PERGAMON

Journal of Structural Geology 23 (2001) 1457–1486

JOURNAL OF STRUCTURAL GEOLOGY

www.elsevier.com/locate/jstrugeo

Zooming on Northern Cap de Creus shear zones

J. Carreras*

Departament de Geologia, Universitat Autònoma de Barcelona, 08193 Bellaterra, Spain

Received 22 September 1999; revised 8 November 2000; accepted 20 December 2000

Abstract

The Northern Cap de Creus shear belt provides fine examples of shear zones developed at different scales in rocks bearing pre-existing penetrative foliations. The shear zones developed under retrograde metamorphic conditions, and they are preferentially located in crystalline schists. Lower grade metasedimentary rocks accommodated deformation by folding. The shear zones are linked in an anastomosing framework with self-similar properties. They are interpreted to have formed as the result of a progressive wrench-dominated deformation affecting anisotropic rocks. A progressive non-coaxial deformation regime is considered responsible for the development of the complex kinematic pattern, wherein shear zones with opposite senses of shear may come to lie in close parallelism. In such a framework, individual shear zones are characterised by marked strain gradients not only across the shear zone, but also along it. These features, together with the geometrical analysis of shear zone marginal domains, indicate that these structures developed with a component of shortening across the shear zones. The shear zones are interpreted to have nucleated as buckling instabilities that gradually evolved to become shear zones approaching the simple shear model. Microstructural and quartz *c*-axis fabric analyses indicate that shear zone formation is associated with strong strain partitioning on all scales.

1. Introduction

The presence of a network of shear zones and associated mylonites at Cap de Creus, Spain, was first described in the early seventies. Today, the area has become classical because of the exceptional outcrop conditions and the abundant structures. Although many papers concerning specific aspects of these shear zones have been published (e.g. Carreras and Santanach, 1973; Carreras, 1975; Carreras et al., 1977), a more general description of the structures is still lacking. To fill up this gap, this study aims to give a general description of the shear zones of the northern Cap de Creus shear belt emphasising their geometrical and kinematical aspects.

To give a picture of these structures, their broader setting in the Variscan basement of the Pyrenees will be given first. This will be followed by progressively zooming in on the northern Cap de Creus shear zones. This study combines already published but scattered information with unpublished details of shear zone geometry. An analysis of papers on structural geology reveals the existence of a gap in information of structures at scales ranging from 1:100 to 1:10000, as most of the information is presented either for the outcrop to microscopic scales or in the range of what might be considered conventional map scales (e.g. 1:10000 to 1:50000). Furthermore, there are many cases in which information from small-scale structures is directly used to infer structures on the crustal scale, skipping intermediate observation levels. Structural self-similarity may be a common situation, but simply excluding the intermediate scales is not justified. There is a risk in extrapolating structures from micro- to macroscales, and this risk increases when determining kinematics based on geometrical relationships at any scale.

The interpretation of the evolution of the Variscan orogeny in the Pyrenees constitutes an example of how contradictory interpretations may arise from the study of same structures by different geologists (see Carreras and Debat (1996) for discussion). These differences concern not only the geotectonic significance of the structures formed during the main event but also later deformations overprinting the main fabric. The mylonite belts associated with ductile shear zones constitute one of these controversial structures. The understanding of the geological significance of shear belts in the Pyrenees requires a detailed structural and kinematic analysis on all scales. This paper aims to present a multi-scale analysis in one of the best exposed mylonite belts. The results not only reveal which of the suggested tectonic settings are compatible with the field evidence, but also provide insight into the general significance of structures associated with shear zones, which might be compared with other settings.

* Tel.: +349-3-581-2284.
E-mail address: jordi.carreras@uab.es (J. Carreras).

0191-8141/01/$ - see front matter
PII: S0191-8141(01)00011-6

2. Shear zones in the Pyrenees

There are various types of fault zones affecting the Variscan basement of the Pyrenees. Because the basement containing Variscan structures was also involved in later geotectonic events, it is not always easy to establish to which single or multiple deformational event a given structure is associated. Fig. 1 illustrates some of the main fault structures on the Pyrenees and the succession of deformational events responsible for the faults affecting the Variscan basement.

Several deformation events of Variscan age leading to localised high strain zones have been recognised. The earliest event appears to be the early Variscan thrusting, where thrust surfaces are more clearly identified by anomalous stratigraphic superposition (Losantos et al., 1986; Raymond, 1986), than for the presence of thrust-related structures or fabrics. In deeper structural levels, high strain zones developed during the Variscan main tectonic event. These zones are contemporaneous with the metamorphic peak and penecontemporaneous with the onset of Variscan magmatism. They develop either gently-dipping or steep-dipping penetrative foliations, there being no consensus on their geotectonic significance. Interpretations vary from extensional (Eeckhout and Zwart, 1988) to compressional (Soula et al., 1986a) or transpressional settings (Carreras and Capellà, 1994; Druguet et al., 1997).

A network of ductile shear zones, mainly affecting crystalline rocks, developed after the metamorphic peak and a significant temperature drop. The Variscan age of these structures, which are the focus of this paper, is still controversial and their geotectonic significance is also a matter of debate. These aspects will be summarised, later. There were three main post-Variscan tectonic events that involved the Variscan basement and which generated or reactivated fault zones:

1. A sinistral strike-slip faulting event played its most important role during the rotation and the SE sliding of the Iberian block with respect to the Eurasian plate (Choukroune and Mattauer, 1978). The north Pyrenean fault is the most representative structure.
2. The main Pyrenean compressional event took place in the Paleogene, causing the Variscan basement and the overlaying cover sequences to be involved in a fold and thrust belt. Although some slices of Palaeozoic basement are clearly involved in this event, especially along the south Pyrenean Zone, it has been a major problem to trace these thrust sheets into the Variscan basement of the Axial Zone. In locations where the thrust fault zone continues in the basement, it consists of crushed rocks with abundant veining. This is a common situation for deformation occurring in the brittle to ductile transition field (Teixell, 1996). Lack of evidence of major offsets of Palaeozoic lithologies and Variscan structures makes it difficult to root the alpine thrust within the Axial Zone (e.g. Soler et al., 1998).
3. Finally, Neogene extensional tectonics produced a system of oblique grabens affecting both the Variscan basement and the cover sequence. These grabens are limited by normal or strike-slip faults that produced predominantly fault gouges (e.g. Julià and Santanach, 1997).

The shear-zone related mylonites are controversial both in age and tectonic significance and have been ascribed to a variety of tectonic events. To some authors these rocks are mainly related to eo-Alpine sinistral strike-slip movement (Soula et al., 1986b), or to Alpine compression (McCaig and Miller, 1986). To others, some might be the result of Alpine reactivation of older Variscan extensional structures (Saint Blanquat, 1993; Delaperrière et al., 1994). A fourth group of authors attribute the fabrics of these rocks exclusively to the Variscan tectonics (e.g. Carreras et al., 1980; Guitard et al., 1980; Evans et al., 1997).

In this study, an attempt is made to discriminate ductile shear zones from other fault-related structures. In spite of the controversies about tectonic significance, it is indisputable that there is a distinct and well defined anastomosed mylonite belt involving the Variscan basement of the Pyrenees that outcrops along the Axial Zone and also along the north Pyrenean Massifs located north of the north Pyrenean fault. This mylonite belt affects crystalline Paleozoic rocks with a homogeneous strain facies and kinematic pattern. The mylonites are associated with shear zones developed under greenschist facies conditions, giving rise to typically fine-grained well-foliated fabrics. In all mylonites, the foliation and associated stretching lineation have a fairly constant attitude (Fig. 2). These mylonites are always associated with discrete shear zones, which affect either previously foliated metamorphic rocks or granitoids bearing previous high-temperature fabrics (Gleizes et al., 1997). These mylonites are referred to as the late Variscan mylonite belt. These mylonites must be distinguished from earlier high-temperature deformation bands, and from all other thrusts or fault-related rocks of Alpine age. Although the possibility of a Pyrenean reactivation of older Variscan structures has been suggested (Delaperrière et al., 1994), field evidence for a widespread Alpine tectonic overprinting is lacking. All mylonitic fabrics developed under a similar kinematic framework and deformation facies.

The shear zones of the Variscan mylonite have predominantly north-dipping attitudes in the Axial Zone (Zwart, 1958; Carreras et al., 1980; Guitard et al., 1980; Saillant, 1982; Casas, 1986) and also in the North Pyrenean Massifs (e.g. in Saint Barthélémy Massif; Passchier, 1982). Only the south-easternmost mylonite belt in Rodes and Roses (Carreras and Losantos, 1982) displays south-dipping attitudes, which is likely the result of Alpine bulk tilting of all Variscan structures along the southern border of the Axial Zone, analogous to the south tilting of Variscan foliations observed on the south boundary of the central Pyrenees Axial Zone (e.g. Hartevelt, 1970).

