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Tectonic framework of the Cabo Ortegal Complex: A slab of lower crust exhumed in the Variscan orogen (northwestern Iberian Peninsula)

A. Marcos*
P. Farias
Departamento de Geología, Universidad de Oviedo, Arias de Velasco s/n, 33005 Oviedo, Spain
G. Galán
Departamento de Geología, Universidad Autónoma de Barcelona,
Edificio C (Sur), 08193 Bellaterra, Barcelona, Spain
F.J. Fernández
S. Llana-Fúnez*
Departamento de Geología, Universidad de Oviedo, Arias de Velasco s/n, 33005 Oviedo, Spain

ABSTRACT

The Cabo Ortegal Complex is a composite allochthonous terrane that was thrust onto the western edge of Gondwana during the Variscan orogeny. It is formed of two main tectonic units: the Upper Tectonic unit, comprising rocks affected by high-pressure (*P*)–high-temperature (*T*) metamorphism, and the Lower Tectonic unit, which represents the resulting suture of the Variscan collision. The suture preserves remnants of strongly deformed and metamorphosed ophiolitic rocks overriding the parautochthon, and the lower Paleozoic sequence of the Ollo de Sapo antiform, regarded as the autochthonous sequence of the Iberian plate. The Upper Tectonic unit is formed by layered ultramafic, mafic, and quartzo-felspathic rocks that were buried at levels in excess of 50 km (~1.56 GPa) before the Variscan collision in a convergent plate boundary within the Rheic ocean domain ca. 490–480 Ma (Early Ordovician). They have been interpreted either (1) as an earlier thinned continental crust, underlain by a lithospheric mantle and oceanic spreading, or (2) as independent terranes, formed in different geodynamic settings (island arc, oceanic). Most structures observed in these rocks are ductile and are associated to their exhumation process. It started with the development of a persistent horizontal foliation in granulite facies conditions, which equilibrated in amphibolite facies conditions ca. 385 Ma, and ended in higher crustal levels with the progressive development of noncoaxial structures, such as east-verging asymmetrical isoclinal folds and thrusts, leading to the emplacement of the Upper Tectonic unit over the Lower Tectonic unit ca. 365 Ma (Late Devonian).

*E-mail, Marcos: marcos@geol.uniovi.es. Current address, Llana-Fúnez: Rock Deformation Laboratory, Department of Earth Sciences, University of Manchester, Oxford Road, Manchester M13 9PL, UK.

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INTRODUCTION

The hinterland of the Variscan orogen in the Iberian Peninsula preserves several allochthonous complexes as klippen, including exotic ultrabasic and basic rocks, orthogneisses, and various metasediments interpreted by some as indicative of a polymetamorphic and polyorogenic evolution (den Tex, 1981; Arenas et al., 1986; Ribeiro et al., 1990; Fernández Suárez et al., 2000) (Fig. 1; Plate 1). The Cabo Ortegal Complex is the easternmost of these structures and defines the suture zone of the orogen. It comprises rock units with entirely different geodynamic settings, ages, and metamorphic and structural evolutions, providing valuable information on the tectonic evolution of the collisional processes.

In the hanging wall of the suture zone, a sequence of crystalline rocks records a metamorphic event of high-pressure (*P*) (~1.5 GPa minimum), high-temperature (*T*) conditions. This sequence is considered the relic of an earlier thinned lower crust (probably with a portion of lithospheric mantle) (Galán and Marcos, 1997), although some authors (Peucat et al., 1990) considered some of the mafic rocks related to a volcanic arc close to a continental margin. Deformation started in shear zones under eclogite facies conditions, but was reequilibrated during their exhumation under retrograde granulite and amphibolite facies conditions. The presence of such lower crustal slabs in relation with thickening processes in collisional orogens has been widely reported in other mountain belts (Windley, 1983; Windley and Tarney, 1986). Examples occur in the Kohistan island arc of the Himalayas (Coward et al., 1982; Coward and Windley, 1986; Bard, 1983), the Ivrea zone of the Italian Alps (Rutter et al., 1993), and the Variscan sutures of western-central Europe (Matte, 1983, 1998).

The origin of the complexes of northwestern Iberia has been a matter of debate in recent years. Some authors suggested an autochthonous "mantle plume" model for their origin (Keasberry et al., 1976; Calsteren, 1977, 1981; Calsteren and den Tex, 1978; Kuijper and Arps, 1983; Kuijper et al., 1985), whereas others consider them as allochthonous terranes emplaced on the Gondwana margin during the Variscan orogeny (e.g., Ribeiro et al., 1964; Anthonioz, 1970; Ries and Shackleton, 1971; Bayer and Matte, 1979; Iglesias et al., 1983; Bastida et al., 1984; Arenas et al., 1986; Matte, 1986; Pérez Estáun et al., 1991; Dallmeyer et al., 1997; Martínez Catalán et al., 1997, 1999). Although at present the allochthonous hypothesis is generally accepted, there is still controversy regarding the origin and metamorphic evolution of the different lithological and structural units.

There are two major groups of hypotheses about the metamorphic evolution of the high-*P*–high-*T* rocks of the Cabo Ortegal Complex: polyorogenic and single-cycle models. Polyorogenic metamorphic models suggest that the eclogite and granulite facies metamorphism occurred during Precambrian time (Vogel, 1967; Anthonioz, 1970; Engels, 1972; Marques et al., 1996), and a strong retrogression in amphibolite facies is related to pre-Variscan and/or Variscan events. Single-cycle metamorphic models propose the subduction or underthrusting of both oceanic (ophiolitic rocks) and continental crust and later obduction of them during the Variscan orogeny (Gil Ibarguchi et al., 1990; Peucat et al., 1990; Arenas, 1991; Schäfer et al., 1993; Marcos and Galán, 1994; Àbalos et al., 1996; Santos Zalduegui et al., 1996; Matte, 1998; Galán and Marcos, 1999, 2000). The eclogite and quasisimultaneous granulite facies metamorphism is related to the subduction or underthrusting process either ca. 480 Ma (Peucat et al., 1990) or ca. 395 Ma (Schäfer et al., 1993; Santos Zalduegui et al., 1996; Ordoñez Casado et al., 1996). Pre-Variscan and Variscan exhumation caused the retrograde metamorphism in amphibolite and greenschist facies. More recent U-Pb dating (Fernández Suárez et al., 2002) indicates that these rocks record a polyorogenic evolution with a high-*P*–high-*T* episode ca. 490–480 Ma and a younger event dated as ca. 390–385 Ma. According to this model, the lower crustal rocks of the Cabo Ortegal Complex would have been exhumed to the surface from >50 km depth (1.5 GPa) in two consecutive tectonic events, separated in time by ~100 m.y. The exhumation of deep crustal rocks (granulites formed in a thickened crust) during two unrelated cycles of tectonism is consistent with the model proposed by Ellis (1987).

Based on this model, the aims of this chapter are (1) to study the ductile deformation in the high-*P*–high-*T* rocks, where pervasive ductile deformation is retrogressive and related to the process of their exhumation; and (2) to assess the effects of the emplacement of the lower crustal section onto the Iberian plate during the Variscan orogeny. A progressive change in the emplacement direction is discussed with geodynamic models of the Variscan orogeny in the Iberian Belt.

GEOLOGIC SETTING

The Galicia-Trás-os-Montes zone (Farias et al., 1987) is the most internal zone of the Variscan Belt in the northwest of the Iberian Peninsula (Fig. 1). It was thrust over Paleozoic metasediments and porphyroids of the neighboring Central Iberian zone, which is represented in northwest Spain by the Ollo de Sapo anticlinorium, a narrow structure built by the interference between large recumbent and late Variscan open subvertical (upright) folds. Two domains have been distinguished within the Galicia-Tras os Montes zone: (1) the Schistose domain (also named parautochthon; Ribeiro et al., 1990), mainly composed of a metasedimentary sequence more than 4000 m thick and probably of early Paleozoic age; and (2) the Allochthonous complexes, tectonically emplaced over the parautochthon and composed of mafic-ultramafic and quartzo-feldspathic rocks with different origins (e.g., oceanic, ophiolitic, and continental). The Schistose domain has been interpreted as the most external sediments of the continental margin of Gondwana (Ribeiro et al., 1990; Martínez Catalán et al., 1997, 1999; Marcos and Farias, 1999).

The rocks forming the Allochthonous complexes are grouped in three major tectonostratigraphic units characterized by their different rock assemblages and metamorphic evolution;

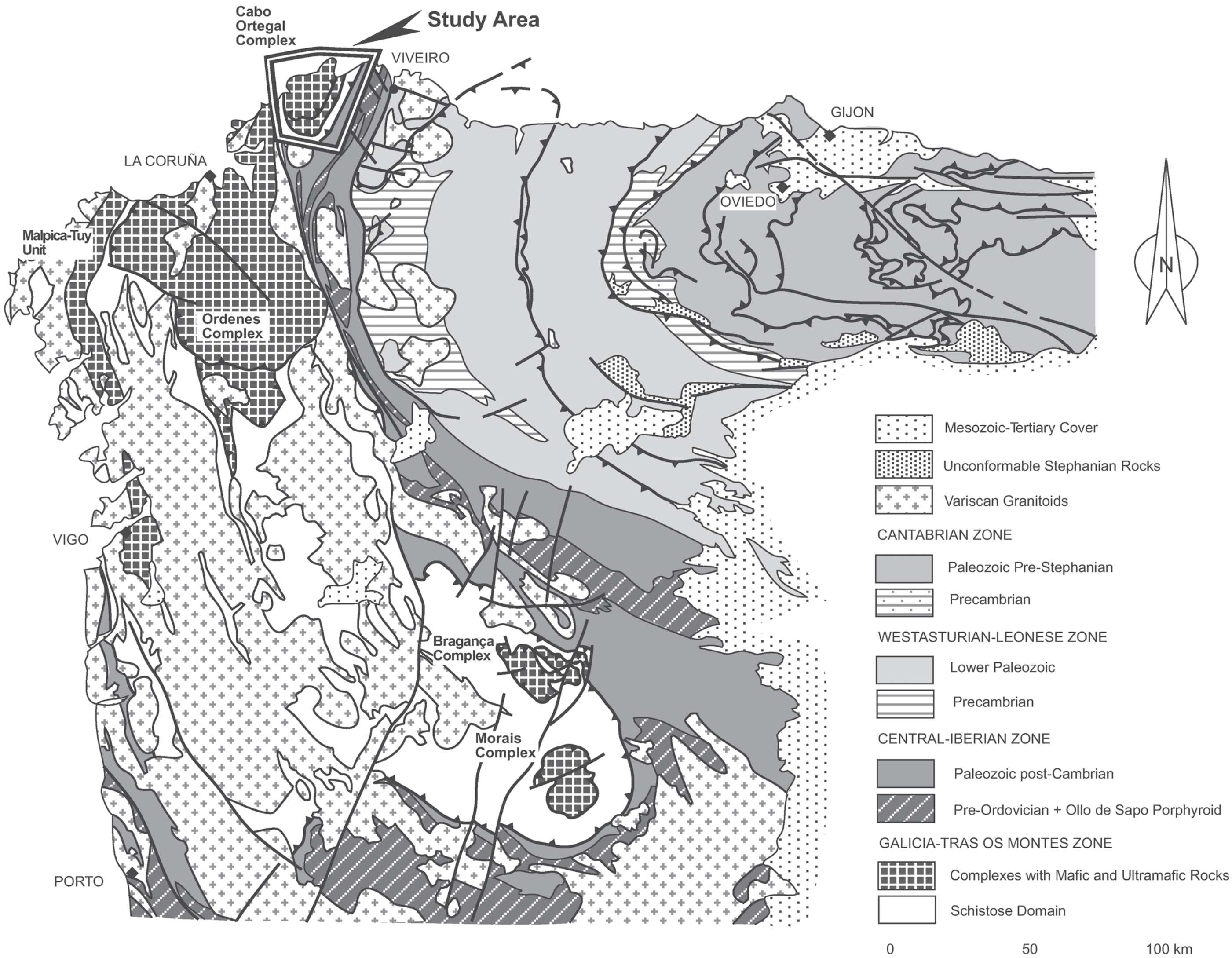


Figure 1. Geological map of northwestern Iberia, simplified after Julivert et al. (1974), Parga Pondal et al. (1982), and Farias et al. (1987). Various allochthonous units and main paleogeographic and structural ensembles of Iberian autochthonous units are shown.

those are, from bottom to top, the Basal, Ophiolitic, and Upper units (Arenas et al., 1986). These units have been correlated with the paleogeographic realms involved in the Variscan collision (Martínez Catalán et al., 1997, 1999). The Basal units are formed of orthogneisses, paragneisses, amphibolites, and schists. These rocks, related to the external continental margin of Gondwana, intruded during an Ordovician rifting event (Ribeiro and Floor, 1987; Martínez Catalán et al., 1996; Marcos and Farias, 1999), and later were involved in a westward subduction episode at the beginning of the Variscan orogeny. They record metamorphism under high-*P* and low- to intermediate-*T* conditions (Arenas et al., 1995; Martínez Catalán et al., 1996). The Ophiolitic units include partially mylonitized amphibolites of basaltic origin and a tectonic melange. They represent the suture zone formed by the closure of the Paleozoic Rheic ocean (Martínez Catalán et al., 1997, 1999). The Upper units are placed in the hanging wall to the suture as an exotic terrane and include high-*P*–high-*T* rocks and mainly intermediate-*P* metasediments, as a whole seen as the remnants of an earlier thinned lower crustal section.

