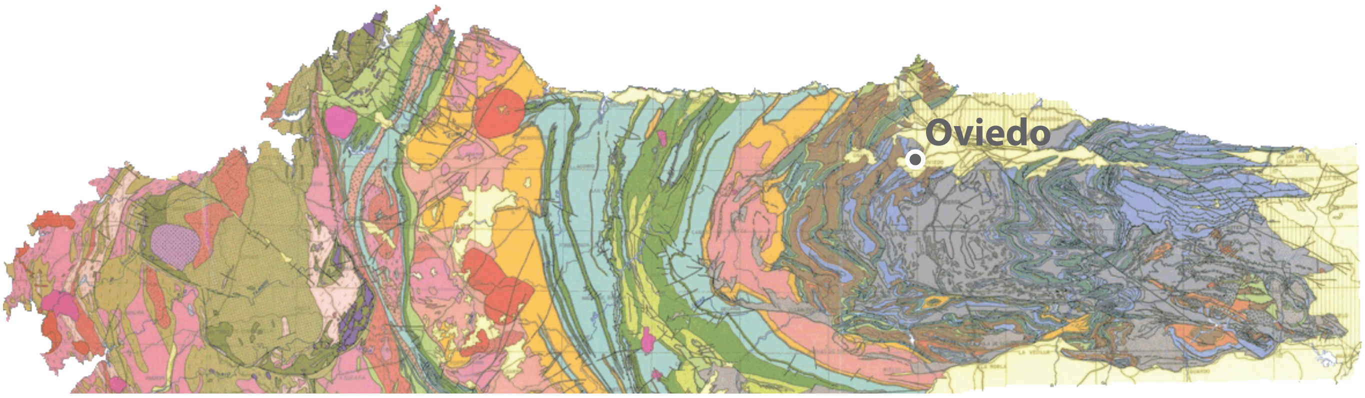

Deformation mechanisms, Rheology and Tectonics

DRT Meeting Oviedo 2011

POST-CONFERENCE FIELDTRIP GUIDE

3-5 September 2011

West Asturian Leonese Zone

Cabo Ortegal

Malpica-Lamego Line

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Cover: **Panoramic view of the contact between the peridotites (above) and the granulites (below) in the Cabo Ortegal Nappe**
Photo by *Sergio Llana-Fúnez*

DRT2011 POST-CONFERENCE FIELDTRIP GUIDE

Contents

Introduction: the Variscan Orogen in NW Iberia

The Variscan-Appalachian Orogen formed during the Devonian and Carboniferous as a result of the collision of Gondwana with Laurentia and other microplates (e.g. Martínez-Catalán et al., 1999). The Variscan collision led to the underplating of Iberia and France below microplates located to the West. At the present level of erosion, rocks belonging to the suture and the upper plate are preserved in klippen in the "allochthonous complexes" (Fig. 1). The cartographic pattern of these allochthonous thrust sheets gives an estimate of the minimum extent of plate overlap. These sheets are a convenient strain marker that indicates the shortening taken up by the upper crust in the hinterland. The Iberian Massif contains an almost complete section across the orogen; both the pre-and post-conference fieldtrips will show different aspects of the deformation associated to distinct parts of the orogen during its development.

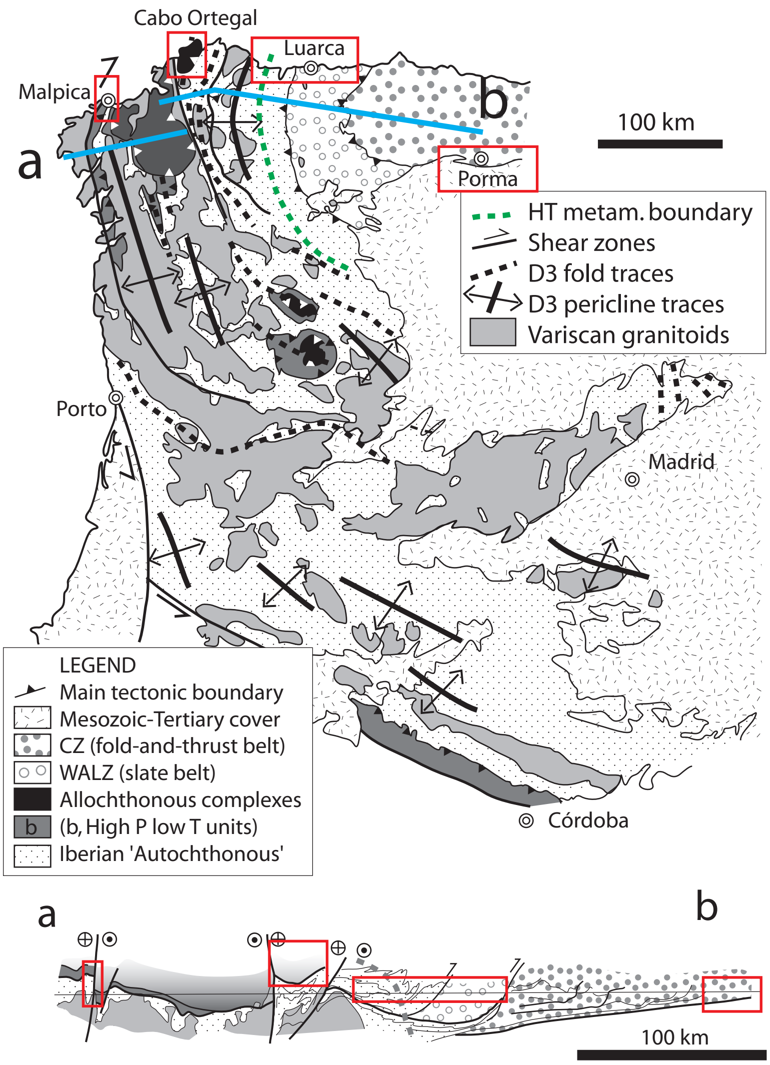

Fig. 1. Main 3D Variscan structures in Iberia: a) Map view (based on Parga Pondal et al., 1982 and Martínez et al., 1988); b) Cross section along the blue lines was redrawn from Pérez-Estáun et al., (1991) and modified to show dip of Malpica-Lamego Line (Llana-Fúnez and Marcos 2007). In green is indicated the eastern boundary of the high temperature/medium pressure metamorphism in the Variscan hinterland. The red squares show the areas visited in this trip and in the pre-conference fieldtrip.

The Variscan belt crops out largely in the western part of the Iberian Peninsula forming the Iberian Massif. According to its sedimentary facies and paleogeography, structural style and deformation, metamorphism and magmatism, the Variscan belt of the northwest Iberian Peninsula may be divided into four different zones (Julivert et al., 1972; Arenas et al., 1986; Farias et al., 1987) with different geological features (Fig. 1). From east to west, that is from the foreland to more internal parts of the belt, these zones are: the Cantabrian Zone (CZ), the West Asturian-Leonese Zone (WALZ), the Central Iberian Zone (CIZ) and the Galicia-Tras-os-Montes Zone (GTOMZ). The Variscan belt is of collisional type and the above-mentioned zones mainly represent the footwall of the suture. The suture, located in the western part of the belt (Galicia-Tras-os-Montes Zone), is marked by ophiolitic rocks beneath basic and ultrabasic complexes called Cabo Ortegal, Ordenes, Morais and Bragança.

The aim of the present fieldtrip is to show tectonic structures formed at different depths within a continental crust involved in a continental collision. Three different areas have been selected as representative of three different structural styles of deformation: the WALZ will show small scale folding affecting large portions of the Iberian crust involved in the Variscan Orogen; the Cabo Ortegal Complex illustrates the strain in lower crustal rocks at eclogites facies conditions and during their exhumation related to Variscan orogenesis; and the Malpica-Lamego Line shows the tectonites developed at the core of large crustal-scale shear zones.

The West Asturian-Leonese Zone is a transitional region between foreland and hinterland regions (Fig. 1). It is made up by a thick, pre-orogenic sequence of sedimentary rocks which includes shallow-water Lower Cambrian to Lower Devonian deposits unconformably overlying Upper Proterozoic terrigenous sediments with turbidite facies. Pervasive internal deformation is conspicuous (Fig. 1), as also is the regional metamorphism which increases progressively towards the west from greenschist to amphibolite facies. Granitic rocks intruded during the Variscan orogenic cycle are abundant in its Western part.

The Central Iberian Zone, in the restricted sense proposed by Arenas et al. (1986) and Farias et al. (1987), is formed by a narrow, arcuate band, between 10 and 65 km wide, known as Ollo de Sapo Antiform or Anticlinorium (Fig. 1). The oldest deposits dated, which are of Lower Ordovician age, lay unconformably over a porphyroid of subvolcanic to volcanoclastic origin, known as Ollo de Sapo Formation, of supposedly Late Cambrian to Early Ordovician age (Montero et al. 2007, 2009, Bea et al. 2007). Deformation, metamorphism and magmatism are comparable to those of the WALZ (Fig. 1).

In spite of the stratigraphic differences among the two zones described and the Cantabrian Zone further to the East, there is enough evidence to suggest that they were part of a single continental margin (Perez-Estaún et al., 1988), probably situated north of the Lower Paleozoic Gondwana landmass (Arenas et al., 1986). The differences in facies and thickness in the Lower Paleozoic succession, and also the scarce record of volcanic activity, can be related to an extensional tectonic activity affecting this margin, which could have given rise to the separation of troughs with a spectacular subsidence (11000 m of Paleozoic sediments in the eastern part of the WALZ after taking their strained state into account).

The Galicia-Tras-os-Montes Zone (Arenas et al., 1986; Farias et al., 1987) represents a huge stack of allochthonous structural units emplaced over the CIZ (Fig. 1). These units exhibit different lithological associations and tectono-metamorphic histories, and are interpreted as displaced terranes accreted during the Variscan orogeny. They can be grouped in two domains: a) the lower one ("Schistose Domain of Galicia-Tras-os-Montes") with Iberian affinities probably was also a portion of the continental margin of Gondwana, and b) the upper one ("Domain of Complexes with mafic and related rocks") consists of ophiolitic and catazonal units and is of more exotic provenance (Fig. 1).

The earlier Variscan structures found in rocks below the suture are grouped into two deformation phases in the literature: D_1 includes the widespread main foliation and kilometric-scale recumbent folds verging to the east, and D_2 includes thrust contacts at the base of the main tectonic units (Matte, 1968; Ribeiro, 1970; Marcos, 1973). Both recumbent folds and thrusts led to bulk thickening of the crust. D_1-related mineral assemblages in the units below the suture and across the internal parts indicate high pressure relative to temperature during deformation. The highest pressures (>2.2 GPa) were reached by rocks belonging to the margin of the Iberian plate (Martínez-Catalán et al., 1996). These high pressure rocks define the coldest geothermal gradient, that indicates the P-T-t burial path during subduction. The age estimated for this high pressure metamorphism, 360 Ma (Bosse et al., 2000; Rodríguez et al., 2003), marks the end of the 'subduction' stage in the Iberian crust.

The oldest large structures that overprint D_1 and D_2 are crustal-scale strike-slip shear zones and km-scale longitudinal pericline folds (D_3). The former cross cut earlier thrusts and follow, with varying degrees of obliquity, the trend of the orogen. Most have a significant dip-slip component (Iglesias & Ribeiro, 1981; Fig. 1). Upright km-scale periclinal folds significantly disrupt the D_1-D_2 fold-thrust pile (Fig. 1). At outcrop scale these sometimes produce pervasive crenulation cleavage. The large Variscan folds have wavelengths on the order of tens of kilometres and amplitudes of several kilometres, with varying aspects ratios that are usually larger than 3 (Fig. 1). Periclinal synforms preserve suture-related rocks and other rocks from the upper plate. Periclinal antiforms expose underlying Variscan migmatites and granites which consistently show medium pressure high temperature metamorphism of regional extent (Martínez et al., 1988). In most cases, quartzo-feldspathic rocks show extensive partial melting with variable amounts of S-type granitic intrusives. The development of all these structures was associated with several pulses of I- (tonalites and granodiorites) and S-type magmatism (two-mica granites) that span from 350 Ma to 290 Ma (Dallmeyer et al., 1997). Strike-slip shear zones, periclines, and 'granitoids' are all essentially contemporaneous.

Towards the east in the foreland fold-and-thrust belt, shortening has been estimated to be at least 150 km based on balanced cross sections (Cantabrian Zone, CZ; Fig. 1) (Pérez-Estáun et al., 1994). The earliest thrusts date from the Middle Pennsylvanian, ca. 310 Ma, and the latest Upper Pennsylvanian, ca. 300 Ma (Julivert, 1978; Marcos & Pulgar, 1982), the latter recently dated as Mosconian and Gzhelian (Merino-Tomé et al. 2009). The amount of shortening accumulated in the neighbouring slate belt (West Asturian-Leonese Zone, WALZ in Fig. 1) may be of the same order of magnitude. However, the latter estimate does not take into account internal deformation within the thrust sheets, which is significant since pervasively deformed slates are the predominant rock type.

DAY 1

Folding in orogens: The West Asturian-Leonese Zone (WALZ)

J. Aller, F. Bastida

This zone is located to the west of the Cantabrian Zone and represents the outermost part of the internal zones of the Variscan belt in NW Iberian Peninsula. The regional metamorphism front is located near the eastern boundary of the zone, and the metamorphic grade increases westwards reaching amphibolite facies in its western part. Ductile deformation is more intense than in the Cantabrian Zone and tectonic foliations are widespread.

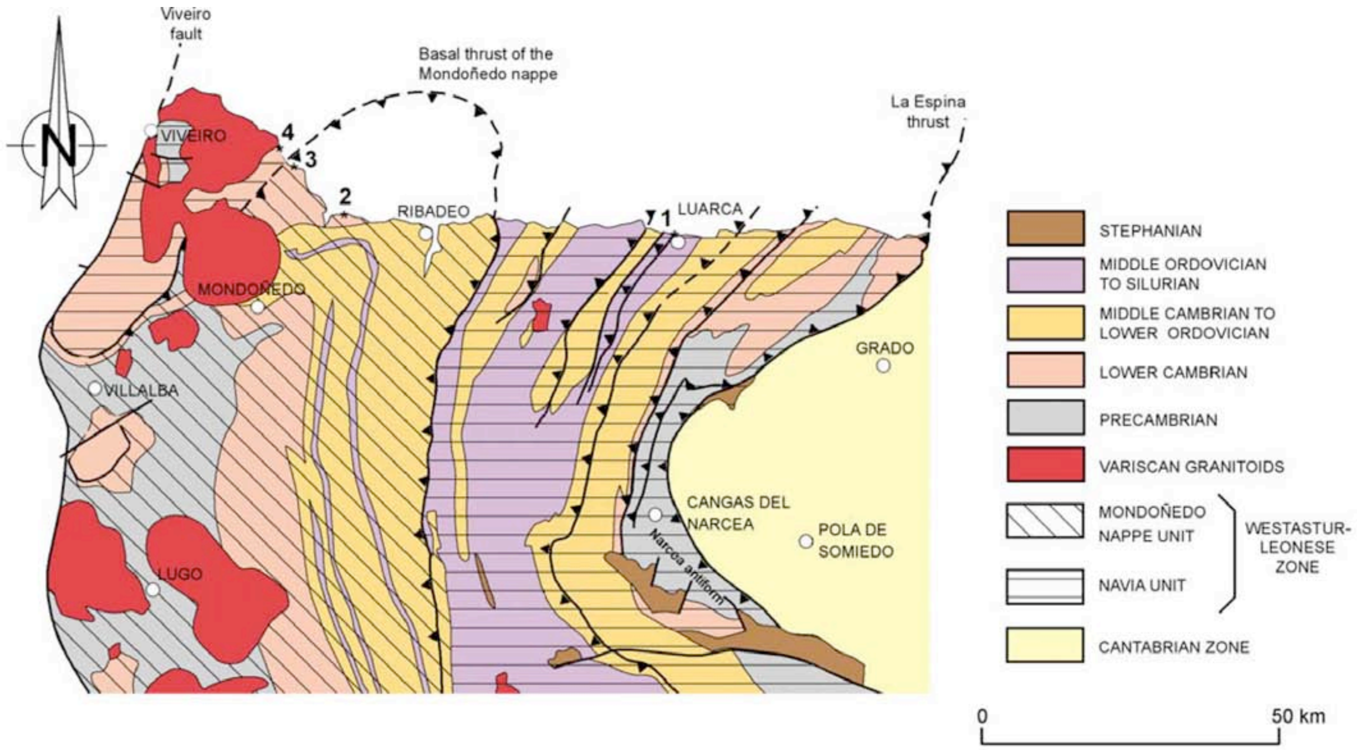

Fig. 3. Geological map and section across the northern WALZ (based on Martínez-Catalán et al., 1990 and Pérez-Estaún et al., 1990). Numbers on the coast line indicate the location of the fieldtrip stops. Letters A to D indicate the location of the cross sections shown in Figure 5.

In the northern part, the West Asturian-Leonese Zone (WALZ) was divided in two large structural units (Marcos, 1973) (Fig. 3): the Navia unit to the east, and the Mondoñedo nappe unit to the west. These two units are separated by an important thrust: the Mondoñedo nappe basal thrust. Outcrops of syn- or post-kinematic granitoids are scarce in the Navia unit, but they occupy a large area in the western part of Mondoñedo nappe unit.

The stratigraphy of the WALZ is characterized by a succession of lower Paleozoic rocks that can reach a thickness of 10 km. This feature marks a noteworthy difference with the Cantabrian Zone (Fig. 4). With the exception of an Early to Middle Cambrian carbonate horizon, the succession is formed in the northern part of this zone by siliciclastic rocks, mainly sandstones and pelites.

