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Anatomy of the Mondoñedo Nappe basal shear zone (NW Spain)

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Abstract—This paper analyses a major shear zone from the Iberian Hercynian belt which forms the basal thrust of the Mondoñedo Nappe. The shear zone developed by ductile deformation under amphibolite facies metamorphic conditions and later by brittle–ductile deformation in greenschists facies. Folds in the shear zone are asymmetric, very tight, 1C or similar class and frequently developing sheath geometries. The sheath folds originated by non-coaxial flow superimposed on earlier irregularities. The fabric of quartzitic rocks in the shear zone changes from bottom to top from ultramylonites through blastomylonitic rocks to non-mylonitic tectonites. *c*-axis fabrics vary across the shear zone, but show a dominant monoclinic symmetry. The blastomylonitic rocks include the fabrics representing the highest temperatures. The main foliation of the schists results from flattening of an earlier foliation, recording occasional microfolds. The use of different kinematic criteria has allowed an analysis of their validity as well as an assessment of movement direction towards the foreland of the orogen.

INTRODUCTION

THE Mondoñedo Nappe basal shear zone is well suited for the study of problems arising in this type of structural association, and has already been the subject of some previous papers (Bastida & Pulgar 1978, Pulgar 1980, Martínez-Catalán 1985, Bastida *et al.* 1986). The shear zone, 3.4 km wide, is located in the northern part of the Iberian Hercynian belt (Fig. 1), presenting a continuous outcrop with strong ductile deformation all along the Cantabrian coast. It is bound at its base by a major thrust developed in brittle–ductile conditions during the last stage of the shear zone deformation history. The shear zone outcrops again, further to the east, in an area closer to the front of the Mondoñedo Nappe (Fig. 1). Here the shear zone is characterized by a brittle–ductile deformation zone not wider than 100 m, above the basal thrust.

GEOLOGICAL SETTING

The shear zone is the lower boundary of a large thrust sheet, the Mondoñedo Nappe, situated in the western part of the Westasturian–Leonese zone, a major tectonic zone in the hinterland of the Iberian Hercynian belt. The Mondoñedo Nappe unit is made up of a stack of recumbent, almost isoclinal, large amplitude E-facing D_1 folds (Fig. 1) with an associated tectonic foliation (S_1). These folds are affected by the nappe basal thrust and the associated shear zone. This shear zone is important in the western part of the nappe, where a large number of shear related E-facing minor folds (D_2) are developed. In this part, S_1 has been flattened (S_{1+2}), crenulated or replaced by a new mylonitic foliation (S_2). Finally, the whole nappe has been folded into a gentle synform in the east and an antiform in the west (D_3), both with a subvertical axial plane and a large wavelength. D_3 minor folds and S_3 crenulation cleavage occur locally. D_3 folds homoaxially overprint D_1 folds producing a Type 3 interference pattern (Ramsay 1967) that can be seen both on the map and in cross-section (Fig. 1).

Uplift due to the D_3 major antiform has resulted in exposure of the thrust and the basal shear zone around the Xistral semi-window (Fig. 1) where amphibolite facies metamorphism was reached (Martínez-Catalán 1979, 1980). It is in this western outcrop of the shear zone, in the section between the coastal localities of Burela and Foz (Fig. 1), that most of this work was focused.

GEOLOGY OF THE CROSS-SECTION

The rocks affected by the shear zone as well as those of the Xistral semi-window, belong mainly to the Cándana Group (Lower Cambrian) (Walter 1966, 1968). Precambrian rocks, consisting of micaschist with common sandstone intercalations and porphyroids outcrop just at the base of the nappe. The Cándana Group consists of four units, only two of which are affected by the shear zone: the lower Cándana formed by coarse-grained feldspathic sandstones and quartzite with intercalated schists, and the middle Cándana formed by micaschist, commonly with some sandstone, and minor black slate, carbonatic and amphibolitic layers.

The macroscopic structure shown in Fig. 2 consists of several very tight D_1 folds, downward-facing due to their location in the western limb of a large gentle D_3 synform. The basal thrust, also in a tilted position, can be seen in the western part of the limb. The ductile deformation of the shear zone is shown by an association of minor structures that developed in the nappe footwall (shaded area in profiles of Fig. 1).

The general features of metamorphism and its chronological relationship with the deformation in the Mondoñedo Nappe have been pointed by out Bastida & Pulgar (1978), Martínez-Catalán (1985) and Bastida *et*

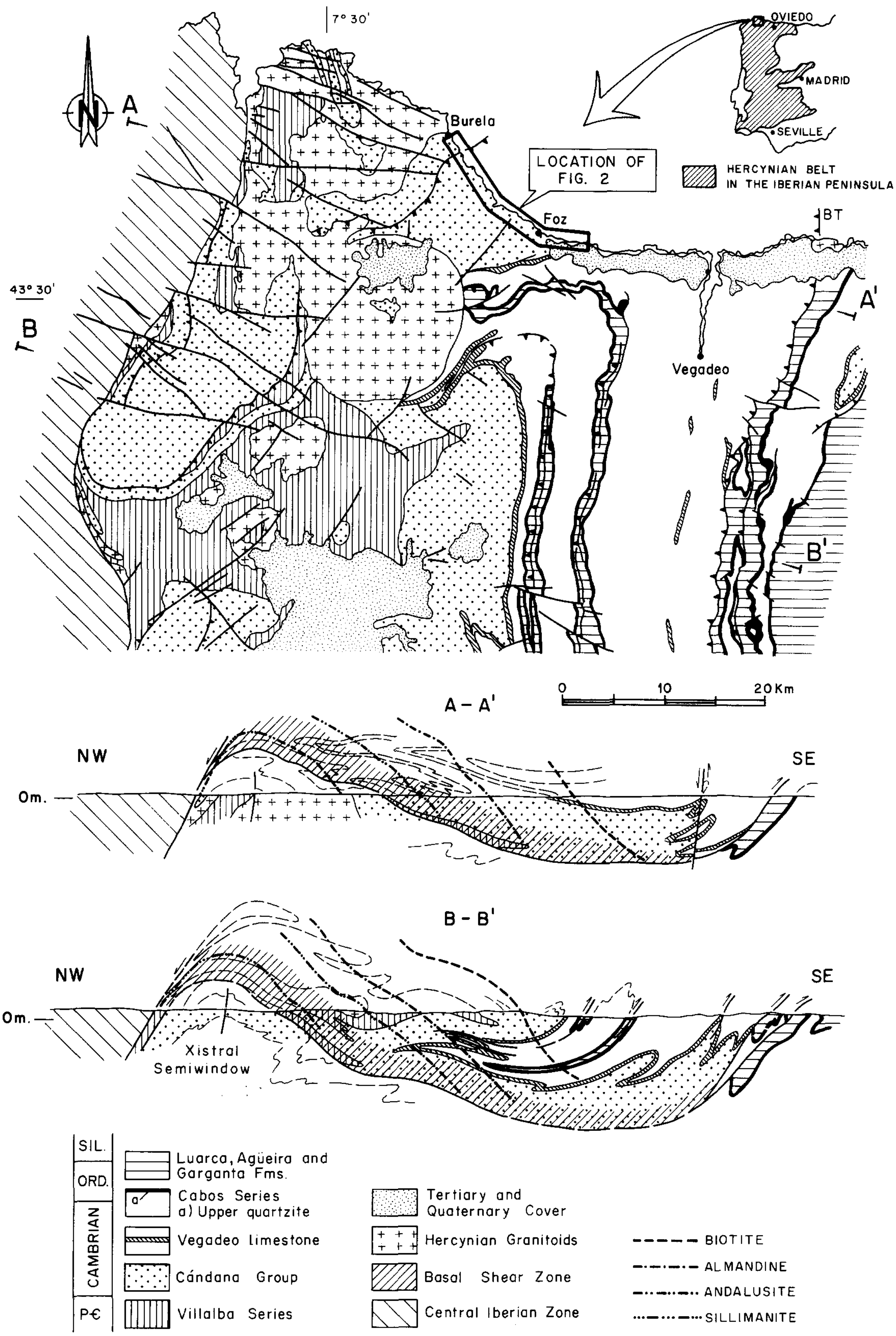

Fig. 1. Geological map and cross-sections of the Mondoñedo area, according to Bastida *et al.* (1986). BT, basal thrust.

al. (1986). The metamorphic zones that occur on the coastal region of the Mondoñedo Nappe unit are shown in Fig. 1. The shear zone sector is situated within the andalusite and almandine zones (Fig. 2). The andalusite zone partly overprints the staurolite zone with relict staurolite occurring locally as residuals inside andalusite porphyroblasts.