All these units are found in the Cabo Ortegal Complex (Vogel, 1967); the closest crystalline klippe to the foreland belt (Fig. 1), but we have grouped them slightly differently for simplification. Thus, we distinguish the Upper Tectonic unit (containing the high-*P*–high-*T* rocks that formed the thinned lower crust) and the Lower Tectonic unit, which includes both the ophiolitic rocks and the basal units and represents the suture of the Variscan collisional belt in northwest Iberia. The thin Rio

Baio thrust sheet (regarded as parautochthonous) separates the Lower Tectonic unit from the underlying lower Paleozoic metasediments of the Ollo de Sapo antiform (Plate 1).

PARAAUTOCHTHON AND AUTOCHTHON OF THE CABO ORTEGAL COMPLEX

The Rio Baio thrust sheet forms part of the Schistose domain of the Galicia-Tras os Montes zone (Farias et al., 1987; Marcos and Farias, 1999). It is composed of a thick succession of siliciclastic metasediments (quartzites, feldspathic sandstones, graywackes, shales) with some interbedded layers of volcanic-sedimentary and volcanic rocks (rhyolites and dacites), all included within the Loiba and Queiroga Series by Marcos and Farias (1999) (Plate 1). This lithostratigraphy, of unknown age, has several differences with respect to the underlying autochthon.

The rocks of the Rio Baio thrust sheet form a monoclinal sequence with scarce folds, but are affected by a slaty cleavage developed under greenschist facies conditions. The basal thrust is parallel to the bedding in the hanging-wall rocks, and is a 50–120-m-thick zone with intense phyllonitization. The phyllonites at the base of the Rio Baio thrust sheet show a weak stretching lineation with a constant northeast orientation and the development of C′ shear bands, which indicate movement in the same direction, top to the northeast. However, the last tectonic pulses produced discontinuous structures in the phyllonites, i.e., centimeter-scale tectonic wedges, with opposite kinematics (top to the southeast). This late sense of movement is in agreement with the cutoff lines of Silurian rocks placed in the autochthon (Plate 1) (Marcos and Farias, 1999).

The Rio Baio thrust sheet is tectonically placed on the lower Paleozoic (Lower Ordovician to Silurian) metasedimentary sequence in the western limb of the Ollo de Sapo antiform, which represents the autochthon of the Cabo Ortegal Complex (Matte, 1968; Pérez Estaún et al., 1991) (Plate 1). Structures developed in these rocks under greenschist and amphibolite facies have been related to two distinct Variscan phases of deformation (Matte, 1968; Capdevila, 1968; Bastida et al., 1984, 1993; Réche et al., 1998). The first phase is represented by tight recumbent folds, from centimeter to kilometer scale, with horizontal north-northeast–south-southwest axes, vergent to the east (Matte, 1968). This folding is associated with axial planar slaty cleavage, which is the regional tectonic foliation in the western limb of the antiform. The structures are overprinted by straight, open folds, with horizontal or slightly north-dipping fold axes, which develop a crenulation cleavage parallel to the axial planes. The basal thrust of the Rio Baio thrust sheet cuts the first folds but is deformed by the later open folds (Marcos and Farias, 1999) (Fig. 2).

LOWER TECTONIC UNIT: THE SUTURE ZONE

The rocks of the Lower Tectonic unit that were initially interpreted as a sequence with ophiolitic affinities (Martínez García et al., 1975; Arce Duarte et al., 1977; Bayer and Matte, 1979; den Tex, 1981; Iglesias et al., 1983; Arenas and Peinado, 1984; Arenas et al., 1986; Arenas, 1988; Ribeiro et al., 1990) are var-

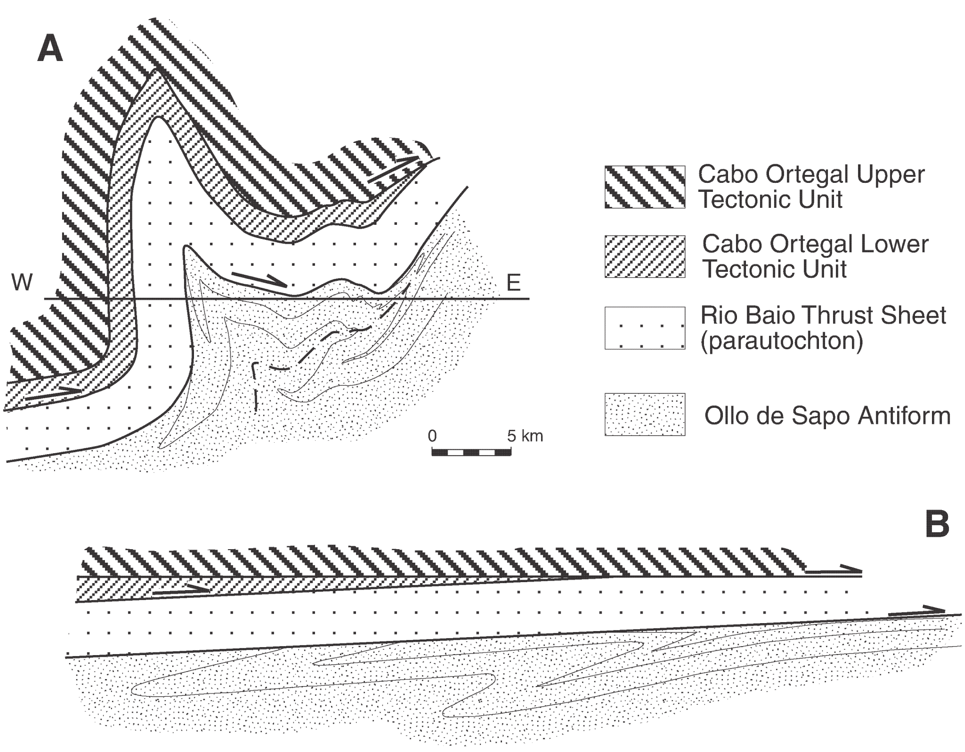


Figure 2. A: Schematic geological cross section of Cabo Ortegal Complex showing relation between allochthonous thrust sheets and their autochthonous sequence. B: Partial restoration of cross-section A (after Marcos and Farias, 1999).

ied lithological elements arranged heterogeneously (see geological maps in Fernández Pompa and Monteserín López, 1976; Arce Duarte et al., 1977; Arenas, 1988). However, detailed geological mapping by Marcos and Farias (1997, 1999) show that these rocks can be grouped in three superposed and separate thrust sheets: the Purrido, Moeche, and Espasante units (Plate 1; Fig. 3). They represent the suture zone of the Variscan orogen in the Cabo Ortegal Complex and separate two distinct terranes: the rocks from the Iberian plate below, and the high-*P* rocks of Upper Tectonic unit above.

Purrido thrust sheet

The ~1100-m-thick Purrido thrust sheet is composed of two formations, the Purrido amphibolites (Engels, 1972; Vogel, 1967; Arenas, 1988, 1991), and the Labacengos ultramylonites

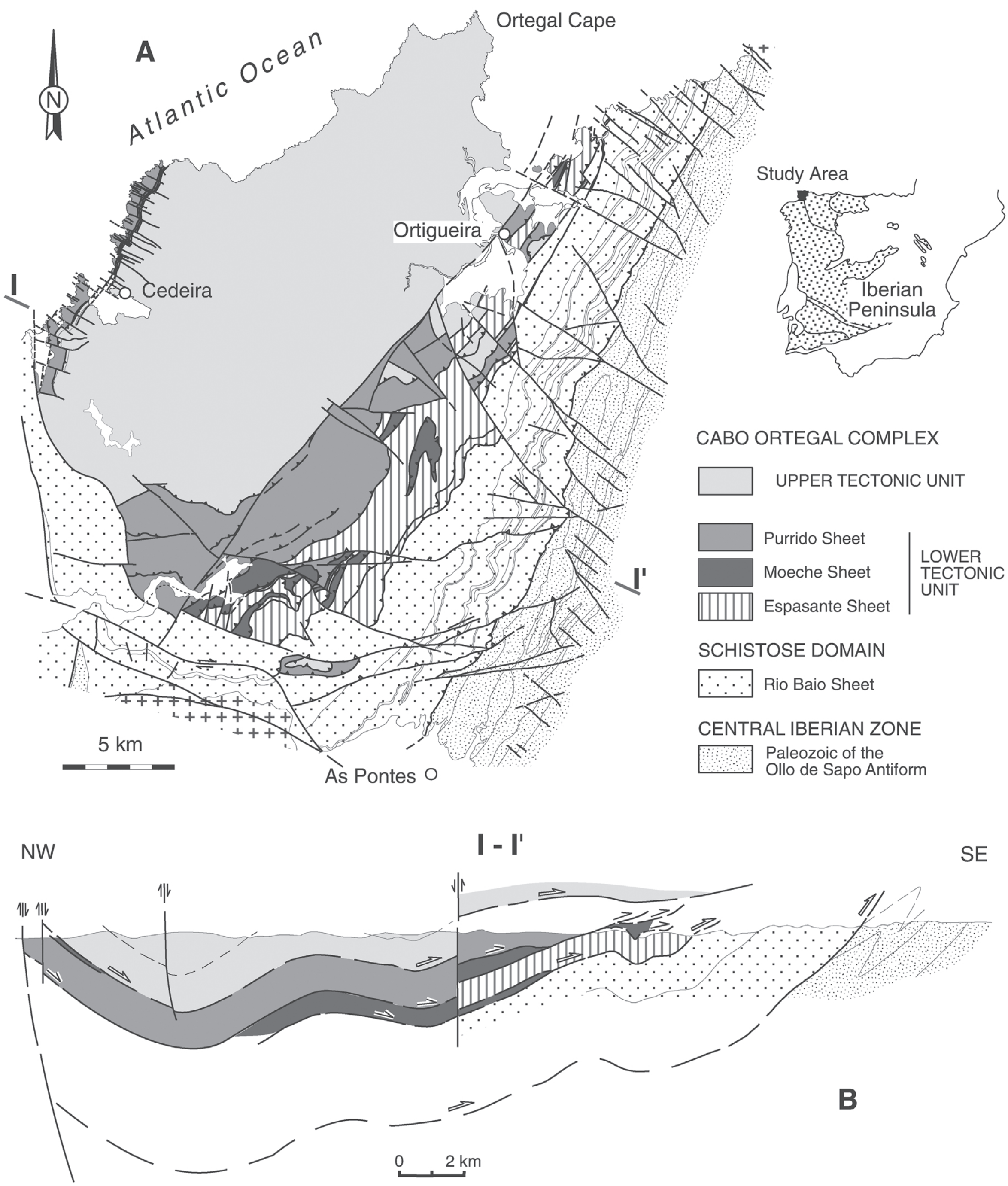


Figure 3. Structure and geological cross section through Lower Tectonic unit (after Marcos and Farias, 1999).

and phyllonites at the base (Marcos and Farias, 1997). The Purrido amphibolites (~300 m thick) form a homogeneous succession of green-grayish amphibolites that have a fine and discontinuous compositional banding defined by plagioclase. The amphibolites are composed of Hbl, Pl, and Ep-Czo, and Grt, Rt, and Ttn are accessory minerals (mineral symbols as in Kretz, 1983); they are probably the metamorphic equivalent of gabbroic rocks (Vogel, 1967). The rocks present a well-developed tectonic fabric defined by the shape-preferred orientation of hornblende, giving the rock fissility (Fig. 4A). The amphibolite facies metamorphism (500–600 °C) was dated by Peucat et al. (1990) as ca. 390 Ma, using $^{40}Ar/^{39}Ar$ in Hbl. The development of the tectonic fabric is probably of the same age.