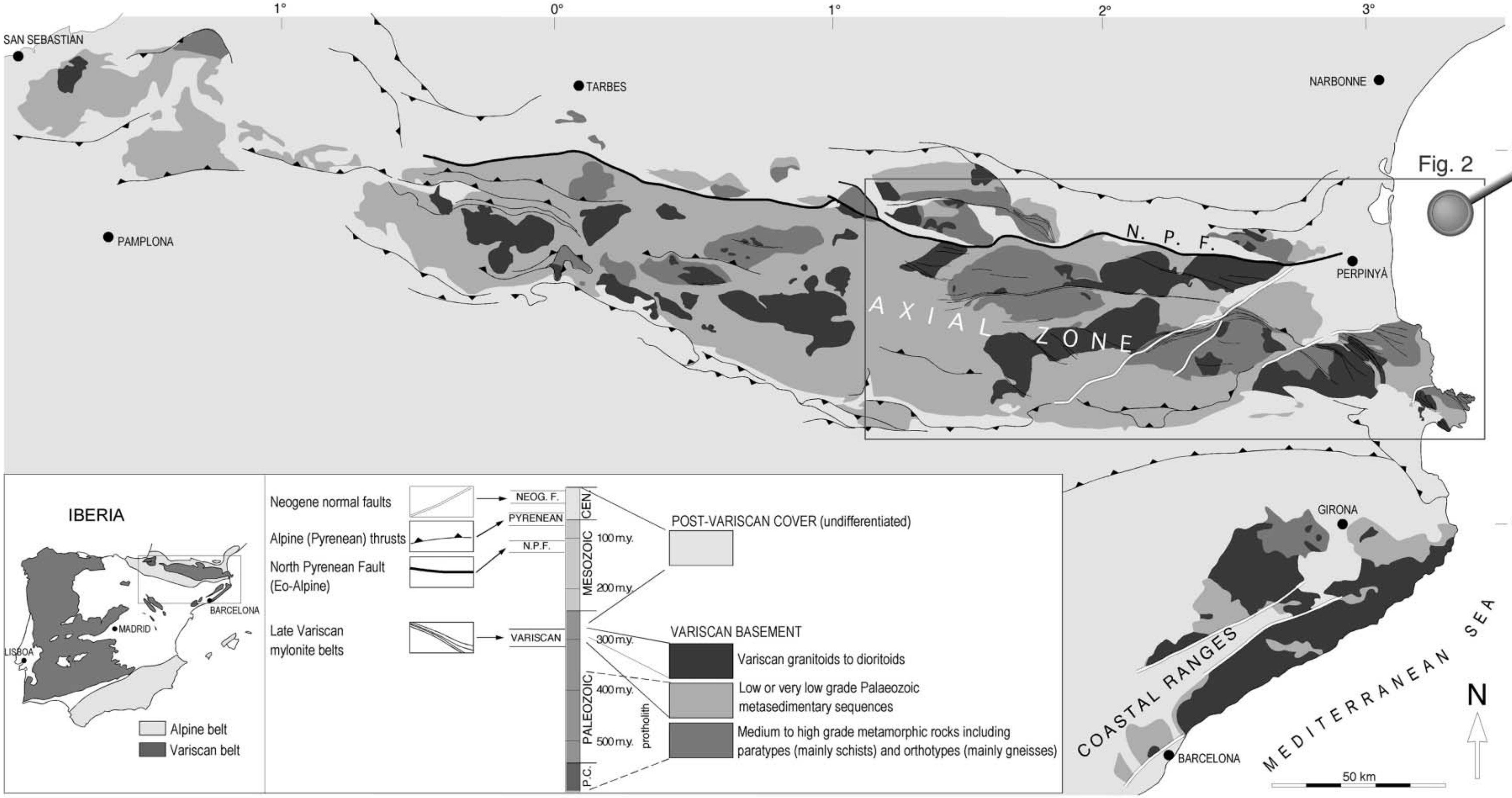


Fig. 1. Ductile and brittle fault zones in the Pyrenees and Northern Coastal Ranges. Chronological indication of each event is shown.

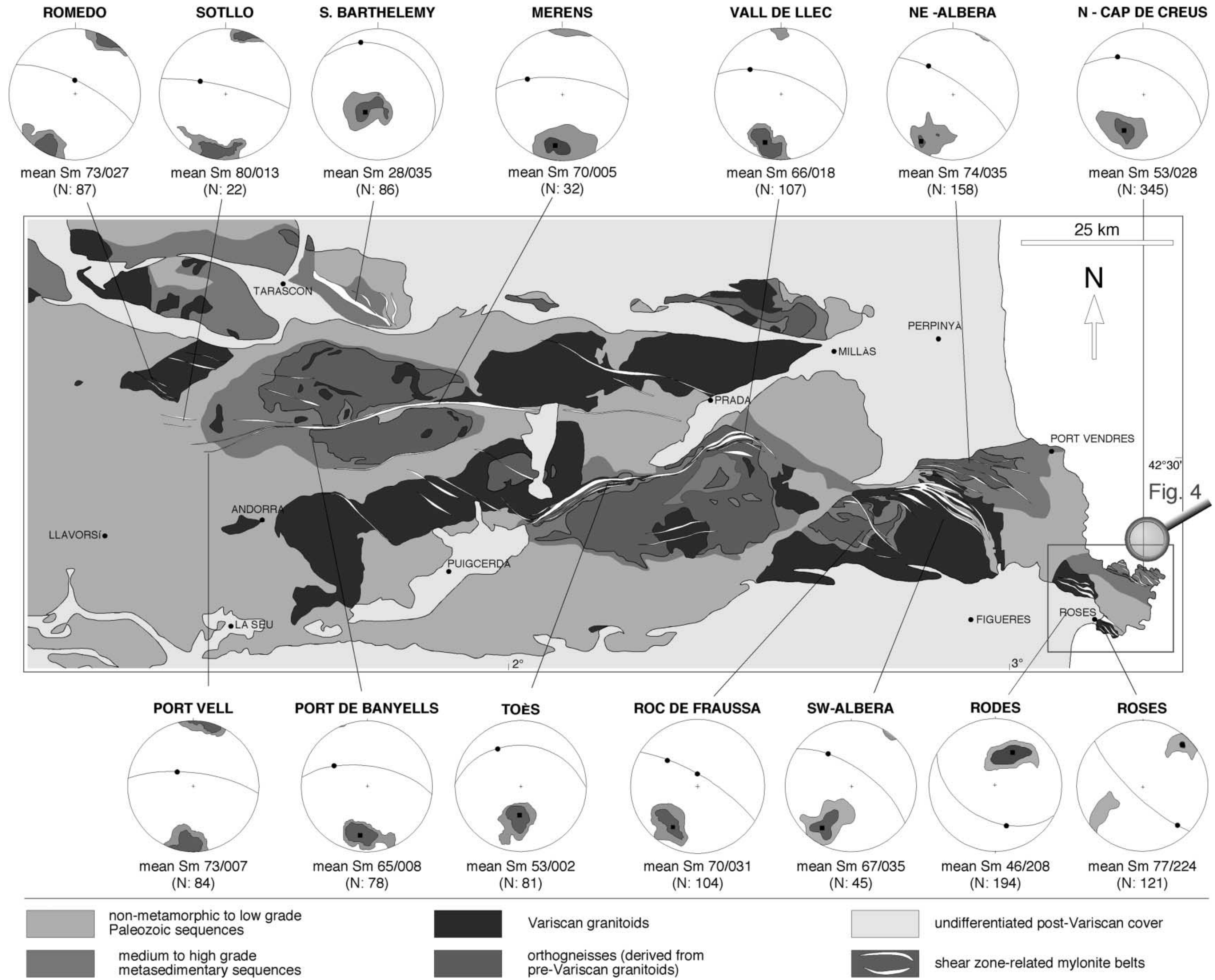


Fig. 2. Ductile shear belts in the Variscan basement of the Pyrenees, and stereograms showing the attitude of the mylonitic foliation (Sm) and the associated stretching lineation (Lm). (Equal area, lower hemisphere, Sm density contours: 2 and 4%.)

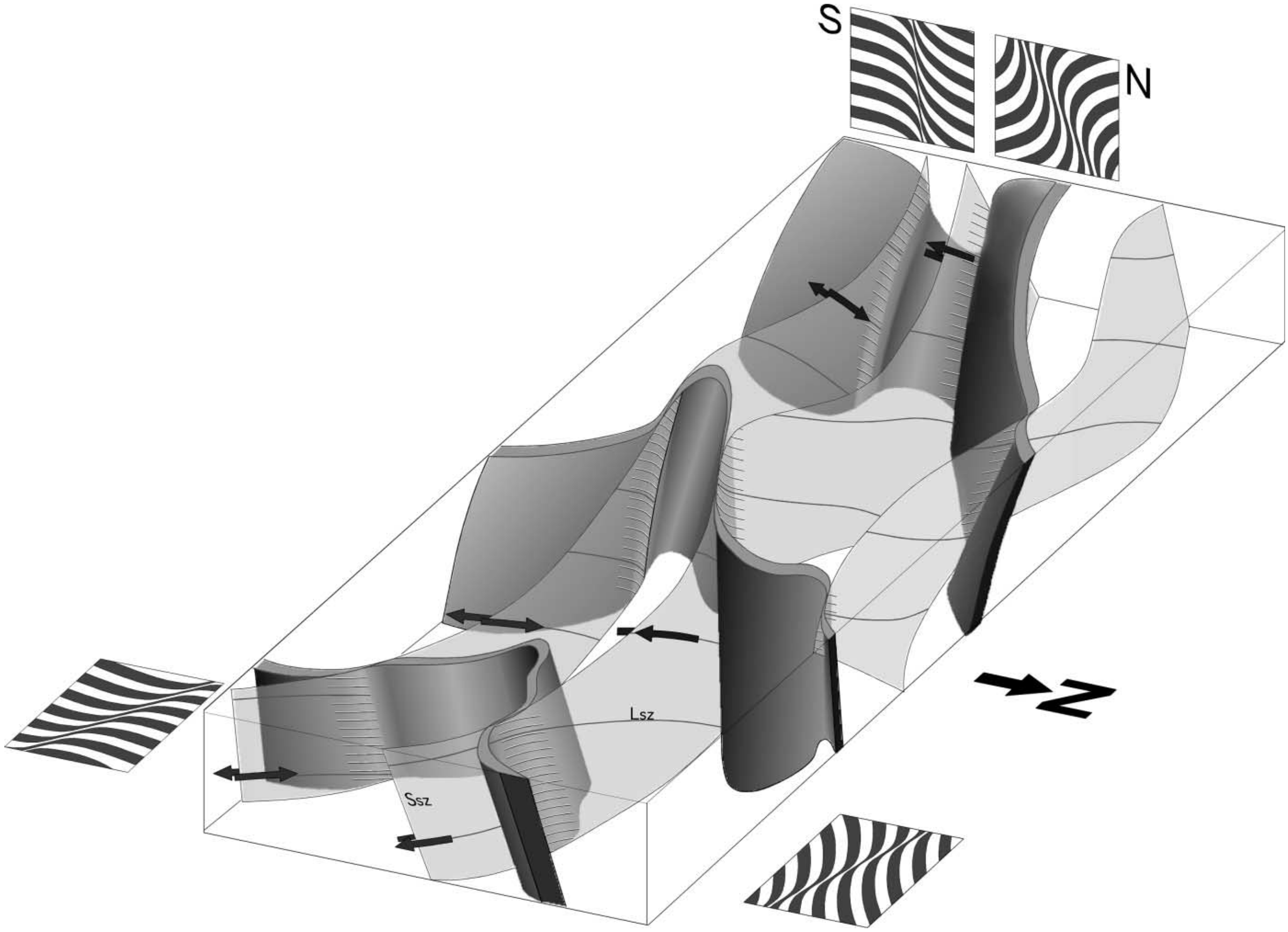


Fig. 3. Idealised sketch showing the three-dimensional arrangement of anastomosed shear belts and associated stretching lineations (Lsz). Sheared layers represent the dominant foliation. Predominant dip-slip occurs in E–W trending zones in the west, while strike slip is prevalent in easterly located, NW–SE trending zones. Note the coexistence of opposite slip senses indicated by the arrows. (Ssz: shear zone plane, Lsz stretching lineations in shear zones).