Three metamorphic episodes can be differentiated in the area of Fig. 2 (Bastida *et al.* 1986): M_1, with a paragenesis containing staurolite and garnet (medium pressure); M_2, with andalusite (rise in temperature and perhaps a pressure drop); and M_3, presenting retrograde metamorphism to greenschists facies. M_1 is pre-D_2. The rise in temperature during M_2 is coeval with ductile deformation in the shear zone (M_2 early syn-D_2). M_3 is late- to post-D_2 (end of ductile deformation in the shear zone and development of the basal thrust) and is related to the drop in pressure and temperature during and after uplift of the nappe rocks.

STRUCTURES IN THE BASAL SHEAR ZONE

The strong ductile deformation associated with the shear zone gave rise to a large number of small-scale D_2 structures, and pre-existing structures (D_1) were modified. D_2 structures consist of folds, foliations and lineations as well as microfabrics, which will be discussed separately.

Folds

Folds are very abundant in the western part of the section, where they are mainly concentrated in four narrow discrete zones (Fig. 2): Punta das Cabras, Punta Riomar and two others with very similar characteristics, that will be considered together (Punta Morago). All three zones are situated along the normal limbs of major D_1 folds. This position is likely to have favoured the development of D_2 folds, since the reverse limbs were likely situated in a stretching position.

The lithology of the folded zones consists of quartzite, feldspar sandstone and micaschist. The quartzite and the sandstone occur in thin layers. In Punta das Cabras and Morago these rocks are more abundant than the schists, whereas in Punta Riomar the amount of quartzite and sandstone is often equivalent to the amount of schist. These lithologies may have favoured the development of D_2 folds, contrasting with the massive quartzite that occurs in the lower Cándana, and the Precambrian or middle Cándana schistose zones.

The folds are E-facing and commonly occur as strongly asymmetric anticline–syncline couples. Small fold trains also occur locally. Folds show varying limb thickness. The D_2 folds are not associated with major folds and preserve the same asymmetry regardless of their location within the D_1 folds. The length of the short limbs ranges in most cases between 20 and 50 cm, measured in the fold profile along the layer between two adjacent hinges.

Fold hinges are SE- to SW-plunging (Fig. 2), with a increased dispersion of plunge directions towards the

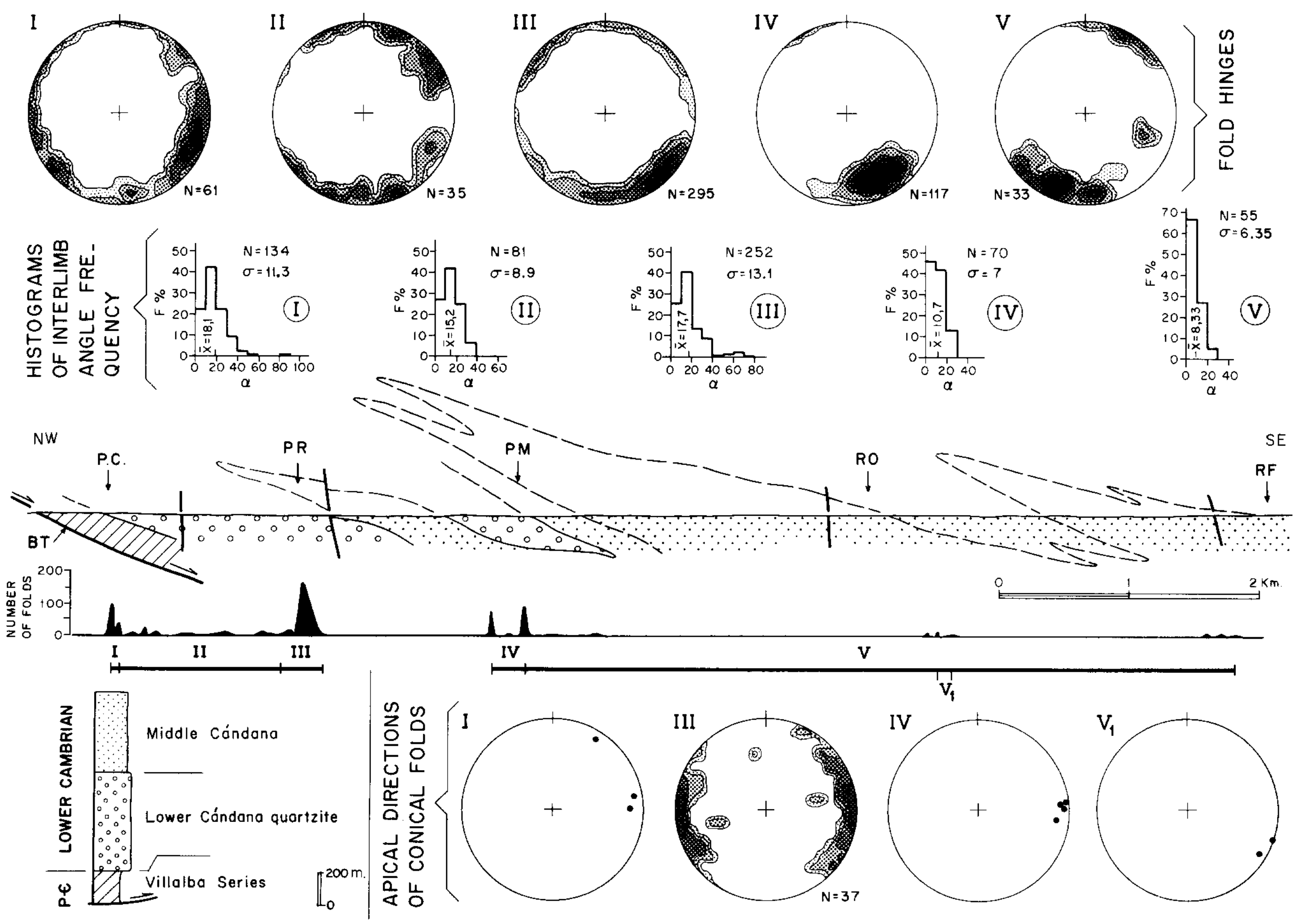

Fig. 2. Cross-section of the Mondoñedo Nappe basal shear zone, showing orientation of D_2 folds, histograms of interlimb angle frequency, number of folds and apical directions of conical folds along the section. Contours: 1, 2, 4 and 8%. Abbreviations: PC, Punta das Cabras; PR, Punta Riomar; PM, Punta Morago; RO, Ría d'Ouro; RF, Ría de Foz; BT, basal thrust.