The Labacengos ultramylonites and phyllonites are made of rocks with the field appearance of phyllites that are affected by a strong and penetrative tectonic foliation (Fig. 4, B and C). The rocks are composed of Qtz, Chl, Act, Ab, Ms, and Bt and derived tectonites from the extreme deformation of the Purrido amphibolites (Arenas, 1983, 1988). This deformation was dated by Dallmeyer et al. (1997) as ca. 364 Ma ($^{40}Ar/^{39}Ar$ whole-rock age).

The Purrido amphibolites have the chemical composition of olivine tholeiitic basalts, whereas the Labacengos ultramylonites and phyllonites have a quartz tholeiitic composition (Arenas, 1988), which points to some chemical changes during mylonitization. Available trace element compositions of these rocks do not discriminate between normal mid-ocean ridge basalt (N-MORB) and within-plate basalt; Arenas (1988) suggested an average enriched-MORB composition for them, although it is uncertain whether it is a primary or a secondary character.

Moeche thrust sheet

This tectonic unit forms a thin, <500-m-thick sheet made of scattered metric to decametric fragments of exotic rocks (e.g., marble, quartzite, basic rocks), but also of native rocks (serpentinized ultramafic rocks), all of them in a matrix of serpentinite, talc-schists, and chlorite-schists. Meer Mohr (1975) found coral fragments, crinoidal debris, and foraminifers not older than Middle Ordovician in the marbles.

All the rocks that form the Moeche thrust sheet are intensely deformed. The serpentinites, talc-schists, and chlorite-schists have ultramylonitic and phyllonitic textures, indicating that deformation took place under greenschist facies conditions. This deformation obliterated previous structures and metamorphism. The structural features of the rocks suggest that they are a tectonic melange (in the sense of, e.g., Hsü, 1968; Raymond, 1984; Cowan, 1985; Rast and Wright Horton, 1989).

Espasante thrust sheet

This 900-m-thick sheet is composed of acid and basic igneous rocks that have been grouped in one lithological unit, named the Somozas orthogneisses and amphibolites (Marcos and Farias, 1997). The lower part of this formation, located close to the basal thrust, is extremely deformed and represented by ultramylonites and phyllonites named the Ramallal phyllonites by Marcos and Farias (1997). The Somozas orthogneisses and amphibolites are ~400 m thick. Quartzo-feldspathic gneisses with ±Ep ± Am predominate in the lower part of the formation, and epidote amphibolites are important in the upper part. This lithological unit includes also dikes of basic rocks (Arenas and Peinado, 1984; Arenas, 1988). According to Fernández Pompa and Monteserín López (1976) and Arenas (1988), the Somozas orthogneisses and amphibolites derive from high-K calc-alkaline and shoshonitic volcanic rocks that are related to a continental margin setting with incipient subduction.

The rocks in the Espasante thrust sheet show a continuous planar tectonic fabric without a clear stretching lineation. The textural features are those of mylonites (sensu lato) developed under a progressive deformation regime, allowing the separation of different deformation stages in the rocks seen as protomylonites, mylonites, ultramylonites, and phyllonites (Fig. 4, D–F). The last two types are geometrically related to the planes of the main basal thrust. Thus, the Ramallal phyllonites represent the last deformational stage in the reactivation of the mylonites and ultramylonites.

In the orthogneisses, Qtz, Pl, Ms, ± Ep-Czo, ± Hbl, ± Grt, ± Stp, ± Aln are stable during mylonitic deformation. In the amphibolites, the most common mineral association defining the main fabric contains Hbl, Pl, Qtz, Ep-Czo, Ms, ± Grt. In ultramylonites and phyllonites only Chl is synkinematic.

UPPER TECTONIC UNIT: THE SLAB OF LOWER CRUSTAL ROCKS

The Upper Tectonic unit is located at the uppermost part of the Cabo Ortegal Complex. It comprises ultramafic, mafic, and quartzo-feldspathic rocks that record high-*P*–high-*T* metamorphic events and polyorogenic deformation. It crops out in a young tectonic graben bounded by rocks of the Lower Tectonic unit and/or the Rio Baio thrust sheet (Plate 1). Within the Upper unit, the reconstruction of structures is partially masked by a system of high-angle normal faults striking N130E dipping to the north-northeast. These faults are several kilometers long, straight, and have vertical separations of hundreds of meters. Due to this late tectonic event, lithological boundaries exhibit frequently discontinuous and complicated cartographic patterns, changing from the footwall to the hanging wall of these normal faults, which were probably formed during the Alpine deformation events in northwestern Iberia (Ferrús, 1998).

The ductile structures of the Upper unit can be related to deformation under prograde granulite and eclogite facies conditions, although the more conspicuous structures result from retrogressive deformation associated with the exhumation processes. The fabric used as reference for the structural reconstruction is a widespread mylonitic fabric, which is observed throughout the Upper unit rocks. This mylonitic fabric is affected by isoclinal folds with axes striking north-northeast–south-

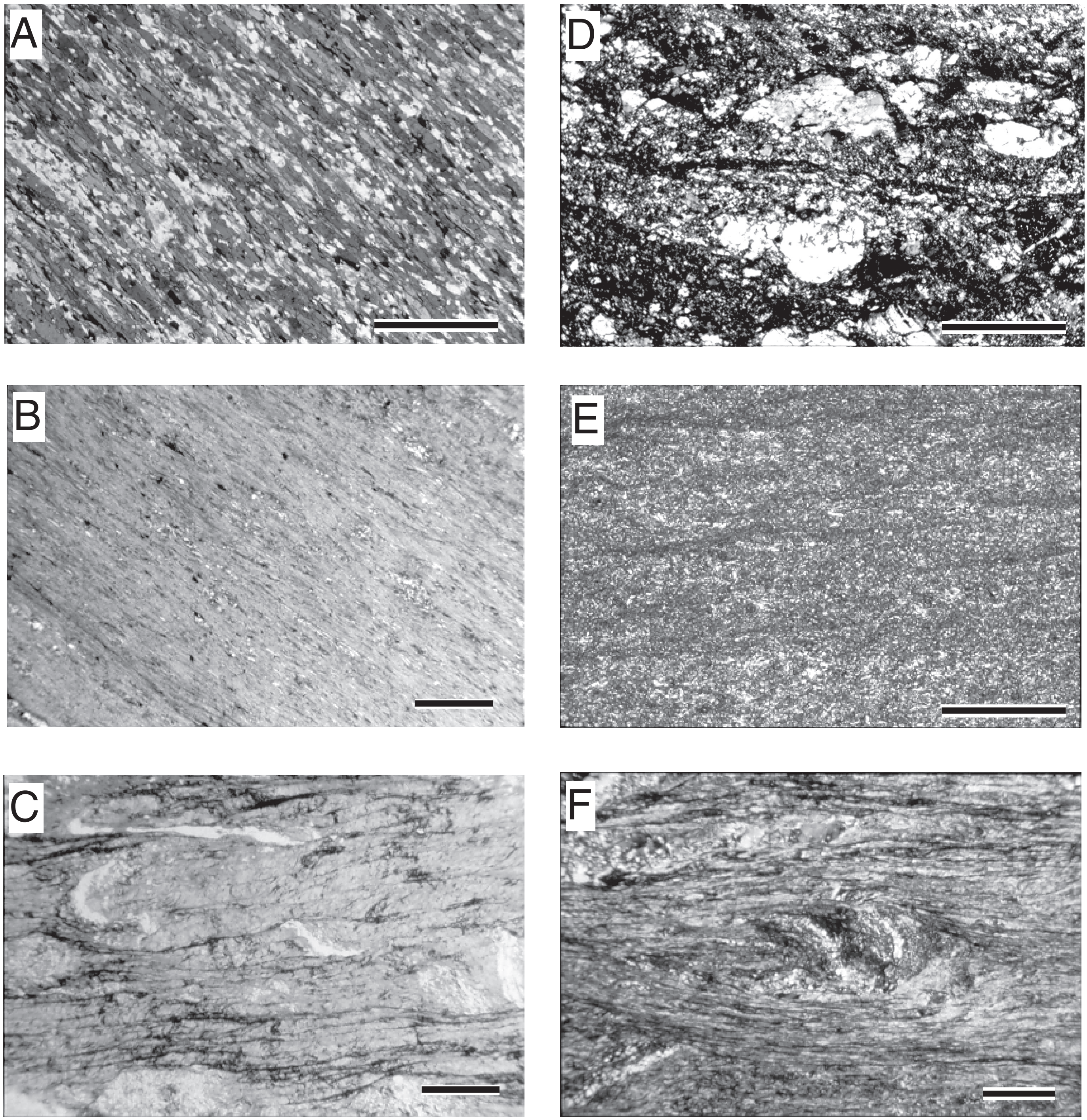


Figure 4. Microstructures in rocks from Lower Tectonic unit in Purrido thrust sheet. A: Tectonic fabric in Purrido amphibolites defined by shape-preferred orientation of Hbl and Pl. B and C: Ultramylonites and phyllonites from Labacengos Formation. Note very fine grain size and its uniform distribution in ultramylonites. Evolution of tectonic fabric in quartzo-feldspathic rocks in Espasante thrust sheet is observed from mylonites (D) to ultramylonites (E) and phyllonites (F). In all photomicrographs, scale bar is 1 mm (figure taken from Marcos and Farias, 1999).

southwest and verging to the east-southeast. The major folds are isoclinal and recumbent and have reverse limbs reaching 6 km; the folds are crosscut by thrusts. Two of them, the named Basal and Upper thrusts, reach several kilometers of displacement. The Upper thrust separates two tectonic sheets: the Cedeira sheet below, and the Capelada sheet above. Both sheets have the same tectonostratigraphic sequence, but are differentiated essentially by their relative degree of deformation and retrograde characteristics, which are stronger in the lower unit, where granulite and eclogite rocks are only found as relics.

The late Variscan refolding phase caused the characteristic elliptical outcrop pattern of the Cabo Ortegal Complex. Open folds with subvertical axial planes and fold axes oriented north-northeast–south-southwest form at all scales, giving Ramsay type 3 intersection patterns with the previous isoclinal folds at outcrop and cartographic scale.

Tectonostratigraphic sequence

Regardless of their detailed metamorphic and deformational history, the rocks that form the Upper Tectonic unit can be divided into three major lithological units: (I) the ultramafic rocks, (II) the mafic rocks, and (III) the quartzo-feldspathic gneisses (see legend in Plate 1).

The ultramafic rocks (I) represent the base of the sequence and are formed of more than 1000 m of alternating serpentinized peridotites (spinel-amphibole peridotites, harzburgites, and dunites) and pyroxenites (websterites and clinopyroxenites) (Vogel, 1967; Girardeau et al., 1989; Gravestock, 1992; Moreno, 1999). They build the core of a huge recumbent fold. The restoration of the structure allows the recognition of a sequence that is formed, from bottom to top, of ~ 300 m layered piroxenites, ~ 300 m dunites (alternating with centimeter to decameter Grt ± Spl pyroxenitic bands) and harzburgites. The origin of the ultramafic rocks is controversial. Girardeau and Gil Ibarguchi (1991) suggested that the peridotites have strong affinities with residual oceanic peridotites and the pyroxenites are intruding magma segregates. Moreno (1999) interpreted this ultramafic-layered complex as the root of an oceanic arc similar to the Himalayan Jijal Complex. The contact with the overlying mafic rock unit (II) is gradational (Galán and Marcos, 1997).