The stretching lineation indicates a dominant of oblique slip, although it ranges from locally dominant dip-slip (e.g. Mérens Fault zone) to strike-slip (e.g. Cap de Creus shear belt) (Fig. 2). The presence of well-developed stretching lineations and the deflection of pre-existing or newly formed foliation at the shear zone margins usually enables the kinematics of a shear zone to be established. Furthermore, shear sense indicators (e.g. shear band cleavage or asymmetric *c*-axes quartz fabrics) are common. A bulk kinematic analysis indicates that relative movement varied from strike-slip to dip-slip (Fig. 3). Strike-slip components are dominant in NW–SE trending shear zones, while dip-slip components are prevalent in E–W trending shear zones. This pattern is not only observed in the mylonite belt as a whole, but also within smaller domains where an anastomosing network of shear zones with different trends exists. There are individual shear zones or sets of shear zones making exceptions to this rule, as for instance in the south edge of the Bassiès granitic massif (Romedo) and Roc de Fraussa Massif (Fig. 2), where NW–SE trending shear zones exhibit anomalous dip-slip movements. On the other hand, while reverse movement predominates in E–W trending, dip-slip shear zones, such shear zones may coexist with contemporaneous and nearly parallel shear zones exhibiting normal displacement. In a similar manner, while dextral strike-slip predominates in NW–SE shear zones, associated zones of sinistral strike-slip can be found. This association of antithetical movements in closely parallel shear zones is schematically shown in Fig. 3. This has been explained as an influence of pre-existing foliations in the nucleation of structures (Carreras et al., 1980).

3. Setting of the Cap de Creus shear zones

The Cap de Creus peninsula is situated in the easternmost Pyrenees. In the peninsula, two WNW–ESE trending late Variscan shear belts exist, forming part of the formerly described mylonite belts affecting the Paleozoic basement of the Pyrenees (Carreras et al., 1997a). The northern belt cuts across medium- to high-grade schists, whereas the southern belt (Roses–Rodes) cuts across Variscan granodiorites

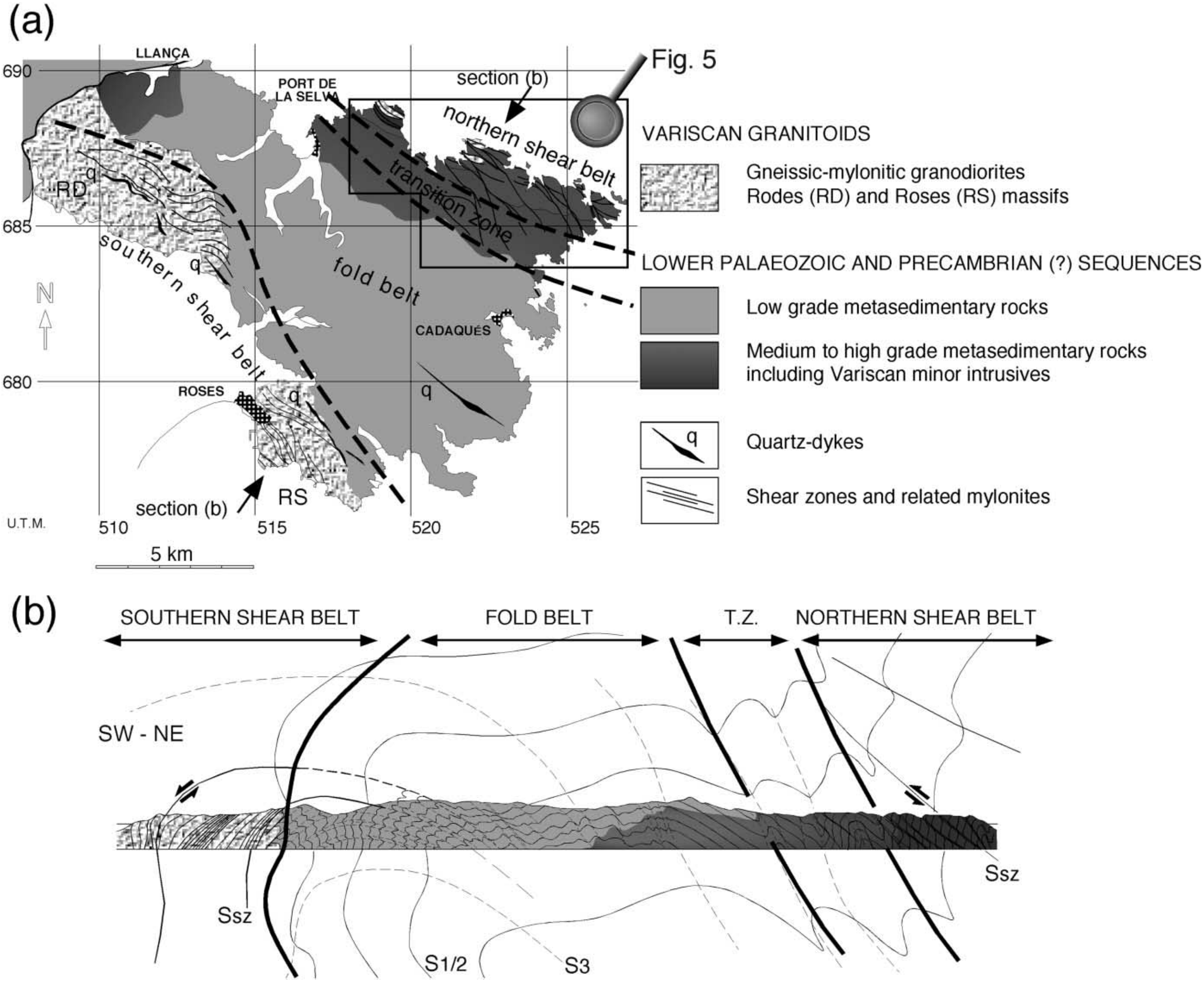


Fig. 4. Shear belts in the Cap de Creus peninsula. (a) Simplified map. (b) Cross-section showing the structure delineated by the trace of the folded dominant foliation ($S_{1/2}$), the axial traces (S_3) of late folds and the shear zones (Ssz).

emplaced in low-grade metasedimentary rocks (Carreras and Losantos, 1982). The two belts are separated by a complex fold belt of low-grade metasediments (Fig. 4a).

The shear belts have similar trends and display dominant strike-slip displacement. However, in the northern shear belt, NE-dipping dextral shear predominates, whereas in the southern belt SW-dips and sinistral strike-slip predominates. This geometry is interpreted as the result of later overturning of structures along the southern border of the Axial Zone. In fact, not only shear zones but also axial planes of Variscan folds display this anomalous southward tilting. This tilting also causes the originally dextral shear zones to outcrop as sinistral ones (Fig. 4b).

The structures in the southern belt (Simpson et al., 1982; Simpson, 1983) display the typical geometry of shear zones developed in rather isotropic rocks with the foliation plane trajectory tracking the *XY* finite strain plane developed in a heterogeneous simple shear (Ramsay and Graham 1970; Ramsay 1980). Thus, the bending axis of the newly formed sigmoidal foliation is always perpendicular to the shear direction. However, the described shear zones of the northern belt are different as they cut across rocks bearing pre-existing foliations. The general nature of shear zones cutting across foliated rocks has been recently described by Carreras (1997). In this paper, this analysis will be extended by examining the structures over a large range of scales at the Cap de Creus shear zones.

3.1. The Cap de Creus northern shear belt

This WNW–ESE trending belt is formed by an anastomosing network of shear zones cutting across medium- to high-grade schist derived from sedimentary rocks sequences of presumably Cambrian or upper Precambrian age (Fig. 4a). The schists underwent a polyphase tectonic history before mylonitisation. They display at least one penetrative foliation (Fig. 5). The dominant foliation prior to mylonitisation is either a predominantly layer-parallel schistosity (S_1) or a transposition foliation (S_2). Metamorphism reached upper amphibolite facies conditions, with localised migmatite development synchronous with the development of the S_2 foliation (Druguet, 1997; Druguet and Hutton, 1998). A widespread intrusion of pegmatitic bodies also took place before shear zone development. Shear zone formation occurred at lower grade metamorphic conditions (greenschist facies) and partial re-equilibritation of minerals is commonly observed in the mylonites, with new growth of white mica, albite, epidote, chlorite and extensive recrystallisation of quartz.

Most shear zones cut obliquely across the earlier foliations (S_1 and/or S_2) in the schists (Figs. 6 and 7), and a marked deflection or rotation of the regional schistosity into close parallelism with the trend of the shear zones is commonly observed (Figs. 8 and 9). Individual shear zones vary in size, but most range in width from a few centimetres to a few tens of meters. Broader zones of decametric width exhibit a composite structure, formed by a network of anastomosing smaller shear zones (Fig. 8) in a similar way as shown by Hippertt (1999). Shear zones are generally discontinuous, joining or dying out, and commonly showing splay-type geometries.

The deflection of the existing foliation indicates in every case the apparent sense of displacement. This information added to the attitude of the stretching lineation determines the movement sense and the displacement direction. Caution is necessary if the sense of movement is to be inferred from the apparent marginal bending of the regional foliation in a two-dimensional section. For particular relative orientations of shear plane and existing foliation, the apparent movement sense in an outcrop surface might be opposite or highly oblique to the actual one (Fig. 10) (Wheeler, 1987; Carreras, 1992; Passchier and Williams, 1996; Carreras, 1997). In most of the mylonite belt, even if the marginal deflection of the pre-existing foliation is not visible, or is unclear or misleading, the sense of movement on each shear zone can generally be deduced from other field kinematic indicators, such as asymmetric lozenges, extensional crenulations or asymmetric folds with fold axes at a high angle to the stretching lineation (Fig. 11).

The geometry of the shear zones varies gradually across the area from north to south as a consequence of the lower crystallinity and, hence, a higher mechanical anisotropy of the schists towards the south at the time of mylonitisation. In the higher grade schists of the mylonite belt, the shear zones are relatively narrow, cutting across domains of unsheared schists. In the lower grade rocks towards the south there is a rather complex late deformation history, with coexisting folds and shear zones forming a complex structural domain. This domain, called transition zone by Carreras and Casas (1987), separates the northern shear zone domain from the late folded domain occupying the central part of the Cap de Creus peninsula (Fig. 4a).

Shear zones developed in the transition are commonly associated with folding. The deflection of the existing foliation towards parallelism with the mylonitic foliation leads to the onset of buckling instabilities, which in low strain domains is expressed as open folds and crenulations in the shear zone margins (Fig. 12). These folds become gradually tightened towards the margins of the shear zone, and completely transposed within the mylonite belt. These tight asymmetrical folds possess fold axes parallel to the stretching lineation and therefore cannot be used as a shear sense criterion.

3.2. Mylonites

The mylonitic equivalents of schists and associated igneous rocks are penetratively foliated and banded (Fig. 13a) as the result of strong deformation of the heterogeneous schists containing segregation quartz nodules and veins, and locally pegmatite dykes or small granitoid bodies (Carreras and Druguet 1994; Carreras et al., 1997b). There are a great variety of mylonite types as a result of the lithological heterogeneity. All types were deformed in conditions that range in space and time from upper greenschist (earliest event in the north) to lower greenschist (latest event and towards the south).

Because of the metagreywacke and metapelitic composition of the host schists, the resulting mylonites are generally quartz–feldspathic and mica-rich. However, the abundance of quartz segregation veins and nodules in the schists makes the presence of thin quartz bands in the mylonites common (Fig. 13a). A third, well-distinguished type is a leucocratic quartz–feldspathic mylonite, derived from the mylonitisation of pegmatites and leucogranites. Minor types are amphibole-bearing mylonites derived either from amphibolites or tonalites.