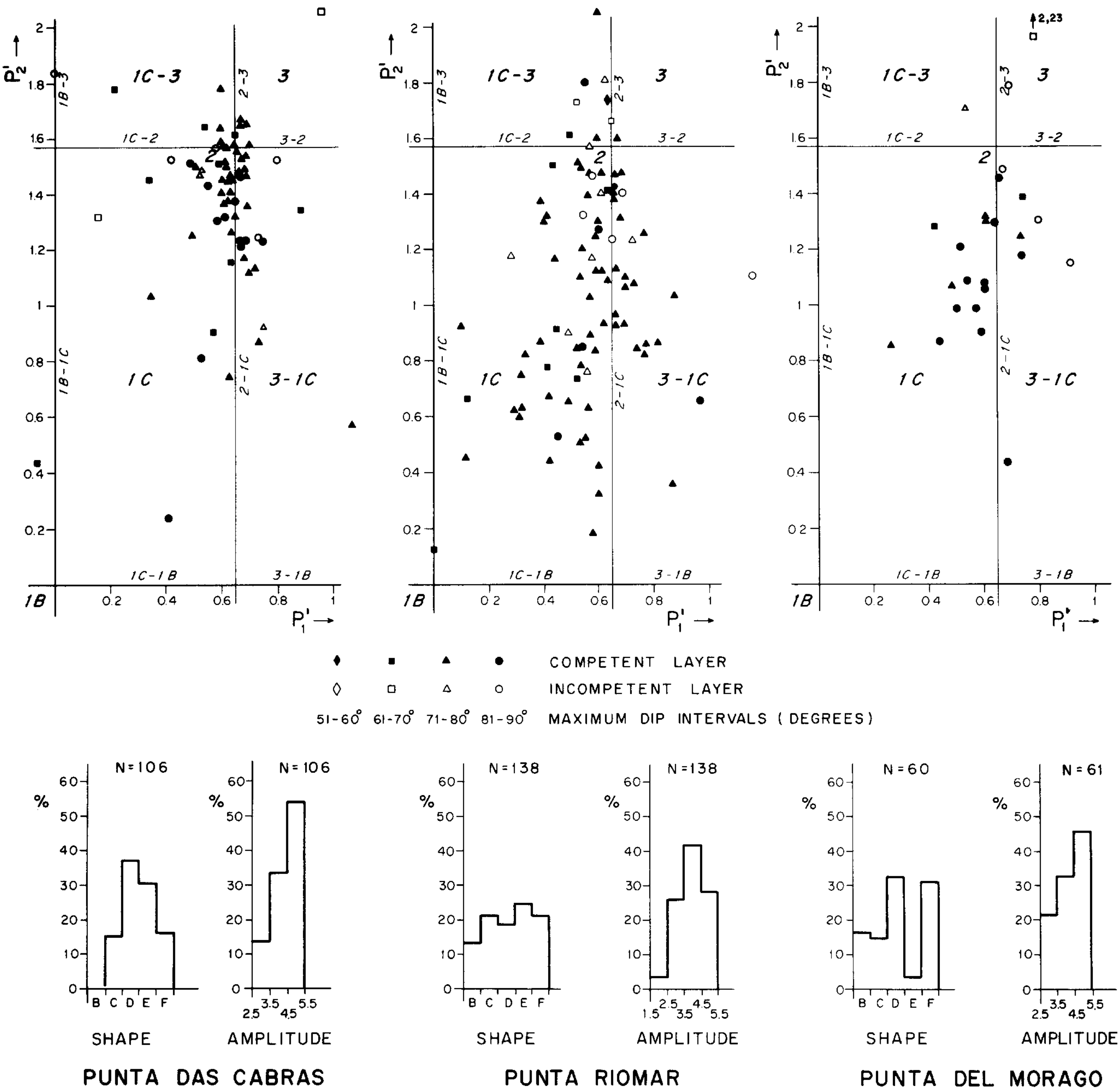

Fig. 3. Classification of fold morphologies on Bastida (1993) diagrams (above) and histograms for Hudleston's (1973) parameters (below) in the three main D_2 fold localities of the shear zone.

base of the shear zone. Curved hinges are common and give rise to conical or sheath folds and to eye-shaped structures. The interlimb angle distribution shows that D_2 folds are very tight to isoclinal (Fig. 2). The fold geometry has been analysed using Hudleston's (1973) classification (Fig. 3). Amplitudes show a higher mode in Punta das Cabras and Punta Morago than in Punta Riomar. There is a great diversity in fold shape. The analysis of fold layers has also been carried out using Ramsay's method (1967), modified by Bastida (1993) (Fig. 3). This analysis shows that in Punta das Cabras the large majority of folds are subsimilar, without geometrical differences between adjacent folded layers of different lithology. Class 1C folds are most common in the other two localities, with alternation of class 3 and 1C folds that imply a difference in mechanical behaviour between layers.

Several mesoscopic Type 3 fold interference patterns (Ramsay 1967), pre-dating D_3, have been recorded in Punta Riomar. At this locality, two superimposed generations of microfolds also deform the previous S_1 foliation. This suggests that progressive superimposed folding took place during the evolution of the shear zone.

Folding mechanisms

The geometry of the folds, together with the type of deformation that affected the shear zone suggest that a significant part of the fold evolution resulted from passive amplification of pre-existing folds or irregularities in non-coaxial flow. Evidence for a passive mechanism operating during at least the later stages of D_2 folding is provided by the broad similarity between quartz c-axis orientation patterns in limbs and hinges of folded quartzite layers. An example is shown in Fig. 4. The presence of sheath folds provides further evidence of non-coaxial progressive deformation (Quinquis *et al.* 1978, Cobbold & Quinquis 1980, Ramsay 1980, Skjernaa 1989). The initiation of D_2 folding would have been enhanced by the orientation of S_0 in the shortening field of deformation, due to obliquity between S_0 and the shear plane (Ramsay 1980, Martínez-Catalán 1985).

The origin of pre-existing irregularities necessary for fold amplification in shear zones has been a matter of discussion (Berthé & Brun 1980, Cobbold & Quinquis 1980, Lister & Williams 1983, Platt 1983, Ghosh & Sengupta 1984). The lack of periodicity in the case of anticline–syncline couples (mainly present in Punta das Cabras) suggests an initial mechanical instability which does not necessarily derive from a buckling mechanism. Rather, it could derive from an instability in the flow, producing local increases in the shear deformation rate (Lister & Williams 1983, Platt 1983). If buckling had taken place it would have stopped operating at a very early stage since its evolution, always implying propagation by wave lateral aggregation, should have involved a periodicity (Biot *et al.* 1961, Biot 1965, Cobbold 1975). In those cases where there is a periodicity and where the behaviour of competent and incompetent rocks is different (mainly Punta Riomar), buckling seems to be the fold initiation mechanism. However, the geometry of the folded layers suggests a remarkable kinematic amplification. Therefore, it seems unlikely for buckling to have given rise, in this zone, to very tight or isoclinal folds as described by Ghosh & Sengupta (1984) in their study of folds of the Kolar Gold Field (India). These observations suggest slightly different folding evolution patterns for the different localities of the shear zone. Passive evolution seems to have played an important role in Punta das Cabras, while in Punta Riomar buckling must have been more important. Punta Morago appears to be intermediate between the two. The differences in evolution may be conditioned by lithological differences and, perhaps, by small differences in pressure and temperature, which are higher towards the base of the nappe and decrease as the deformation continues.

As regards fold hinge orientations, the asymmetric character of most projections (Fig. 2) with respect to the direction of shear suggests a slight initial obliquity between S_0 and the shear plane (Martínez-Catalán 1985). The gradual approach of the projection maxima in sectors IV, III and I (Fig. 2) to the shear direction and the increase in dispersion suggest stronger deformation of the plane containing the hinges toward the basal part of the shear zone (Sanderson 1973, Williams 1978).

