreement with in situ measurements.

in its reverse limb and an underlying recumbent syncline, before continuing eastward. This syncline corresponds to the Mos syncline, which crops out in the southern part of the MTU (Figures 3 and 10). The Noia orthogneiss occupies

the reverse limb shared by the Fervenza and Mos recumbent folds, whereas the Vigo orthogneiss depicts the hinge zone of the syncline. Both gneiss bodies are separated, either because they correspond to different plutons or due to N-S large-scale *boudinage* of a single massif. Another band of orthogneisses in the normal limb of the Mos syncline, namely the Mosende orthogneiss (Figure 3), would then occupy a position equivalent to the Borneiro massif in the northern Malpica-Tui Complex.

[54] The integrated structure shows that the whole train of recumbent folds in the MTU belongs to the normal limb of the Carrio anticline, which also folds the two tectonic sheets of the basal units in the Órdenes Complex (Figure 10). However, whereas S_2 is the axial planar foliation to the recumbent folds in the MTU, the same foliation is folded by the Carrio anticline and points to an eastward propagation of the recumbent folding. Although S_2 is folded by the Carrio anticline, a new foliation related to the Lalín-Forcarei thrust developed across the Carrio anticline, whose amplification nucleated the Lalín-Forcarei thrust at its reverse limb [Martínez Catalán *et al.*, 1996]. This oblique fabric did not develop in the MTU, thus supporting the structural position proposed for the folds of the MTU.

[55] The integration of the regional D_2 recumbent folds shows two main orthogneissic bands, corresponding to flattened and stretched batholiths, the Fervenza and Borneiro massifs (the later would include the Mosende and Santiago orthogneisses). Both orthogneissic bands share the same lithologies and geochemistry [Rodríguez Aller, 2005], have similar preserved igneous textures suggesting shallow emplacement [Rodríguez Aller, 2005; Díez Fernández, 2011], and intrude similar metasedimentary rocks [Díez Fernández *et al.*, 2010]. Their vertical layout, which extends over several tens of kilometers along the D_2 structure, could be explained by (1) two different but close granitic batholiths being strongly flattened, stretched and superposed by simple shear, and (2) a tectonic duplication of the crustal level where a single granitic suite intruded. Figure 11a shows a sketch of the distribution of the orthogneisses before D_2 , and suggests two possible mechanisms to explain the hypothetical duplication: crustal-scale thrusting or a giant syncline. Both mechanisms would have occurred within the continental subduction channel and fit the ductile deformation recorded in the basal units during D_1 .

6. Exhumation of the Basal Units

6.1. Structural Aspects

[56] D_2 mylonites in the MTU are restricted to the Fervenza massif and the upper half of the Borneiro massif, while in the Órdenes Complex mylonitization affected the Santiago, Carboeiro and Bermés orthogneisses (Figure 10). These

Figure 5. (a) Map showing the axial traces of the two main fold systems around the Fervenza reservoir. UTM coordinates. (b) Attitude of the late subvertical folds inferred from the structural analysis of the regional foliation. Each plot includes foliation planes (black circles), their poles (small dots), the cylindrical best fit (orange circle), and the fold axis obtained (large dot). The arrows in Figure 5a represent the mean direction of D_4 - D_5 fold axes. Numbers inside the arrows indicate the sectors (F1 to F8) used for the stereographic plots. (c) Sketch of the D_4 - D_5 structure and main faults in the northern half of the Malpica-Tui Complex.