Mylonites exhibit a well-marked stretching lineation (Fig. 13b), especially in quartz–mylonites where corrugated lineations arise from the stretching of surface irregularities such as in the contact of quartz nodules with the enclosing schist. Thus, the mylonitic lineation commonly consists of a stretching lineation defined by the alignment of elongated mineral crystals and also on an intersection lineation. Quartz–feldspathic mylonites derived from sheared pegmatites and granitoids may exhibit a stretching lineation defined by alignment of prismatic minerals (e.g. tourmalines in pegmatites). This planar–linear mylonitic fabric is accompanied by a strong preferred crystallographic orientation of minerals. Microstructures and quartz *c*-axis fabrics from these mylonites have been described by Carreras (1974), Carreras et al. (1977), Carreras and García-Celma (1982), García-Celma (1982), Norton (1982) and García-Celma (1983).

Mylonitic foliation and compositional banding are commonly disturbed by the presence of folds and/or extensional crenulation cleavages (Fig. 11). Folds are developed mainly in quartz and quartz–feldspathic bands while extensional crenulations occur in mica-rich mylonites

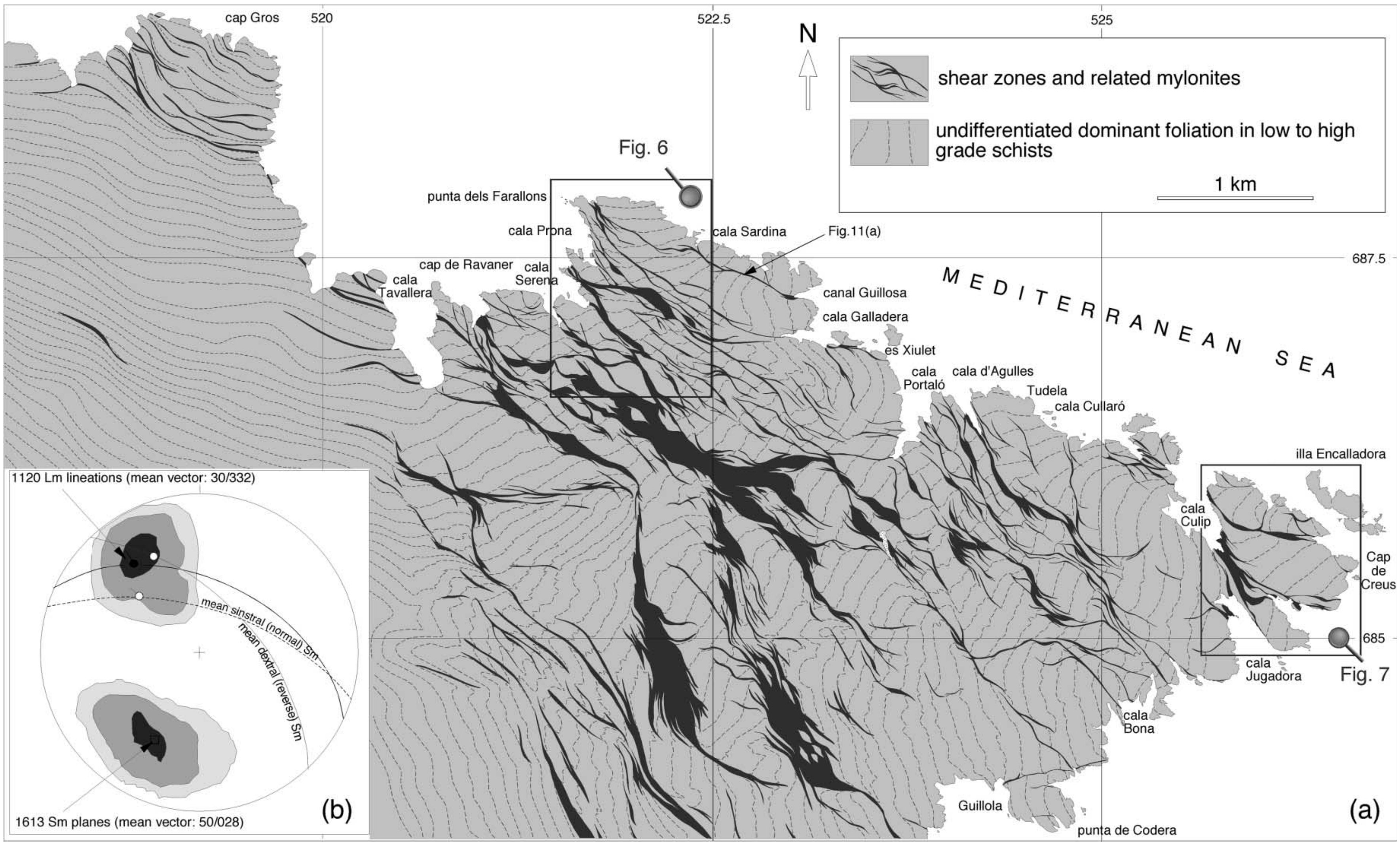


Fig. 5. (a) Map of the Northern Cap de Creus shear belt. (b). Equal area, lower hemisphere stereogram showing the attitudes of the mylonitic foliation (Sm) and the associated stretching lineation (Lm). Mean orientations of Sm and of Lm, for 133 dextral and 89 sinistral shear zones, have also been indicated.

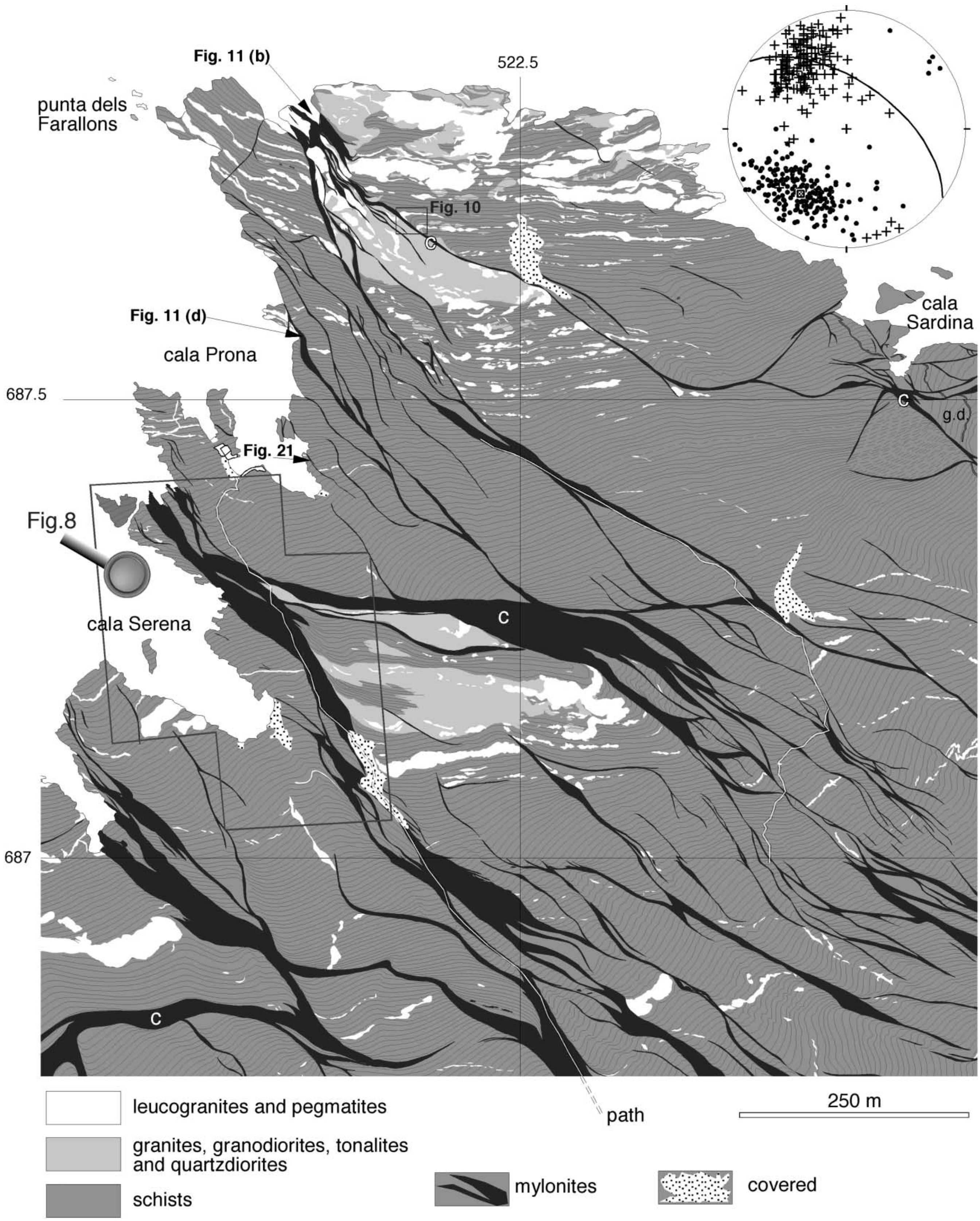


Fig. 6. Map of the Cala Serena–Punta dels Farallons area. This domain is characterised by predominant dextral shear zones cutting across the WSW–ENE trending S_2 dominant sub-vertical foliation. Shear zones labelled with 'c' display anomalous or complex kinematic behaviour. Key to simbols: g.d. is a gabbro dyke, stereogram shows 238 poles to mylonitic foliation (dots), and 179 stretching lineations (crosses). The great circle denotes the mean mylonitic foliation plane.