The observation of geological maps and sections across the WALZ indicates the existence of three main deformation phases (Marcos 1973) (Fig. 3):

First deformation phase (D_1)

Folds vergent to the foreland, with an associated tectonic foliation (S_1), were formed during this phase. These folds are overturned in the Navia unit and recumbent in the Mondoñedo nappe unit. The latter is formed by a stack of large recumbent folds whose overturned limb length, measured between two adjacent hinges, can exceed 10 km. The size of D_1 folds is smaller (the length of overturned limbs ranges between 1 and 4 km) in the Navia unit. The interlimb angle is lower and the bulk shortening related to folding is higher in the Mondoñedo nappe unit than in the Navia unit.

Second deformation phase (D_2)

Shear zones and thrusts affecting D_1 folds were produced during this phase. These structures resulted from tangential tectonics and were probably caused by progressive deformation that initiated during the first deformation phase. However, new structures, superimposed on D_1 structures, originated during D_2. The most outstanding thrust originated during this phase is the Mondoñedo nappe basal thrust, which separates the two major units of the WALZ. It crops out in the middle part of the WALZ and has a N–S trace (Fig. 3); it is folded by a later gentle fold (D_3) and appears again in the north-western part of the zone.

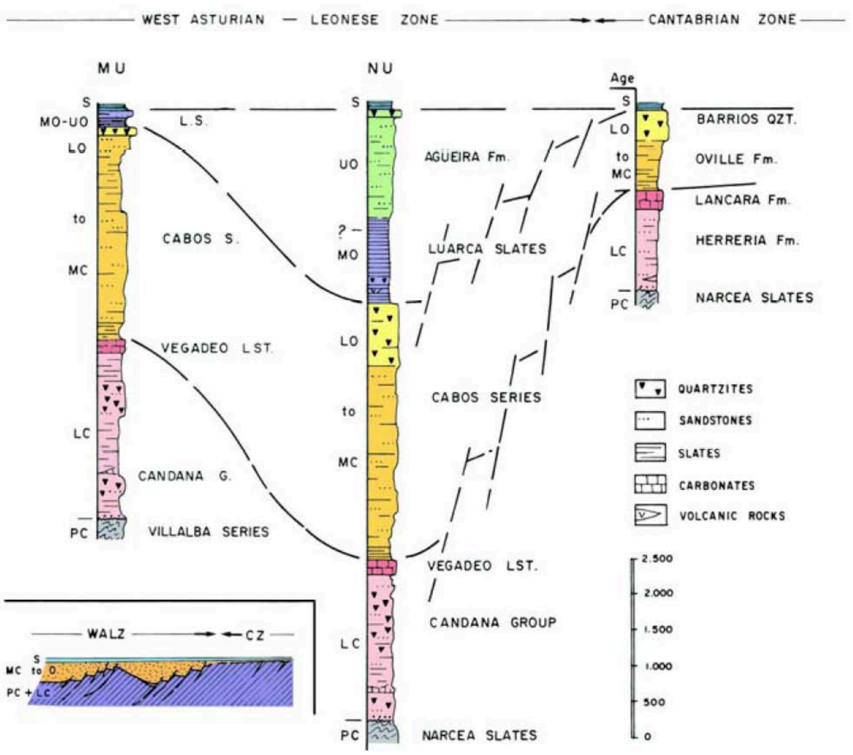

Fig. 4. Synthetic stratigraphic sections for the WALZ (MU, Mondoñedo nappe unit; NU, Navia unit) and for the lower Paleozoic succession that crops out in the western Cantabrian Zone (ZC) (after Pérez-Estaún et al., 1990). Ages: PC, Precambrian; LC, Lower Cambrian; MC, Middle Cambrian; LO, Lower Ordovician; MO, Middle Ordovician; UO, Upper Ordovician; S, Silurian.

In the eastern outcrop, the Mondoñedo basal thrust exhibits a narrow brittle-ductile shear zone, in which ductile deformation is represented by minor folds with usually curved hinges and an associated crenulation cleavage (S_2). Brittle deformation is represented by tension gashes filled in with quartz and minor faults with slickensides. In the western outcrop, the Mondoñedo basal thrust has a wide shear zone associated with a thickness greater than 3 km. The structures more common in this shear zone are asymmetric minor folds with usually curved hinges, a mylonitic foliation with a mineral lineation, schistosity often associated with micro-folds (S_2 or S_{1+2}) and shear bands.

Third deformation phase (D_3)

Gentle to open asymmetric folds were formed during this deformation phase. They are upright folds, or folds with axial surfaces dipping steeply to the E, and are nearly homoaxial with respect to D_1 folds. D_3 folds have an associated crenulation cleavage (S_3) well developed in the metapelites. The development of D_3 folds is different in the two main units that crop out in the northern portion of the WALZ.

In the Mondoñedo nappe unit only two major D_3 folds exist: a wide synform in the eastern part and a narrower antiform in the western part. They are gentle folds and their wavelength equals the width of the structural unit. Minor D_3 structures are very scarce in this unit. In the Navia unit, major D_3 folds are more abundant; their wavelength is smaller and their amplitude is greater than those in the Mondoñedo nappe unit. In this unit, the minor D_3 structures, that is folds and crenulation cleavage, have not a uniform distribution and they concentrate along short, gentler dipping limbs of the major folds.

The superimposition of D_3 folds on D_1 folds gave rise to Ramsay's type 3 interference patterns. The resulting geometry can be recognized on both the geological map and the cross sections (Figs. 3 and 5). In the Navia unit, these patterns have a hook-like shape (Fig. 5), so that the gentler dipping, normal limbs of D_1 folds are clearly folded by D_3 structures, whereas the overturned, steeper dipping limbs of D_1 folds underwent mainly rotation.

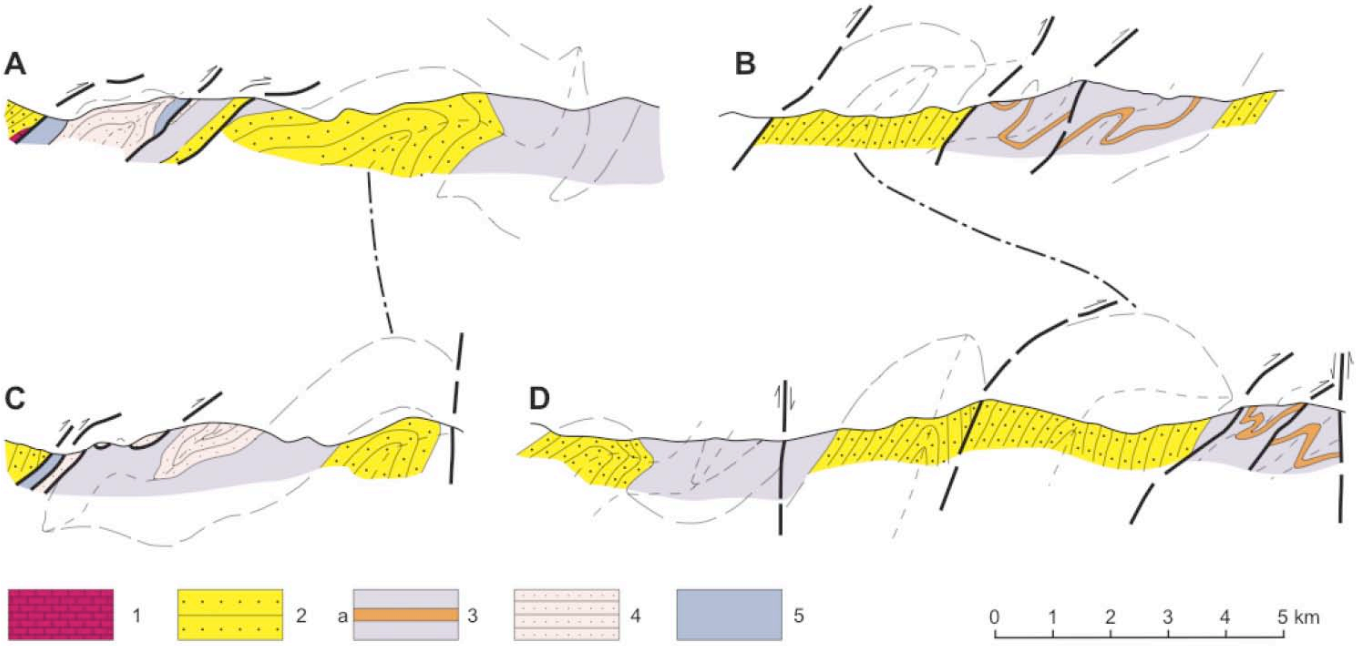

Fig. 5. Geological sections across the Navia unit showing the geometry of the interference patterns produced as a result of the superimposition of D_1 and D_3 folds (after Pulgar, 1980). The D_1 axial traces are deformed by D_3 folds. 1, Vegadeo limestone; 2, Cabos Series; 3, Luarca slates (a, Sabugo quartzite); 4, Agüeira Formation; 5, Silurian slates. The location of the cross sections is shown in Figure 3.

In addition to the structures formed during the deformation phases described above, other structures occur, such as normal faults, near-horizontal kink bands and transverse open folds. Usually, the two first types of structures have a local distribution, resulted from a near-vertical compressive stress and have been interpreted as late-Variscan structures.

STOP 1.1: **Luarca beach**

This beach is an excellent locality to visualize structures formed during the third Variscan deformation phase (D_3) in the West Asturian-Leonese Zone. The outcrop is constituted by Ordovician rocks that belong to the upper member of the Luarca Formation. This member is formed by a uniform series of black slates whose thickness is greater than 500 m. Usually, in these rocks, bedding is obliterated by a well developed S_1 slate cleavage and cannot be observed. Quartz veins, generally concordant with the S_1 foliation, are common. The rocks found in this area underwent a low grade regional metamorphism in the green schist facies (chlorite zone).

The structure of this area is shown in Fig. 6, in which we can see the asymmetry of the D_3 folds. The short limbs are subhorizontal and define bands whose thickness ranges between 5 and 25 m. The long limbs dip more than 60º and define bands whose thickness can reach 50 m.

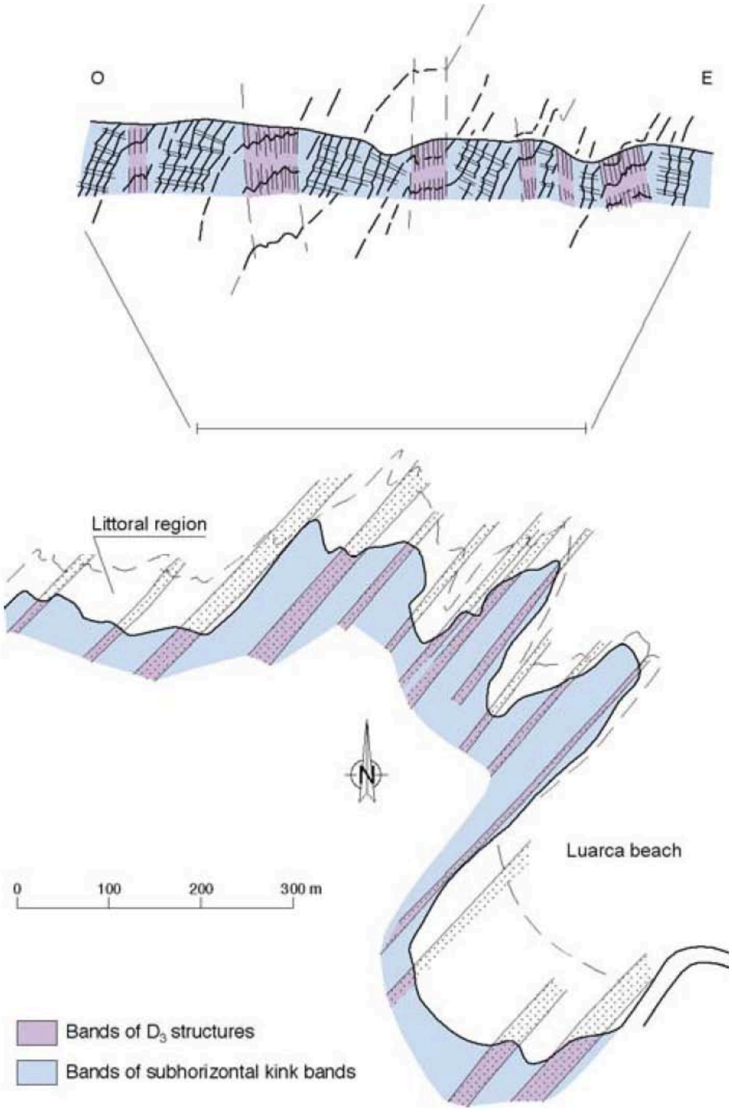

Fig. 6. Geological cross section and map corresponding to the stop 1 (Luarca beach and surrounding areas) showing the distribution in bands of the structures formed during D_3 and the later near horizontal kink bands (after Pulgar, 1980).

The short limbs underwent an important D_3 deformation that gave rise to minor folds commonly affecting the quartz veins. The wavelength of the minor folds is of several centimetres or decimetres. The limbs have a steeply dipping crenulation cleavage S_3, commonly involving development of a white-dark tectonic banding that reflects changes in quartz contents and is the result of pressure solution processes. The long limbs do not develop minor D_3 structures and usually exhibit subhorizontal or gently dipping kink-bands formed during a late-Variscan episode in which the maximum compressive stress was almost vertical.

The non uniform distribution of minor D_3 structures described above also occurs at larger scales, so that the metric folds described are located in the short gentler dipping limbs of major asymmetric D_3 folds. These limbs define macroscopic bands whose thickness ranges between 500 and 2000 m (Pulgar, 1980).

STOP 1.2: **Benquerencia beach section**

This cross section (Fig. 7) is a nice example of how a structure can be reconstructed using geometrical relationships between bedding and tectonic foliation together with cross bedding. It is located in the Mondoñedo nappe unit (western part of the West Asturian-Leonese Zone), which is formed by a stack of recumbent folds (D_1 structures) of large dimensions (the length of the overturned limb, measured between two adjacent hinges, is 10 km or even more), offset by a basal thrust with a thick shear zone associated. D_1 folds are folded by homoaxial gentle D_3 folds giving rise to a Ramsay's type 3 interference pattern.

The main folds in this section were formed during the D_1 deformation. They are parasitic folds located on the normal limb of a major anticline (Foz-Tapia anticline) and developed on sandstones and phyllites of the Cándana group (Lower Cambrian), whose metamorphic grade corresponds to the biotite zone. The folds have an associated S_1 foliation well developed in all the rocks. The structure shown on the section consists of a recumbent anticline and a syncline tilted as a result of D_3 deformation. Microgranite dykes cutting across D_1 folds can be observed in several points along the cross section.

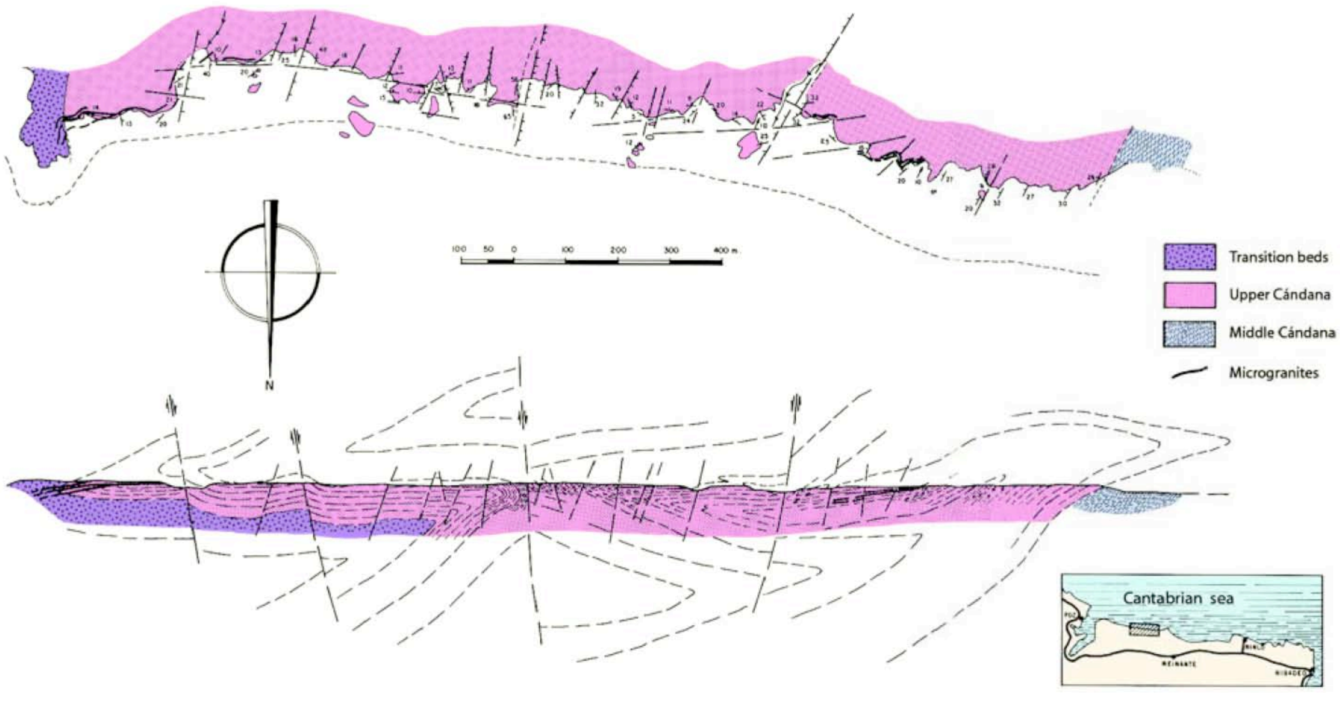

Fig. 7. Geological section across the Benquerencia beach (after Bastida and Pulgar, 1978).