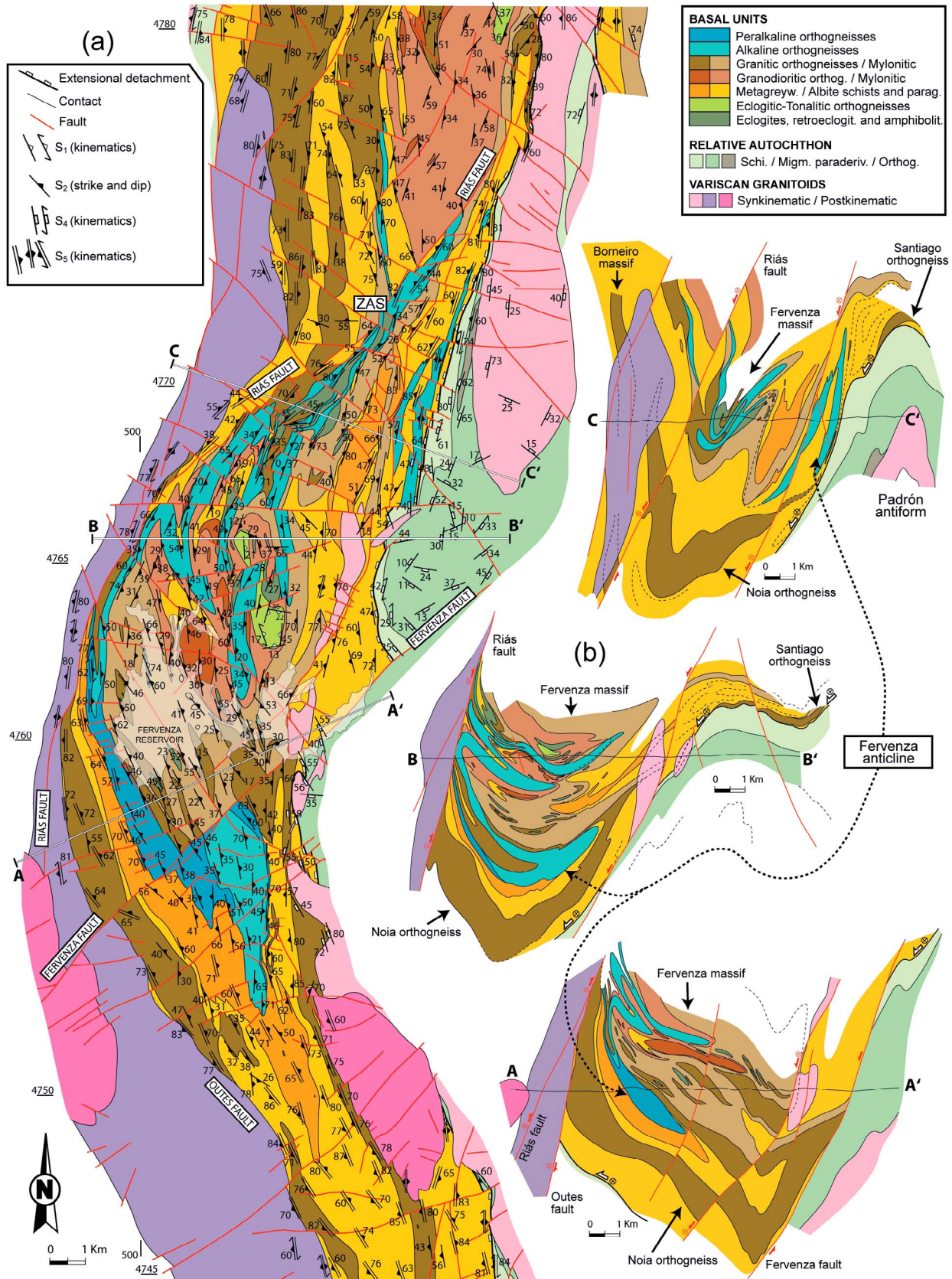


Figure 7. (a) Geological map of the Malpica-Tui Complex around the Fervenza reservoir. The attitude of the main foliations is shown. UTM coordinates. (b) Cross sections.

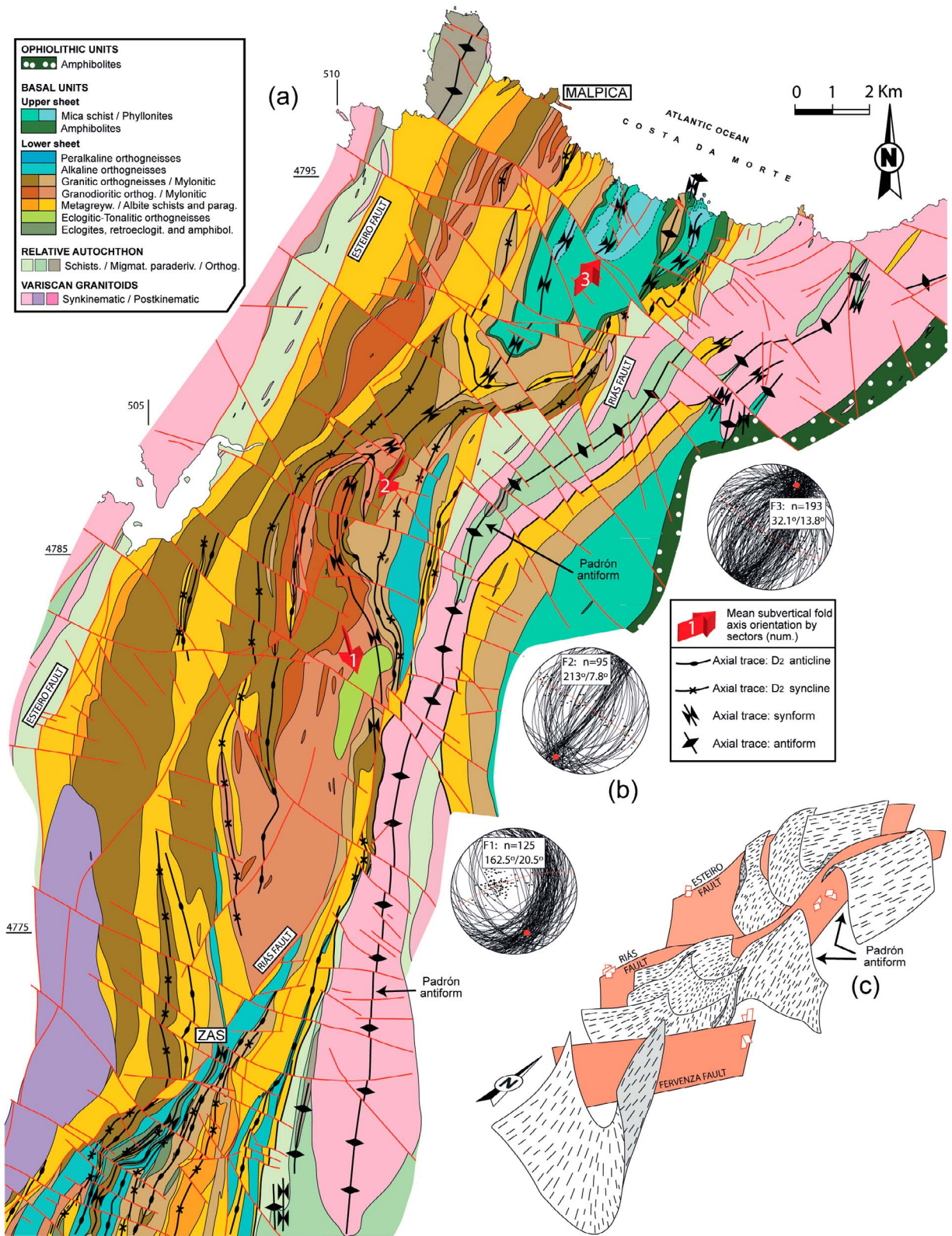


Figure 8. (a) Map showing the axial traces of the two main fold systems in the northern part of the Malpica-Tui Complex. UTM coordinates. (b) Attitude of late folds inferred from the structural analysis of the regional foliation. Each plot includes foliation planes (black circles), their poles (small dots), the cylindrical best fit (orange circle), and the fold axis obtained (large dot). The arrows in Figure 8a represent the mean direction of D₄-D₅ fold axes. Numbers inside the arrows indicate the sectors (F1 to F3) used for the stereographic plots. (c) Sketch of D₄-D₅ folds and main faults.