Lineations

A well developed mineral lineation (L_m) commonly occurs on the mylonitic foliation along most of the shear zone, mainly in its lower part (Fig. 5). The lineation is defined by the dimensional preferred orientation of micas, especially biotite, and quartz aggregates. The lineation plunges gently towards the east, and is usually folded by later D_2 structures. This folding is especially evident in Punta Riomar. In the upper part of the shear zone, the most frequent lineation is an intersection

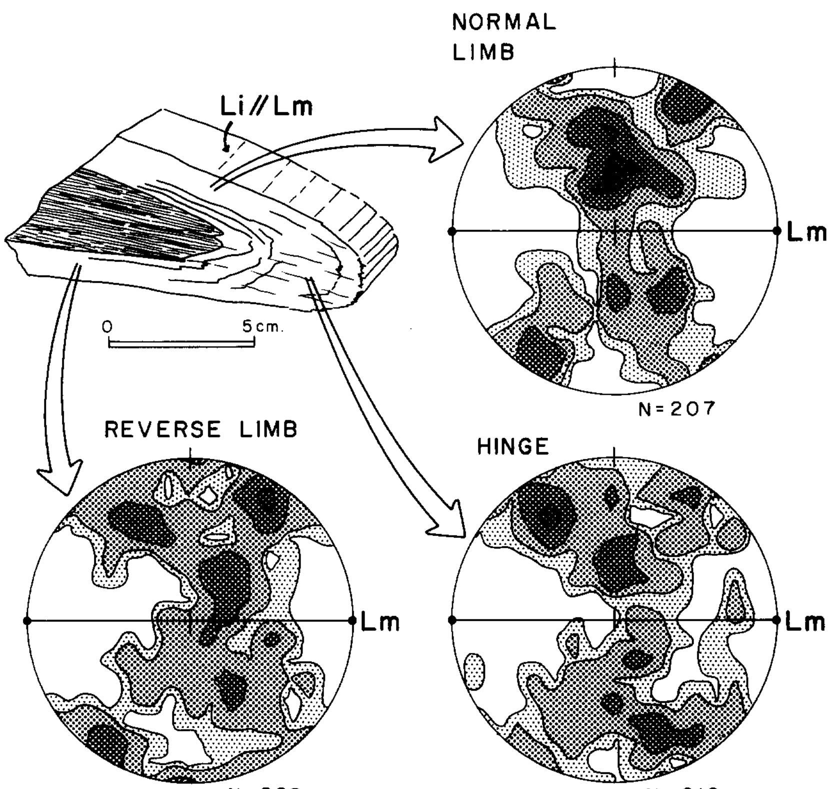

Fig. 4. Quartz *c*-axis fabrics for limbs and hinge of a minor D_2 fold. Locality 162 (Fig. 6).

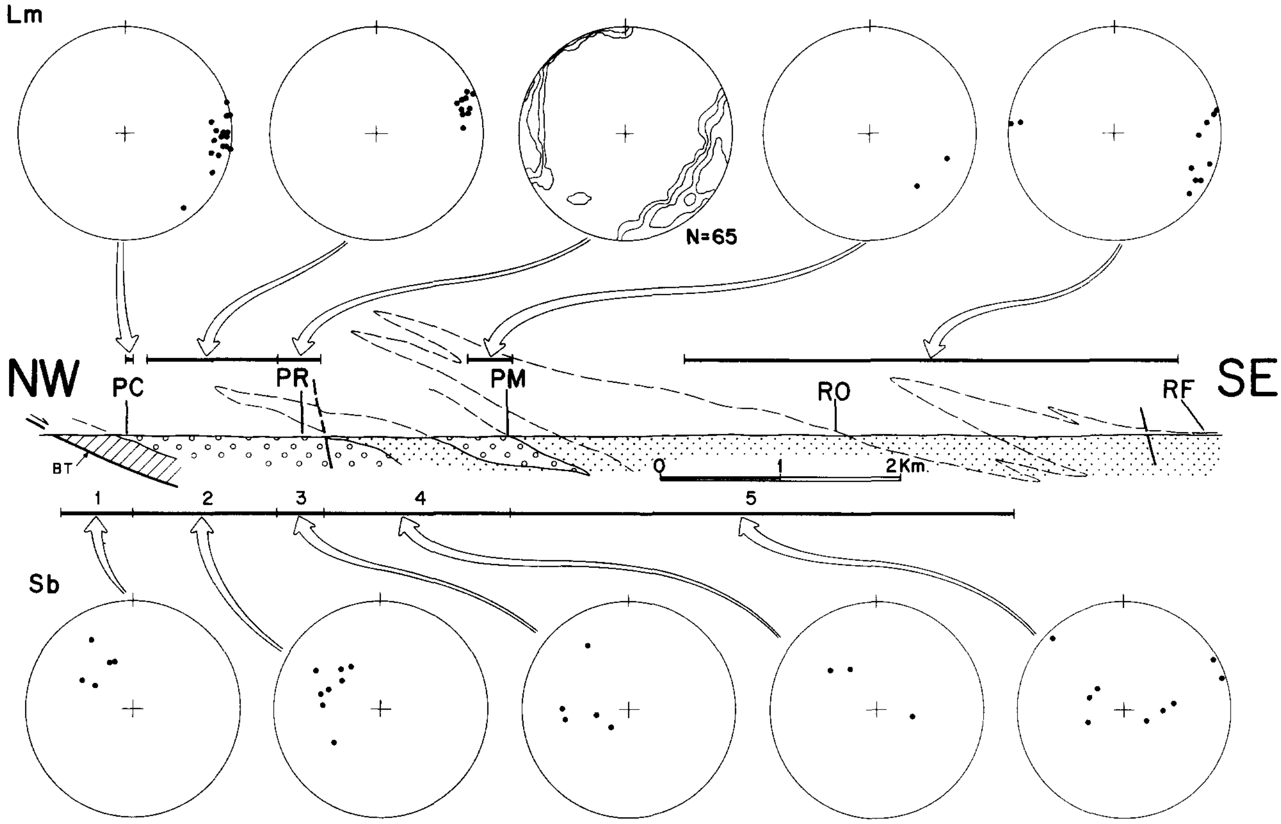

Fig. 5. Distribution of L_m (above) and shear band poles (below) along the shear zone. Contours: 1, 2, 4 and 8%. Abbreviations and ornamentations as in Fig. 2.

lineation between the regional foliation S_2 and S_0, which is parallel to D_2 fold hinges.

Quartzite microstructures

With quartz microstructures it is possible to distinguish several areas of mylonitic development along the shear zone. There is also a marked change in microstructure development upwards, away from the basal thrust.

In quartzitic layers near to the basal thrust (Fig. 6) the foliation is defined by small grains (Fig. 7a) with straight grain boundaries and a strong shape preferred orientation. Intracrystalline deformation microstructures are scarce. These features seem to indicate a total and fast dynamic recrystallization resulting in an overall decrease in grain size. The microstructure of these rocks is that of an ultramylonite.

A mylonitic foliation is also developed in Punta Morago. Here, small equidimensional recrystallized grains constitute a dominant matrix in which elongate old grains with a strong preferred shape orientation define the foliation. To the west of Punta Morago, the transition to a mylonitic foliation can be seen as weakly oriented old grains become strongly oriented. Quartz ribbons are the cores of the old grains that remain unrecrystallized.

Above the ultramylonites of the basal part, intracrystalline deformation features such as deformation bands and subgrains are well developed. On the whole, the grain size is larger than in the more basal quartzite and the foliation is less well developed (Fig. 7b). The larger grain size together with the less well developed preferred grain shape orientation seem to indicate a lower stress, at least during the final stages of microstructural development, than in quartzite near the basal thrust. Further to the east (higher in the shear zone), quartz grains show a weak to absent preferred grain shape orientation. The main foliation is defined by a small amount of long mica crystals with a grain shape orientation (Fig. 7c). Usually, these samples are predominantly coarse grained with irregular grain boundaries, but not strongly lobate or serrated. Grain size decreases towards the eastern part of the section. Subgrains and deformation bands are abundant and a minority of small recrystallized grains with very few deformation microstructures also occurs. The quartzite commonly exhibits a weak foliation oblique to the main one, which formed by local dynamic recrystallization (Fig. 7c). Similar microstructures, although finer-grained, have been presented by Lister & Snoke (1984, figs. 4e and 10d), who interpreted them as recrystallized type II *S–C* mylonites. Well-oriented biotite ribbons (long and thin mica fish in Lister & Snoke 1984) suggest intense deformation. All these rocks on top of the basal ultramylonites can be considered as blastomylonites, in the sense defined by Sibson (1977).