STOP 1.3: **Punta das Cabras, the Mondoñedo Nappe basal shear zone**

Many minor D_2 structures, developed within the Mondoñedo Nappe basal shear zone, can be observed in this locality. The eastward dip of both the bedding and the basal thrust is a consequence of a generalized tilting due to D_3 deformation in this area (Figs. 3 and 8). The rocks that crop out in this locality are quartzites and micaschists, and belong to the basal part of the Cándana Group (Lower Cambrian). These rocks were metamorphosed in the andalusite zone conditions. This zone partly overprints the staurolite zone defined by relicts of this mineral inside andalusite porphyroblasts. The rise in temperature during the development of andalusite was coeval with ductile deformation in the shear zone. A retrograde metamorphism to greenschists facies conditions occurred late- to post-D_2 (end of the ductile deformation and development of the basal thrust).

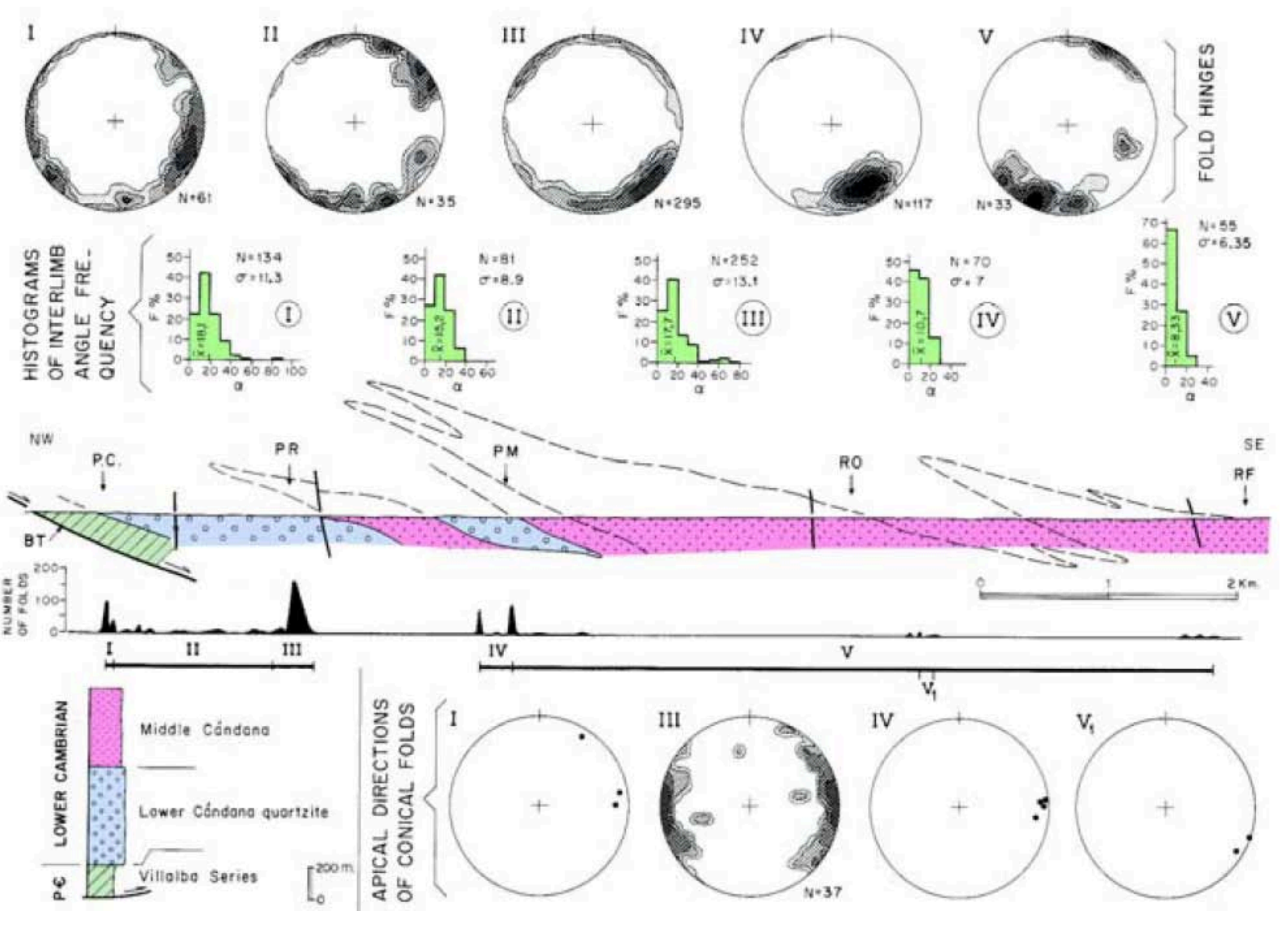

Fig. 8. Section across the Mondoñedo nappe basal shear zone showing orientation of D_2 folds, histograms of interlimb angle frequence, number of folds and apical directions of conical folds along the section (after Aller and Bastida, 1993). Contours: 1, 2, 4 and 8%. PC, Punta das Cabras; PR, Punta Riomar; PM, Punta Morago; RO, Ría d'Ouro; RF, Ría de Foz; BT, basal thrust.

Minor tight to isoclinal folds, not associated with major folds, are abundant in the western part of the shear zone and they are concentrated in four narrow discrete zones, one of them in the Punta Das Cabras (Fig. 8). The folds are E-facing and commonly occur as strongly asymmetric anticline-syncline couples. They are subsimilar folds and their hinges have a large dispersion of plunge directions (Fig. 8) in the lower part of the shear zone. Curved hinges are common and are responsible for conical or sheath folds and eye-shaped structures. The quartzites exhibit a blastomylonitic microstructure with a well developed mineral lineation (L_m) that plunges gently towards the E. A schistosity with scarce microfolds developed in the micaschists. Small extensional bands are also common.

In order to visualize the displacement direction in the shear zone, Figure 9 shows the direction of the first eigenvector of the Bingham's distribution for several criteria in different sectors of the shear zone. The first eigenvector of the distribution of the mineral lineation shows a direction between N75E and N115E, i.e., with a dominant eastern component. Other criterion is the direction contained in bedding planes (S_0) perpendicular to the intersection between S_0 and the shear bands; its distribution of direction ranges between N65E and N140E. Fold hinges show a great dispersion of directions; only in the Punta Das Cabras, where the deformation is probably greater than in other localities, the direction is approximately E–W. Finally, the direction of the first eigenvector of the apical directions of conical folds, defines a fan of directions between N70E and N115E. All these data suggest an approximately eastward direction of displacement of the shear zone.

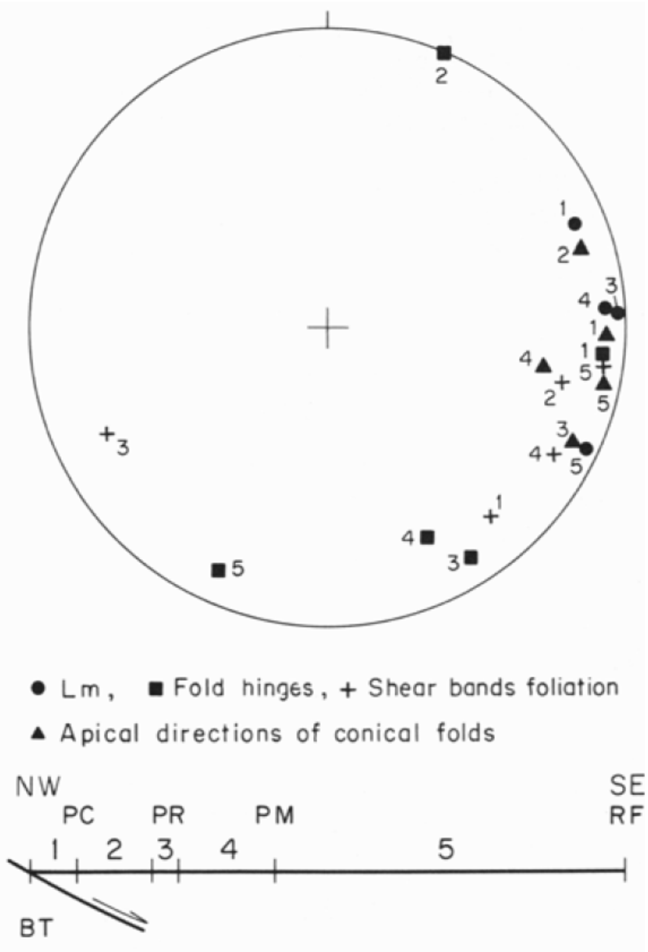

Fig. 9. Pattern of the fi rst eigenvector obtained from the distributions of the movement direction inferred from different criteria along the shear zone (after Aller and Bastida, 1993). Abbreviations as in Figure 8.

STOP 1.4: **Burela D_1 folds**

A train of minor metric D_1 folds crops out in the south-eastern part of the Burela Harbour along more than 1 km of coast (Fig. 10). They are developed in sandstones and sandy metapelites of the lower part of Cándana Group (Lower Cambrian). The folds are tight, overturned and vergent to the E. Commonly, they have associated boudinage and joints perpendicular to the axial surfaces.

Together with the D_1 folds, some small open folds appear with subhorizontal axial surfaces that resulted from subvertical compression subsequent to the first deformation phase. Another effect of this compression was the development of a subhorizontal or moderately dipping tectonic foliation that cuts the limbs of the D_1 folds and obliterates partially the S_1 foliation. In addition, the boudins are not slender structures and they look shortened and thickened as a consequence of this compression.

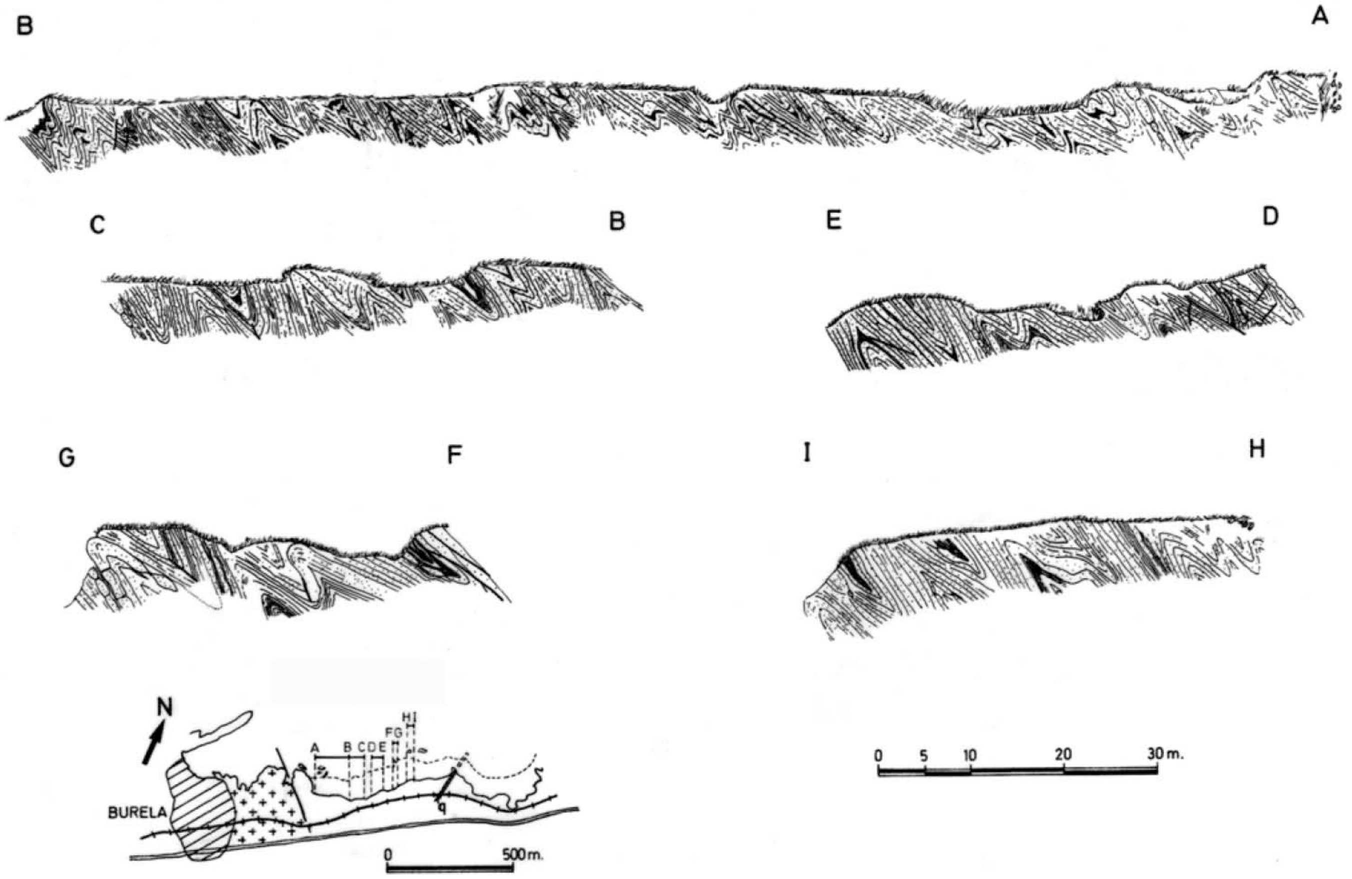

Fig. 10. Detailed cross sections showing the D_1 folds geometry in Burela (after Bastida, 1980).

DAY 2

High Pressure and High Temperature Cabo Ortegal Nappe

A. Marcos, F.J. Fernández, S. Llana-Fúnez

The Cabo Ortegal Complex (Vogel, 1967) is a composite of allochthonous sheets that were thrusted onto the western edge of Gondwana during the Variscan Orogeny (Ries and Shackleton, 1971; Ribeiro, 1974; Arenas et al., 1986). It is formed by an Upper Tectonic Unit, the Cabo Ortegal Nappe (CON) (Marcos, 1998), made up of rocks that recorded a metamorphic event of high pressure and high temperature (Vogel, 1967), and a Lower Tectonic Unit, subdivided in three thrust sheets and made up of rocks equilibrated in amphibolite facies or lower metamorphic conditions.

The rocks that form the lower units come from different and distinct geodynamic settings: i) E-MORB basalts in the Purrido Thrust Sheet (Arenas, 1988); ii) tectonic melanges in the Moeche Thrust Sheet; and iii) calcalkaline arc volcanics in the Espasante Thrust Sheet. This set of sheets would have been emplaced over the passive margin of Gondwana during the collision that gave rise to the Variscan belt and would represent the suture zone of this collisional belt. A thin thrust sheet, the Para-autochthonous, separates the Lower Tectonic Unit, above, from the metasediments of the Lower Paleozoic sequence in the Ollo de Sapo Antiform, which are situated below (Marcos and Farias, 1997, 1999).

The lithological sequence: age and metamorphism

The rocks that form the CON can be divided, in three major ordered lithological units: 1) the ultramafic rocks, 2) the mafic rocks, and 3) the quartzo-feldspatic gneisses (Fig. 11 and Map). Unit 1 represents the base of the sequence and it is formed by more than 600 m of alternating serpentinized peridotites and pyroxenites (Vogel, 1967; Girardeau *et al*., 1989). The contact with the neighbouring mafic rock unit (2) is gradational (Galán and Marcos, 1997). Three members can be distinguished within the mafic rocks. The *lower member* is formed by massive or weakly foliated ultramafic-mafic rocks (50-100 m thick) which pass gradually into layered mafic rocks (*middle member*, 400 m thick) by increasing the amount of plagioclase (Galán and Marcos, 1997). Sharp contacts separate the middle member from the overlying *upper member*, made of massive mafic rocks (100-200 m), and the latter from the quartzo-feldspatic gneisses (Unit 3, more than 600 m thick). In the proximity of this contact, the gneisses include meter-scale lenses of eclogites and mafic rocks, and show evidences of migmatization.

Considering the nature of petrologic and geochemical features in the lithological sequence described above and its relation with the underlying rocks, the section has been interpreted as a portion of lower crust, formed as a consequence of ensialic extension or deriving from a passive margin (Galán and Marcos, 1997).

The absolute age of the protoliths and the H*P*-H*T* metamorphic imprint in the rocks has been a matter of debate (see review in Ábalos and Aranguren, 1998). Recent data (Fernández-Suárez *et al*., 2000) indicate that the rocks in the CON record a polyorogenic evolution, with an H*P*-H*T* episode at ca. 490-480 Ma (Early Ordovician) and a younger Early Devonian event dated at ca. 390-385 Ma. The age of the late deformation in greenschists facies conditions would occur at 365 Ma (Late Devonian) (Dallmeyer *et al*., 1997).

The metamorphic evolution of the mafic rocks supports the previous H*P*-H*T* regime, followed by the development of different structures related to exhumation and equilibrated in consecutive HP granulite, amphibolite and greenschist facies which, as a whole, define an isothermal decompression-type *PT* path. The estimated metamorphic peak conditions range from 842-884°C for 1.4-1.6 GPa (Bacariza-type eclogites, Galán and Marcos, 1999) to 770-800°C for 1.4-1.7 GPa (Concepenido-type eclogites, Mendia, 1996).

The deformation sequence

The outcrop of the high-grade rocks of the CON is limited by the Atlantic Ocean to the N and by high-angle normal faults convergent to the S, E and SW (Map and Fig. 11). The tectonic graben in which the CON occurs is bounded by the Lower Tectonic Unit rocks to the E and by the Para-autochthonous rocks to the SW. The normal faults cut all previous structures and are probably formed during the Alpine deformation events in NW Iberia. Within the CON, the reconstruction of structures is partially masked by a system of high-angle normal faults striking N130E that sink the northern block. These faults, several kilometres long, are straight and have vertical separations of hundreds of meters. Due to this late tectonic event, lithological boundaries exhibit frequently discontinuous and complicated patterns in the maps, changing from the footwall to the hangingwall of the mentioned normal faults.