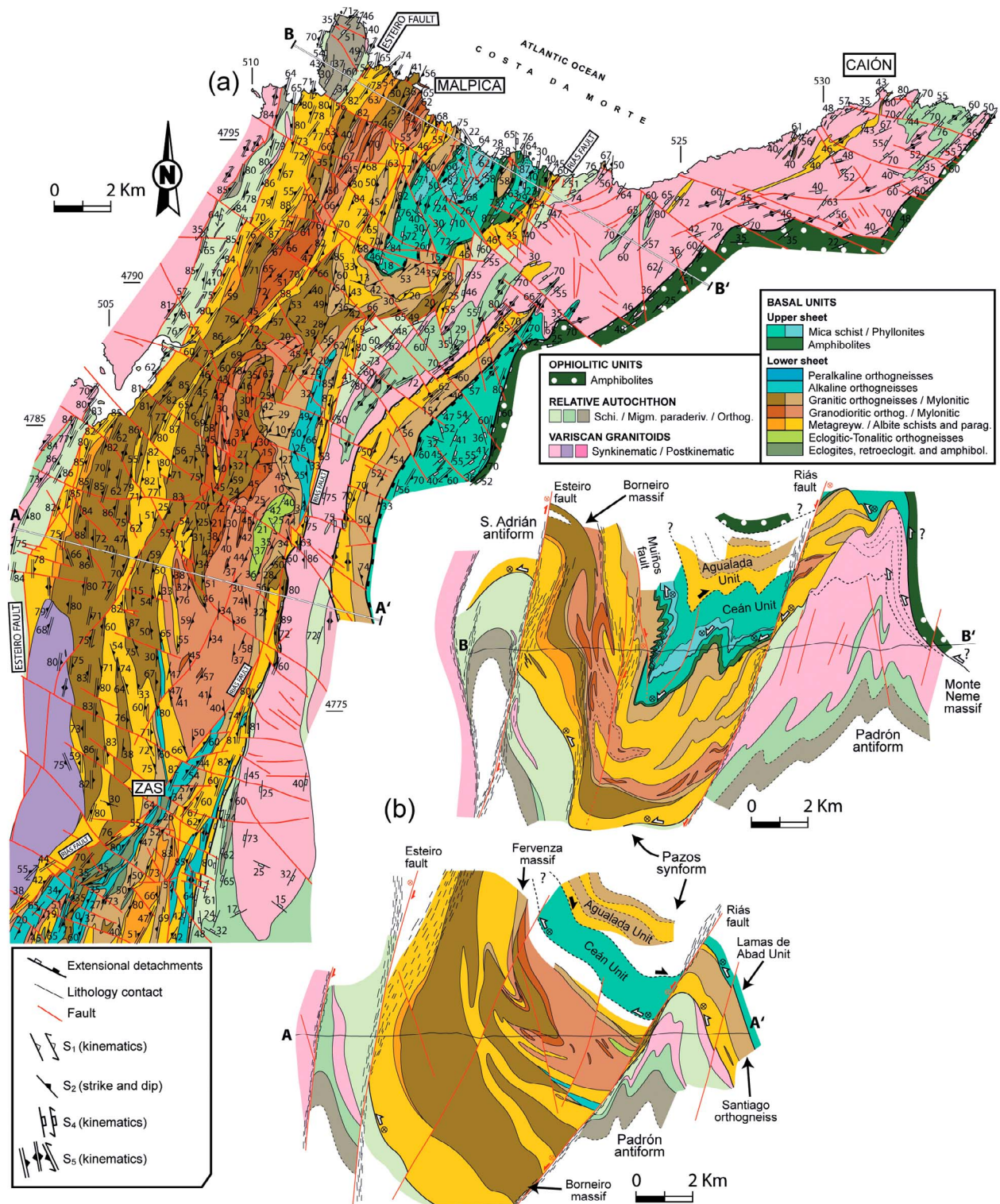


Figure 9. (a) Geological map of the northern part of the Malpica-Tui Complex. The attitude of the main foliations is shown. UTM coordinates. (b) Cross sections.

mylonites represent high-strain domains irregularly distributed through the gneissic massifs, being interpreted as discrete shear zones. Altogether, D₂ mylonites constitute a crustal-scale ductile thrust system developed within the lower tectonic sheet. This regional structure, hereafter the Fervenza thrust,

is the first one developed during exhumation, being the main responsible for the internal duplication of the lower sheet.

[57] The broadest mylonitic zone occurs in the normal limb of the Fervenza anticline, but mylonitization does not affect its reverse limb, suggesting a coeval development of