Three members can be distinguished within the mafic rocks (II). The lower member is formed by massive or weakly foliated ultramafic-mafic granulites (50–100 m thick), including the "hornblende pyrigarnites" (Grt, Cpx, ± Hbl, ± Pl < 5%, ± Qtz, Rt, ± Aln, ± Czo, ± Ap) and the "hornblende plagiopyrigarnites" (Grt, Cpx, ± Hbl, Pl, ± Qtz, Rt, ± Aln, ±Czo, ± Ap) of Vogel (1967). Some of these rocks lack Pl in equilibrium with Grt-Cpx and others show relict Ky and/or Zo, suggesting an earlier eclogite facies metamorphism (Galán and Marcos, 2000). Their composition is essentially ultrabasic-basic and equivalent to Fe-Ti rich (picro) basalts-(mela) gabbros (Galán and Marcos, 1997). In the middle member (400 m thick), these rocks alternate with garnet amphibolites (Grt, Hbl, Pl, Qtz, ± Bi, Ilm, Ttn, Czo, Ep, ± Aln) retrogressed from granulites, the composition of which is basic to intermediate. Both types of rocks define a centimetric to metric layering in the less deformed outcrops. Layers of garnet hornblende schist, garnet tonalitic gneisses, and scapolite-rich and carbonate-rich mafic granulites are also present in much subordinated proportion. These two members of mafic rocks are interpreted in two different ways. Peucat et al. (1990) favored either a continental volcanic arc margin or a continental rifting (attenuated crust), and Galán and Marcos (1997) favored the latter hypothesis. Galán and Marcos (1997) interpreted these two members as derived from a layered gabbro-type protolith related to continental within-plate magmatism in an extensional setting. In the latter model, both the ultramafic rocks (II) and the mafic rocks (II) would be part of a thinned lower crust. However, the westernmost outcrop band of this mafic middle member (Plate 1) has some peculiar lithological features: there are observed layered ± garnet amphibolites throughout, retrogressed after granulites (Vogel, 1967), and metagabbros and amphibolites after metagabbros, which were previously affected by granulite-amphibolite facies metamorphism in less severe conditions (Vogel, 1967; Gil Ibarguchi et al., 1990). The geochemistry of these rocks was correlated to N- and transitional (T)-MORB by Peucat et al. (1990).

A sharp contact separates the middle member from the upper member made of massive mafic rocks (100–200 m). The upper member is basically formed of eclogites, but fine-grained dark garnet amphibolites, retrogressed after eclogites-granulites (Vogel, 1967), are also present, especially in the northwestern outcrops of the Cabo Ortegal Complex (Plate 1). The most common type of eclogite contains Omp, Grt, Zo, Qtz, Amph, and Rt (Concepenido-type eclogites of Vogel, 1967), but kyanite-bearing eclogites (Vogel, 1967; Gil Ibarguchi et al., 1990; Mendia, 1996) and Fe-Ti–rich eclogites are also observed (Mendia, 1996). The Ky-bearing eclogites are mainly exposed as a thin band, <25 m thick, along the contacts. All the eclogites are more or less retrogressed in granulite-amphibolite facies conditions. Van Calsteren (1978) proposed a continental quartz-normative tholeiite parentage for the eclogites on the basis of major and trace element geochemistry. However, most authors agree on a N-MORB protolith for the common eclogites (Bernard-Griffiths et al., 1985; Peucat et al., 1990; Gravestock, 1992; Mendia, 1996).

The last unit (III) is composed of >600 m of quartzo-feldspathic gneisses in sharp contact with the eclogites. Near this contact the augen gneisses are fine or medium grained, composed of Qtz, Pl, Grt, Bt, Ms, and Ky (Vogel, 1967), and have a characteristic compositional banding resulting from changes in the amount of mafic minerals. They belong to the Banded Gneiss Formation of Vogel (1967) and include blocks or pods, lenses, and bands of eclogites, more or less retrogressed, and subordinate mafic and ultramafic rocks, giving the appearance of a block-in-matrix formation, in the sense of Raymond et al. (1989). Metamorphic mineral associations in both mafic rocks and host gneisses indicate a high-*P*–high-*T* metamorphism (Vogel, 1967; Gil Ibarguchi et al., 1990), causing partial melting of the gneisses. The thickness of these lower gneisses is ~150 m.

The rest of the unit (III) is formed of two-mica layered quartzo-feldspathic gneisses (often with Grt, St, and Ky), chemically and mineralogically similar to the Banded Gneiss Formation, but with different appearance in the field. The Cariño Gneisses and most of the Chímparra Gneisses of Vogel (1967) are included. They are interpreted as metamorphic graywackes (Vogel, 1967; Fernández and Marcos, 1997) with calc-silicate and mafic inclusions. The mafic inclusions have mineral assemblages formed in amphibolite facies in the Cariño Gneisses and in granulite facies in the Chimparra Gneisses, which were also affected by migmatization, as was the Banded Gneiss Formation (Vogel, 1967). Available geochemical data for some of the mafic inclusions indicate T-MORB compositions with important contributions of continental material (Peucat et al., 1990), which suggests that they intruded in an extensional continental setting.

Metamorphism and age

In general, the metamorphic evolution of both the ultramafic (I) and mafic (II) rocks is similar. The ultramafic rocks record a first episode of high-*P*–high-*T* metamorphism at ~800 °C, 1.6 Gpa, followed by a pervasive amphibolitization at >750 °C (Girardeau and Gil Ibarguchi, 1991). The metamorphic evolution of mafic rocks suggests a first high-*P*–high-*T* metamorphic event, followed by the development of different structures related to exhumation processes. They consecutively equilibrated in high-*P*–high-*T* eclogite-granulite, amphibolite, and greenschist facies, defining a decompression-type *P*-*T* path (Mendia, 1996; Galán and Marcos, 2000). The estimated metamorphic peak conditions, based on thermobarometric methods, range from 838–876 °C, ≥1.2–1.4 GPa (in the Bacariza-type eclogites within lower and middle mafic member; Galán and Marcos, 1999, 2000) to 770–800 °C, ≥1.4–1.7 GPa (common Concepenido-type eclogites within the upper mafic member; Gil Ibarguchi et al., 1990; Mendia, 1996).

Although the metamorphic evolution of the quartzo-feldspathic gneisses (III) is less studied, available thermobarometric data also indicate a high-*P*–high-*T* event for some of them. The metamorphic evolution of the quartzo-feldspathic gneisses in close contact with the upper mafic rocks is similar to them, as inferred from their eclogitic mafic inclusions. Some of these mafic inclusions, with plagioclase already coexisting with Grt-Cpx, record 880–620 °C, 1.35 ± 0.13 Gpa (Gil Ibarguchi et al., 1990), comparable to values estimated for the host gneisses (800–700 °C, 1.60–1.50 Gpa) (Basterra et al., 1989; Fernández Rodríguez, 1997). Other types of gneisses from the same unit (III) (e.g., the Cariño Gneisses) only present evidence of metamorphism under amphibolite facies conditions, subsequently retrogressed in greenschist facies (Vogel, 1967).

The absolute age of the protoliths and of the high-*P*–high-*T* metamorphism are a matter of debate (see reviews in Ábalos et al., 1996; Martínez Catalán et al., 1999). Peucat et al. (1990) consider an Early Ordovician age, ca. 490–480 Ma, for both the protholiths of the lower and middle mafic member and for the high-*P*–high-*T* event in eclogite and granulite facies. Ordoñez Casado (1998) agreed on this age for the protoliths and obtained a comparable age (507–473 Ma) for the protoliths of the Concepenido eclogites (i.e., for the upper mafic member). However, Ordoñez Casado and others (Schäffter et al., 1993; Santos Zalduegui et al., 1996) estimated the age of the high-*P*–high-*T* event in eclogite and granulite facies as Early Devonian (ca. 405–380 Ma). Schäffter et al. (1993) and Ordoñez Casado (1998) also provided similar ages for the high-*P*–high-*T* episode in the quartzo-feldspathic gneisses (397–391 Ma). However, more recent data, from U-Pb dating on zircon, monazite, titanite, and rutile (Fernández Suárez et al., 2002) indicate that the rocks in the Upper Tectonic unit record a polyorogenic evolution with a high-*P*–high-*T* episode ca. 490–480 Ma (Early Ordovician) and a younger Early Devonian event dated ca. 390–385 Ma. The age of the late deformation in greenschists facies conditions would occur ca. 365 Ma (Late Devonian) (Dallmeyer et al., 1997).

Structures associated with high*-P*–high-*T *metamorphism

Most of the structures found in the high-*P*–high-*T* rocks of the Upper Tectonic unit are developed during exhumation under retrogressive metamorphic conditions. However, preferred orientation of omphacite in eclogites and granulites (Engels, 1972; Godard and van Roermund, 1995; Ábalos, 1997) and the development of ductile shear zones (Marcos and Galán, 1994) under high-*P* conditions provide evidence of intracrystalline plasticity of clinopyroxene caused by deformation during the earlier high-*P*–high-*T* metamorphism. The crystallographic preferred orientation of omphacite from the Concepenido-type eclogites (Upper Member of the mafic rocks unit) show the distribution of c[001] axes in girdles and b[100] in maximum, indicating the development of planar fabrics (S-type according to Helmsteadt et al., 1972), which suggests that coaxial flow is predominant during high-*P* deformation (Engels, 1972; Godard and van Roermund, 1995; Ábalos, 1997). Ábalos (1997), however, reported local noncoaxial components during the deformation of omphacite, relating the crystallographic preferred orientation fabrics with the structural reference framework in the rock, i.e., the foliation and the lineation. However, at the regional scale the lineation in omphacite scatters in the foliation plane (Mendia, 1996) and does not provide a consistent orientation for noncoaxial deformation, suggesting that deformation is predominantly coaxial (Llana-Fúnez et al., 2001). This is not incompatible with local noncoaxial components at thin-section scale, because deformation is probably more heterogeneous and might be influenced by the presence of rigid garnets. The development of these microstructures could be related to the crustal thickening produced by pre-Variscan continental collision.

Deformation during exhumation and emplacement of the Upper Tectonic unit

The first structure that appears is a planar generalized tectonic fabric (RD1), which at the map scale is parallel to the lithological boundaries. It is a mylonitic foliation, sometimes showing a gneissic banding, which started to form in granulite facies conditions and finally equilibrated in amphibolite facies conditions (Fernández, 1993; Marcos and Galán, 1994; Fernández and Marcos, 1996) (Fig. 5). At the outcrop scale, the foliation shows an anastomosing aspect nucleated around pods in which the rocks preserved a previous migmatic fabric, or a clear and completely formed planar fabric. In places a disrupted or truncated foliation is observed, which suggests the development of more than one consecutive set of mylonitic fabric during the same progressive deformation event (Fig. 6A). A heterogeneous component of noncoaxial strain is indicated by some nonpervasive, asymmetrical elements, such as folds and sigma and delta microstructures (Fig. 6B). A detailed analysis of the macrostructures and microstructures associated with the mylonitic foliation in quartzo-feldspathic gneisses leads us to propose a heterogeneous coaxial flow regime during the extensive ductile deformation. This model is supported by the lack of a well-developed stretching lineation, the orthorhombic symmetry of quartz *c*-axis textures, and the absence of a consistent sense of shear inferred from kinematic criteria (Fernández, 1993; Fernández and Marcos, 1996) (Fig. 7). Persistent foliations not related to any particular type of tectonic structure (e.g., a detachment) are common in granulites. Mechanisms restricted to the lower crust, such as regional coaxial flow, repeated transposition, and advective or convective flow, have been proposed by several authors for the development of these foliations (see Dirks et al., 1997).