Non-mylonitic tectonites occur predominantly in the eastern half, and also above the shear zone (Fig. 6). The boundary with the preceding group is at the andalusite isograd. Non-mylonitic tectonites also appear in the footwall. The quartz preferred grain shape orientation is

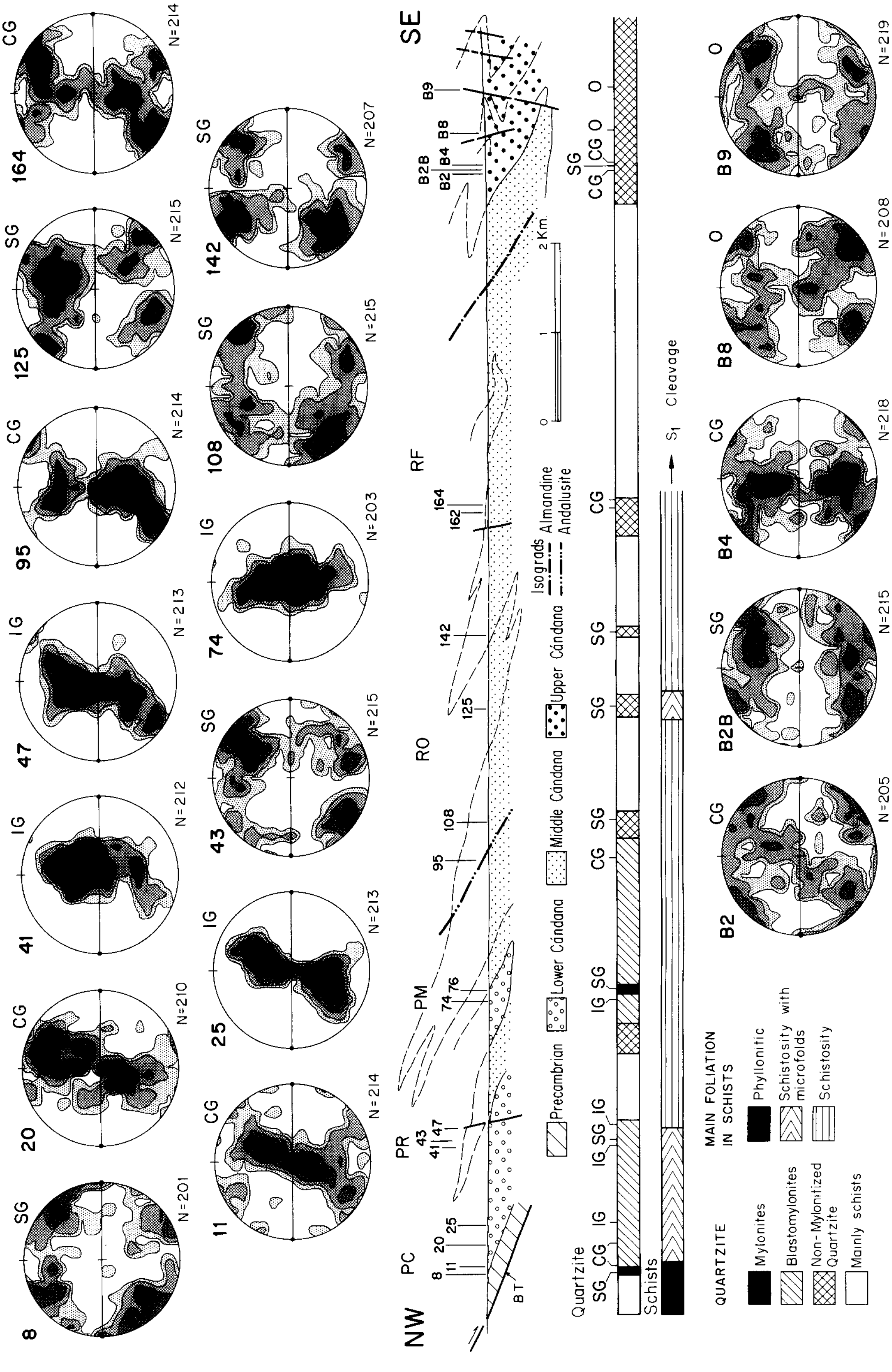

Fig. 6. Quartz *c*-axis fabrics (lower-hemisphere, equal-area projections; contours: 0.5, 1, 2, 4 and 8%), showing locality and fabric type (foliation is the vertical E–W plane and L_m is horizontal), isograds, quartzite tectonite types and schist foliation types along the shear zone. Abbreviations as in Fig. 2.

weak and crystal-plastic deformation features in the grains vary, though in most cases deformation bands and undulose extinction are common (Fig. 7d). These rocks differ from the blastomylonitic rocks in that they do not show the microstructural features of *S–C* mylonites.

Main foliation (S_{1+2} *or* S_2) *in micaschists*

Micaschists in the shear zone have a high quartz content which usually gives a heterogeneous domainal character to the foliation. It is possible to distinguish two microstructural types: phyllonitic foliation and schistosity. Phyllonitic foliation (S_2) occurs at the base of the shear zone (Fig. 6) and is formed by very fine-grained minerals; biotite has a strong grain-shape preferred orientation and quartz domains have a mylonitic character. Schistosity presents scarce microfolds mainly in two zones (Fig. 6). In Punta Riomar, the main foliation associated with D_2 microfolds is crenulated by late D_2 microfolds.

Shear band foliation

Small extensional shear bands within the schists are common in the shear zone (Fig. 5). Their spacing ranges from several decimetres to less than 1 cm. Locally they appear as very short cleavage domains associated with gentle folding. They also occur as mesoscopic bands, which locally become small-scale normal faults that can also be interpreted as riedels. The angle between bedding and shear bands ranges in most cases between 20° and 40°.

Conjugate sets of shear bands are locally present. In such cases, one of the sets is clearly dominant. Shear bands commonly dip east or southeastwards (Fig. 5). However, there are also cases where the shear bands dip westwards, presenting a reverse sense of movement. Field observations suggest a good correlation between the shear band orientation and their position on different limbs of the major D_1 folds. In the normal limb the bands dip east or southeastwards, while in the reverse limbs they dip westwards. This means that the orientation of the anisotropy controls the shear band orientation. If shear bands are formed in some geometric relationship with the shear direction as several authors have proposed (Platt 1984, fig. 1, Weijermars & Rondeel 1984), the orientation of the shear bands can be explained by deformation partitioning (Lister & Williams 1983) (Fig. 8). In reverse limbs, the anisotropy represented by bedding is oblique to the shear direction and the shear bands can be related to slip or flow parallel to the anisotropy due to strain partitioning. If this is so, major D_1 folds would have gone on tightening during the late stages of D_2 owing to a flexural-flow or flexural-slip component generated at the reverse limbs. At the same time, the reverse limbs would have become thinner because of a coaxial component of flow.

Shear bands give a lenticular or sigmoidal microstructural aspect to the main foliation. During shear band development biotite and chlorite, and less commonly quartz, crystallized along the shear plane. The former have a well-developed preferred grain-shape orientation. Quartz commonly forms fine-grained aggregates that define very narrow bands parallel to the shear bands, enhancing the mylonitic character of the rock. The presence of chlorite implies that the shear bands developed during retrograde metamorphism that can be correlated to the M_3 episode. These shear bands also affect D_2 folds and late D_2 biotite porphyroblasts. According to this, shear band formation represents a very late episode in the formation of the Mondoñedo Nappe basal shear zone.

Quartz c-axis fabrics

The quartz *c*-axis orientation in quartzite samples from the shear zone and from adjacent zones has been analysed using a universal stage. Figures 4, 6 and 10 show the diagrams obtained for the shear zone and for the rocks that immediately overlie it, while Fig. 9 illustrates the results for samples from the footwall. According to the position of maxima and the symmetry relative to the foliation (Lister 1977, Schmid & Casey 1986), the following types of fabrics can be observed in these figures.