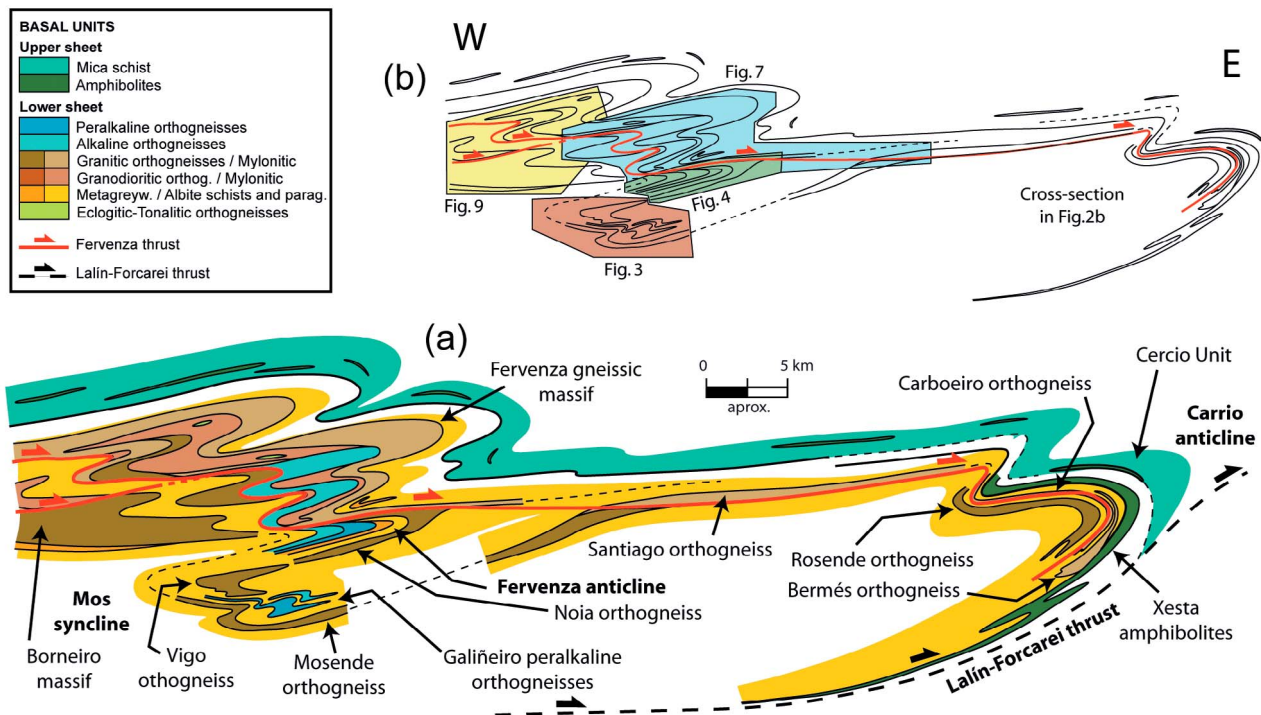


Figure 10. Idealized reconstruction of the regional recumbent fold structure built from the cross sections in Figures 3, 4, 5, and 9.

the fold and the ductile thrust, which, in turn, would have sheared the normal limb in its hanging wall, and rotated the reverse limb in its footwall. However, thrusting was followed by the development of the train of recumbent folds in the normal limb of the Fervenza anticline, as indicated by folding of the mylonitic orthogneisses.

[58] The Mos syncline must be coeval with the Fervenza anticline. The fact that they were not refolded during D_2 indicates that they nucleated during thrusting and evolved during the generation of the subsequent train of recumbent folds (Figures 11c and 11d).

[59] The Fervenza thrust concentrated the deformation and tectonic flow during the first stages of exhumation. The regional strain partitioning depicts a top-to-the-east CS megastructure, where the Fervenza thrust is the C plane and the axial surface to the Fervenza anticline is the south plane (Figure 11c). The length of the reverse limb suggests a minimum tectonic transport of 12–17 km, but considering the displacement accumulated in the mylonites, the total tectonic transport would probably be in the range of several tens of kilometers.

[60] Later on, motion was transferred from the Fervenza thrust to a new in-sequence structure, the Lalín-Forcarei thrust (Figure 11d). Amplification/flattening of preexisting recumbent folds and nucleation/propagation to the east of new ones occurred along the hanging wall of the Lalín-Forcarei thrust.

[61] Development of regional-scale recumbent folds can be explained by a combination of shortening and simple shearing [Sanderson, 1982; Ramsay et al., 1983; Martínez Catalán, 1985; Dietrich and Casey, 1989; Ez, 2000]. Ramsay et al. [1983] proposed shortening along the hanging

wall of a large-scale thrust to explain such structures in the Helvetic nappes. However, this model requires ramps cutting across the competent layers that will nucleate the folds. Staircase geometry is possible for the Lalín-Forcarei thrust, but no ramps have been identified. The staircase option is more typical of foreland thrust belts. As pointed out by Cooper and Trayner [1986], obliquities between thrusts and competent layers may result when thrusts cut previously formed folds developed under the same shearing process. The standard model of nappe folds developed by shearing of the reverse limb of anticlines [Hatcher and Hooper, 1992] matches the Lalín-Forcarei thrust geometry. Moreover, continental subduction zones tend to rise driven by buoyancy forces between crust and mantle [Ernst et al., 1997; Burov et al., 2001]. This causes progressive horizontalization and exhumation of the subduction wedge by horizontal axis rotation, while tangential tectonics is still active, eventually carrying layers of contrasted lithologies to lie in the shortening field and triggering folding.

[62] The Ceán and Lamas de Abad units do not preserve major pre- D_4 structures. Therefore, D_2 ductile regional thrusting and large recumbent folding cannot be described in the upper sheet of the basal units. However, the sequence of microstructures, the compatibility of the model with the metamorphic evolution, and the fact that the contact between the Lalín and Cercio units is folded in the hinge zone of the Carrio anticline suggest that the two tectonic sheets share a comparable Variscan evolution (Figure 11).