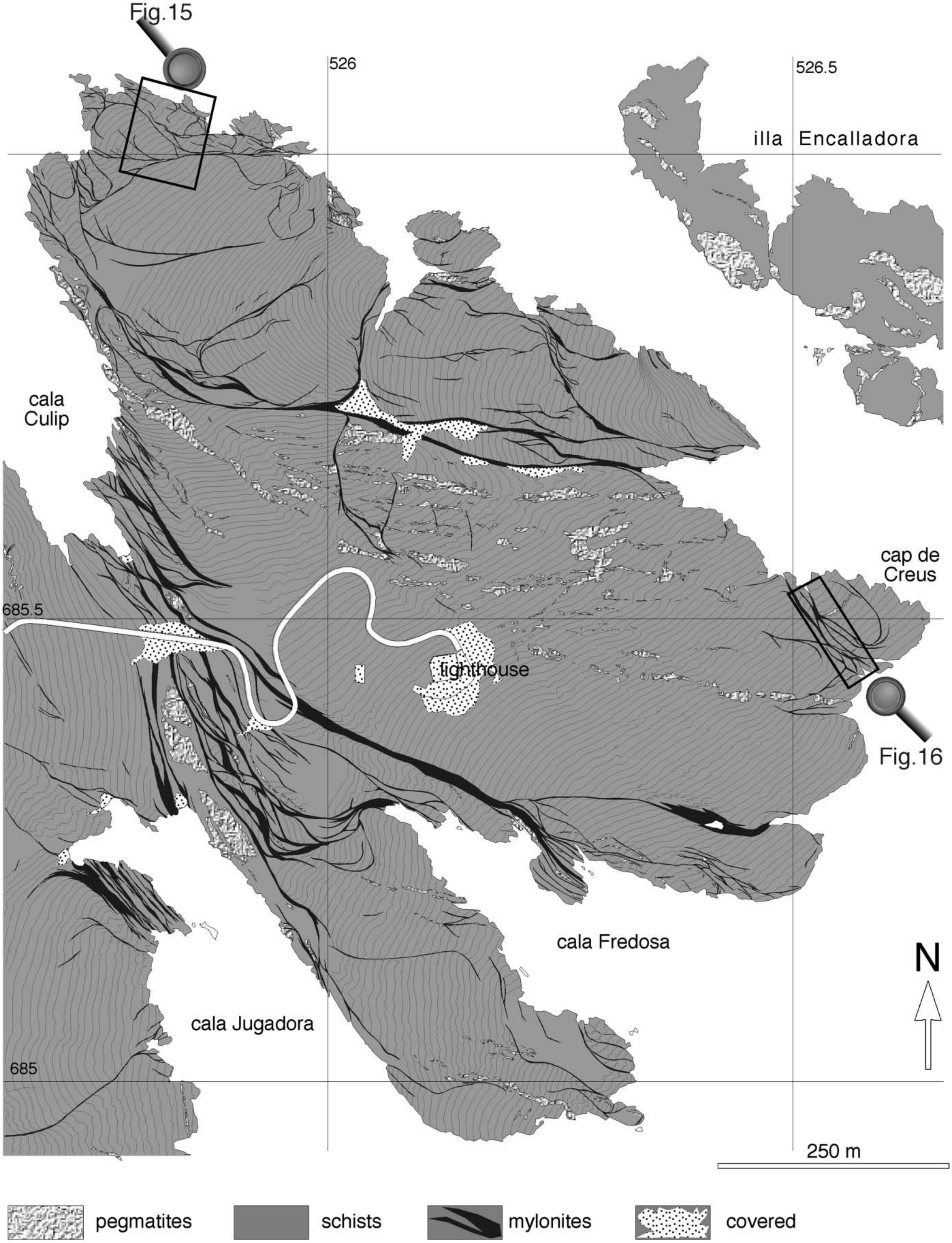


Fig. 7. Map of the Cap de Creus area. This domain is characterised by the predominant conjugate dextral shear zones cutting across the SW–NE to N–S trending S_1-layer parallel foliation.

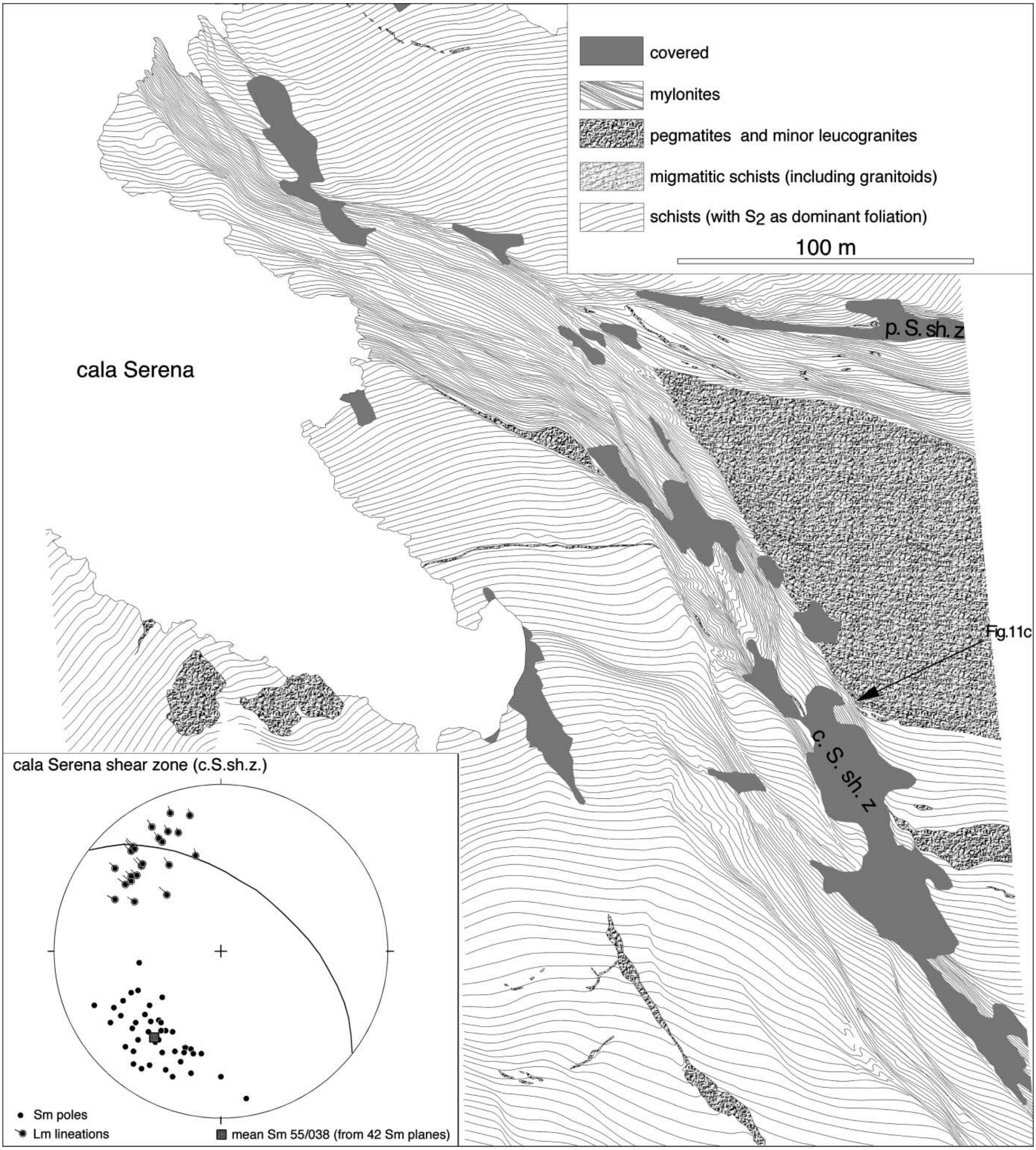


Fig. 8. Cala Serena shear zone (c.S. sh. z.), a typical NW–SE trending anastomosed shear zone with predominant dextral shearing and a minor reversal component. Prat de Serena shear zone (p.S. sh. z.) on the upper right corner of the figure is a kinematically complex shear zone. Note the extensional crenulation-shaped arrangement of the mylonitic foliation in the c.S. sh. z. Stereogram only shows c.S. sh. z. data. Mapping performed on an aerial photograph and contains topography distortions.

(phyllonites). The development of these internal structures is restricted to shear zones where there is evidence (e.g. oblique stretching lineations, changes in deflection sense across or along the shear zone) of a complex kinematic history.

4. Bulk structure and kinematics of the northern Cap de Creus shear belt

Two features are important in assessing the significance

Fig. 9. Outcrop photograph showing a double deflection of the S_2 foliation into parallelism with two adjacent shear zones at Cala Prona shear zone. Structural analysis of this shear zone is shown in Fig. 21. Sm: mylonitic foliation. Width of view: 2.1 m

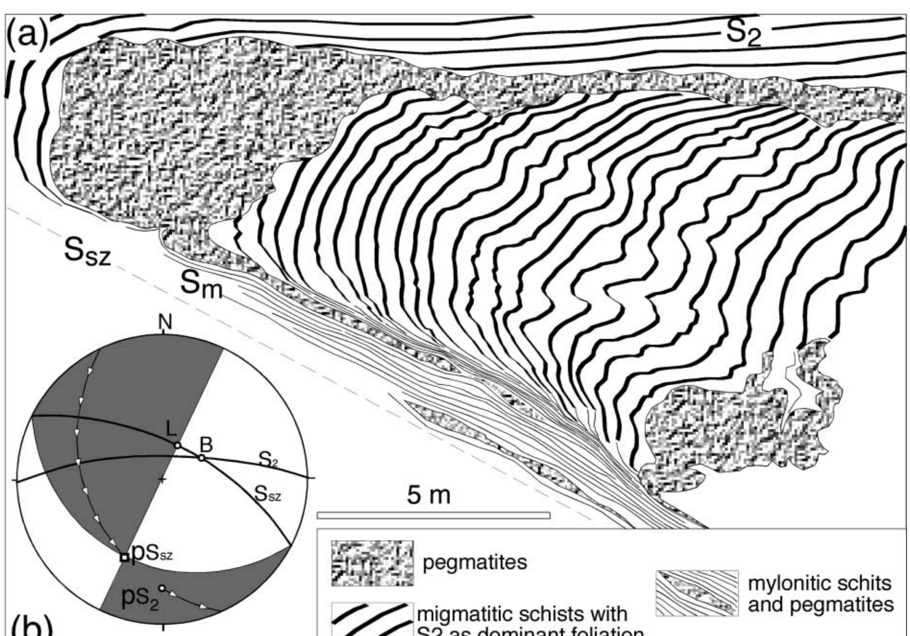


Fig. 10. (a). Detailed map of an apparent sinistral deflection of the S_2 foliation and pegmatite dykes in a dip-slip shear zone. (b). Stereographic plot showing the relation between geometrical and kinematic elements in this shear zone: the angle between the bending axis and the shear direction is about 10°. Foliation with poles plotted in the shaded area will develop apparent sinistral deflections in plane view.