Locally, in mafic rocks of La Capelada sheet, the main mylonitic fabric (RD1) is affected by centimeter- to decameter-size discrete shear bands where blastomylonitic biotite gneisses (Galán and Marcos, 1997) are formed (Fig. 8). These tectonites, mapped by Vogel (1967) as "doubtful paragneisses," locally enclose centimeter blocks of ultramafic-mafic granulites and garnet amphibolites in a chaotic arrangement. These blastomylonitic gneisses contain Grt, Am, Pl, Scp, and Czo porphyroclasts set in a very fine grained mylonitic matrix made of Bi, Pl, Qtz, and Ttn.

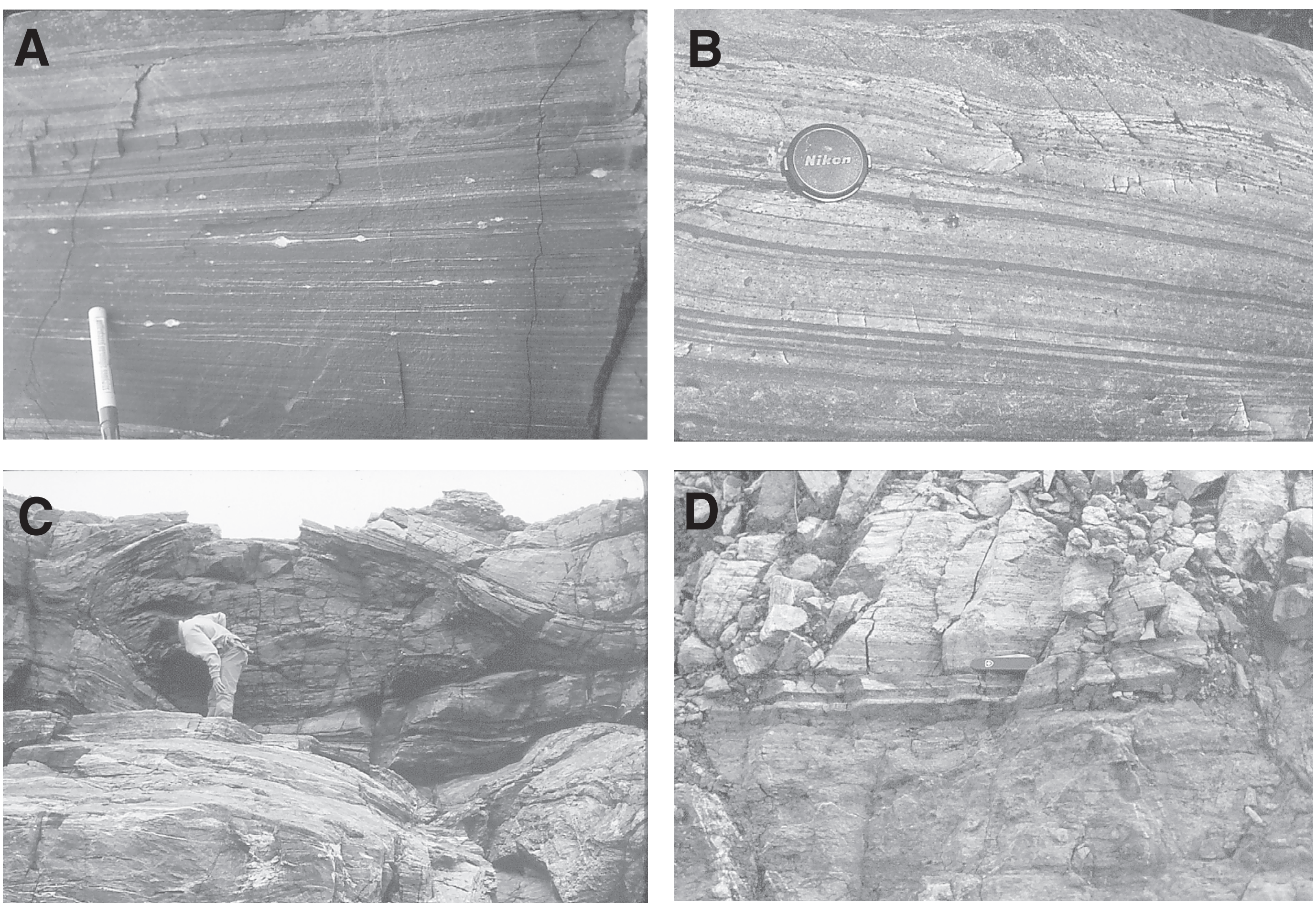


Figure 5. Field aspects of mylonitic foliation RD1. Planar fabric is shown in quartzo-feldspathic gneisses (Ría de Cedeira, immediately east of Punta Chirlateira) (A) and in mafic rocks (B) from Upper Member (garnet amphibolites, Tarroiba boulder beach). White part of the marker in A is 10 cm and diameter of lens cap in B is about 5 cm. C: In mafic granulites of Middle Member, main foliation surrounds pods of pyriagarnites. West of San Andrés de Teixido. D: Contact zone between foliated mafic (below in picture) and ultramafic rocks in reverse limb of recumbent fold of La Capelada is shown. Knife is 6 cm in length. West of Miranda hill.

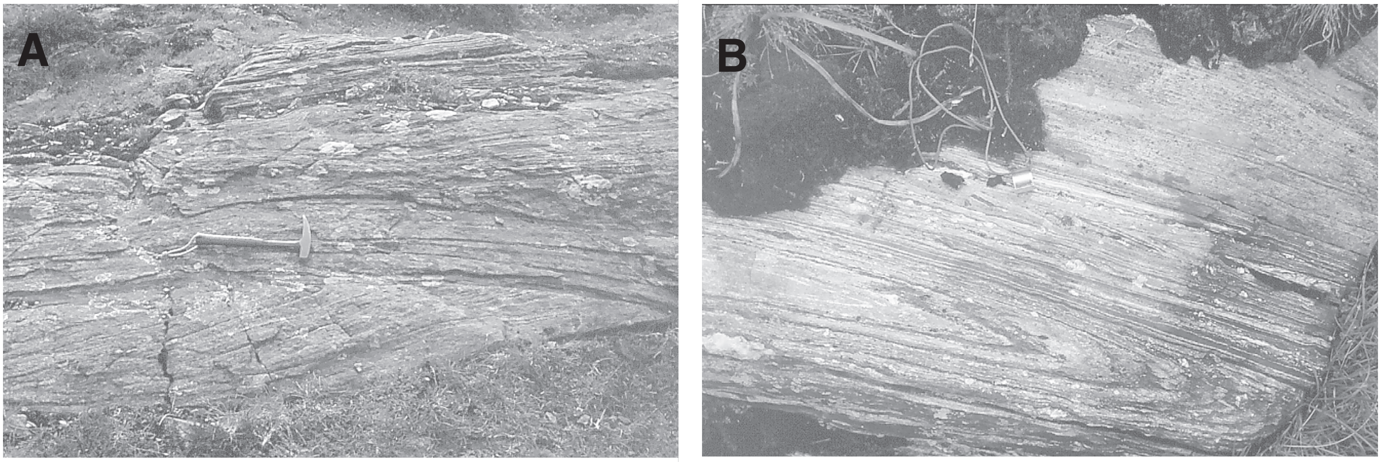


Figure 6. Truncated foliations (A) and intrafolial folds (B) in relation to main mylonitic fabric RD1 (mafic granulites). In A, hammer is about 32 cm in length. In B, magnifier is 2 cm wide. Outeiro hill.

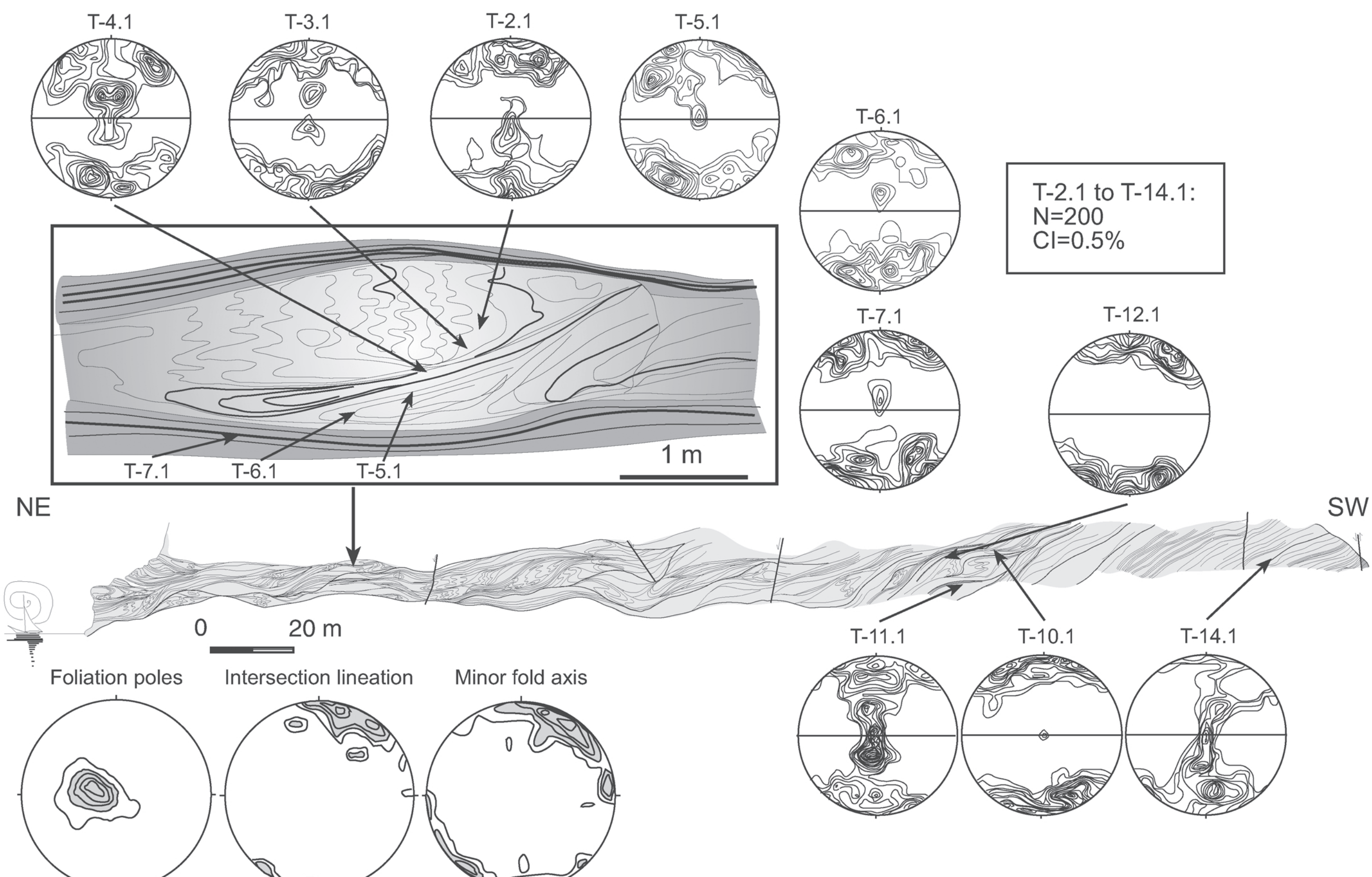


Figure 7. Structural transition in quartzo-feldspathic gneisses from homogeneous planar mylonitic fabrics to inhomogeneous anatomizing pattern is observed in Tarroiba cliff. Intersection between foliation and compositional banding in gneisses (Chímparra Gneisses) is predominant lineation observed in field, having same orientation as fold axes of minor folds. Quartz *c*-axis fabrics in these rocks have orthorhombic symmetry, indicating that heterogeneous coaxial flow is predominant during quartz deformation. Only sample T-14.1, collected from southwest sector, has asymmetric girdle and deviates from this common pattern. Half-opening angles of small-circle girdles vary between 20° and 40°. All plots are Schmidt lower hemisphere projection (CI is contour interval).