6.2. Petrological Aspects

[63] Differential burial during subduction creates pressure gradients, which may be modified during exhumation

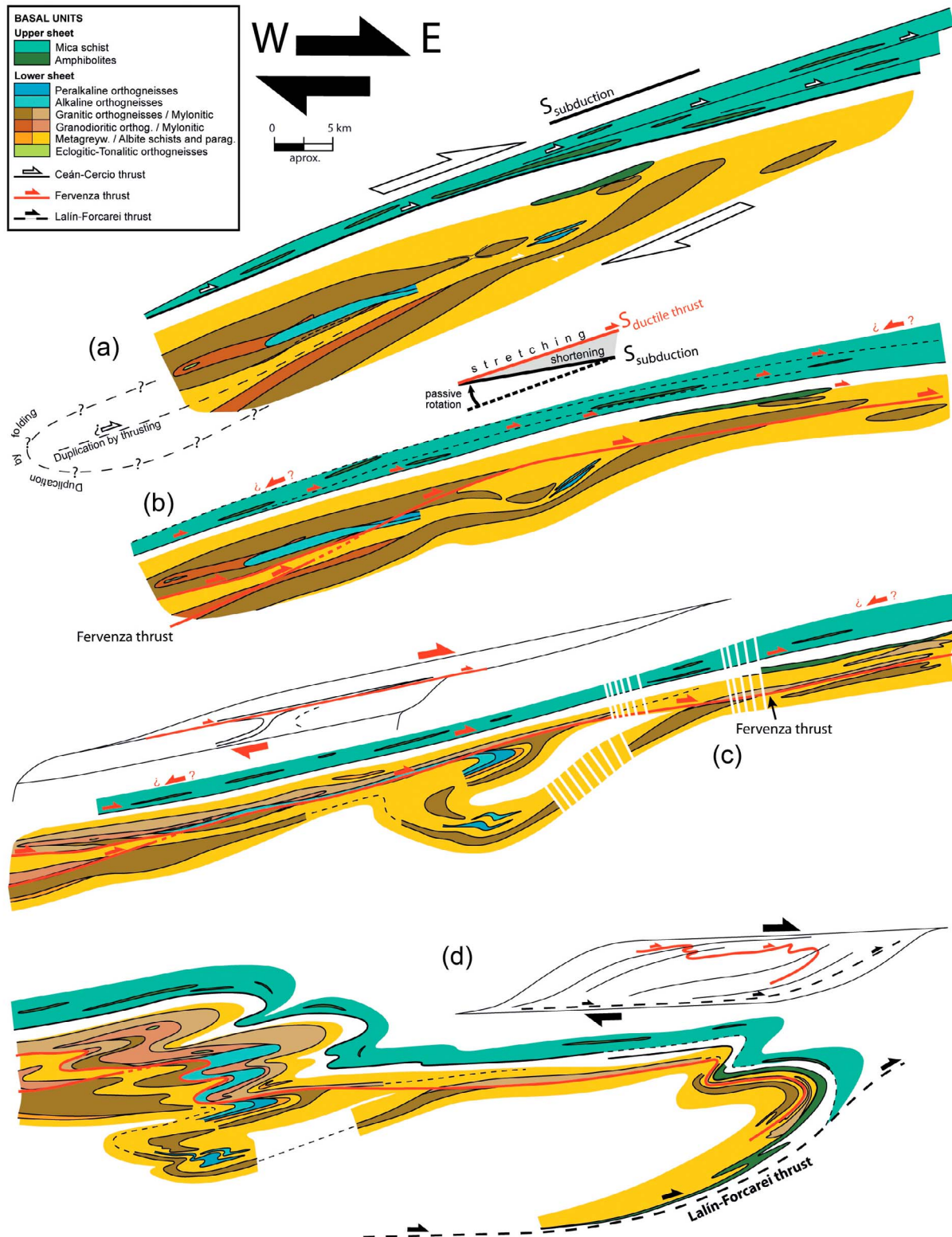


Figure 11. Structural evolution from subduction to early exhumation of the basal units. (a) End of subduction stage. (b) Development of the Fervenza ductile thrust and nucleation of a pair of recumbent folds in the footwall. (c) Passive rotation of the reverse limb, fold amplification, and stretching of the anticline hinge zone while the thrust propagated to the foreland and affected the Santiago orthogneiss. (d) Translation is transferred to the Lalín-Forcarei thrust, which developed a train of recumbent folds along its hanging wall that propagated toward the foreland. Figures 11c and 11d include sketches showing the distribution of deformation.

[Peacock, 2003]. The two HP domains in which the MTU has been classically divided, but whose limit had not been established until now, represent both the hanging wall (more P) and footwall (less P) of the Fervenza thrust.

[64] Maximum peak pressure in the hanging wall to the Fervenza thrust (z26 kbar, garnet-omphacite-phengite [Rodríguez *et al.*, 2003]), compared to peak pressure in the thrust mylonites to the east (Santiago Unit: 15–16 kbar, garnet-omphacite [Rubio Pascual *et al.*, 2002]), suggests the existence of a pressure gradient indicating west directed subduction in present coordinates [Martínez Catalán *et al.*, 1996]. The pressure gradient can be linked to the different positions in the accretionary wedge. E-W horizontal distance between the points where peak pressure estimations were made is about 13 km. To this figure, we must add the shortening related to the train of recumbent folds. The present length of their limbs is around 35 km (Figures 3, 7 and 10), which cannot be considered the original distance but a maximum estimation due to flattening and stretching undergone during D₂, D₄, and D₅.

[65] To sum up, the distance between the Fervenza and Santiago D₂ mylonites before folding did not exceed 25–30 km. Assuming 30° as a maximum inclination of the subducting slab [Alcock *et al.*, 2005], 30 km of distance along slab represents a depth difference of 15 km, i.e., 5 kbar in peak pressure difference, which does not fit the actual difference of around 10 kbar. Moreover, Alcock *et al.* [2005] estimated a dip greater than 20° to be unlikely for this continental subduction zone. Assuming 20° of dip, the initial distance along slab for a pressure difference of 10 kbar would have been approximately 90 km. Therefore, the distance calculated from the pressure difference (10 kbar) and dip of subduction plane are two or three times the estimated original separation before recumbent folding. Much of the pressure difference should be explained by exhumation along the Fervenza thrust, which becomes, hence, quantitatively necessary.