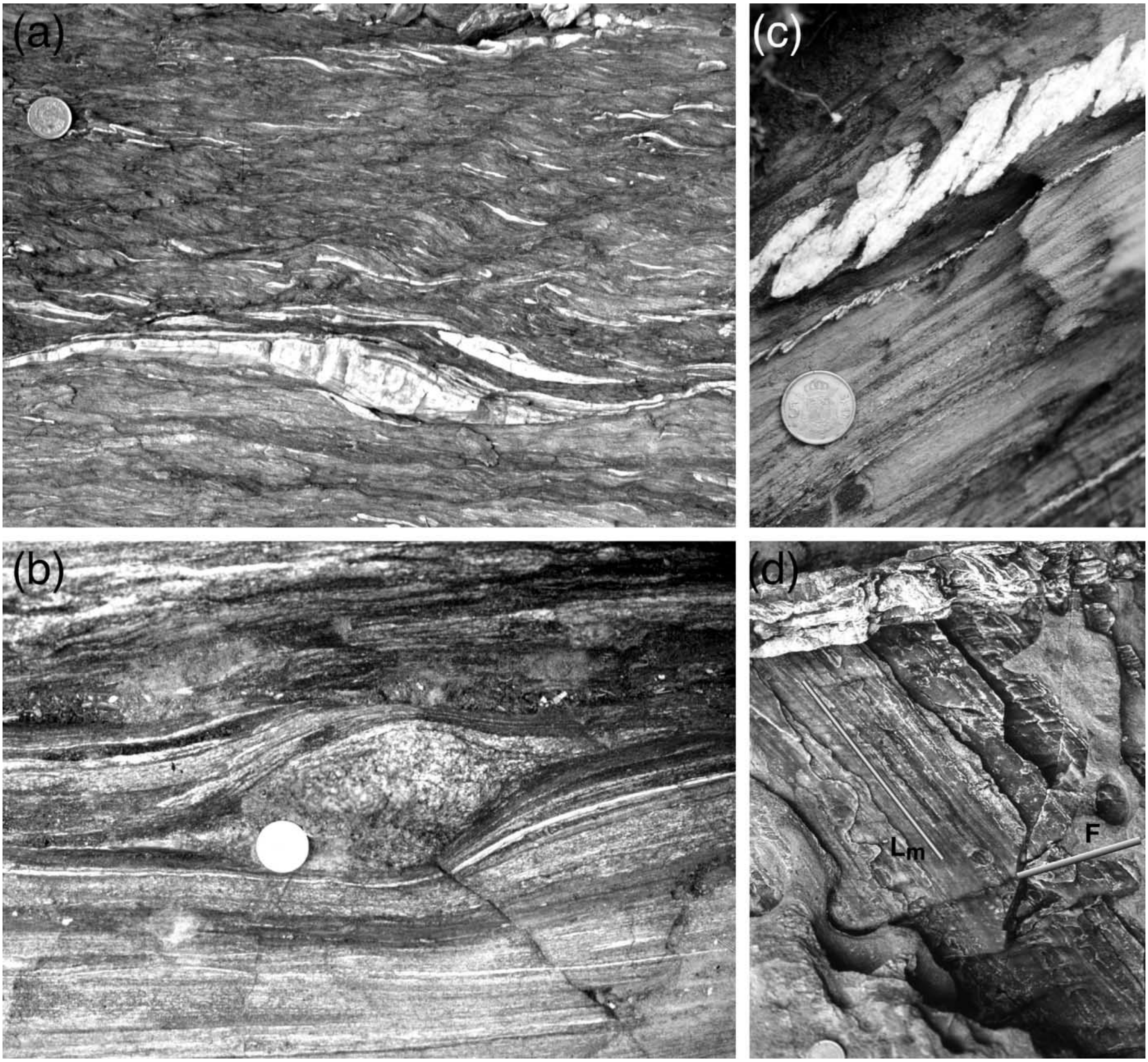


Fig. 11. Minor structures developed in the shear zone interior. (a) Extensional crenulation cleavage developed in a phyllonitic mylonite from a shear zone with a complex kinematic history (plane view, North to the top, width of view: 37 cm). (b) Asymmetric granitic pod in a mylonite affecting migmatites (plane view, north to the top, width of view: 32 cm; see Fig. 6 for location). (c) Intramylonitic folds developed on a quartz–feldspathic band and minor sized folds developed in thin quartz bands (vertical section view NNW on the left, width of view: 12 cm; see Fig. 8 for location). (d) Intramylonitic folds developed on a quartz band with axes (F) oblique to the stretching lineation (view looking down the mylonitic foliation plane, width of view: 21 cm; see Fig. 6 for location).

of shear zones of the northern Cap de Creus shear belt: (i) the bulk kinematics of the anastomosing pattern of shear zones in which individual zones with different shear senses coexist, and (ii) the transition southwards from a typical shear zone belt into a complex fold-shear belt.

4.1. Anastomosing patterns of shear zones

Shear zones are commonly arranged in anastomosing patterns (e.g. Ramsay and Allison, 1979; Bell, 1981; Davidson et al., 1995; Corsini et al., 1996; Hudleston, 1999). The anastomosing network of shear zones forming the northern Cap de Creus shear belt is most visible in the NE corner of the Cap de Creus peninsula, because of the high obliquity between pre-existing foliation and mylonitic foliation. Towards the west, the shear zones trend in close parallelism to the pre-existing foliation, making their identification more difficult (Fig. 5).

A global analysis reveals that while foliation shows trends varying from W–E to NNW–SEE with a general N-to-NE dip, the stretching lineation shows a single, well-marked maximum towards the NNW (Fig. 5b). Thus,

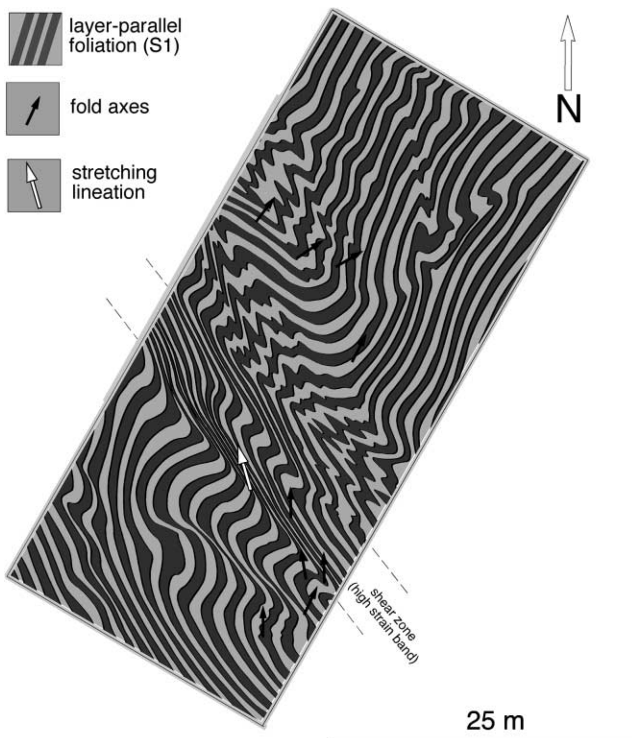


Fig. 12. Shear related folds developed in the transition zone (see Fig. 4a). Relatively less crystalline schists show an increasing fold closure towards the shear zone. These folds affect the S_1 layer-parallel foliation and form in shear zones affecting highly anisotropic rocks.

in E–W trending shear zones the lineation is indicating predominant dip-slip movement, whereas in NW–SE trending shear zones the lineation indicates strike-slip movement, predominantly dextral. A well-defined relationship between pitch and dip direction is evident in Fig. 14.

In spite of the predominance of dextral-reverse shear zones, sinistral-normal shear zones are also very common. These sinistral shear zones trend close to E–W (Fig. 15) but adjacent or coalescent parallel shear zones showing an opposite sense of shear also occur (Fig. 16).

There is no significant difference in orientation of the stretching lineation between shear zones of opposite senses of shear. This means that, regardless of the orientation of the shear zone and its shear sense, there is a fairly constant attitude of the stretching lineation. The mean orientation of the lineation, close to 330/30, is similar to the mean orientation of lineations in mylonites from other domains of the Pyrenean Axial Zone (322/38).

Not all shear zones are strictly coeval, as some E–W trending zones are deflected by the NW–SE ones. However, no clear overprint relationship supporting the existence of distinct shearing events has been found. The E–W trending shear zones commonly display complex deformation histories as reflected by the local presence of slightly differently oriented stretching lineations and abundance of internal minor structures (i.e. intramylonitic folds and widespread extensional crenulation cleavage). This suggests that the anastomosing network formed over a period of time. The E–W shear zones always record the longer and more complex kinematic pattern, while most shear zones trending NNW–SSE record a clear steady dextral movement.

Although the above general statements about the relative age of the shear zones can be made, there is not an absolute rule of timing, as indicated by the presence of associated and likely coeval, dextral and sinistral shears as shown in Figs. 15 and 16. Considering the northern shear zone belt as a whole, the angle between the mean planes of each shear zone set is about 25°. Stretching lineations are not normal to the intersection line between these two planes, but oblique (about 70°) to it (Fig. 5b). If the two sets are considered as conjugate, with the obtuse angle bisected by the maximum shortening direction (Ramsay and Huber, 1987), it follows that the shear zones might have formed under an overall NNE shortening direction, with a moderate southerly plunge. However, the large predominance of dextral-reverse shear zones and the fact that the existing foliation probably plays a significant role in determining the initial orientation of the shear zones, suggests a more complex situation. In addition, it is likely that complex shear zone arrangements arise in part as a result of local accommodation due to space compatibility problems as shear zones die longitudinally out in unsheared domains (Coward, 1976; Ramsay, 1980; Simpson, 1983; Hudleston, 1999).

The anastomosing arrangement of the shear zones gives rise to lozenges of unsheared rock that are strongly prolate fusiform-shaped in three dimensions, their maximum elongation parallel to the stretching lineation. (Fig. 17). Such lozenges exist from kilometric sizes downwards, but they occur most commonly in metric to decametric sizes. The presence of these lozenges of all sizes reflects well the existing pattern of self-similarity. Because the mylonitic foliation strike changes around the lozenges, while the stretching lineation remains fairly constant in orientation, the pitch of the lineation around an individual lozenge varies according to a relation (Fig. 17c) equivalent to that found for the bulk northern Cap de Creus shear zones as a whole (Fig. 14).

The geometry and kinematics of the structures is clarified by zooming into selected domains of the Cap de Creus northern mylonite belt (Figs. 6 and 7, with enlarged details in Figs. 15 and 16). These areas, besides including most of the spectacular shear zones, provide two examples of different shear zone arrangements, likely motivated by different initial orientations of the pre-existing foliation. In the western example (Cala Serena–Punta dels Farallons area, Fig. 6) the pre-existing foliation (S_2) trends WSW–ENE and the NW–SE trending dextral shear zones prevail. On the other hand, in some domains of the eastern example (Cap de Creus, Figs, 7 and 15) the pre-existing foliation (S_1) strikes N–S, and a more complicated pattern of conjugate dextral and sinistral shear zones is formed. This suggests that the orientation of the pre-existing foliation has played an important role in the initiation and evolution of the shear zones as shown by Cobbold et al. (1971), Cosgrove (1976,

Fig. 13. Outcrop photographs of mylonites. (a) Banded mylonite with quartz bands originated by stretching of quartz-segregation nodules and veins enclosed in the precursor schists (north to the top, width of view: 34 cm). (b) Stretching lineation in mylonitised pegmatite marked by the alignment of feldspar porphyroclasts and stretched tourmaline crystals (mylonite foliation plane lies parallel to the surface view, width of view: 30 cm).

1989) or Williams and Price (1990) in experimentally deformed artificially layered materials.

A two-dimensional interpretation is given, which is a simplification of the three-dimensional deformation and kinematic pattern of the northern Cap de Creus shear zones (Fig. 18). The assumption is made of a bulk-wrench or transpressive-wrench geotectonic regime (*sensu* Harland, 1971) in this crustal domain during development of the shear zones. This assumption is justified by the large predominance of dextral-reverse shearing, the development of shear zones and folds involving a broadly NW–SE shearing and the evidence for a minor shortening component normal to the shear zones (see Section 5.3). Assuming either a strictly dextral strike-slip regime or a slight transpressive regime does not significantly modify the proposed interpretation.