(1) A random fabric (R) (sample 301).

(2) Fabrics with orthorhombic symmetry (O). One of the samples (302) presents a maximum subperpendicular to the foliation and two other samples (B-8 and B-9) present maxima that are symmetrical with respect to the vertical N–S plane.

(3) Monoclinic small-circle girdle fabrics (SG) (samples 8, 43, 108, 125, 142 and B-2B). The best developed maximum (or pair of maxima) forms a smaller angle with the foliation than that of the other maxima.

(4) Monoclinic type I crossed girdle fabrics (CG). The line linking the inclined maxima can be orthogonal to the foliation (samples 20, 76, 95, 162 and B-4) or oblique (samples 11, 300 and B-2). In some cases, one of the branches of the crossed girdle is poorly developed.

(5) Monoclinic incomplete single girdle fabrics (IG) (samples 25, 41, 47 and 74). These have a sigmoidal trend with two inclined maxima, except for sample 74 which has a single centred maximum. There is a complete transition between this fabric and type CG.

Quartz *c*-axis fabrics along the shear zone display a variable degree of *c*-axes dispersion, with R and O fabrics having the largest amount of dispersion and IG the least. The amount of dispersion increases away from the core of the shear zone (Figs. 6 and 9) and corresponds to a gradation from higher to lower strain, and higher to lower metamorphic grade.

Some of the quartz *c*-axis fabrics from outside the shear zone (R and O types), and therefore not affected by D_2, show a symmetry consistent with a coaxial deformation (cf. Raleigh 1965, Tullis *et al.* 1973, Etchecopar 1977, Lister & Hobbs 1980). *c*-axis fabrics from inside the shear zone have a monoclinic symmetry (SG, CG and IG types) consistent with rotational strain due to the

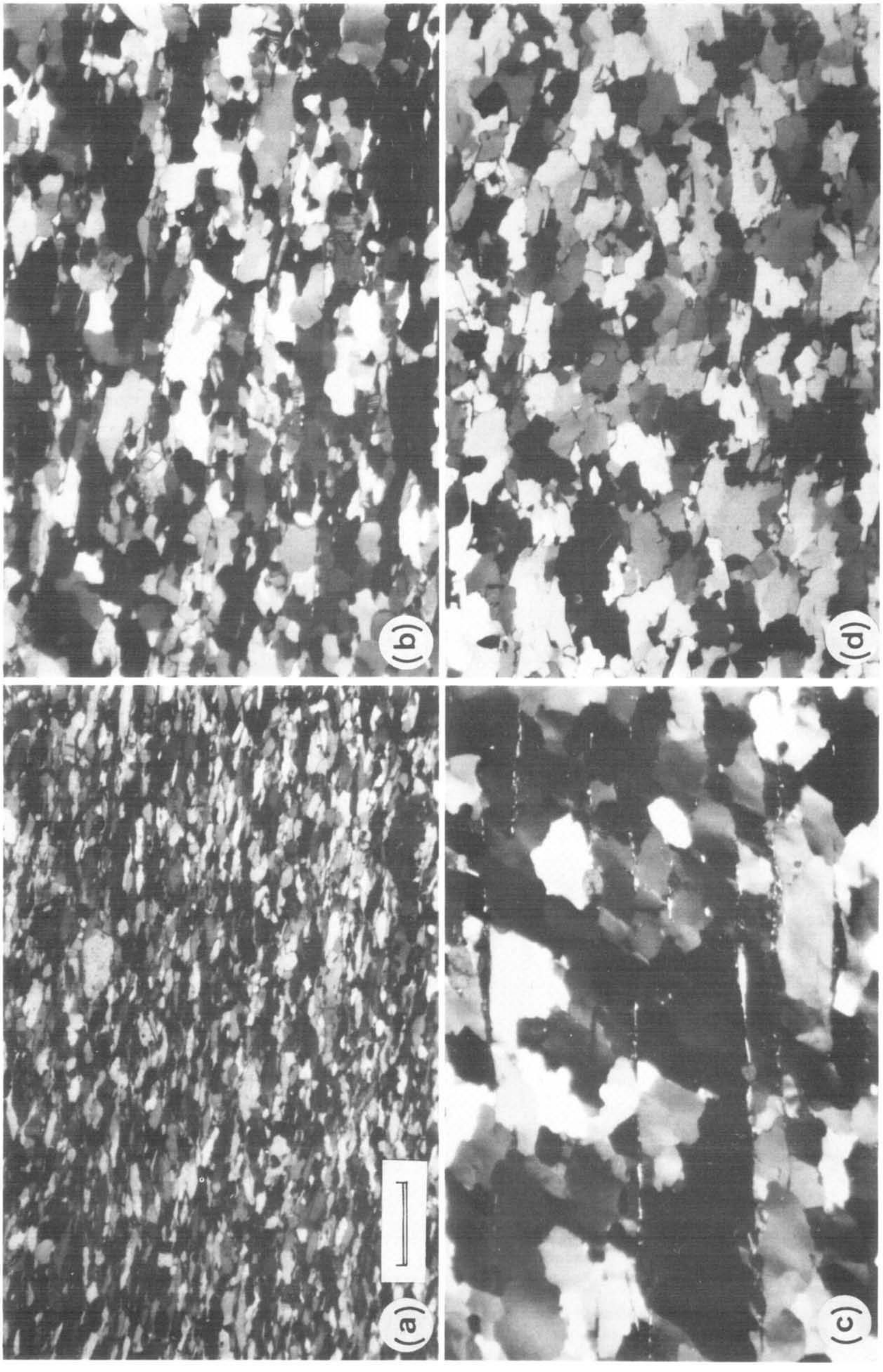

Fig. 7. Different types of quartzitic tectonites. (a) Ultramylonites in the basal part. (b) Blastomylonites in Punta das Cabras. (c) Blastomylonites (recrystallized S–C mylonites) from the lower half of the shear zone; note the thin, long and preferentially oriented micas and the weak oblique foliation in quartz. (d) Non-mylonitic quartzite from the upper part of the shear zone. All sections normal to the foliation and parallel to L_m. Scale bar for photographs = 0.5 mm.

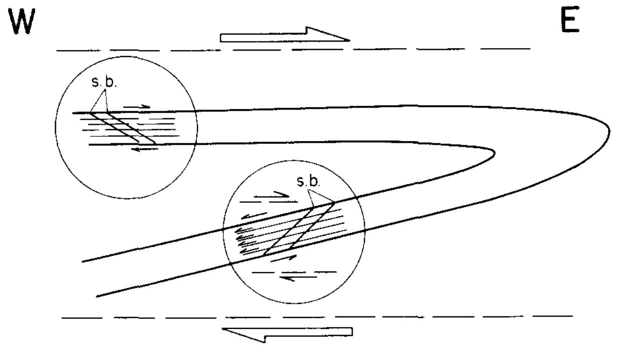

Fig. 8. Partitioning of simple shear deformation component in the reverse limb of D_1 folds to form shear bands (s.b.) with the opposite movement sense to the ones in the main shear zone (see text).

simple shear component (cf. Bouchez & Pecher 1976, 1981, Etchecopar 1977, Lister & Hobbs 1980, Simpson & Schmid 1983). However, there are some samples outside the shear zone that also have a monoclinic symmetry (SG and CG types). Samples B2B and B4 present fabrics very similar to those in the shear zone, and can thus be interpreted as local manifestations of D_2.

Peripheral maxima in the position found in our samples are indicative of low-temperature basal slip, whereas peripheral and inclined or centred maxima may indicate basal and rhombohedral or prismatic slip at medium temperature, all along an ‹*a*›-axis (Nicolas & Poirier 1976, Vernon 1976, Bouchez 1977, Nicolas *et al.* 1977, Bouchez & Pecher 1981, Hobbs 1985). Inclined or centred maxima suggest higher temperature ‹*a*› slip on rhombohedral or prismatic planes (Bouchez & Pecher 1981, Hobbs 1985). The different slip systems active in quartz across the shear zone agree with the temperature changes recorded by the metamorphic assemblages (Fig. 6).