[66] Figure 12 shows the position of the basal units within the subduction channel and the pressure distribution at the end of the subduction stage and early exhumation. It supports the verisimilitude of the thrusting based on structural arguments. Given the typical thermal structure of subduction wedges (Figure 12c) [Minear and Toksöz, 1970; Yamato *et al.*, 2007; Warren *et al.*, 2008; Beaumont *et al.*, 2009], the development of a thrust at the onset of the exhumation would produce accretion of relatively cold material in the footwall, as well as initial decompression and cooling in the hanging wall [Davy and Gillet, 1986; Messiga and Scambelluri, 1991]. This mechanism accords well with the existence of posteclogitic glaucophane in the metabasites of the MTU.

[67] The shallower sections of the lower tectonic sheet did not reach eclogite facies but blueschists facies conditions, as described for the Forcarei Unit [Martínez Catalán *et al.*, 1996]. Eclogite facies relicts in the southern MTU rule out a possible correlation of the reverse limb of the Mos syncline with that of the Carrio anticline, thus reinforcing the structural correlation depicted in Figure 10.

[68] A contrasted HP thermobaric record exists between the lower and upper tectonic sheets. Peak pressure in the lower sheet is about 26 kbar [Rodríguez *et al.*, 2003], while in the upper sheet it ranges between 17 and 19 kbar (garnet-glaucophane-chloritoid-phengite [Arenas *et al.*, 1995; López-

Carmona *et al.*, 2010]). Therefore, a normal regional gradient is defined for the whole Malpica-Tui Complex. Before the Variscan cycle, the lower sheet represents a domain placed closer to the Gondwana mainland than that represented by the upper sheet [Díez Fernández *et al.*, 2010]. The lower sheet was accreted beneath the upper sheet during the eo-Variscan subduction (Figures 11 and 12), although the present contacts between them are extensional detachments. The nature of the original contact is uncertain, though the lithostratigraphy of the outer part of the Carrio anticline (Lalín, Forcarei and Cercio units) suggests a transition between the lower and upper sheet compatible with a distributed shearing.

7. Model for the Contractual Exhumation of the Basal Units

[69] The Fervenza thrust accounts for the exhumation of crustal material concentrated along the upper half of the subducted slab. This thrust is the major structure developed in the subduction-exhumation channel and represents forced tectonic backflow between the Gondwanan continental crust and the base of the Laurussian lithosphere during the transition from a subduction to a collisional regime (Figures 13a and 13b). Continuation of convergence created a new contractional shear zone, the Lalín-Forcarei thrust, which forced the deactivation of tectonic flow within the subduction-exhumation channel and can be considered the beginning of the continental collision. It is related to the nucleation and propagation to the foreland of a train of recumbent folds that transported the subduction-exhumation channel onto inner domains of Gondwana (Figure 13c). The Lalín-Forcarei and subsequent thrusts did not create HP gradients at their footwalls. Such thrusts represent accretion to the orogen (underthrusting) of thicker and more buoyant continental lithosphere, and they might have been accompanied by normal faulting and extension above to compensate crustal thickening and tectonic extrusion [e.g., Chemenda *et al.*, 1995].

[70] The regional structures depicted by the Fervenza and Lalín-Forcarei thrusts have striking similarities with the flow patterns obtained from numerical and analogical modeling on the exhumation of high- to ultrahigh-pressure rocks [e.g., Gerya *et al.*, 2002; Warren *et al.*, 2008; Beaumont *et al.*, 2009]. The overriding of the upper sheet together with the internal duplication of the lower sheet during subduction (Figure 11a) and the early exhumation via the Fervenza thrust (Figures 11b and 11c) would account for a forced clockwise return flow within the subduction channel. At the front of the orogenic wedge, the Lalín-Forcarei thrust and the train of recumbent folds (Figure 11d) conform a nappe-fold geometry, which continued clockwise ductile circulation of coherent slices at crustal levels.

[71] Exhumation combined active and passive mechanisms. The active mechanisms included the Fervenza and Lalín-Forcarei in-sequence thrusts, whereas the passive mechanism was the progressive horizontal rotation of the subduction-exhumation channel induced by buoyancy forces and underthrusting. Tangential stress related to convergence was important, but vertical stress related to gravitational forces too. This might have produced a progressive transition from a simple shear dominated regime, to a pure shear dominated one. The gravitational component could have