The fabric used as reference for the structural reconstruction is a widespread mylonitic fabric which is observed throughout the CON rocks. This mylonitic fabric is affected by isoclinal folds with fold axes striking NNE-SSW and verging to the SSE. The major folds, isoclinal and recumbent, have reverse limbs that reach up to 6 km and are cut by thrusts. Two of these thrusts, the named Basal and Upper Thrust, reach several kilometres offset. The late Variscan refolding gives the whole CON the characteristic elliptical outcrop pattern. Open folds with vertical axial planes and fold axes oriented WNW-ESE form at all scales. The intersection relations with previous isoclinal folds give place to Ramsay type 3 interference patterns at outcrop and map scale (Map and Fig. 11).

Prograde deformation

The structural record in the rocks of the CON can be related with deformation in prograde conditions in granulite and eclogite facies, although the more widespread and extensive ductile structures result from retrogressive deformation associated with the exhumation of the rocks. The development of crystallographic preferred orientation in clinopyroxenes in eclogites and granulites (Engels, 1972; Godard and Van Roermund, 1995, Ábalos 1997, Llana-Fúnez et al., 2005) occurred simultaneously to high pressure metamorphism, at around 380 Ma.

The widespread amphibolite facies overprint and local greenschist facies shear zones have younger ages, <360 Ma (Dallmeyer et al, 1997), which correspond to crustal thickening produced by the Variscan continental collision (Marcos and Galán, 1994).

Retrograde deformation

Most of the structures found in the H*P*-H*T* rocks of the CON are defined by a mineral assemblage in amphibolite-facies conditions and are thus considered to have developed during their exhumation. The first structure that appears is the generalized tectonic fabric (D_1), planar, which at the map scale is parallel to lithological boundaries. This foliation affects all rock types and is taken as the structural reference frame in them. It is a mylonitic foliation, sometimes a gneissic banding, which started to

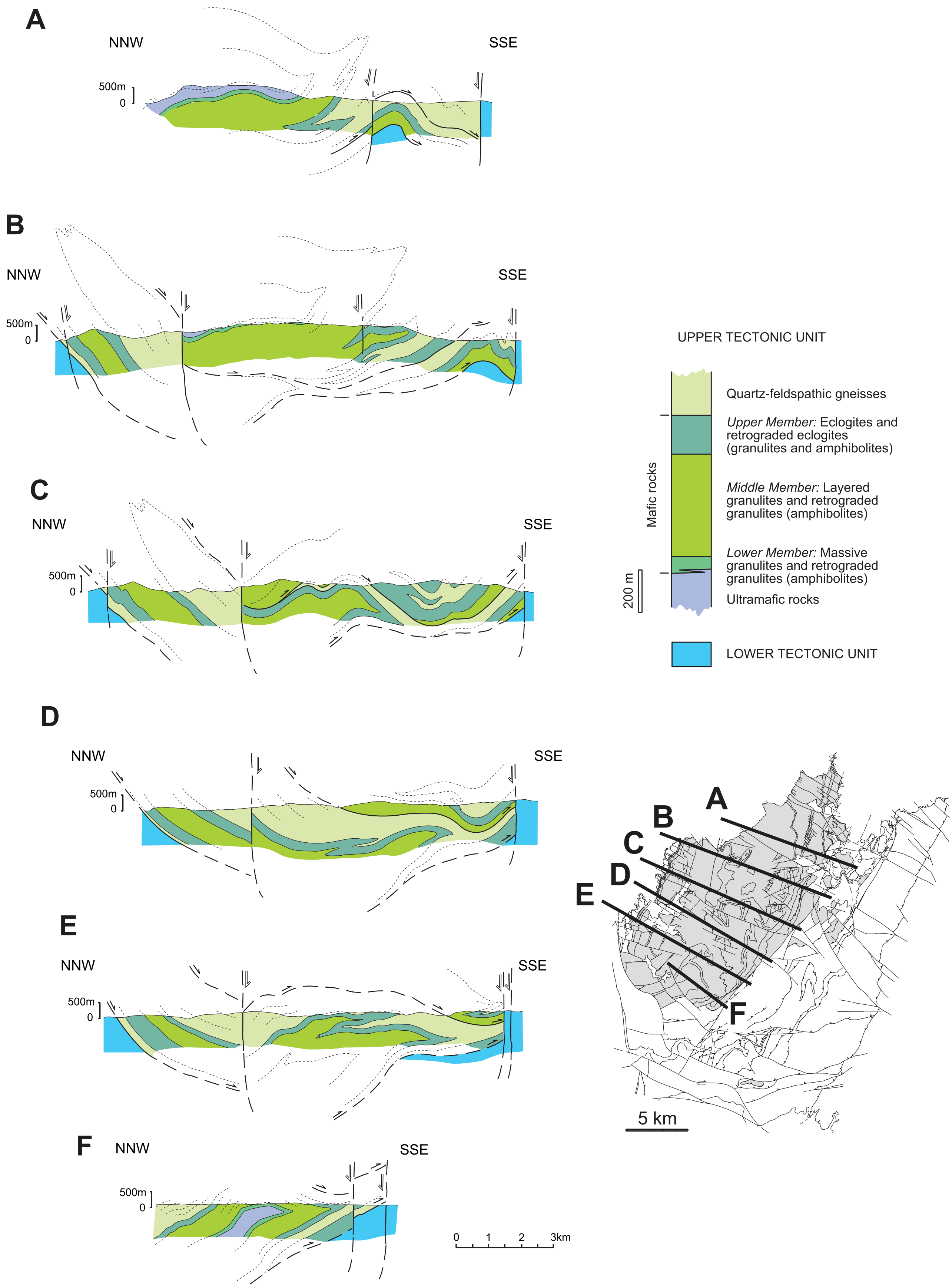

Fig. 11. Geological sections through the Cabo Ortegal Complex (after Marcos et al. 2002).

develop in granulite facies conditions and was finally equilibrated in high grade amphibolite facies *PT* conditions (Fernández, 1993; Marcos and Galán, 1994; Fernández and Marcos, 1996). At the outcrop scale, the foliation may show an anastomosing aspect nucleated around pods, in which the rocks preserved the previous (migmatic) fabric, or a clear and completely formed planar fabric. In some cases, disrupted or truncated foliations are observed, which may indicate the development of more than one consecutive mylonitic fabric during the same event of deformation. In relation to the generalized tectonic fabric, isolated intrafoliar isoclinal folds are observed.

A detailed structural analysis of the macro- and micro-structures associated with the mylonitic foliation in quartzo-feldspatic gneisses leads us to propose a heterogeneous coaxial flow regime during the extensive ductile deformation (Fernández and Marcos, 1996). This model is supported by the lack of well-developed mineral lineation (Fig. 12), the orthorrombic symmetry of quartz *c*-axis textures, and the absence of consistent sense of movement inferred from kinematic criteria (Fernández, 1993; Fernández and Marcos, 1996). The distribution of c-axis maxima in omphacite in eclogites also supports a generalized strain geometry characterized by flattening (Llana-Funez et al., 2005).

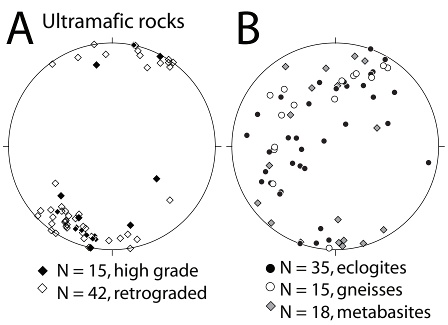

Fig. 12. Stereographic plots of lineations in (A) ultramafic rocks, distinguishing primary assemblages (black diamonds) from retro- grade ones (open diamonds) and (B) eclogites, quartzofeldspathic gneisses and metabasites (black circles, white circles, and gray-filled diamonds, respectively). From Llana-Fúnez et al. (2004)

The foliation is affected by similar-type asymmetric folds with low interlimb angles (tending to isoclinal folds) with axes oriented NNE-SSW and vergence to the SSE (D_2 folds). The major folds are isoclinal and recumbent and their reverse limbs reach kilometric scales (more than 6 km) (Map and Fig. 11). Minor folds are of decametric size and generally asymmetric. The refolding by these folds of previous intrafoliar ones gave place to type 3 interference patterns of Ramsay and sometimes to type 2 patterns. In the field, no axial plane cleavage is observed, although with the handlens or under the microscope a slight orientation of hornblende (in amphibolites) or phylosillicates (in gneisses) is seen in some D_2 hinges. The lack of an associated crenulation cleavage in relation to the D_2 folding makes it difficult to establish the exact *PT* conditions at which deformation microstructure was produced. However, their similar geometry and the strong flattening that they indicate, imply a general ductile behaviour of rocks which would suggest that the deformation microstructure was heavily reset in amphibolite facies conditions. In consequence, a significant reworking of D_1 foliation (main tectonic fabric) is to be expected in fold limbs.

The major D_2 folds are cut by thrusts, two of them reaching kilometric size (Map and Fig. 11). One of them, the Basal Thrust, places the CON over the lower units. This thrust 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 marked by strongly sheared and serpentinized ultramafic rocks, probably derived from one of the lower units (Marcos and Farias, 1999). The shear zone, where minor structures such as thrusts, folds and crenulation cleavage are formed, is in some locations wider than 50 m and it is developed mainly on the footwall rocks. The minor folds are asymmetric, reach metric scale and have curved hinges, allowing to deduce an approximate WNW-ESE orientation for the tectonic transport. The other major thrust, the Upper Thrust, imbricates the rocks of the CON producing two superimposed tectonic sheets (the Cedeira Sheet, above, and La Capelada

Sheet, below). These sheets have the same tectonostratigraphic sequence but are differentiated by the degree of retrogradation, which is stronger in the lower unit, where granulite and eclogite rocks are only found as relics. The Upper Thrust shows a complicated cartographic pattern due to refolding during D_4 deformation and, basically, to the late faulting. Its general features are similar to the ones found in the Basal Thrust. The structural relations between the Basal and Upper Thrusts and the existence of klippen of La Capelada Thrust Sheet over the underlying Lower Tectonic Units (Marcos and Farias, 1999), indicate that this latter thrust was the youngest to be formed, following an inward sequence of emplacement.

The following deformation episode, D_4, is related to the general refolding event that gives the Cabo Ortegal Complex its typical elliptical cartographic shape. The D_4 folds form at all scales; they are open folds with subvertical axial plane and fold axis oriented WNW-ESE. In some cases, they show axial crenulation cleavage and a mineral lineation, parallel to fold axes, which is defined by chlorite. The distribution of D_4 structures is rather heterogeneous. The interference between D_2 and D_4 folds produces interference patterns of type 3 of Ramsay, both at cartographic and outcrop scale.

Kinematic interpretation of the CON structures

The predominant orientation for lineations (mineral orientation and/or intersection) associated with the main foliation (used as kinematic reference frame) in the rocks of the CON is SSW-NNE. In several localities, the more clearly defined lineation is an intersection lineation between the foliation plane and a previous compositional banding; the mineral orientation lineation, when present, is parallel to this intersection. In sections normal to the foliation different kinematic criteria are found (d and s porphyroblasts, intrafoliar folds, C' shear bands) which do not indicate, at outcrop scale, a common shear sense. It seems to suggest that the mylonitic foliation is developed in a general non-coaxial fl ow which strongly departs from the simple shear fl ow and which is closer to coaxial deformation.

The general kinematic interpretation is even more complicated due to the superimposition of the structures formed during D_1 and D_2. The consequences on the main foliation of the development of isoclinal folds in ductile conditions is not easy to evaluate. On this point, it should be mentioned that the more clear lineation found in the rocks is defined by the orientation of hornblende, which could grow during D_1 as well as during D_2. In any case, a simple geometric effect has to be considered: deformation of the D_1 lineation by isoclinal and homoaxial isoclinal folds leads to the observation of opposite senses of shear in normal and reverse limbs if the kinematic criteria are seen in sections parallel to lineation (and therefore to isoclinal fold axis). In these conditions, the study of kinematic criteria requires an extremely careful work.

The structures originated during D_2 and D_3 events can be considered the result of progressive deformation that led to the fi nal emplacement of the CON rocks to its present position. Both the vergence of the D_2 folds and the sense of shear in D_3 thrusts indicate a general tectonic transport of the CON towards the ESE. This movement is consistent with the geometry of the structures developed in the underlying Paleozoic metasediments which constitute the Autochthon of the nappe (Matte, 1968; Pérez Estaún *et al*., 1991).

The Cabo Ortegal Nappe represents a thinned portion of lower crust that was exhumed during the convergence associated with the formation of Variscan Orogeny. From the available geochronological and metamorphic data, we conclude that the rocks of the CON reached granulitic conditions during their burial at the stage of collision, then experienced a pervasive retrogression and final emplacement in middle-upper crustal levels over the tectonic sheets that formed the lower units.

STOP 2.1: **Basal thrust and Candelaria amphibolites at Cedeira** *(alternative Cabalón boulder beach)*

The first locatity is found to the west of the Cedeira harbour. In this first stop we will briefly introduce you to the geology of the Cabo Ortegal Complex (COC). As can be seen in the cliffs further to the north and oriented roughly N-S, the main foliation in the mafic rocks of the Upper member is dipping to the E. The COC is a klippe; therefore the high-grade rocks come from above as we see them now, they represent relics of a larger nappe which was emplaced towards the E in the Variscan Orogeny.

Here we can see the boundary between the Purrido Thrust Sheet and the Upper Unit of the Cabo Ortegal Complex (Carreiro Thrust). Near this thrust the Purrido Amphibolites are intensely folded and are in contact whith quartz-feldsphatic mylonitic gneisses (Chímparra Gneisses).

The Purrido Amphibolites (ca. 300 m) form a homogeneous succession of green-grayish amphibolites that present a fine and discontinuous compositional banding defined by plagioclase. The amphibolites are composed of Hbl, Pl and Ep-Czo; with Grt, Rt and Ttn as accessory minerals (from now, mineral symbols as in Kretz, 1983), and 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. The metamorphism in amphibolite facies (500-600 °C) has been dated by Peucat et al. (1990) at ca. 390 Ma, using $^{40}Ar/^{39}Ar$ in Hbl. The development of the tectonic fabric is probably of the same age. The Purrido Amphibolites have composition of olivine tholeiitic basalts (Arenas, 1988). Available trace element compositions of these rocks do not discriminate between N-MORB and WPB (Arenas, 1988). This author suggests an average E-MORB composition for them, although it is uncertain whether it is a primary or a secondary character.

Near the contact, the gneisses are fine- or medium grained composed of Qtz, Pl, Grt, Bt, Ms and Ky (Vogel, 1967), and present a characteristic compositional banding, which results from changes in the amount of mafic minerals. They 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. Metamorphic mineral associations in both mafic rocks and host gneisses indicate a H*P*-H*T* metamorphism (Vogel, 1967; Gil Ibarguchi et al., 1990), causing partial melting of the gneisses. The thickness of these lower gneisses is of the order of 150 m. The rest of this unit is formed of two-mica layered quartzo-feldspathic gneisses (often with Grt, St and Ky). They are interpreted as metamorphic greywackes (Vogel, 1967; Fernández and Marcos, 1997) with calc-silicate and mafic inclusions. Available geochemical data for some of the mafic inclusions indicate T-type MORB compositions with important contribution of continental material (Peucat et al., 1990), which suggests that they intruded in an extensional continental setting.

STOP 2.2: **Mafic granulites at the Outeiro hill**: **main foliation and high grade fined grained shear bands**

We walk for approximately 1 km from La Capelada track through a pine forest until we reach the top of the Outeiro hill.

The aim of this stop is to observe and discuss the development of the mylonitic foliation in mafic granulites and garnet amphibolites (retrograde after granulites) of the lower member.

The outcrop is formed by coarse- to medium-grained garnet amphibolites made of garnet, hornblende, plagioclase, quartz, ± biotite, ilmenite, sphene, ± allanite, clinozoisite and epidote. Clinopyroxene, scapolite, kyanite, zoisite and rutile are rare (< 5%) and are found as relict phases. Garnet always shows more or less developed coronas of plagioclase, amphibole and ± ilmenite. Locally, these rocks followed partial melting during the previous granulite facies metamorphism (Marcos and Galán, 1994). The resultant partial melts are trondhjemites that either crystallised *in situ* or were removed forming centimetre-wide veins crosscutting the mafic granulites (Galán and Marcos, 1997). Layers and tectonic inclusions of ultramafic (pyrigarnites and hornblende pyrigarnites after Vogel, 1967: garnet, clinopiroxene, ± hornblende, plagioclase, ± quartz, rutile, ± allanite, ± zoisite, ± apatite) and mafic granulites (± hornblende plagiopyrigarnites; Vogel, 1967) can be seen enclosed in the garnet amphibolites.

The rocks include also blastomylonitic biotite-gneisses (Fig. 13) which are located in shear bands, from scant centimetres to several metres wide, deforming the garnet amphibolites. They enclose porphyroclasts (2-5 mm) of garnet, hornblende, plagioclase, clinozoisite with zoisite and allanite cores and, less frequently, scapolite, all of them in a very fine grained mylonitic matrix of plagioclase, quartz, biotite, epidote and ilmenite.