Within the framework of the assumed tectonic setting, a crustal domain with pre-existing sub-vertical schistosity would be subject to an instantaneous N–S shortening induced by the bulk NW–SE shearing. Crystalline schists would accommodate this shearing by developing local instabilities, which evolve into shear zones leading to a highly heterogeneous strain distribution. The initial orientation of these instabilities would be governed by the same principal crenulations or folds that develop within anisotropic media (Cobbold et al., 1971; Cosgrove, 1976; 1989; Williams and Price, 1990). In domains where the existing foliation is parallel to the instantaneous stretching direction, two distinct sets of instabilities would form (Figs. 15 and 18), while in domains with foliation oblique to this direction (Figs. 6 and 18) one set will predominate. With progressive shear, these instabilities would evolve into shear zones. While one set is suitable to accommodate progressive shear, the other would be subjected to a reverse sense of shear at a given stage. Also, the angle between those two sets would decrease until dextral and sinistral shear become

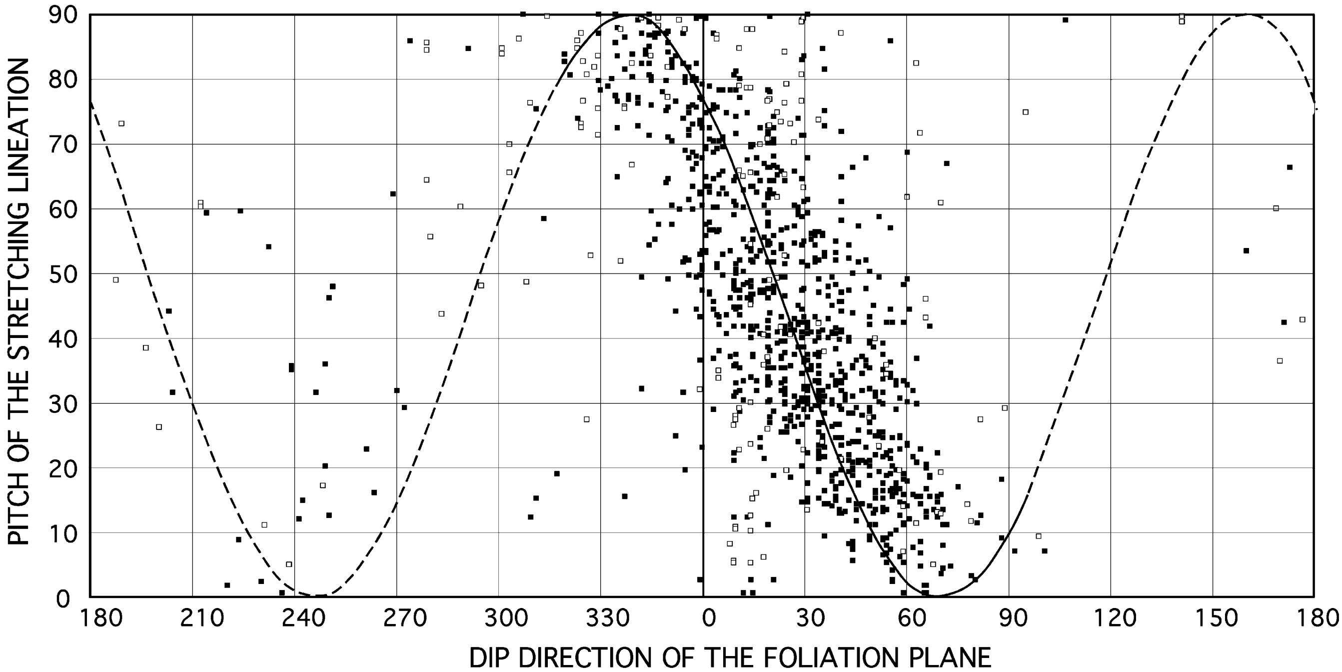


Fig. 14. Pitch/dip correlation diagram. The dip direction of each mylonitic foliation plane is plotted against the pitch of the associated stretching lineation. Black dots correspond to lineations with western dip components, while open circles correspond to lineations with eastern dip components. A good correlation exists between NE dipping strike slip shear zone and NNW dipping dip-slip shear zones. In shear zones with different orientations the pitch/dip correlation is poor (dashed line).

nearly parallel, as shown in Fig. 16. In advanced stages of shearing, only the dextral set would be active and newly nucleating dextral shear zones would form. Continued dextral shear in zones that were always dextral would be responsible for the meter-scale extensional crenulations found in some shear zones (e.g. Fig. 8).

4.2. Association between shear zones and folds at a regional scale

Another remarkable characteristic of the Cap de Creus mylonite belt is that the anastomosing patterns of shear zones pass transversally and laterally into domains of intense folding. This is associated with partial to total transposition of the earlier fabric and, where the shear zones appear as high-strain transposition bands, this occurs with no clearly defined shear sense (Fig. 19). These structural style transitions occur as the shear zones progressively affect the lower-grade metamorphic rocks, resulting in a lower crystallinity and a higher mechanical anisotropy. This was interpreted (Carreras, 1975; Carreras et al., 1980; Carreras and Casas, 1987) as reflecting the different ways through which rocks accommodate regional strain depending on their ability to develop either buckling instabilities (less crystalline schist being more mechanically anisotropic) or ductile shearing (medium-high grade schist and associated igneous rocks, which are more susceptible to strain localisation). At Cap de Creus, there is a gradual change in tectonic style, from highly crystalline rocks close to the northern shore of the peninsula towards less crystalline schists in the central part. The bulk deformation associated with NW–SE wrenching along discrete shear zones is transferred upwards into broader deformation bands with transitional boundaries (Fig. 19). It is assumed that this structural domain experienced a heterogeneous non-coaxial shortening with a N–S directed instantaneous shortening direction (Fig. 20). Thus, as in similar models proposed by Graham (1978) and Iglesias and Choukroune (1980), oblique folding would form at shallow crustal levels together with deep-seated shearing. Progressive deformation leads to: (i) axial plane and fold axis rotation towards a NW–SE trend, and (ii) lateral and longitudinal shear zone propagation, which, in the transition zone, interferes with the early developed folds.

5. Individual shear zones

The individual shear zones are typically sinuous and display heterogeneous strain distributions. As a consequence, transverse profiles are highly variable and strong longitudinal strain gradients exist. Examples of shear zones dying out longitudinally are common (Fig. 21).

Shear zone boundaries show varied geometries, some smoothly deflecting the pre-existing foliation (Figs. 9 and 21); others display sharp boundaries and crosscutting relationships with pre-existing structures. Changes in the boundary may occur along the same margin of a shear zone.

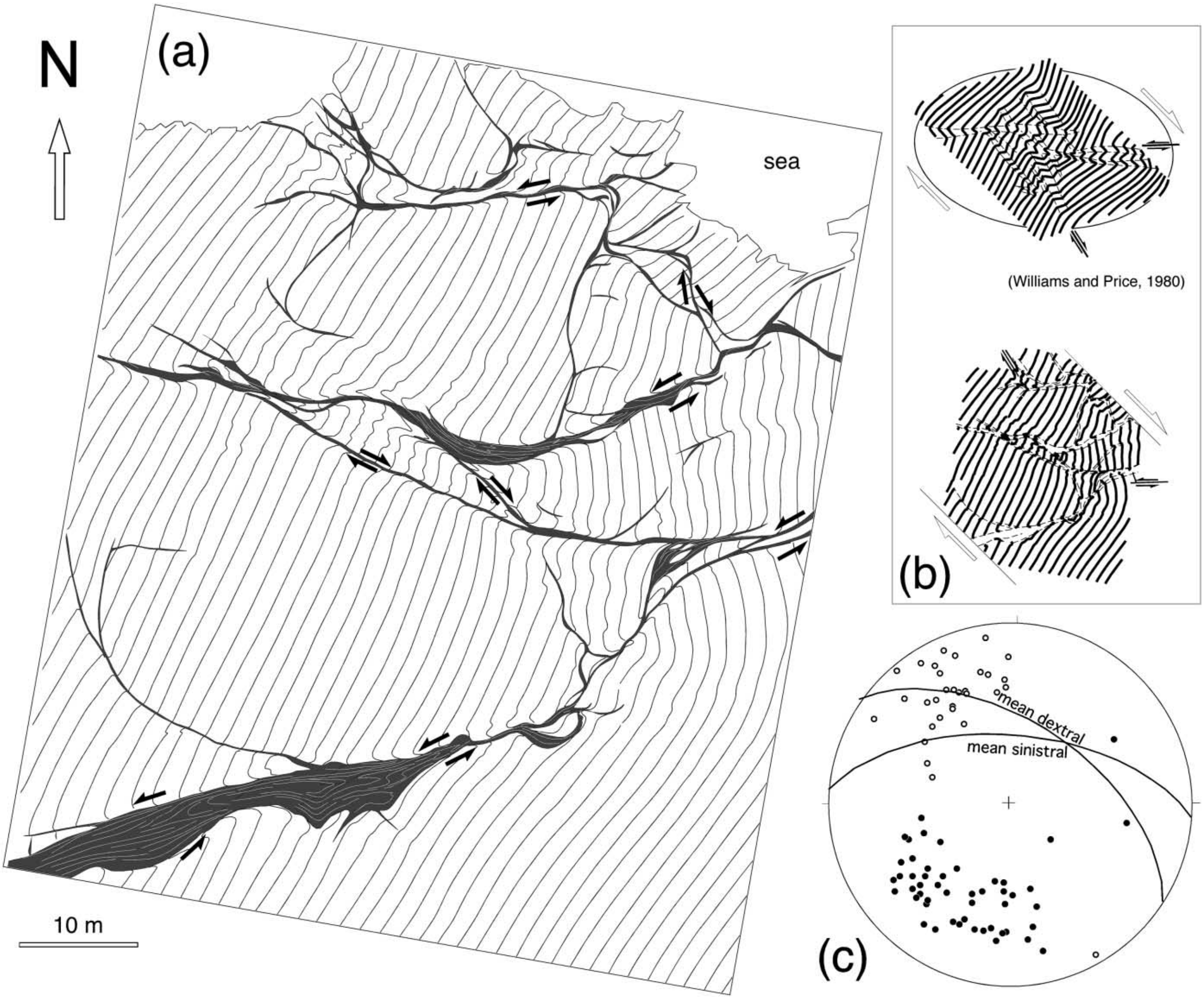
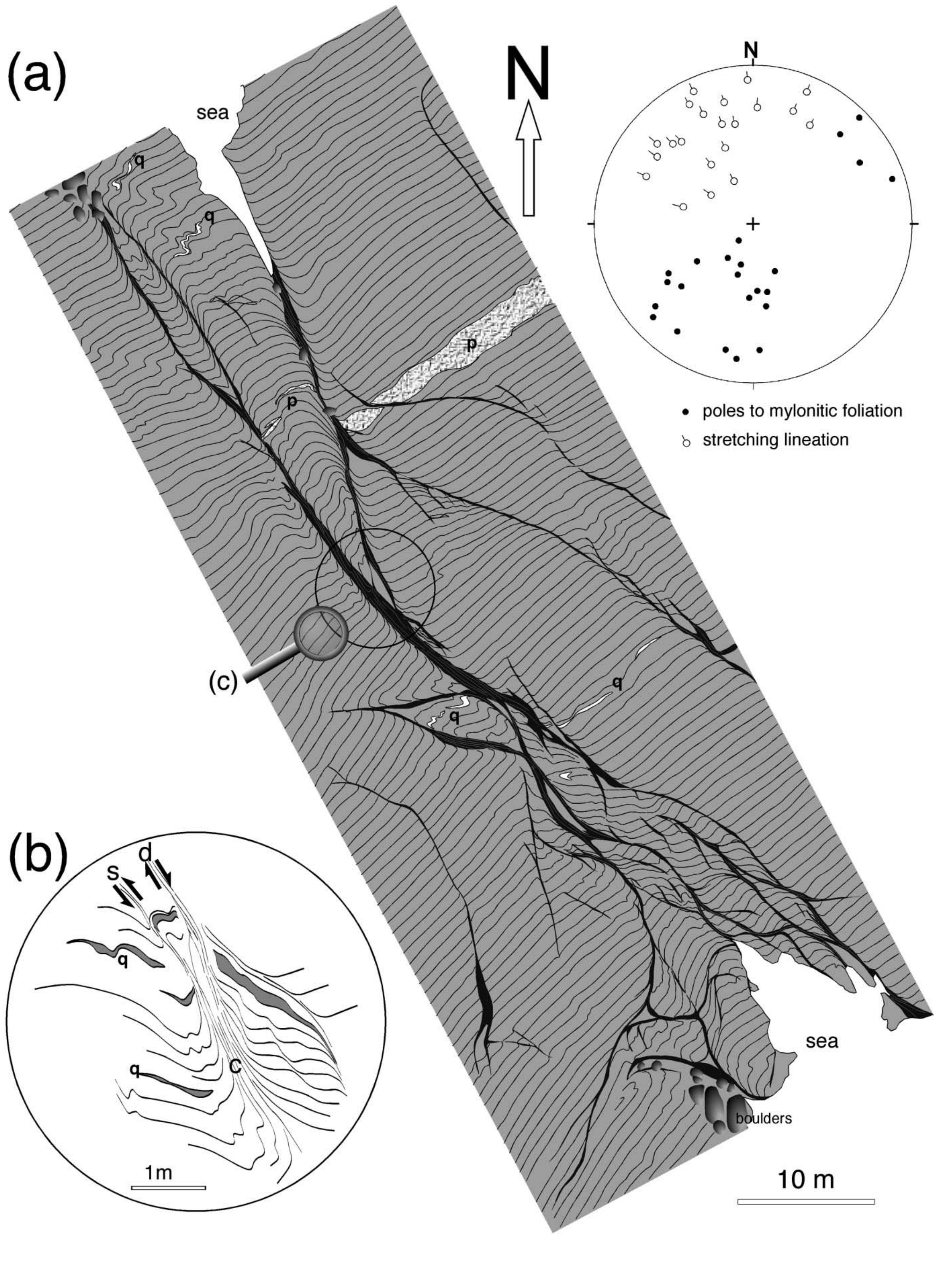