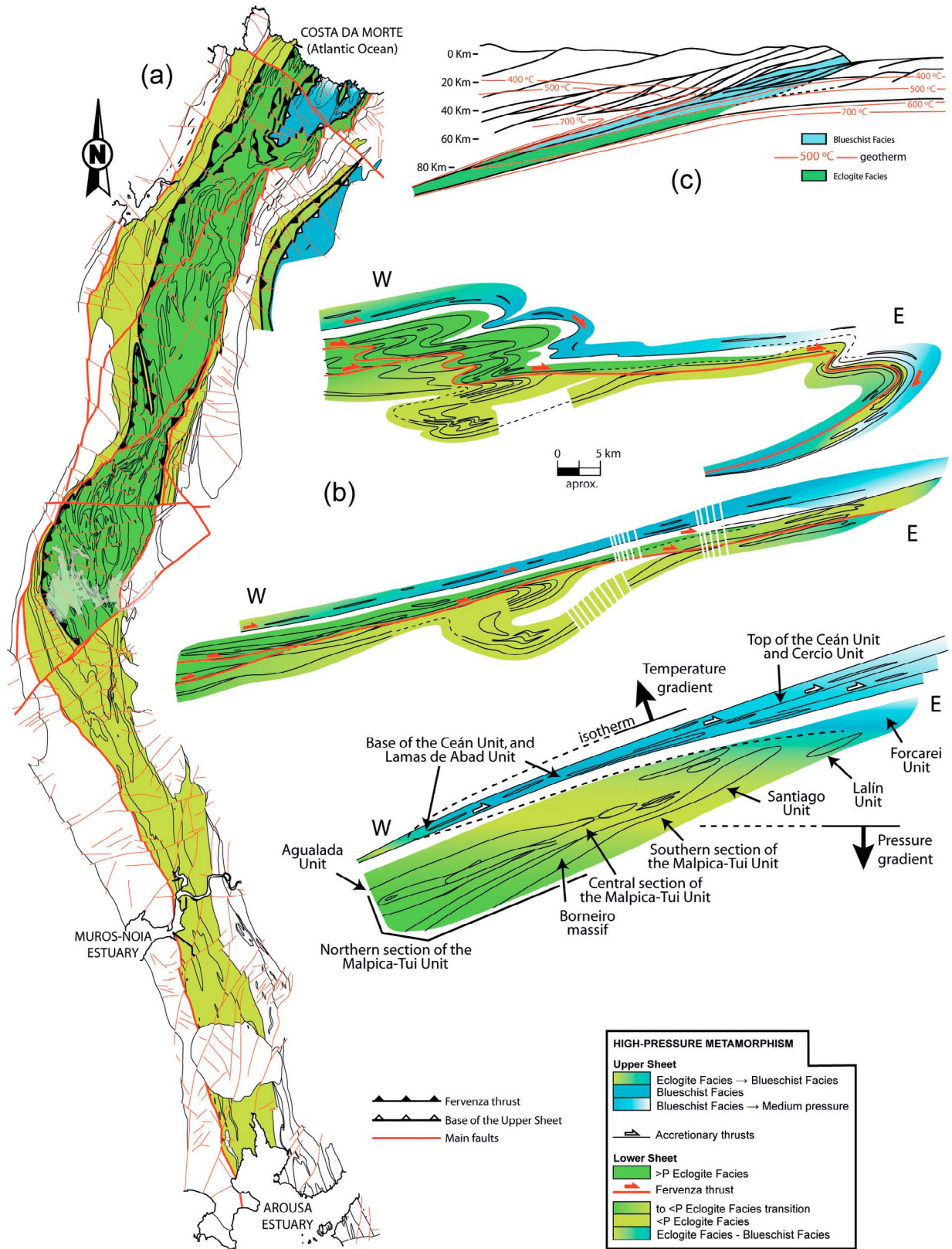


Figure 12. (a) Regional distribution of high-pressure domains in the central and northern sections of the Malpica-Tui Complex. The base of the Fervenza thrust and the detachment below the Upper Sheet are shown. (b) Distribution of HP domains during the structural evolution shown in Figure 11. (c) Sketch of the continental subduction zone showing the distribution of metamorphic facies and the deflection of isotherms.