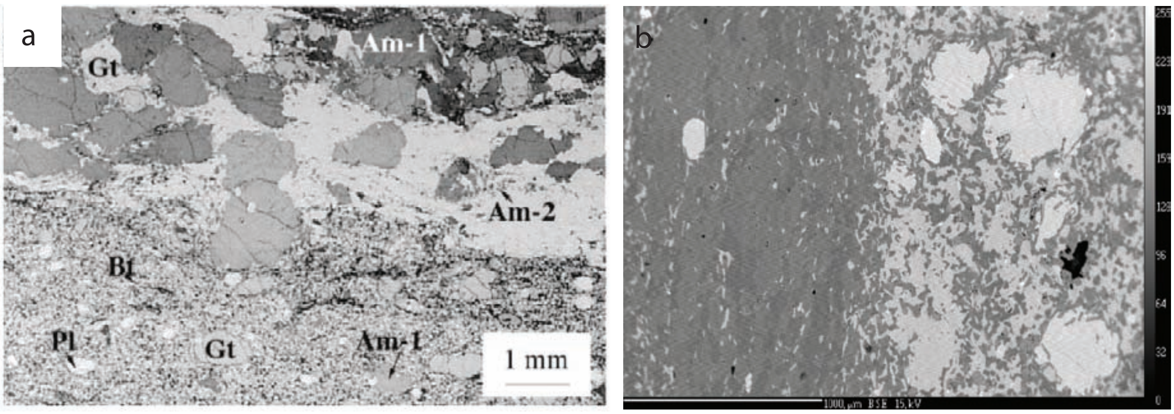

Fig. 13. a) Biotite blastomylonitic gneiss formed from garnet amphibolite in a microshear band; garnet, Am-1 type amphibole and plagioclase appear as porphyroclasts in the gneiss AMP-441, La Capelada unit (Galán and Marcos 2000). b) Back scattered electron micrograph of the contact between a quartzo-feldspathic layer in the granulites (left half of the image) and a biotite blastomylonitic gneiss (right side of the image).

The garnet amphibolites show a pervasive mylonitic foliation, which started to develop at granulite facies *PT* conditions and equilibrated finally at high-grade amphibolite facies *PT* conditions (Marcos and Galán, 1994; Galán and Marcos, 1997, 2000) (Fig. 14). The deformation associated with this fabric is characterized by dynamic recrystallisation of hornblende, plagioclase, quartz and epidote. At outcrop scale, the foliation shows two relevant aspects: an anastomosing pattern surrounding mafic granulites and ultramafic inclusions or a clearly planar pattern. In some cases, truncated relations between sets of foliated rocks are common and are interpreted as having developed during the same process of mylonitic deformation (Fig. 15). Also, in relation to foliation development, isolated intrafoliar folds are observed (Figs. 15 a, b).

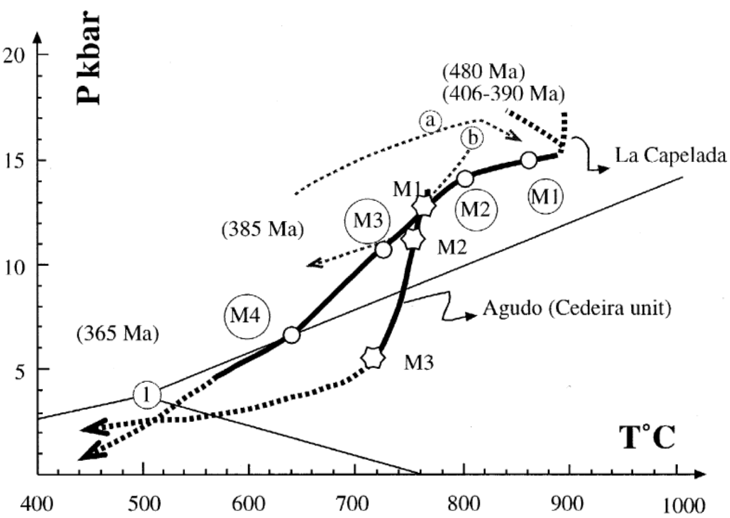

Fig. 14. P–T–t paths of the HP granulites in La Capelada and Cedeira unit. Numbers indicate estimated ages in Ma for the M1–M2, M3 and M4 metamorphic stages in La Capelada unit. (a) and (b) are the prograde and the retrograde paths (formation of clinopyroxene-plagioclase symplectites), respectively of the neighbouring Concepenido eclogites determined by thermobarometric methods (Mendia, 1996). (1) Alumino-slicate polymorph stability fi elds after Salje (1986). Taken from Galán and Marcos (2000).

Associated with the fabric there is a weak lineation (Fig. 12). The final stages of fabric development include the formation of metric-scale folds that bend the foliation (Fig. 15 c). In most cases, these folds are nucleated around inclusions of mafic granulites and ultramafic rocks; it is in these particular cases where the lineation is better seen .

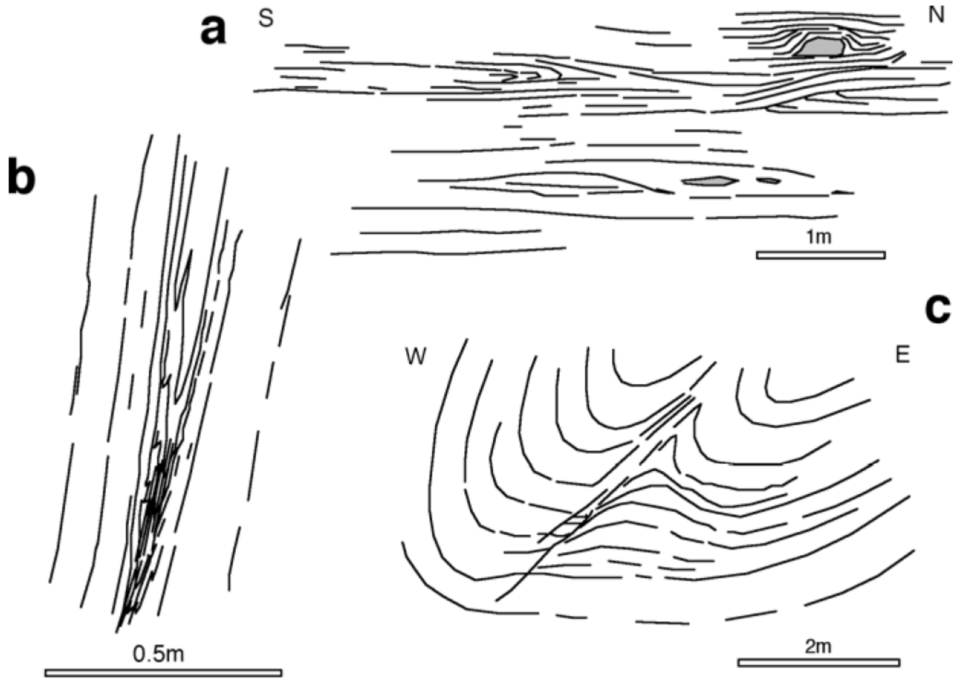

Fig. 15. Field sketches showing some of the stages during progressive foliation development in mafic rocks (garnet amphibolites and granulites) at the locallity of the Outeiro hill. In a) the planar mylonitic foliation surrounds inclusions made of mafic granulites and shows truncating relations between sets of foliated rocks. In b) a week crenulation cleavage is preserved in hinges of isoclinal intrafoliar folds within the strongly foliated parts of garnet anphibolites. Some of the open folds affecting the mylonitic fabric c) seem to have been grown at the end of the foliation development, as they show similar petrology in strongly sheared limbs as well as in the less deformed hinge zones.

STOP 2.3: **Banded granulites at the Cruceiro do Curutelo (San Andrés de Teixido)**

On either side of the village of San Andrés de Teixido we have the chance to see two different parts of the mafic unit within the Cabo Ortegal Nappe, in the literature is also referred to as the Bacariza unit. At the view point coming from the South, before the descent to San Andrés, we find the mafic and ultramafic members of this unit. At the view point to the north of San Andrés, at the Cruceiro do Curutelo, we find the member of the Bacariza richer in plagioclase. The small stone cross at this second view point marks the way of pilgrimage to San Andrés.

At this second view point, we find rocks that are rich in plagioclase, with bulk composition ranging from intermediate to acid. Although the rocks were equilibrated at high pressure and high temperature (Fig. 14), they are not in eclogite facies (Galán and Marcos 2000). Pyrigarnites are still seen, normally preserved in pods and surrounded by plagioclase-rich layers. In the literature, this is named as the layered or banded granulites.

Plagioclase has a substantial effect on the petrophysical properties of the host rocks where it grows. First because it lowers both P and S wave velocities, but also because it reduces seismic anisotropy (Fig. 16). Reported CPO in plagioclase in highly deformed rocks is not too strong, in the current case, substantially weaker than the CPO of clinopyroxene, the mineral phase that replaces (Fig. 16). Mafic compositions will produce limited amount of plagioclase, according to bulk composition constraints, but plagioclase can be extracted by mechanical processes and concentrated elsewhere, as it may be the case in the current stop.

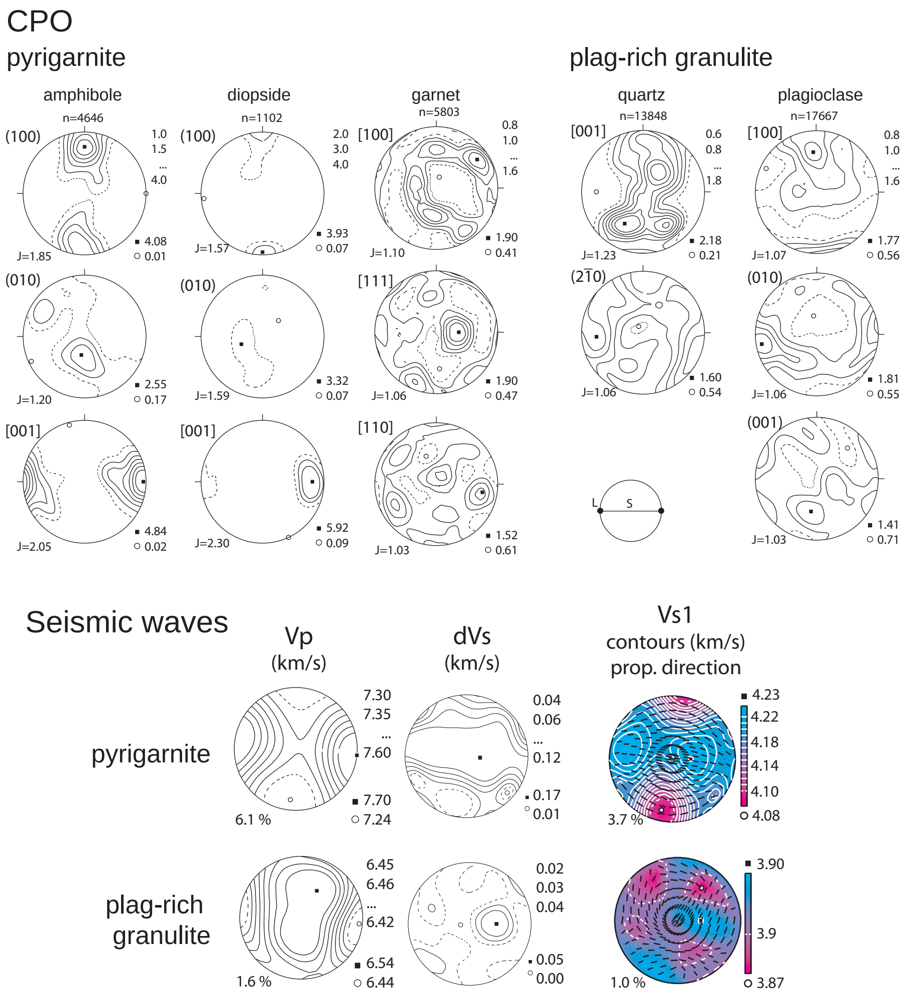

Fig. 16. CPO of rock-forming mineral phases in pyrigarnites and plag-rich plagioclase. Seismic velocities of two typical rocks from the Bacariza unit: ultramafic rocks (pyrigarnites) and layered granulites. Velocities are calculated from integrating cyrstallographic orientation of rock-forming minerals, mineral properties (elasticity tensor), mineral proportions and density at target conditions (in the figure 25ºC and 600 MPa). Modified from Llana-Fúnez and Brown, in review).

STOP 2.4: **Layered upper mantle: ultramafic rocks at the Vigía Herbeira**

Observation of ultramafic rocks around the Mt. Herbeira top, near the bus.

The ultramafic rocks represent the base of the sequence and are formed of more than 1000 m of alternating serpentinized peridotites (spinel-amphibol peridotites, harzburgites and dunites) and pyroxenites (websterites and clinopyroxenites) (Vogel, 1967; Girardeau et al., 1989; Gravestock, 1992; Moreno et al., 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 layered pyroxenites (ca. 300 m), dunites (alternating with centimetric to decametric Grt ± Spl pyroxenitic bands) (ca. 300 m) and harzburgites. The origin of the ultramafic rocks is controversial. Girardeau and Gil Ibarguchi (1991) suggest that the peridotites have strong affinities with residual oceanic peridotites and the pyroxenites are intruding magma segregates. More recently, Moreno et al. (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 is gradational (Galán and Marcos, 1997).

The websterites in the dykes are the rock type least affected by serpentinization, which is widespread in the ultramafics. They were sampled for seismic velocity measurements under pressure (Brown et al 2009). The velocities are slightly lower than expected, but with 7.55 and 7.95 can be regarded as "mantelic".

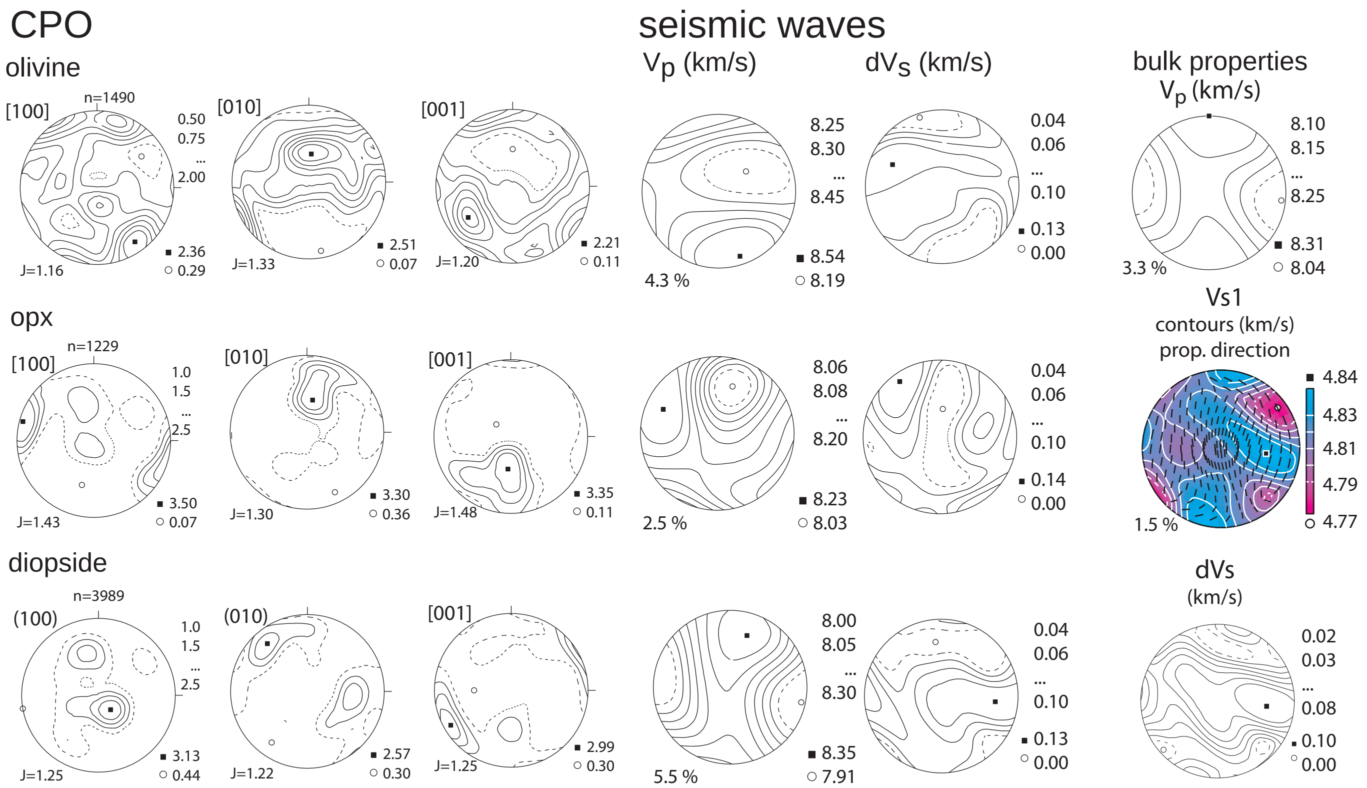

Fig. 16. Seismic velocities of two typical rocks from the Bacariza unit: ultramafic rocks (pyrigarnites) and layered granulites. Velocities are calculated from integrating cyrstallographic orientation of rock-forming minerals, mineral properties (elasticity tensor), mineral proportions and density at target conditions (in the figure 25°C and 600 MPa).