The distribution of *c*-axis maxima in CG fabrics, in the context of a retrograde metamorphism, suggests that these fabrics may be associated to a drop in temperature. This agrees with *c*-axis orientations of old and new grains in a mylonite from Punta Morago (Fig. 10). New recrystallized grains have an increased dispersion of *c*-axes, associated with the activation of basal slip.

Although there is no one-to-one correspondence between microstructural and *c*-axis orientation types, a correlation of the two features of the rock fabric has been established (Fig. 6). R and O fabric always correspond to non-mylonitic tectonites. SG fabric commonly occurs in non-mylonitic tectonites, although it is also found in basal ultramylonites and, rarely, in blastomylonites. CG fabric presents a wider microstructural variety since it occurs in all the microstructural types described here; this probably means that it has a transitional character between SG and IG types. Finally, IG type is recorded only in blastomylonites.

SHEAR ZONE MOVEMENT DIRECTION

Two types of criteria have been used: those that do not provide information on the movement directional sense like the mineral lineation, shear band foliation, fold hinges and conical fold apical directions, and those that do supply data about the movement directional sense like fold asymmetry and facing, oblique foliation by dynamic recrystallization in quartzite and quartz *c*-axis distributions. In order to visualize the movement direction, the direction of the first eigenvector of Bingham's distribution (Whitten 1966, Cheeney 1983) has been projected in Fig. 11 for several criteria in different sectors along the shear zone.

Movement direction indicators

The mineral lineation (L_m) is folded in some cases by D_2 folds. Hence, it does not record all the D_2 deformation. In addition, the folding of L_m causes its dispersion, which can be mainly observed in Punta Riomar (Fig. 5, localities 40–47). However, quartz *c*-axis fabrics show symmetries that generally agree with the position of this lineation, indicating that L_m agrees with the movement direction in late stages of the deformation. The first eigenvector of L_m distributions has a direction between N75°E and N115°E (Fig. 11), that is, with a dominant eastward component.

The relationship between the position of the shear bands and the type of major fold limbs (D_1) on which they develop implies that it is difficult to use the shear bands as a shear-sense indicator. However, shear band formation is related to shear zone deformation and therefore provides information about its movement direction during the late stage at which shear bands formed. Since S_0 approximately coincides with the shear plane, the direction contained in S_0 which is perpendicular to the intersection between S_0 and the shear bands was used as a direction movement indicator. The first eigenvectors of the corresponding distributions have a fan-like distribution whose directions range between N65°E and N140°E (Fig. 11).

Fold hinge directions cannot be considered here as a good criterion to determine the shear direction, because the corresponding eigenvectors do not agree with those of the rest of the criteria, except for sector 1 (Punta das Cabras) (Fig. 11). The cause of the hinge directions dispersion has been discussed above.

Geometric and experimental models indicate that apical directions of sheath folds can be used to deduce shear direction regardless of means of fold nucleation (Cobbold & Quinquis 1980, Ramsay 1980). The directions of the first eigenvector of conical folds apical directions for different domains of the shear zone define a fan of directions between N70°E and N115°E (Fig. 11).

Shear-sense indicators

The strongly asymmetric D_2 folds always have a facing with an eastward component. This provides strong evidence for eastward shear sense for conical folds apical directions. Oblique foliations by quartzite dynamic recrystallization are common in our samples (Fig. 7c). This criterion (Means 1981) mostly indicates an eastward

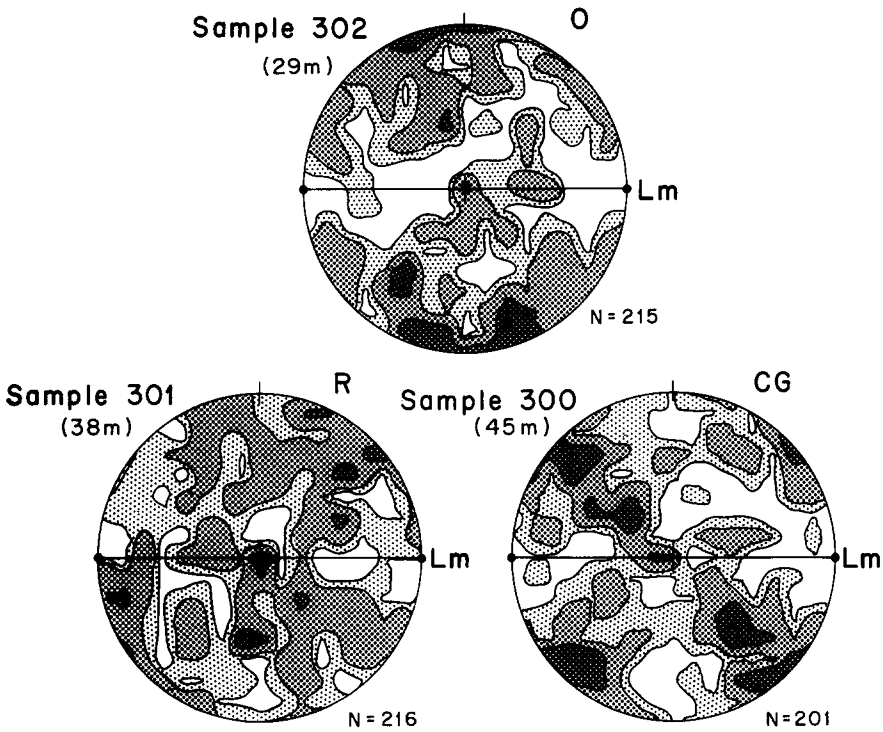

Fig. 9. Quartz *c*-axis fabrics (lower-hemisphere, equal-area projections; contours: 0.5, 1 and 2%) in the footwall of the Mondoñedo Nappe basal shear zone. Numbers refer to locality and distance in metres below the basal thrust.

movement directional sense. Asymmetry of the main girdle is the most suitable criterion in quartz *c*-axis distribution diagrams in this shear zone (criterion 2 in Simpson & Schmid 1983). In most cases, the asymmetry indicates a shear sense with a dominant eastward motion of the hanging wall. Other criteria, such as those using the fabric skeleton or the leading edge (criterion 3 in Simpson & Schmid 1983) usually indicate the same shear direction.

The kinematic criteria are, in general, in good agreement with a movement direction between N75°E and N115°E for the shear zone. Figure 11 shows that only two points corresponding to folds from two sectors are deviant. However, the distribution of kinematic criteria

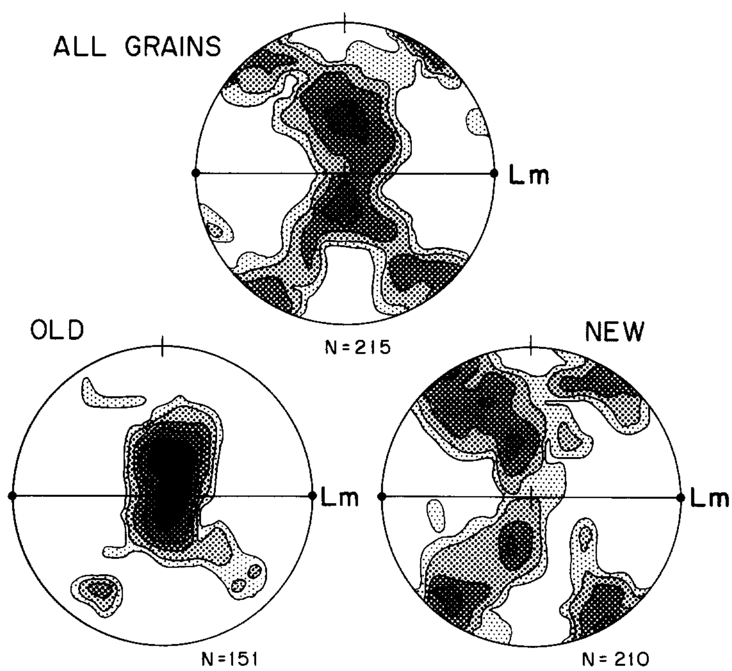

Fig. 10. Quartz *c*-axis fabric for a quartzite sample from Punta Morago (locality 76, Fig. 6). Fabrics for old and new grains are shown separately for the same sample. Contours: 0.5, 1, 2, 4 and 8%.