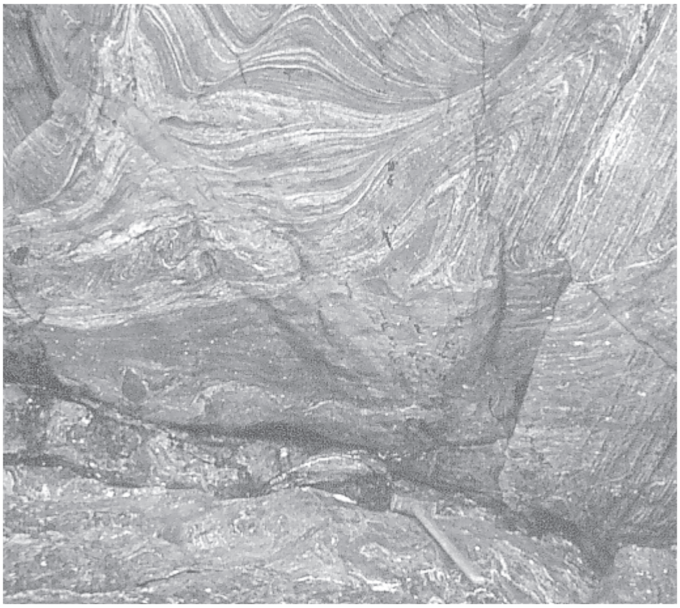


Figure 8. Biotite blastomylonitic gneisses in meter-size discrete shear zone (lower part of picture). Hammer handle is 60 cm in length. West of San Andrés de Teixido.

The general foliation is affected by asymmetric similar folds (RD2 folds), trending to isoclinal folds, with low interlimb angles. Their axes are oriented north-northeast–south-southwest and their vergence is to the east-southeast. The major folds are isoclinal and recumbent and their reverse limbs reach kilometric scale (>6 km) (Plate 1; Fig. 9). Related minor folds are of decametric size and generally also asymmetric (Fig. 10). Refolding of previous intrafoliar folds by these RD2 folds resulted in type 3 interference patterns of Ramsay (1967) and sometimes type 2 patterns. In the field, no axial plane cleavage is observed, although a slight orientation of hornblende (in amphibolites) or phyllosilicates (in gneisses) can be seen in some RD2 fold hinges. The lack of foliation associated with RD2 folding makes it difficult to establish the exact *P-T* conditions at which deformation occurred. However, both the similar geometry of folds and their strong flattening imply a generally ductile behavior of the rocks, which suggests deformation under amphibolite facies conditions. In consequence, there would be a relevant reworking of the RD1 foliation in the limbs of the RD2 folds. The inferred reworking of the main previous foliation during this folding stage is confirmed by the development of a stretching lineation parallel to the RD2 fold axes. The colinearity of lineation and fold axes points to folding having occurred in a constrictional environment.

The major RD2 folds are crosscut by thrusts (RD3), two of them reaching kilometer size (Plate 1; Fig. 9). One of them, the Basal thrust, places the Upper Tectonic unit over the Lower Tectonic unit. It is parallel to the reverse limb of a major recumbent fold and gives rise to the thrusting of high-grade gneisses over the Purrido amphibolites. In its westernmost section, it is highlighted by strongly sheared and serpentinized ultramafic rocks, probably belonging to the Moeche thrust sheet (Marcos and Farias, 1999). The related shear zone, with associated minor structures such as thrusts, folds, and crenulation cleavage, is in some locations >50 m thick and is developed mainly in the footwall rocks. The minor folds are asymmetric, reach meter scale, and have curved hinges that indicate the direction of the tectonic transport, approximately east-southeast. The Upper thrust imbricates the rocks of the Upper Tectonic unit and produces the stacking of the La Capelada tectonic sheet over the Cedeira sheet. The geometrical relationships between the Basal and Upper thrusts and the existence of klippen of the La Capelada sheet over the underlying Lower Tectonic unit (Plate 1) point to a relative younger age for the Upper thrust, which would be consistent with an inward sequence of emplacement of these thrust sheets (Marcos and Farias, 1999).

The following RD4 deformation episode is related to the general postnappe refolding that affected both the allochthonous and autochthonous rocks at all scales and gave the Cabo Ortegal Complex its typical elliptical cartographic shape. The RD4 folds are open with subvertical axial planes oriented west-northwest–east-southeast. In places they show axial crenulation cleavage and a mineral lineation parallel to fold axes defined by chlorite. The distribution of RD4 structures is heterogeneous. The interference between RD2 and RD4 folds produces the type 3 patterns of Ramsay (1967), both at cartographic (Fig. 11) and outcrop scale. Contemporaneously or subsequent to the folding, some discrete north-south–oriented shear zones, equilibrated in greenschist facies, appear and usually reactivate previous structures. These shear zones are subvertical, with left-lateral shear sense. They are probably related to other major structures in northwest Iberia characterizing intracontinental postcollisional deformation, such as the Valdoviño fault (Iglesias Ponce de León and Choukroune, 1980) (see Fig. 1).

Kinematic interpretation of the main tectonic fabric. Most stretching and/or intersection lineations associated with the main tectonic RD1 foliation in the Upper Tectonic unit rocks show a constant south-southwest–north-northeast orientation. Kinematic criteria, such as delta and sigma porphyroblasts systems and intrafoliar folds, do not indicate a consistent sense of shear at outcrop scale. This seems to suggest that the mylonitic foliation is developed in a general noncoaxial flow (Fernández Rodríguez, 1997; Fernández and Marcos, 1996).

The general kinematic interpretation is difficult because of the superimposition of RD1 and RD2 structures. Thus, it is not easy to evaluate the effects on the main foliation caused by the development of (RD2) isoclinal folds in ductile conditions. In addition, the best developed lineation in the rocks is defined by the hornblende and plagioclase, which could grow during RD1 or RD2. Interpretation of kinematic criteria must be made with caution. For example, one assumption might be that RD1 lineation is intimately related to the noncoaxial component of deformation and indicates the direction of tectonic transport (Ábalos et al., 1996). However, it must be kept in mind that deformation of RD1 lineation by homoaxial isoclinal folds results in opposite sense of shear in normal and reverse limbs.

KINEMATICS OF THE SUTURE-ZONE THRUST SHEETS

RD2 and RD3 structures can be considered the result of progressive deformation that led to the emplacement of the Upper unit. Both the vergence of the RD2 folds and the sense of shear in RD3 thrusts indicate a general tectonic transport of the nappe toward the east-southeast (Marcos and Farias, 1999). This is consistent with the geometry of the structures developed in the underlying sheets (Lower Tectonic unit, Rio Baio thrust sheet) and in the Paleozoic autochthonous sequence (Fig. 11).

The lower tectonic sheets of the Cabo Ortegal Complex that form the suture zone are, from top to bottom, the Purrido thrust sheet, the Moeche sheet, and the Espasante thrust sheet (Plate 1; Fig. 3). The internal structure of the Moeche thrust sheet is that of a tectonic melange. A mylonitic foliation is developed parallel to the boundaries of the Purrido and Espasante sheets, and no folds have been observed in relation to the fabric. In both sheets, ultramylonites and phyllonites occur toward the bottom as the result of the extreme deformation associated with their emplacement (Fig. 4). The main textural change in the rocks is the grain-size reduction. The transition between mylonites and ultramylonites is gradational and corresponds to a band of ductile shear. This observed reworking of the mylonites and ultramylonites near the basal thrusts (i.e., basal induced shear) implies the development of a new foliation simultaneously with the general rock retrogression. The new fabric is either a slaty cleavage or a closely spaced crenulation cleavage, with synkinematic Chl and opaque minerals in the cleavage plane. Previous mylonites and ultramylonites are preserved in some lenticular bodies or pods surrounded by the new foliation, suggesting a relatively late formation of the basal thrust. Some of the structural features already mentioned, e.g., the development of a tectonic fabric parallel to the sheet boundaries, the changes of textures approaching the lower parts, and the lack of folds associated with the main fabric, provide evidence that these thrusts represent major tectonic boundaries and support the existence of a basal resistance to forward flow (Rathbone et al., 1983; Ramsay and Lisle, 2000).

Available data on the kinematics of the different thrust sheets are scarce and most of them are from the Purrido thrust sheet. In this sheet, a well-developed amphibole lineation is consistently orientated at outcrop scale, but its orientation differs from place to place, from N30E in the western coast to N40E–N60E in the southernmost inland exposures. However, there are several kinematic indicators, such as rotated porphyroblasts and C′ shear bands, that point to displacement of the hanging wall toward the northeast. In the Labacengos ultramylonites and phyllonites, also within the Purrido sheet, both the mineral lineation and the curved hinges of meter-scale folds indicate tectonic transport to N70E. The last structures (phyllonites, folds with curved hinges) formed in relation to the thrust movement under greenschist facies conditions, indicating tectonic transport to the southeast.

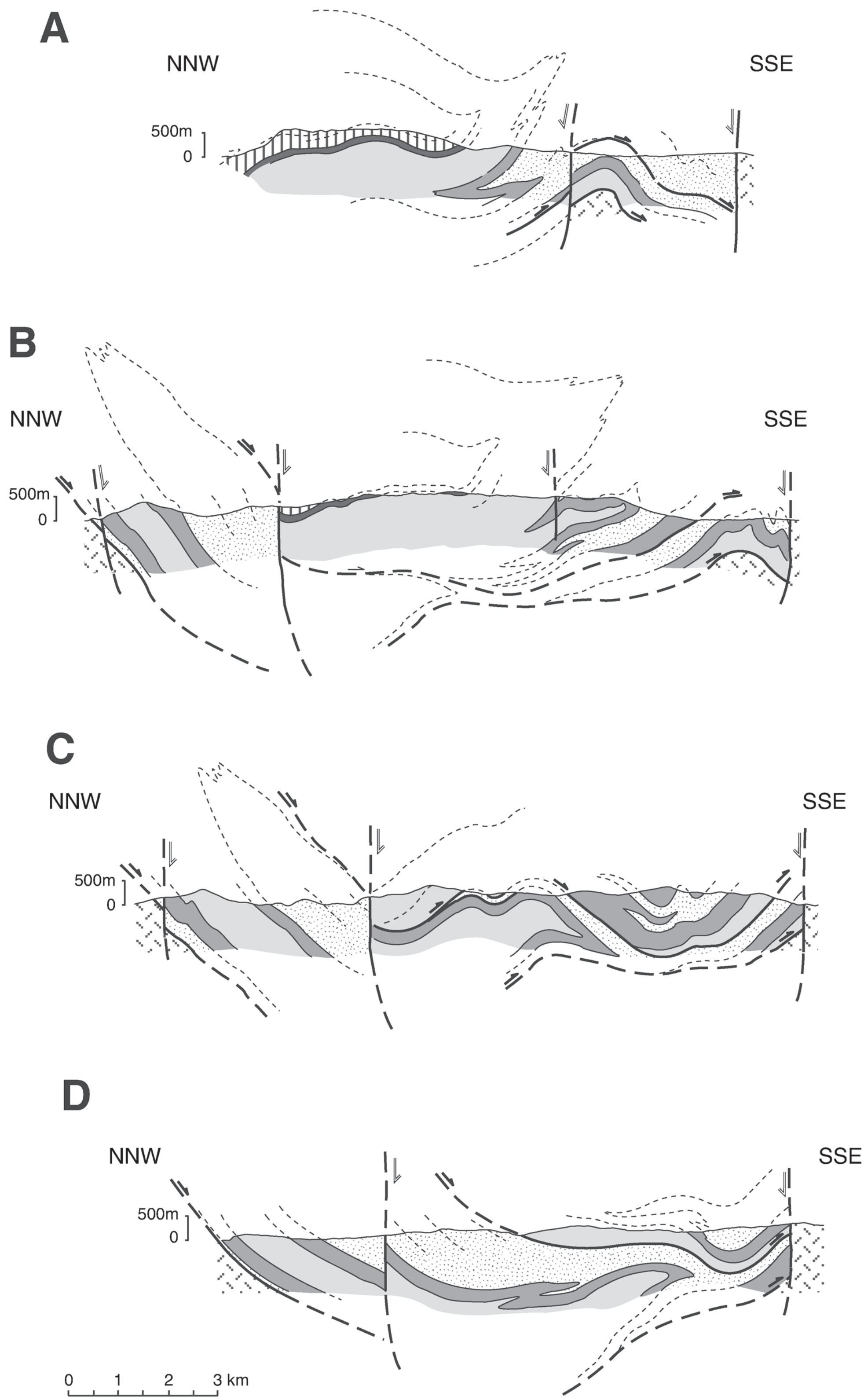
In summary, during the ascent and emplacement of the thrust sheets under retrograde metamorphic conditions from amphibolite to greenschist facies, a progressive clockwise rotation of transport direction is recorded in the tectonites (Marcos and Farias, 1999), in a similar way to the changes or rotations of overthrust shear that were reported in the Helvetic nappes of the Alps by Dietrich and Durney (1986).