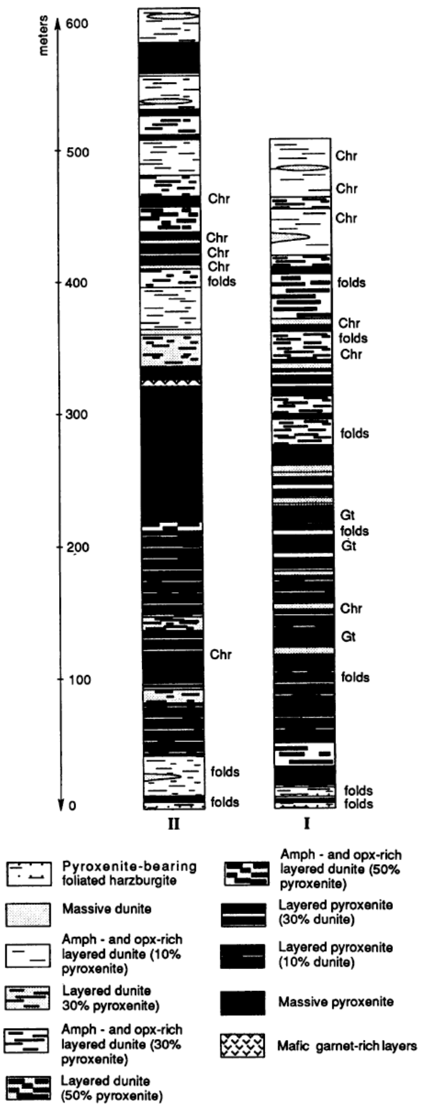

Fig. 17. Synthetic sections through the largescale pyroxenite layer of the Herbeira massif (Cabo Ortegal). Chr, chromite enrichments; folds, highly deformed rocks showing intrafolial folds; Gt, garnet-rich layer; opx, orthopyroxene; amph, amphibole. Section I corresponds to section A-B on Fig. 1; Section II has been picked up about 500 m north of Section 1. Taken from Girardeau et al (1991)

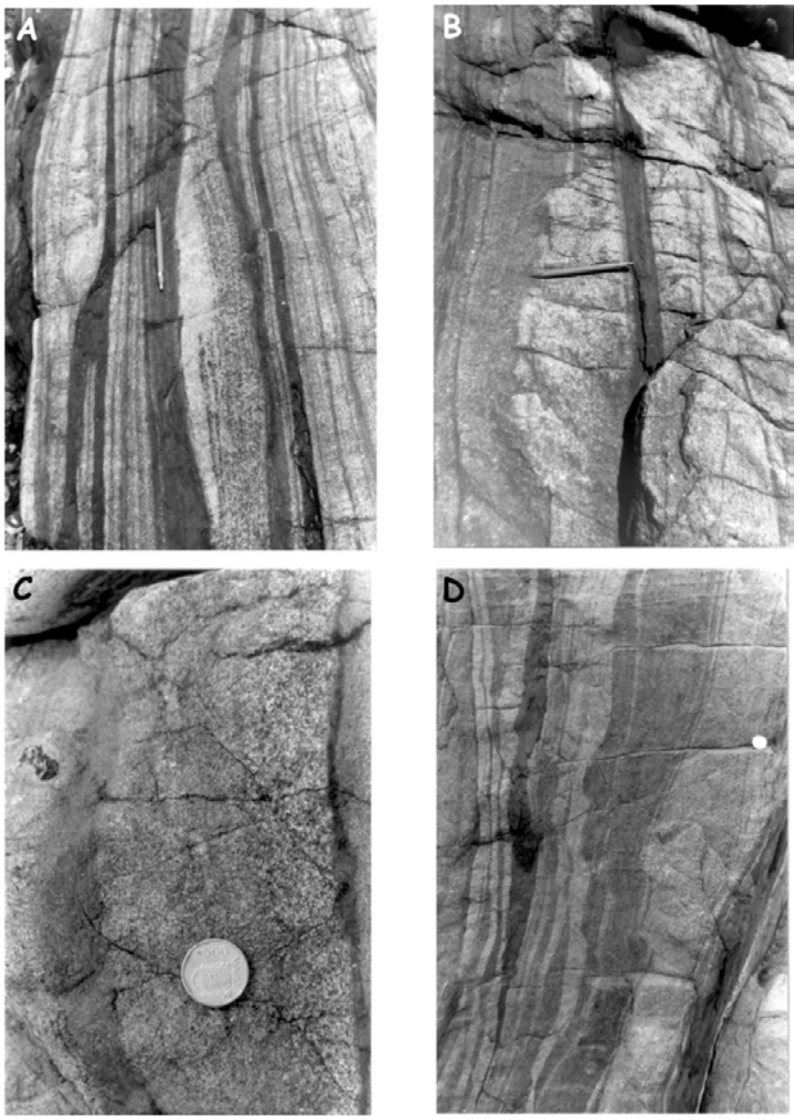

Fig. 18. A to D) peridotite (dark) and pyroxenite (light) cumulates in the ultramafic rocks of the Herbeira Fm. Normal faults similar to those shown in B and D usually are decorated with garnet.

STOP 2.5: **Eclogites at Punta dos Aguillons (Cabo Ortegal)**

Observation of massive eclogites along the road towards the Cabo Ortegal lighthouse (Punta dos Aguillons).

The Upper Member of the mafic rocks (100-200 m) is basically formed of eclogites, but also of fine-grained dark garnet amphibolites, retrogressed after eclogites-granulites (Vogel, 1967), especially in the northwestern outcrops of the Cabo Ortegal Complex. 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 have also been observed (Mendia, 1996). The Ky-bearing eclogites are mainly exposed as a thin band of less than 25 m thickness, along the contacts with the under- and overlying rocks. All the eclogites are more or less retrogressed in granulite- amphibolite facies conditions. From major- and trace element geochemistry Van Calsteren (1978) proposed for the eclogites a continental quartz-normative tholeiite parentage. However, most authors lately agree on a N-type MORB protolith at least for the common eclogites (Bernard-Griffiths et al., 1985; Peucat et al., 1990; Gravestock, 1992; Mendia, 1996).

The most common omphacite CPO in Cabo Ortegal is an S-type, with point maxima for b[010] and girdle distributions for c[001] (Fig. 19). Sub-maximum within this girgle distribution in c[0001] tends to coincide with the lineation in the rock (e.g. Fig. 19). However, not always the lineation can be clearly recognised in the field.

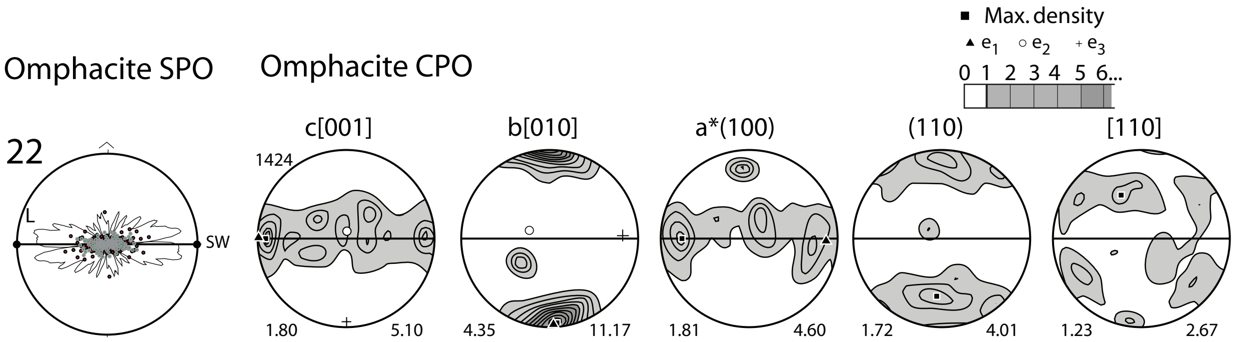

Fig. 19. Omphacite shape preferred orientations (SPO) and crystallographic preferred orientation (CPO) in an eclogite sample from the Cabo Ortegal lighthouse. The SPO was measured in a thin section cut directly and parallel from the blocks used for CPO measurements. CPO is represented as contoured pole figures of five relevant crystal directions or poles, equal area projection of lower hemisphere, contour levels 1, 2, . . . multiples of uniform distribution, shaded above 1, maximum position marked (black square). First eigenvector (related to largest eigenvalue) in c[001], b[010] and a*(100) pole figures is shown (black triangle), also second and third eigenvectors for c[001] (open circle) and b[010] (cross). Numbers below each figure indicate pole figure J (left) and maximum value (right); the numbers of orientation data and area covered are indicated for each sample. Sample 22 in fig. 5 in Llana-Fúnez et al. (2005).

Omphacite lineations and omphacite c[001] concentrations scatter in geographical coordinates (Fig. 20) and do not show a single preferred orientation that would indicate a predominant direction of extension during deformation in eclogite facies (Llana-Fúnez et al. 2005). An overall regional strain geometry dominated by coaxial strain is consistent with neighbouring rocks, particularly in gneisses (Fernández and Marcos, 1996).

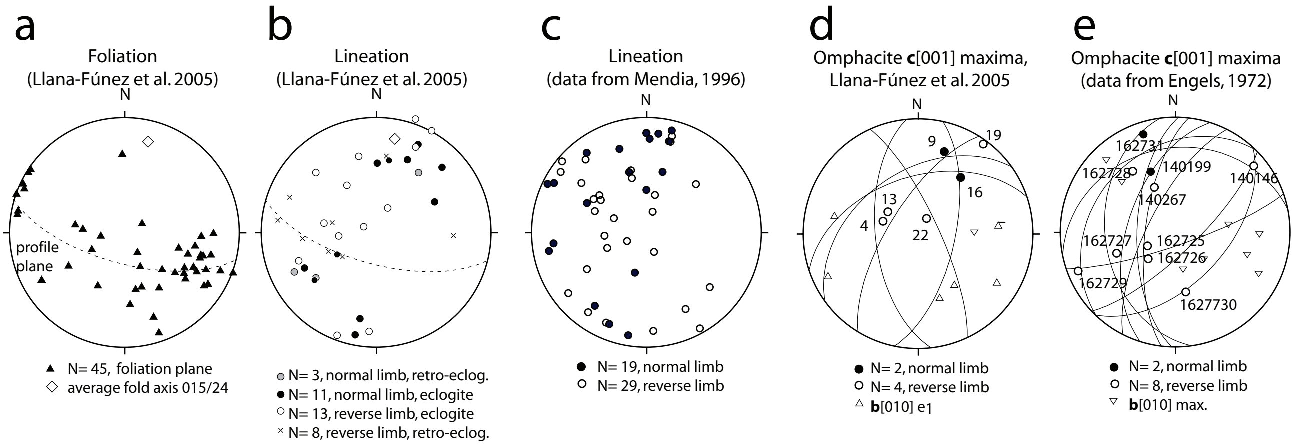

Fig. 19. Equal area projection in geographic coordinates of structural elements: foliation and lineation measured in the field, and fabric indicators derived from CPO patterns. Foliation poles (a) and lineations (b) in the Concepenido eclogites; indicated are the common fold axis (diamond) and the profile plane (short dashed great circle). The lineation data are distinguished according to their structural position and their defining mineral assemblage (grey and black circles are eclogite facies, open circles and crosses are HP granulite facies). (c) Lineation data in eclogites from Mendia (1996) taken from her maps for comparison. (d) For our samples we used the first eigenvector of the c[001] pole figures to represent the extension direction (balls) and the first eigenvector of the b[010] pole figures to represent the pole to the flattening plane (triangles). (e) Comparable data to (d) from Engels (1972), visual estimations after his fig. IV-3, where foliation poles are derived from b[010] point maxima (triangles) and balls mark maxima in the c[001] pole figures (note that these are not strictly identical to first eigenvectors).

Eclogites in Cabo Ortegal show low measured (Brown et al. 2009) and calculated seismic anisotropy (Ábalos et al. 2004), both in the case of P- and S waves (Fig. 21). It can be highlighted that the maximum direction of P waves is contained within the foliation but is consistently oblique to the lineation (Fig. 21).

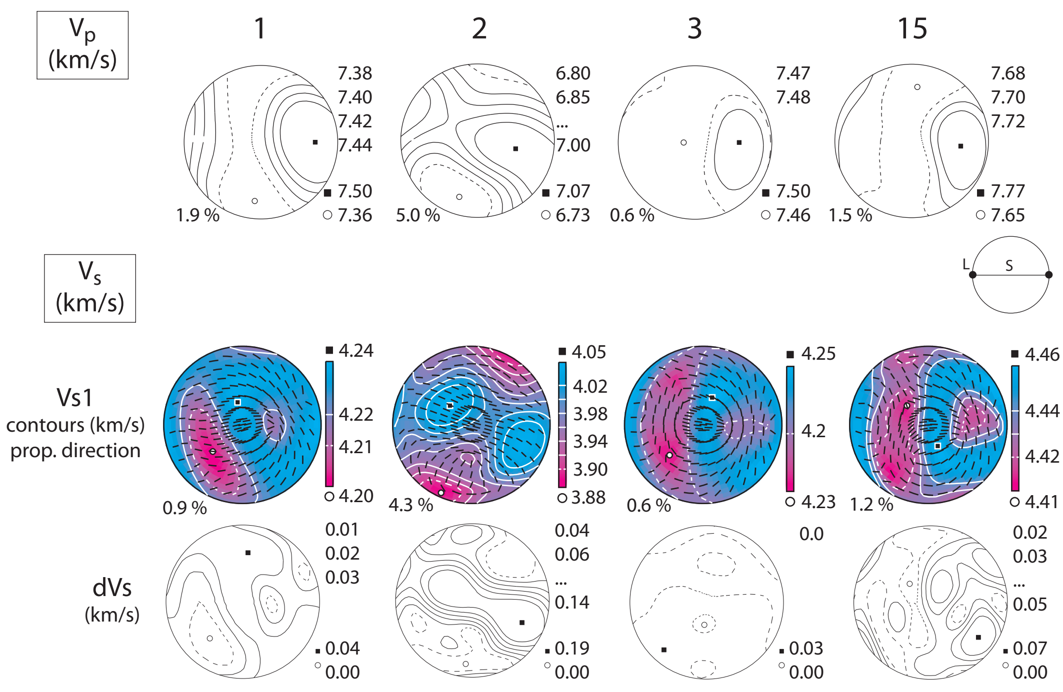

Fig. 21. Calculated seismic velocities in Concepenido eclogites from the Cabo Ortegal Complex. Procedure is similar to Fig. 16. Note that maximum velocity of P waves in all samples is oblique to the lineation, although still contained in the foliation plane. This is due to physical properties of dominant omphacite. Taken from Llana-Fúnez and Brown (in review).

STOP 2.6: **Quartzofeldespathic gneisses at the Masanteo peninsula**

The overturned sequence of quartzofeldespathic gneises progresses from metasedimentary gneisses to migmatitic gneisses toward the contact with the eclogites or the amphibolites boundaries of the Cedeira or the Capelada sheet, respectively (Map 1). Locally Vogel (1967) defined the so-called Chímparra gneisses within the Cedeira seet, or Banded gneisses within the Capelada sheet, both migmatitic gneisses; whereas the metasedimentary gneisses, named Cariño gneisses, only outcrops within the Capelada sheet. The migmatitic gneisses consist mainly of garnet and kyanite quartzofeldespathic gneisses that contain inclusions of eclogite, mafic granulites and calc-silicate rocks (Vogel, 1967; Gil Ibarguchi et al., 1990; Fernández, 1997). Composition is heterogeneous and thickness variable (> 200 m), with gradational contacts between layers locally enriched in garnet or amphibole. Intercalations of leucocratic garnet-bearing orthogneisses, coronitic metagabros and dioritic dykes evidence also an older magmatism. Migmatitic gneisses included variably retrogressed fine-grained eclogite blocks and boudins of different size and shapes. The PT-values (1.5 GPa and 725 °C) were obtained for the eclogite blocks as well as for the migmatitic host-gneisses (Gil Ibarguchi et al., 1990; Fernández, 1997) (Fig. 22).

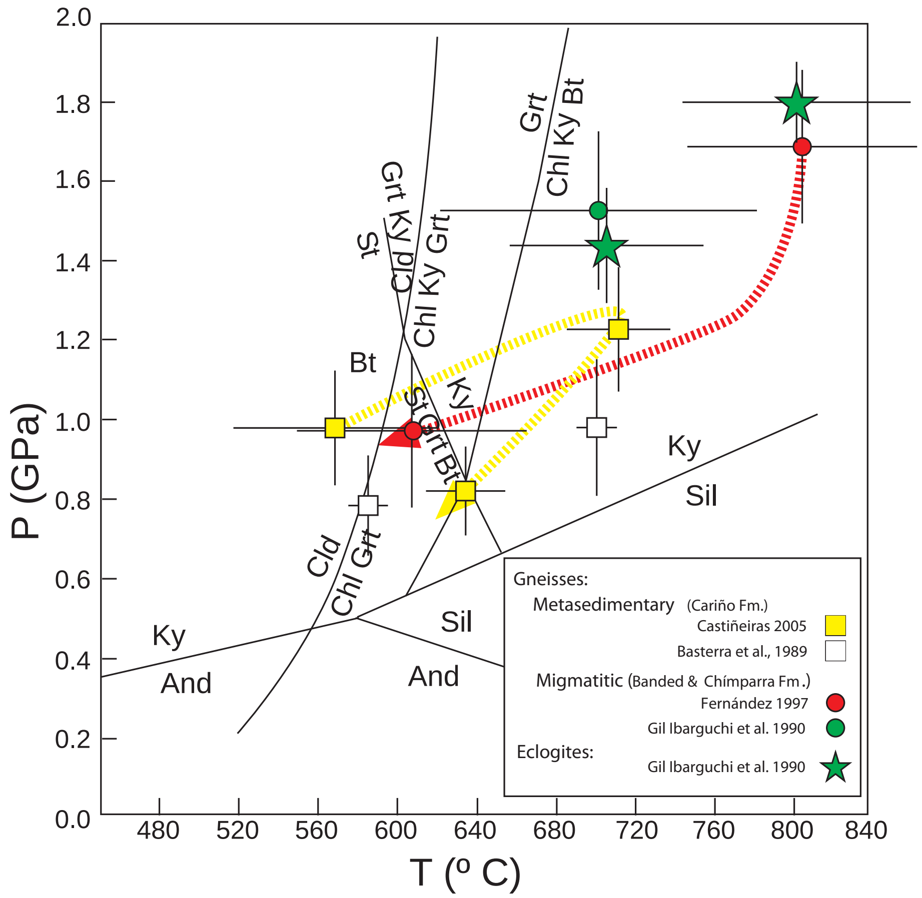

Fig. 22. P-T diagram is showing representative data obtained into the quartzofeldespatic gneisses and eclogites of the Cabo Ortegal nappe by different calibrations and authors. In addition, two possible P-T paths are suggested. (Fernández et al. in prep)

The term "metasedimentary gneisses" is used because the rocks only present minor evidences of partial melting and preserve an apparent sedimentary compositional layering consisting of pelitic and psammitic sequence intercalations. They also include boudins of strongly retrograded garnet-bearing amphibolites and within some of the pelitic layers can be found relicts of staurolite (Vogel, 1967; Castiñeiras, 2005). Almost all of the above evidences suggest slightly lower peak of pressure conditions for the metasedimentary gneisses according with the range of the values obtained (0.8-1.2

GPa and 600-700 °C) (Basterra et al., 1989; Castiñeiras, 2005) (Fig. 22). However, the geochemistry of both metasedimentary and migmatitic gneisses suggests the same sedimentary setting with an affinity to Phanerozoic greywackes (Fernández and Marcos, 1997).