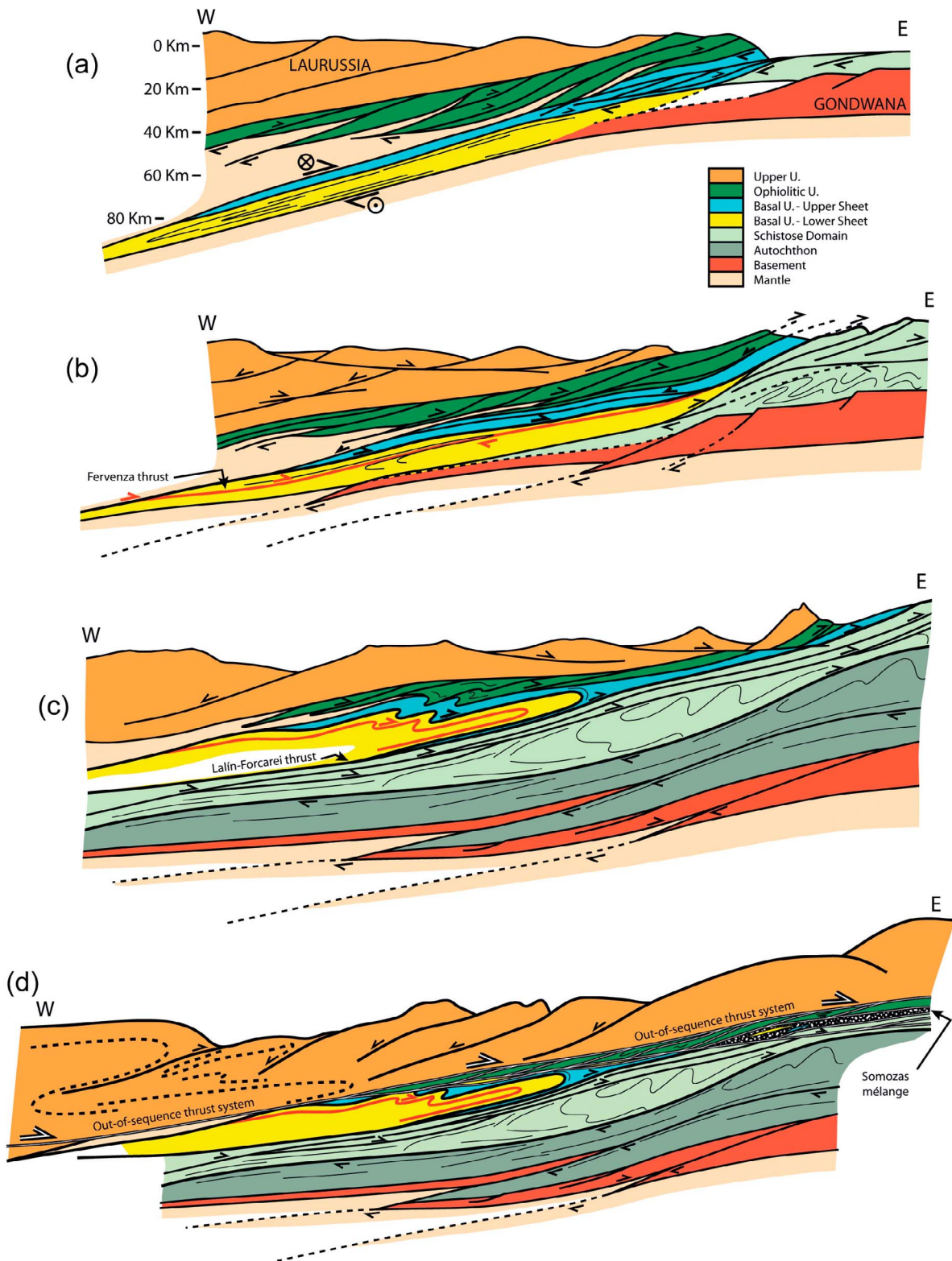


Figure 13. Tectonic evolution of the Variscan subduction channel preserved in NW Iberia. (a) Architecture during subduction. Stacking of the lithostratigraphic record of Gondwanan continental edge. (b) Start of deep exhumation activated by underthrusting and initial accretion of the Gondwanan autochthon (Fervenza thrust). First recumbent fold generation during exhumation. (c) Full imbrication of the Gondwanan autochthon and second generation of recumbent folds (Lalín-Forcarei thrust). (d) Out-of-sequence thrusting and translation of peri-Gondwanan terranes, ophiolitic units, and slices of the subduction-exhumation wedge toward the orogenic front.

played a decisive role in the amplification of recumbent folds and development of a penetrative S_2 axial planar foliation [Bucher, 1963; Hudleston, 1977].

[72] The axial plane foliation to the train of recumbent folds in the MTU ranges in age between 340 and 350 Ma [$^{40}\text{Ar}/^{39}\text{Ar}$ in muscovite and phengite, Rodríguez *et al.*, 2003], which accords well with the slightly younger estimated age of the Lalín-Forcarei thrust and the Carrio anticline (340 Ma [Martínez Catalán *et al.*, 1996; Dallmeyer *et al.*, 1997]). The age of the Fervenza thrust can be constrained in the range 350–370 Ma, since the HP metamorphism is dated at 380–370 Ma (Rb-Sr in micas [Van Calsteren *et al.*, 1979; Santos Zalduegui *et al.*, 1995], $^{40}\text{Ar}/^{39}\text{Ar}$ in phengite and paragonite and apparent ages in biotite [Rodríguez *et al.*, 2003], U-Pb in metamorphic zircon rims [Abati *et al.*, 2010]).

[73] The out-of-sequence thrust system is less inclined than the Lalín-Forcarei thrust. Their relative chronology [Martínez Catalán *et al.*, 2002] accounts for the progressive horizontalization of the reference structural planes through time. The out-of-sequence thrust system was the response to superimposed convergent shortening during the Variscan collision, which forced the transference of ophiolitic (Rheic) and peri-Gondwanan (Laurussia) terranes from the previously built accretionary prism to the Gondwana mainland (Figure 13d).

8. Conclusions

[74] The basal units of the NW Iberia allochthonous complexes are pieces of the outermost continental margin of northern Gondwana, which was subducted at the onset of the Variscan Orogeny during the Upper Devonian (380–370 Ma) beneath an accretionary wedge built during the closure of the Rheic Ocean. Two sedimentary sequences deposited in the outer continental margin and their associated igneous rocks were heterogeneously flattened, stretched and metamorphosed in the subduction channel, while a tectonic fabric (S_1) developed. The subduction process stacked the lithostratigraphic record in two tectonic sheets. No discontinuity formed during the subduction has been found between both tectonic sheets, although it could be hidden by the present extensional detachments separating them. However, no emplacement of older onto younger rocks occurred.

[75] The metamorphic conditions and the relative position of the two tectonic sheets have permitted to reconstruct the thermal architecture of the subduction channel. Differential burial during subduction established a lineal pressure gradient, while deflection of the geotherms drove relatively cold material to reach depths at which other parts of the subduction channel reached higher temperatures. The lower sheet recorded an initial prograde eclogite facies metamorphism in the more pressurized zones, and progressive colder conditions (blueschist facies metamorphism) toward the shallower zones. The upper sheet shows a blueschist facies evolution.

[76] The transition from subduction to exhumation was accomplished by the development of a ductile thrust within the subduction channel during the Famennian and Tournaisian (370–350 Ma). The Fervenza thrust placed the deepest

domains of the lower sheet onto shallower domains, causing mylonitization in its hanging wall, and forcing the development of a pair of recumbent folds in its footwall. The thrust caused the duplication of the lower sheet and carried piggyback the upper sheet. It was developed at HP conditions and brought deep domains of the subduction channel at its hanging wall near shallower and colder domains of the footwall.

[77] The exhumation continued during the Tournaisian and Viséan (350–340 Ma) with the development of a train of recumbent folds that propagated progressively to the east, while those formed before were amplified and flattened. During their propagation, the largest fold, the Carrio anticline, developed the Lalín-Forcarei thrust (circa 340 Ma) at its reverse limb. The train of recumbent folds developed a penetrative axial planar foliation (S_2) under amphibolite to greenschist facies conditions.

[78] The Fervenza and Lalín-Forcarei thrusts drove the exhumation and subsequent emplacement of the basal units over the parautochthonous Schistose Domain, transferring the continental subduction channel on top of more internal domains of the Gondwana margin, which was heated and moderately pressurized.

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J. Abati Gómez and R. Arenas Martín, Departamento de Petrología y Geoquímica, Universidad Complutense/Consejo Superior de Investigaciones Científicas, D-28040 Madrid, Spain.

R. Díez Fernández and J. R. Martínez Catalán, Departamento de Geología, Universidad de Salamanca, E-37008 Salamanca, Spain. (georuben@usal.es)

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