CONCLUSIONS

The Cabo Ortegal Complex is a composite allochthonous terrane that was thrusted onto the western edge of Gondwana during the Variscan orogeny. It is formed of two main tectonic units: the Upper Tectonic unit and the Lower Tectonic unit, each of which includes several sheets. Another thin thrust sheet (the Rio Baio sheet or parautochthonous sheet) separates the Lower Tectonic unit from the underlying metasediments of the lower Paleozoic sequence in the Ollo de Sapo antiform, which is considered the autochthonous sequence of the Iberian plate.

The Upper Tectonic unit includes ultramafic rocks, mafic rocks classified into three members, and quartzo-feldspathic gneisses, all of them affected by high-P–high-T metamorphism in eclogite and granulite facies. Most structures observed in both mafic rocks and quartzo-feldspathic gneisses are related to their exhumation and concomitant retrogression. An exception would be the preferred orientation of omphacite in eclogites and granulites, retrogressed after eclogites. This orientation indicates predominately coaxial flow during the earlier stages of the high-P–high-T event, probably caused by crustal thickening during pre-Variscan continental collision. During the exhumation, these rocks developed a persistent horizontal foliation (RD1) in a constrictional environment, that started to form in granulite facies and was finally equilibrated in amphibolite facies conditions (Table 1). Later, also in amphibolite facies conditions, the rocks recorded the progressive development of noncoaxial structures, such as east-verging asymmetrical isoclinal folds and thrusts. These thrusts led to the emplacement of the Upper Tectonic unit over the Lower Tectonic unit and of both of them over the parautochthon and autochthon (Fig. 11).

The Lower Tectonic unit is believed to represent the resulting suture of the Variscan collision and the closure of the Rheic ocean. Available geochemical data indicate that the rocks in each sheet are derived from different geodynamic settings: E-MORB for the Purrido rocks, tectonic melange for the Moeche sheet, and high-K calc-alkaline volcanics for the Espasante sheet. All of them were equilibrated in amphibolite or greenschist facies conditions. Deformation is concentrated at the bottom of each sheet, indicating that their boundaries constitute major tectonic limits. Kinematic criteria provide a sense of tectonic transport for these thrusts toward the east, but show a clockwise rotation from northeast to southeast in progressively shallower



A
NNW
SSE
500m
0
B
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0
C
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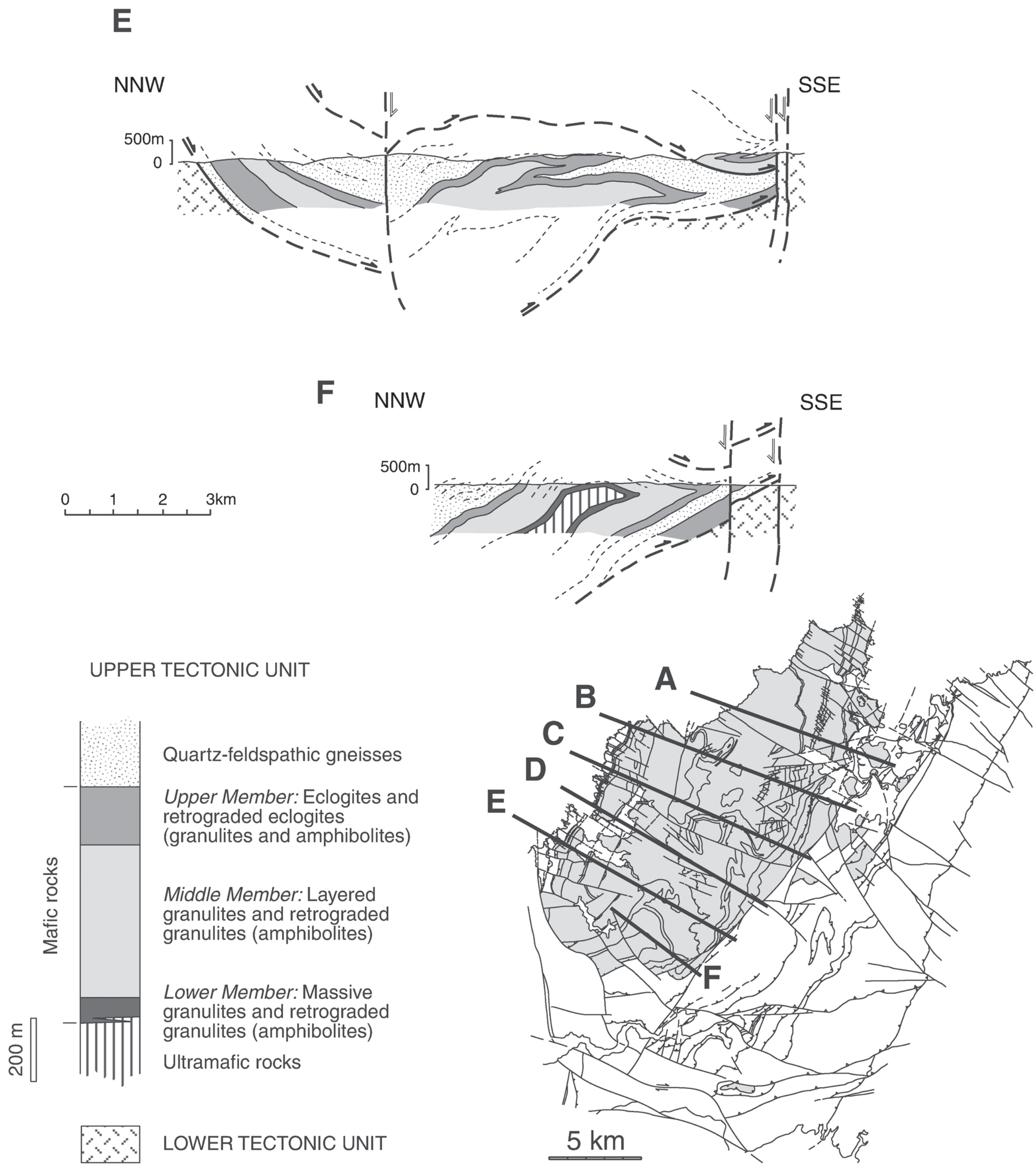

Figure 9 (on this and facing page). Geological sections across Upper unit (Marcos et al., 2000).

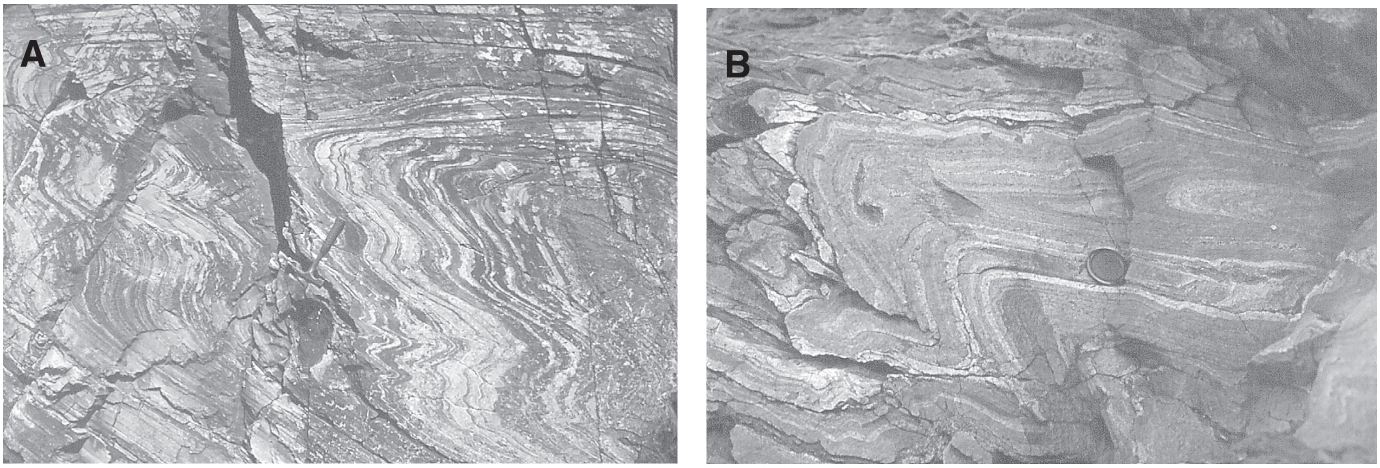

Figure 10. A: RD2 folds in quartzo-feldspathic gneisses affect previous mylonitic fabric. Hammer is 28 cm in length. Punta do Carreiro. B: Interference pattern between intrafolial folds, related to development of main tectonic fabric, and RD2 folds is shown. Diameter of the lens cap as a scale is about 5 cm. Masanteo Peninsula.

deformation conditions. Kinematic criteria in the Rio Baio sheet also indicate an earlier tectonic transport (top to the northeast), changing later toward the southeast.

An attempt to integrate the various tectonothermal events observed in the Cabo Ortegal Complex and underlying units is presented in Table 1. However, this type of modeling is poorly defined, among other reasons, by the fact that available radiometric ages show a high degree of scatter and contradictions. This must be regarded as normal due to the limitations of the dating methods and to the complex mineral reactions during the different tectonothermal events. Thus, the age of the high-*P*–high-*T* metamorphic event remains as a major question (Table 1). One set of data points to Early Ordovician, while the other set of data point to Early Devonian. Both ages are U-Pb dates, the former

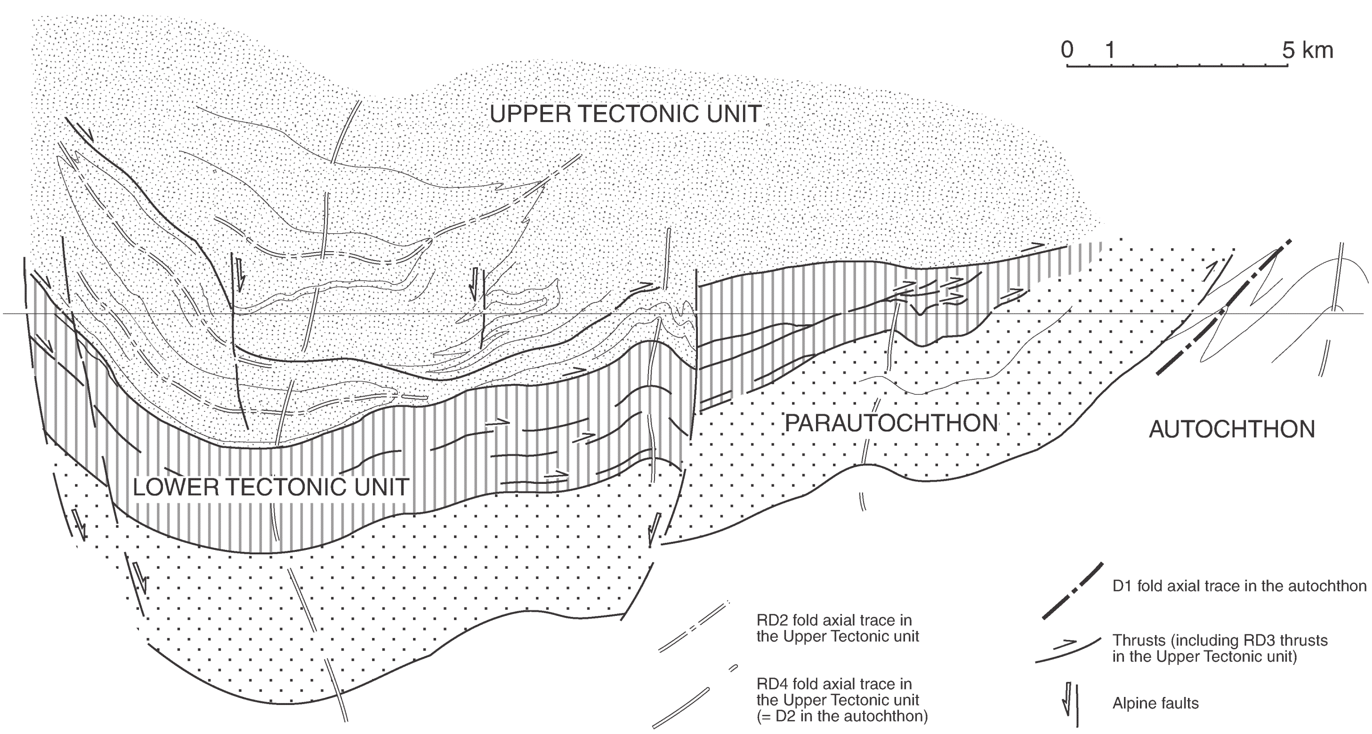

Figure 11. Composite section (approximately west-east) showing reconstruction of suture zone in Cabo Ortegal Complex.