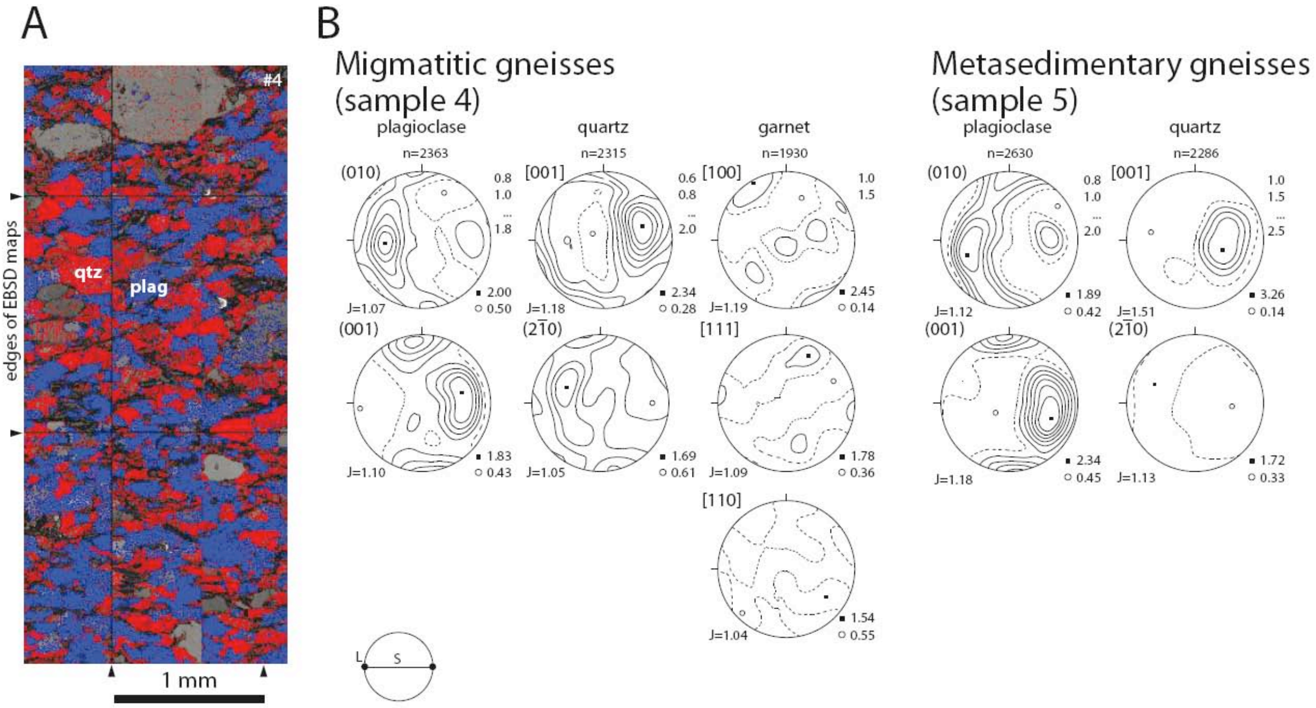

Fig. 23. Representative fabric (A) and crystallographic preferred orientation (B) of quartz, plagioclase and garnet, formed during the development of the main mylonitic foliation into the quartzofeldespathic gneisses (modifield from Llana-Fúnez and Brown *in review*)

Along Masanteo peninsula stop (stop 2.6) we can observe the structure at the oucrop scale and the structural relationships within the eclogites and quarztofeldespathic gneisses of the HP-HT unit of the Cabo Ortegal nappe. The main structure is a Variscan, a southeast-facing antiform, which refolds a recumbent syncline and the associated high-grade structures. This is overprinted by Alpine faults. In detail, the structure has been restored using the main mylonitic foliation as the structural reference. This fabric is a gneissic foliation defined by a compositional layering, migmatitic or metasedimentary in origin, heterogeneously deformed by a mylonitic process. The main foliation was initially developed under eclogite- HP-granulite conditions and re-equilibrated under amphibolite facies conditions. Fig. 23 shows two examples of quartz and plagioclase CPO patterns formed during this process. At the outcrop scale, mylonitic high-strain zones usually present intrafoliar folds, rotated porphyroblasts, and small inclusions of mafic and calc-silicate rocks. The lineation is mainly defined by the intersection between compositional layers and the mylonitic foliation, but also by the orientation of minerals and kinematic markers. In despite of a dominant N15E strike of both lineations (Fernández and Marcos, 1996, Abalos et al., 2003), the stereonet patterns always show a wide misorientation that are not consistent with the intensity of the mylonitic deformation (Fig. 24). In addition, the kinematic markers are generally symmetric, and where it is possible to deduce shear sense it can be found opposite senses of movements (Fernández, 1997), consistent with a general bulk coaxial deformation.

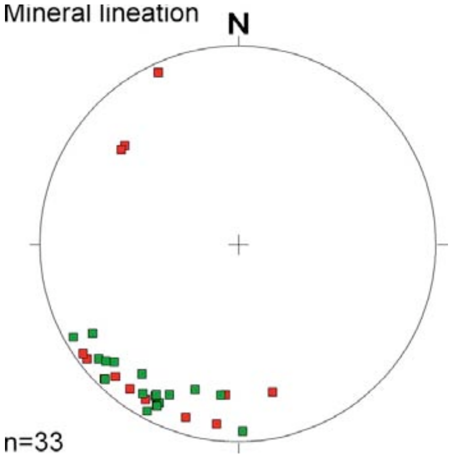

Fig. 24. Amphibole (green squares) and garnet (red squares) orientations defined over the main mylonitic foliation of the migmatitic gneisses of the Masanteo peninsula, near the contact to the basal eclogites and retrograded eclogites.

Blocks of eclogite within the migmatitic gneisses occur frequently along minor ductile detachments next the boundary between the migmatitic gneisses and the massive eclogite unit of Cabo Ortegal. The detachment contains minor structures, such as asymmetric rootless folds, rotated porphyroblasts, sheath folds, complex mantled structures and oblique shear bands (C and C' types) developed in a phyllonite matrix. According to the quartz and omphacite fabric patterns, almost all shear sense criteria related to this tectonic event indicate flattening kinematics.

A ductile thrust is well exposed along the northern shoreline of the Masanteo peninsula, juxtaposing migmatitic gneisses over a transitional layer of biotitic gneisses, near the boundary to the eclogites (Fig. 25).

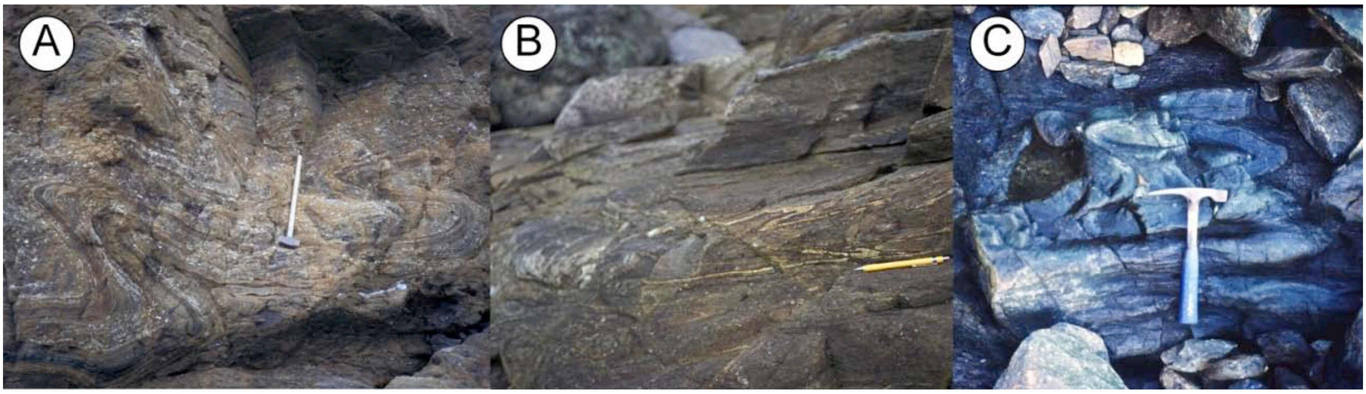

Fig. 25. structures related to ductile thrust formed between the eclogites and retrograded eclogites and the migmatitic gneisses of the Masanteo peninsula. (A) asymmetrical trend of metric folds facing SE; (B) sheath folds pointing to SE; (C) sheath folds pointing to NE.

Finally, the IP-HT metasedimentery gneisses are exposed toward the eastern boundary of the Masanteo peninsula. The main foliation is folded defining asymmetric minor folds related to a recumbent syncline facing to the SE. An extensional ductile detachment is displaced toward the NW the recumbent syncline over the migmatitic gneisses (Fig. 26).

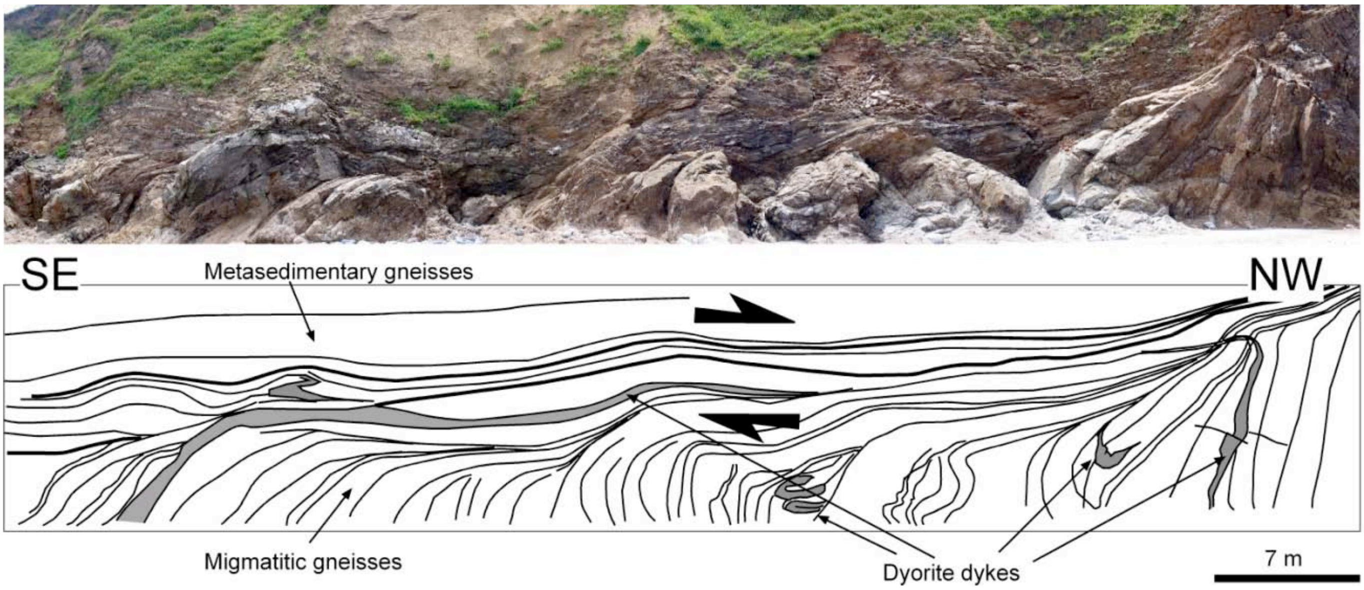

Fig. 26. Extensional detachment exposed on the cliff of the Serrón beach (Masanteo peninsula) that are juxtaposing the metasedimentary gneisses over the migmatitic gneisses within the Cabo Ortegal nappe.

DAY 3

Crustal-scale shearing: The Malpica-Lamego Line

S. Llana-Fúnez, A. Marcos, M.A. López-Sánchez

The Malpica-Lamego Line (MLL) is a deformation zone in the Variscan belt of NW Iberia (NW Spain and N Portugal) that runs parallel to the chain for at least 275 km, is delineated along 200 km of strike length by I-type granodiorite plutons (Fig. 27). The present level of exposure shows several types of structures, from micro- to macro-scale, formed at mid-crustal levels (Fig. 28). The restoration of the lithostratigraphy and the previous structure at both sides of the shear zone implies a significant dip-slip displacement to fit the distribution of geological markers and suggests a tectonic event prior to the strike-slip deformation, and thus a poly-phase history for the MLL (Llana-Fúnez and Marcos 2001).

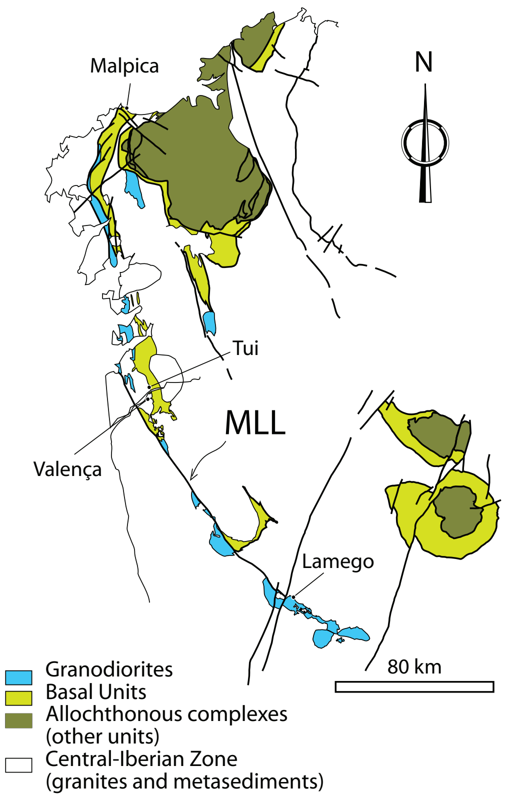

Fig. 27. Schematic geological map of Western Iberia illustrating the extend of the Malpica-Lamego Line.

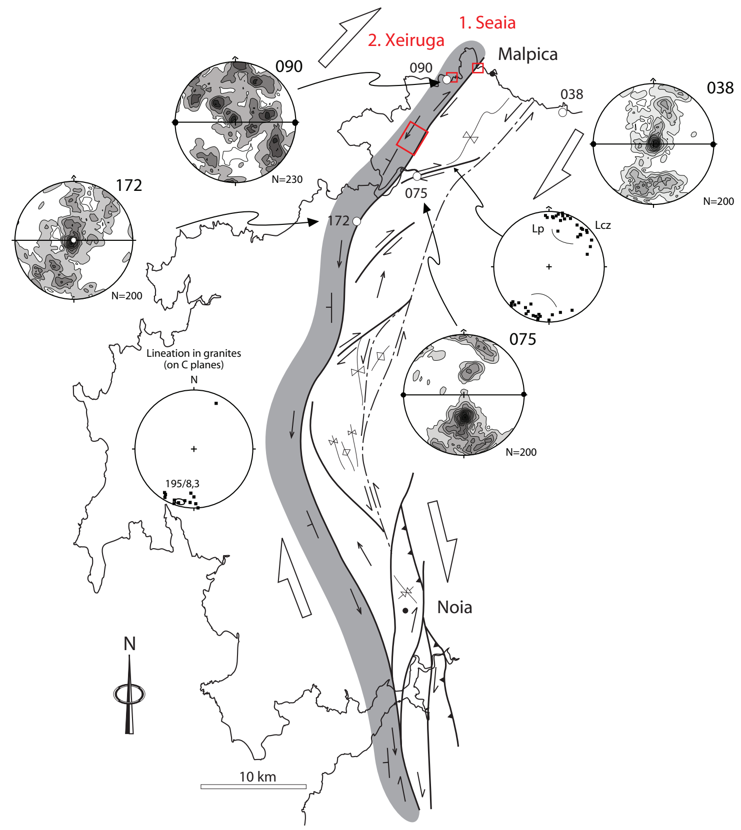

Fig. 28. Sketch of structures associated to the MLL in northern Spain. Boxes in red indicate the location of stops in this trip. Points in the stereonets show the trend of mineral lineation in granites (left) and in oblique shear zones (right). Contoured stereonets show distribution of c-axis in quartz from samples in the MLL (090 and 172) or associated to MLL structures (075). Sketch map is taken from Llana-Fúnez (2001).