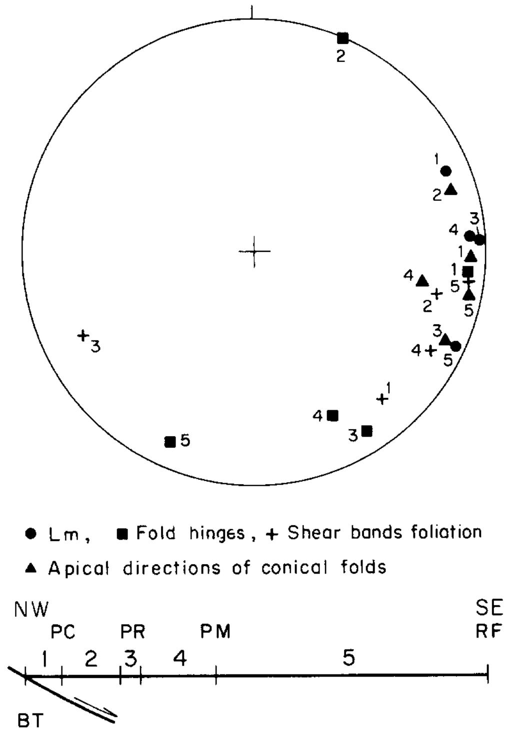

Fig. 11. First eigenvector pattern of the movement direction distributions deduced from different criteria along the shear zone. Abbreviations as in Fig. 2.

in Fig. 11 does not allow deduction of temporal or spatial systematic changes in the shear direction.

SHEAR ZONE STRUCTURAL MODEL

The Mondoñedo Nappe basal shear zone shows evidence for strong ductile deformation which is mainly due to a simple shear component approximately parallel to the normal limbs of earlier major D_1 folds. Evidence for this deformation is given by the development of several types of structures that occur at meso- and microscopic scales and that imply a general movement towards the foreland (eastwards). On the whole, these structures display some spatial variation that suggests a decrease in D_2 strain towards the upper part of the shear zone. However, this decrease presents some remarkable anomalies due to the existence of bands with a large number of folds (Fig. 12). Apart from ductile deformation, it is worth mentioning that the shear zone is cut at the base by a major brittle thrust nearly parallel to the shear plane.

The shear zone developed in an area with a pre-existing metamorphic gradient (M_1). During an early period of ductile deformation in the shear zone, an M_2 episode developed. The characteristics of the D_2 fabric are influenced by the intensity of this metamorphism. The M_2 isograds are later cut by the basal thrust during retrograde metamorphism (M_3) associated with uplift of the nappe.

The distribution of structures within the shear zone allows a division into three parts: basal, middle and upper. The basal part (Fig. 12) is characterized by the presence of phyllonites and some ultramylonites with SG type quartz c-axis fabrics. This implies strong ductile deformation in a late stage at low temperature near to the basal thrust related to retrograde metamorphism. The middle part, whose eastern boundary coincides with the andalusite isograd (Fig. 12), is characterized by the presence of blastomylonites and dominant IG and CG quartz c-axis fabrics, which suggest a deformation in amphibolite facies. Another notable feature is the pres-

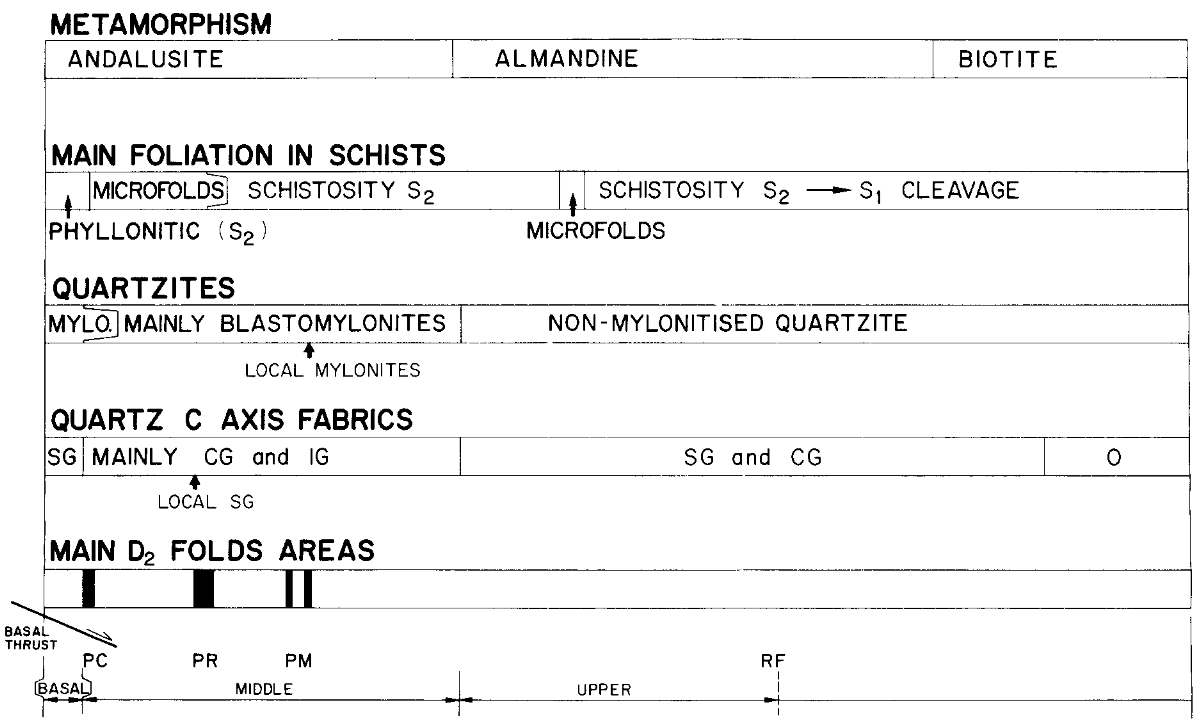

Fig. 12. Scheme of the Mondoñedo Nappe basal shear zone showing position of its basal, middle and upper parts, according to variations in metamorphic grade, foliation type, quartz c-axis fabrics and distribution of D_2 folds. Abbreviations as in Fig. 2.

ence of three bands with peculiar structural characteristics. The first band is located in Punta das Cabras, where a large amount of subsimilar folds appear. The second band (Punta Riomar) contains a large number of D_2 folds with variable geometries (Fig. 4) and interference patterns of Type 3; L_m is frequently folded and quartz *c*-axis fabrics are SG here. The deformation of this band appears to have a long history that includes late movement during retrograde metamorphism. The third band (Punta Morago) contains a large number of folds and mylonites with quartz ribbons; mylonitization here took place during retrograde metamorphism although earlier than in the previous band. The upper part of the shear zone (Fig. 12) is featured by the presence of non-mylonitic quartzite and by SG and CG *c*-axis fabrics, indicating deformation at a temperature that was on the whole lower than in the middle part. This is consistent with data on the metamorphism.

There is no clear upper boundary to the shear zone, but a gradual transition into rocks that do not display shear zone textures. The boundary has been located at Ría de Foz (Fig. 12). However, some of the typical characteristics of the shear zone, like the blastomylonitic microstructures and the main foliation with microfolds in the schists disappear further to the west. Some other features still exist more to the east of Ría de Foz, like some quartz *c*-axis fabrics.

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