TABLE 1. GENERALIZED TECTONIC HISTORY OF THE CABO ORTEGAL COMPLEX

Ages (in Ma)	Events
490–480 (Early Ordovician) or 390 (Early Devonian) ?	Subduction or underthrusting of a thinned mafic crust and metasediments in a convergent setting within the Rheic Ocean, clearly before or during the Eo- Variscan collision. H*P*-H*T* metamorphism in eclogite and granulite facies, reaching temperatures of 770–876 °C and minimum pressure of 1.4 Gpa. Partial melting of some of the mafic rocks and quartzo-feldspathic gneisses. Orientantion of omphacite could be related to this episode. It indicates coaxial flow, which is likely related to crustal thickening.
<480 ?	Partial exhumation and reequilibration, in granulite facies, of lower crustal rocks up to the base of a normal continental crust (35–40 km; ca. 1.2 GPa). Development of a conspicuous horizontal foliation (RD1) in a constrictional environment. This RD1 foliation was initially formed in HP granulite facies, but later reequilibrated in amphibolite facies conditions (720–600 °C, P: 1.25–0.50 Gpa).
390–385 (Early-Middle Devonian)	Laurentia-Gondwana Collision (Variscan orogeny). Formation of non-coaxial structures in the Upper Tectonic Unit rocks (RD2 recumbent folds) in a constrictional environment. Reworking of previous RD1 foliation in the fold limbs. Strain partitioning: RD3 thrust onset from about 15 km (ca. 0.5 GPa).
ca. 365 Ma (Late Devonian–Early Carboniferous)	Formation of ultramylonites and filonites related to RD3 thrust in the rocks of the Lower Tectonic Unit. First Variscan deformation event in the autochthon (Ollo de Sapo Antiform).
ca. 315 Ma (Late Carboniferous)	General refolding yielding folds with vertical axial planes (RD4) and subvertical shear zones with predominant strike-slip displacement.
Late Oligocene–Early Miocene	Alpine faulting.

on multigrain or single-grain fractions and the latter on single crystals, using sensitive high-resolution ion microprobe.

This contradiction is a crucial point because an Early Ordovician high-*P*–high-*T* event implies the exhumation of these rocks in two orogenic cycles, while a single orogenesis would account for the second age. Structures clearly related to such an important tectonic event during the Early Ordovician have never been reported in the neighboring areas. Ellis (1987) discussed that high-*P* granulites, formed as a result of doubly thickened crust and therefore similar to rocks from the Upper Tectonic unit, could have been exhumed in two orogenic events. The pressure-temperature-time (*P-T-t*) path of such granulites would present an important isobaric cooling (~400 °C), after the metamorphic peak, at pressures equivalent to a normal lower crust (~1.2 Gpa). This does not seem to fit the available *P-T-t* paths for the eclogites and granulites of the Upper Tectonic unit. In both rocks, the *P-T-t* paths show both decreasing *P* and *T* or clear isothermal decompression. However, if these rocks were uplifted to the base of a normal crust and stayed there for a period of ~100 m.y., they would have been reequilibrated to conditions of intermediate-*P* granulites. Such mineral associations have never been reported in the rocks of the Upper Tectonic unit. Eclogites were retrograded either to eclogites or to high-*P* granulites and later to amphibolite facies.

The likely age of the onset of retrogression in granulite and amphibolite facies conditions is 390–385 Ma. This seems to be supported by $^{40}Ar/^{39}Ar$ data on amphibole, which yield a similar age (385 Ma) (Peucat et al., 1990; Dallmeyer et al., 1997). However, temperatures for the first stages of the retrograde path are only slightly lower with respect to the peak metamorphic conditions, and if the closure temperature of zircons for the U-Pb system is within this narrow range or slightly lower, it would be difficult to distinguish the peak age from the age of the earlier retrograde stage. Therefore, 390–385 Ma would be a minimum age for the earlier high-*P*–high-*T* metamorphic event. In summary, a further detailed study is needed in order to understand the meaning of all the radiometric ages in relation to other geological data from the Cabo Ortegal Complex.

From Early to Middle Devonian time onward, exhumation of the Upper Tectonic unit is correlated to the Variscan orogeny (sensu stricto) and continental collision. This led to final emplacement of the Upper Tectonic unit in upper crustal levels over the Lower Tectonic unit. The rocks recorded the progressive development of noncoaxial structures, east-verging asymmetrical isoclinal folds, and thrusts. These thrusts caused the final emplacement of the Cabo Ortegal Complex over the parautochthon and autochthon ca. 365 Ma.

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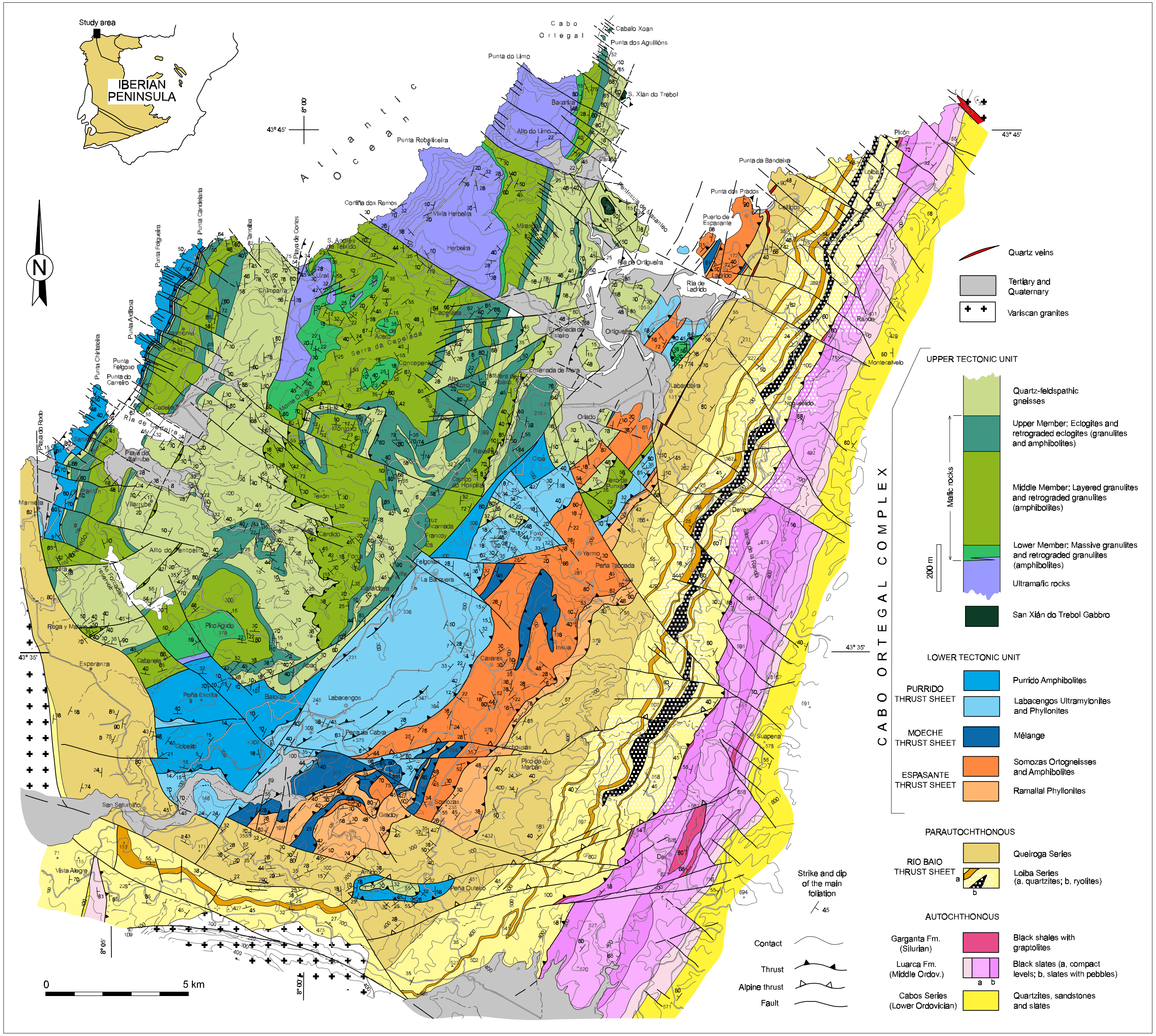

CABO ORTEGAL COMPLEX
Study area
IBERIAN PENINSULA
Atlantic Ocean
Cabo Ortegal
Quartz veins
Tertiary and Quaternary
Variscan granites
UPPER TECTONIC UNIT
Quartz-feldspathic gneisses
Upper Member: Eclogites and retrograded eclogites (granulites and amphibolites)
Middle Member: Layered granulites and retrograded granulites (amphibolites)
Lower Member: Massive granulites and retrograded granulites (amphibolites)
Ultramafic rocks
San Xián do Trebol Gabbro
Mafic rocks
200 m
LOWER TECTONIC UNIT
PURRIDO THRUST SHEET
Purrido Amphibolites
Labacengos Ultramylonites and Phyllonites
MOECHE THRUST SHEET
Mélange
ESPASANTE THRUST SHEET
Somozas Ortogneisses and Amphibolites
Ramallal Phyllonites
PARAUTOCHTHONOUS
RIO BAIO THRUST SHEET
Queiroga Series
Loiba Series (a, quartzites; b, ryolites)
AUTOCHTHONOUS
Garganta Fm. (Silurian)
Black shales with graptolites
Luarca Fm. (Middle Ordov.)
Black slates (a, compact levels; b, slates with pebbles)
Cabos Series (Lower Ordovician)
Quartzites, sandstones and slates
Strike and dip of the main foliation
Contact
Thrust
Alpine thrust
Fault
0
5 km
43° 45'
43° 35'
8° 00'
8° 05'
Cabalo Xoan
Punta dos Aguillóns
S. Xián do Trebol
Punta do Limo
Punta Robaliceira
Vixía Herbeira
Herbeira
Cortiña dos Remos
Punta da Bandeira
Punta dos Prados
Puerto de Espasante
Ría de Ortigueira
Ría de Ladrido
Ortigueira
Serra da Capelada
Labacengos
Peña Tallada
Insua
Peña da Cabra
Peña Outeiro
Vista Alegre
San Saturniño
Esperanza
Punta Candelaria
Punta Fulgueira
Punta Ardilosa
Punta Chirlateira
Punta Felgoso
Punta do Carreiro
Praia do Rodo
Ría de Cedeira
Cedeira
Alto do Pantorillo
Pico Agudo
Piçón
Loiba
Montecalvelo
Devesos
Suapena