The MLL affects previous structures by which high pressure and ophiolitic rocks were exhumed and emplaced on the Iberian plate during earlier deformation phases (Fig. 27). Correlation and reconstruction of the stratigraphy of these sheets or tectonic units at both sides of the shear zone allows a preliminary estimate of the accumulated vertical and horizontal offsets after the tectonic activity of the fault. The value of the separations, of crustal-scale proportions, reaches a maximum 15 km of vertical offset that decreases gradually to the south. The structural record found in the rocks indicates a strike-slip regime that, in general, does not fit the geometry of the offsets. We suggested that the MLL went through two different stages during the same orogenic cycle: a first dip-slip episode, a reverse faulting event, overprinted by a later strike-slip reactivation (Llana-Fúnez and Marcos 2001).

STOP 3.1: **Praia de Seaia (Malpica)**

We leave Malpica on the AC-4307 towards Ponteceso. In the town of Seaia we turn right following the signs to the school, the capilla de San Andrés and the picnic area at the top of the hill. If going by bus, we park once we reach the school. If going by car we turn right towards the car park by the beach (to our right). The car park is only a few hundred metres away from the school.

Low tide is predicted at 17:04 for the 5th September 2011.

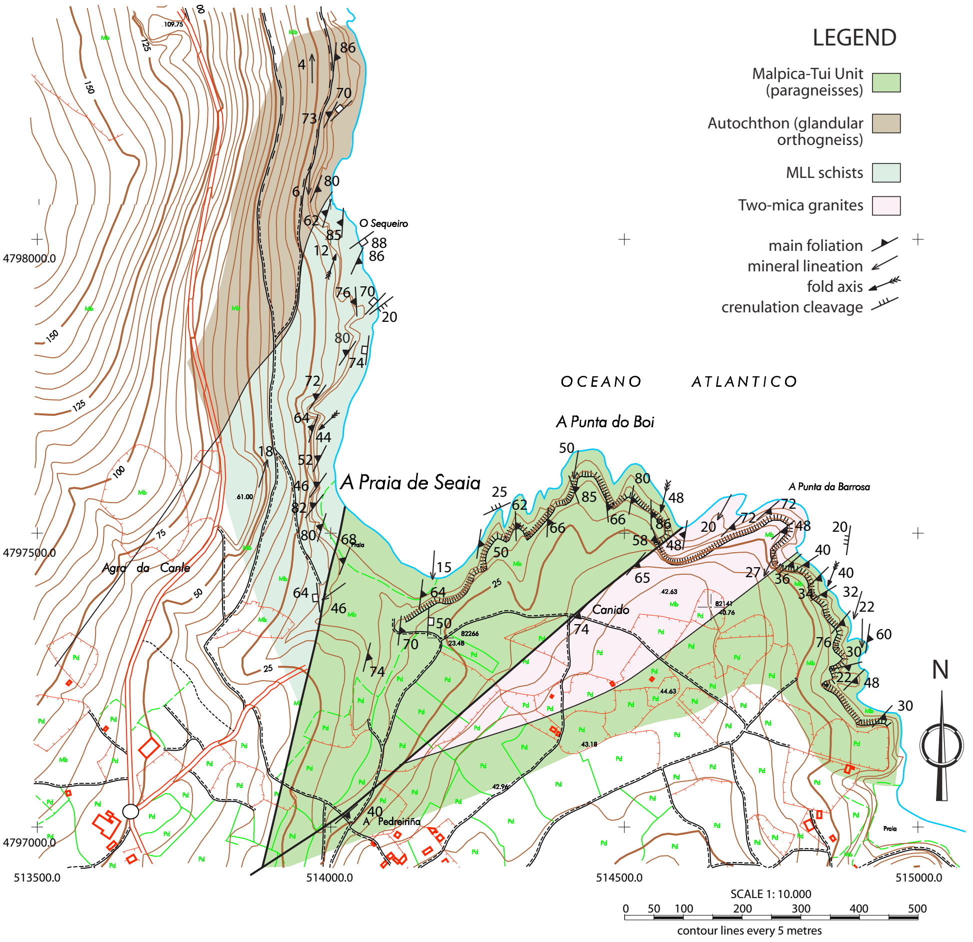

Fig. 29. Geological map of the coastline around Seaia beach (Llana-Fúnez 2001).

The coast line around Malpica is the northernmost exposure of the MLL and perhaps the best to observe deformation structures associated with the activity of this major crustal-scale shear zone because exposure is nearly continuous. Due to late NW directed faults, two sections through the MLL deformation zone are seen, in the Seaia beach (stop 3.1) and in the Seiruga beach (stop 3.2) (Fig. 28). In less than a kilometre across the deformation zone we can walk from the footwall (here, the Malpica-Tui unit) to the hangingwall of the shear zone (here, the authochthonous of the complexes).

The MLL is defined along much of its length in the Spanish segment by subvertical and highly deformed schists, on map view not exceeding in general the km in width. In the case of the Seaia beach, the transit occurs within hundreds of metres and it fairly well exposed (Fig. 29). The core of the shear zone in the Seaia beach, west of Malpica, is defined by phyllonites and strongly deformed schists and paragneisses (Fig. 30).

On the eastern side of the beach, metasediments from the basal units, belonging to the Malpica-Tui unit, are exposed (Llana-Fúnez & Marcos 2002). These are albite-rich quartzo-feldspathic paragneisses. Significant amount of albite blasts enclose a mineral assemblage that in other similar units of the allochthonous complexes of NW Iberia indicate equilibrium conditions at high pressure and low temperature (Arenas et al. 1995, Martínez-Catalán et al. 1996).

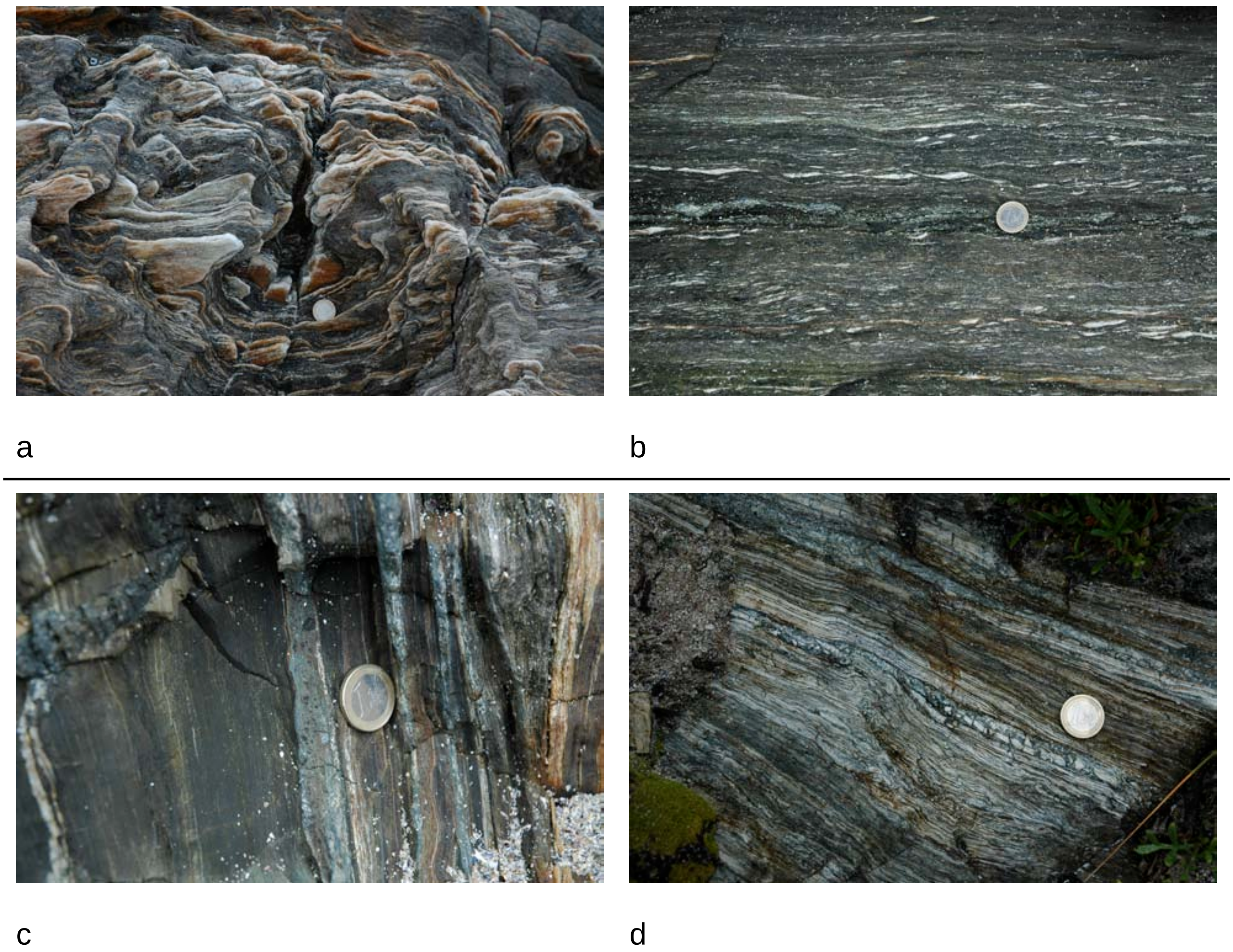

Fig. 30. (a) The schists exposed at the contact between the Malpica-Tui unit and the rocks further to the west, belonging to the authochthon, show a significant amount of quartz veins. The veining may be evidence that the deformation zone associated with the MLL was rich in fluid activity during deformation. (b) The core of the deformation zone is characterised by phyllonites, the amount and thickness of veins is also reduced to mm. In (b) dextral kinematics is indicated by shear bands, affecting the veins. (c, d) Development of cataclasites in the core of the MLL deformation zone. (c) shows at least three separate cataclasites in relation to three separate fracturing events of the phyllonites (green matrix). (d) shows a cataclastic band of similar characteristics but developed in rocks rich in quartz and feldspar.

To the Western side of the beach, subvertical well foliated and strongly deformed schists can be seen, together with some fine grained quartzo-feldspathic gneisses. These rocks show abundant veins which in cases constitute a significant fraction of the rock (Fig. 30a). Adjacent to the schists with the veins, there are a few metres of phyllonites, where vein thickness is very much reduced to few mm and where grain size reduction is also very significant (Fig. 30b). The rocks with finer grain size and with higher content in quartz and feldspar do show evidences of brittle fracturing during deformation and several strands of cataclasites have been found (Fig. 30c). The cataclasites of mm in width present fragments of host rock surrounded by a fine-grained and green to dark matrix (Fig. 30c,d).

The aim of the stop at the Seaia beach is to walk across the deformation zone associated to the MLL, starting in the footwall, the Malpica-Tui unit, and finishing close to the hangingwall (the glandular gneisses). The kinematics of the shear is quite clear in this locality, so the main topic for discussion will be the role of fluids during deformation.

STOP 3.2: **Praia de Seiruga (Malpica)**

We return to the AC-4307 towards Ponteceso and continue driving until we reach the town of Beo. Here we stop at the end of the village. To access the cliffs along the coastline we need to follow the last paved road we find on the right hand side (see Fig. 31). It will be narrow, this road can be done by car, but not by bus. The walk will be 1-2 km until we reach the Punta Galiana, the first headland east of the Xeiruga beach. There are no tide constrains for this stop, alternative for 3.2 if the tide is up.

We then returned to the AC-4307 towards Ponteceso. At the first junction we turn right. The road will go uphill for a few hundred metres towards Barizo. At a relative sharp left turn we take the little street going downhill in the direction of the beach (which we will see to our right). Low tide is expected at 17:04 for the 5th September 2011.

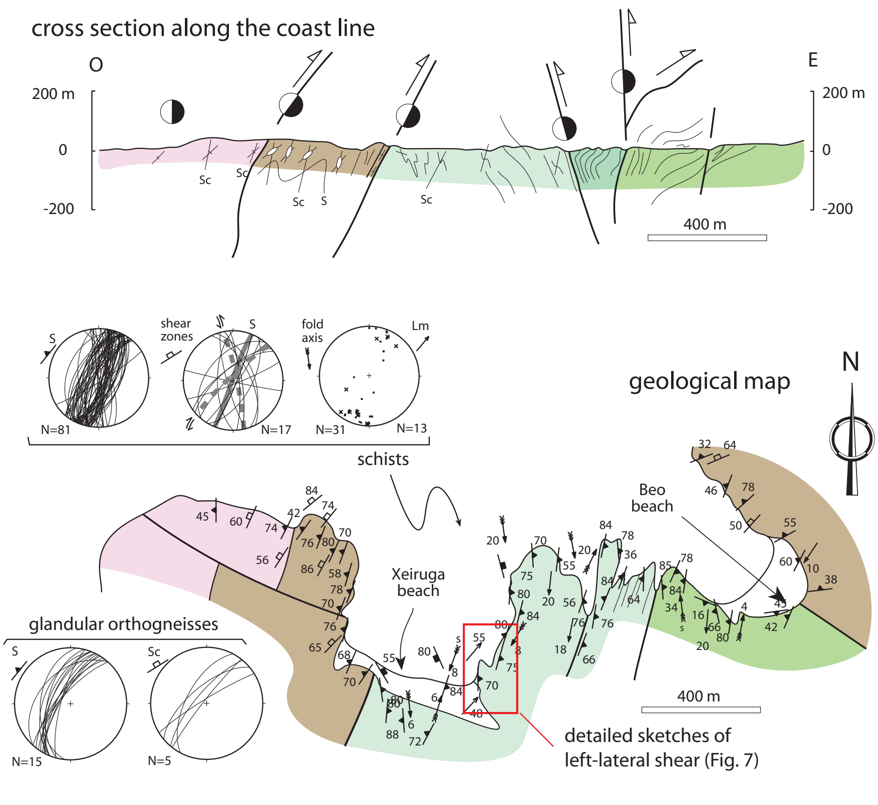

Fig. 31. Geological map of the coastline around the Seiruga beach. Based on Llana-Fúnez and Marcos (2001). Same legend as in Fig. 29.

The section at the Esteiro or Seiruga beach exposes again the main part of the shear zone, although in this case the high strain rocks (phyllonites) are not seen. To the west, we find the glandular gneisses of San Adrian and the two mica granites and to the east after few hundreds of metres we find the albite-rich metasediments of the Malpica-Tui unit. The latter contact is neat, since the paragneisses stand out from the mostly schistosse rocks of the MLL deformation zone, however no obvious change in strain is seen across the contact.

Along the cliffs to the east of the Seiruga beach there are some structures of interest. Some of the shear zones affecting the schists have a left-lateral sense of movement, opposite to the main sense of shear for the MLL (Fig. 32). Unlike the left-lateral structures that can be found to the east of the Malpica-Tui unit which show increase fluid activity and a clear brittle character, here they are similar to the dextral structures. The other aspect of interest in the area are the biotite-rich granitoids which we related in a previous work with the granodioritic intrusions that characterise the MLL further south (Llana-Fúnez & Marcos 2001). These are found at the Punta Galiana (Fig. 33).

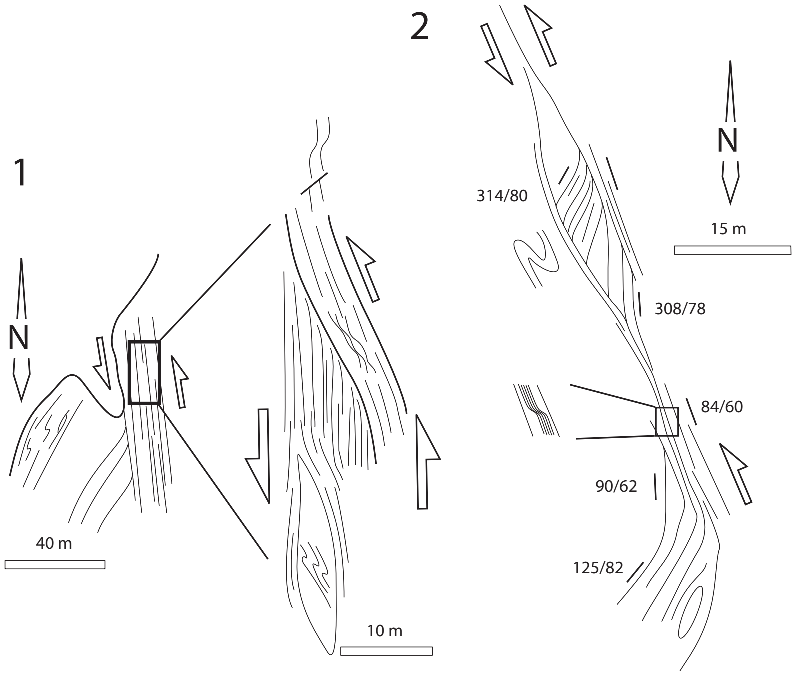

Fig. 32. Left-lateral shear zones affecting schists at the core of the MLL deformation zone, east of the Seiruga beach.

Fig. 33. Dykes and irregularly shaped intrusions of granodiorite can be found around Punta Galiana. Further south they may be related with larger bodies of granodiorite which intrude along the MLL.

The hightlight of the stop at Seiruga, to the East (particularly around Punta Galiana), are the granitoids intruding the schist at the core of the MLL deformation zone (Fig. 33). Here the intrusives seem to post date deformation. Further south, granodiorites intruding along the MLL show a magmatic fabric that may be indicative of forceful intrusion, indirectly giving a time constrain for late tectonic activity (this is discussed further in Llana-Fúnez & Marcos 2001). Within the schists, several generations of foliations sets are superimposed one on top of another without apparent change in metamorphic conditions (see field sketches in Fig. 32). Some of this overprints have left lateral sense of shear, opposite to the main in MLL.

To the Western end of the Seiruga beach the schists are in contact with glandular gneisses and two mica-granites. The latter intrusives can appear either deformed or with only a slightly developed fabric.

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