Fig. 15. (a) Detail map of a domain in the Cap de Creus area (Fig. 7) showing a set of variably trending minor shear zones with dextral and sinistral senses of shear. This area shows conjugate shear zones closely symmetric relative to the trend of the pre-existing foliation. Mapping performed on an aerial photograph and may contain topography distortions. Note in (b) the similarities between this pattern and the experimental conjugated structures developed by Williams and Price (1990) by simple shearing of anisotropic materials. Finite strain ellipse is also shown. (c). Stereographic lower hemisphere, equal area plot of the mylonitic foliation planes and associated stretching lineations.

A main characteristic of shear zones developed in foliated rocks is the variable relationship between geometry and kinematics, depending on the initial orientation of kinematic vectors and structures. The pre-existing foliation plane rotates towards parallelism to the shear zone around an axis that might form a variable angle to the shear direction. As a result, anastomosing networks formed by slightly differently oriented shear zones cut across domains with a variable orientation of the pre-existing foliation. In consequence, a wide spectrum of angular relationships can be found (e.g. fig. 10.4 in Carreras, 1997). If deflection of the pre-existing foliation is the result of passive shearing, the resulting geometry will be cylindrical with a rotation axis dependent on the relative orientation of foliation and shear plane, but independent of shear direction. In the Cap de Creus shear zones, however, foliation deflection was non-passive, as indicated by the development of minor structures as the pre-existing foliation was deformed. These minor structures are folds and crenulations or extensional crenulations or both, and their orientation is governed by the sectional strain ellipse (Ramsay, 1980; Skjernaa, 1980). These structures rotate with increasing shear strain.

An analysis of shear zone boundaries reveals that slight departures from the ideal simple shear model occur. Using the pre-existing foliation as markers, in a large section of a shear zone, an increase in shear zone parallel thickness (i.e. the thickness between two selected pre-existing foliation surfaces measured parallel to the shear zone) is observed (Fig. 21c). This increase is most apparent in low strain domains (Carreras, 1997). The bulk thickness increase is inhomogeneous and is


(a)
sea
q
q
p
p
N
poles to mylonitic foliation
stretching lineation
(c)
q
q
(b)
d
s
q
q
c
1m
sea
boulders
10 m

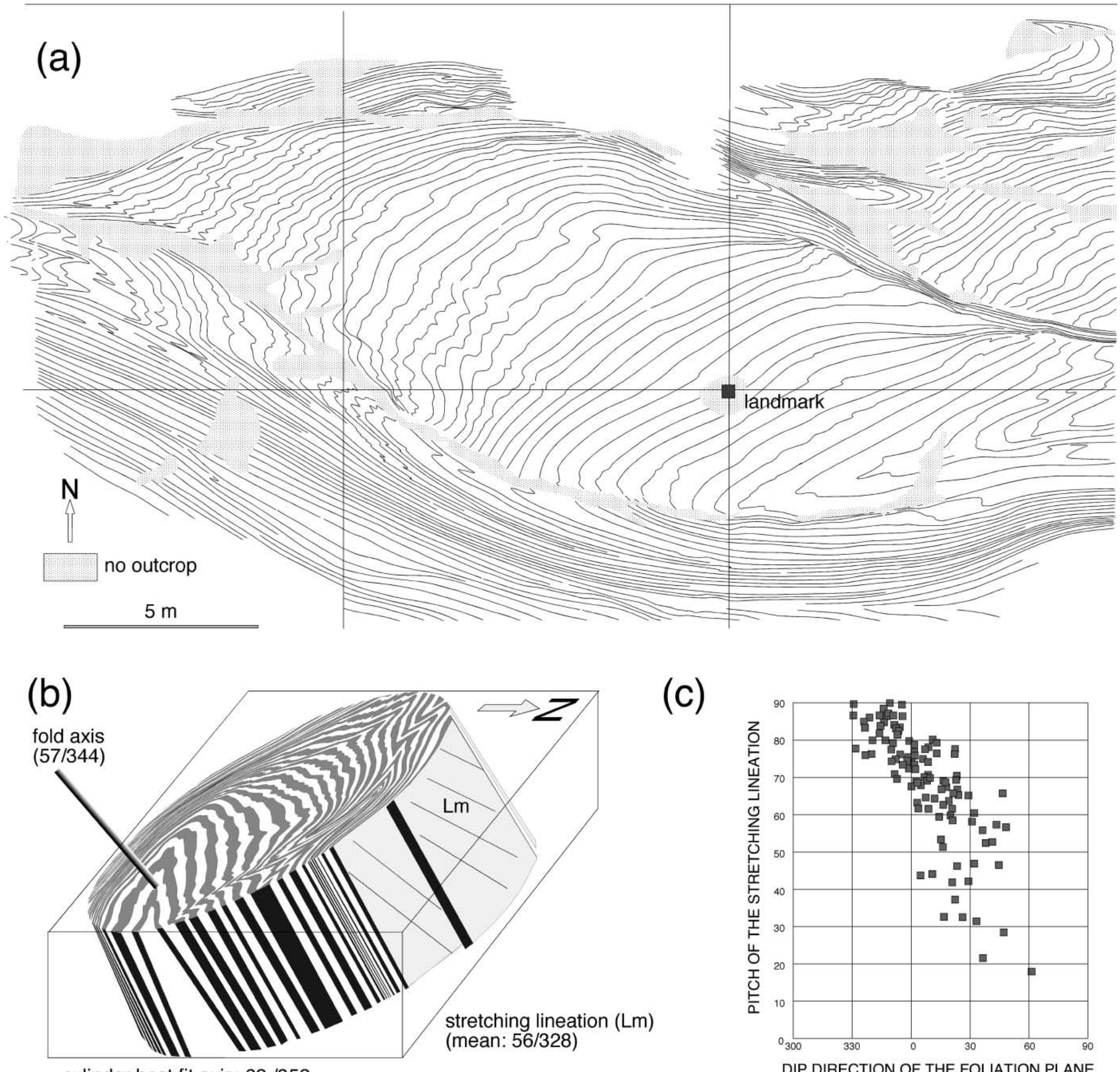


Fig. 17. Structure of a schistose lozenge trapped in an E–W trending dip-slip shear zone. (a) Detail map of the structure. Note the development of folds affecting the pre-existing foliation which becomes isoclinal and transposed towards the lozenge tails, and the wrapping mylonitic foliation. (b) Sketch showing the close parallelism of fold axes and stretching lineation. The whole structure is slightly conical and broadens downwards. (c) The pitch/dip correlation follows a similar pattern to that shown in Fig. 14, the prevalence of higher pitches, indicating a dominant dip-slip shearing.

most marked between surfaces bounding metagreywacke layers, while in metapelitic layers a decrease in shear zone parallel thickness may sometimes occur. This situation is reflected by the zigzag pattern of the isogons (Fig. 21c). This geometry suggests that, at shear zone margins, deflection of the layer parallel foliation is controlled by the most competent layers and, to keep strain compatibility, an antithetical layer parallel flow along mica-rich planar domains occurs.

Fig. 16. Detail map of a domain in the Cap de Creus area (Fig. 7) containing sets of subparallel minor shear zones with dextral and sinistral movements. Note that the coalescence of antithetical shear zones produces an odd marginal geometry shown in enlarged box figure (b). This framework may represent a more advanced stage than the one described in Fig. 15, as the conjugate shear zone continuously rotated towards parallelism. Dextral shearing predominates in the merging set. Mapping performed on an aerial photograph and may contain topography distortions.

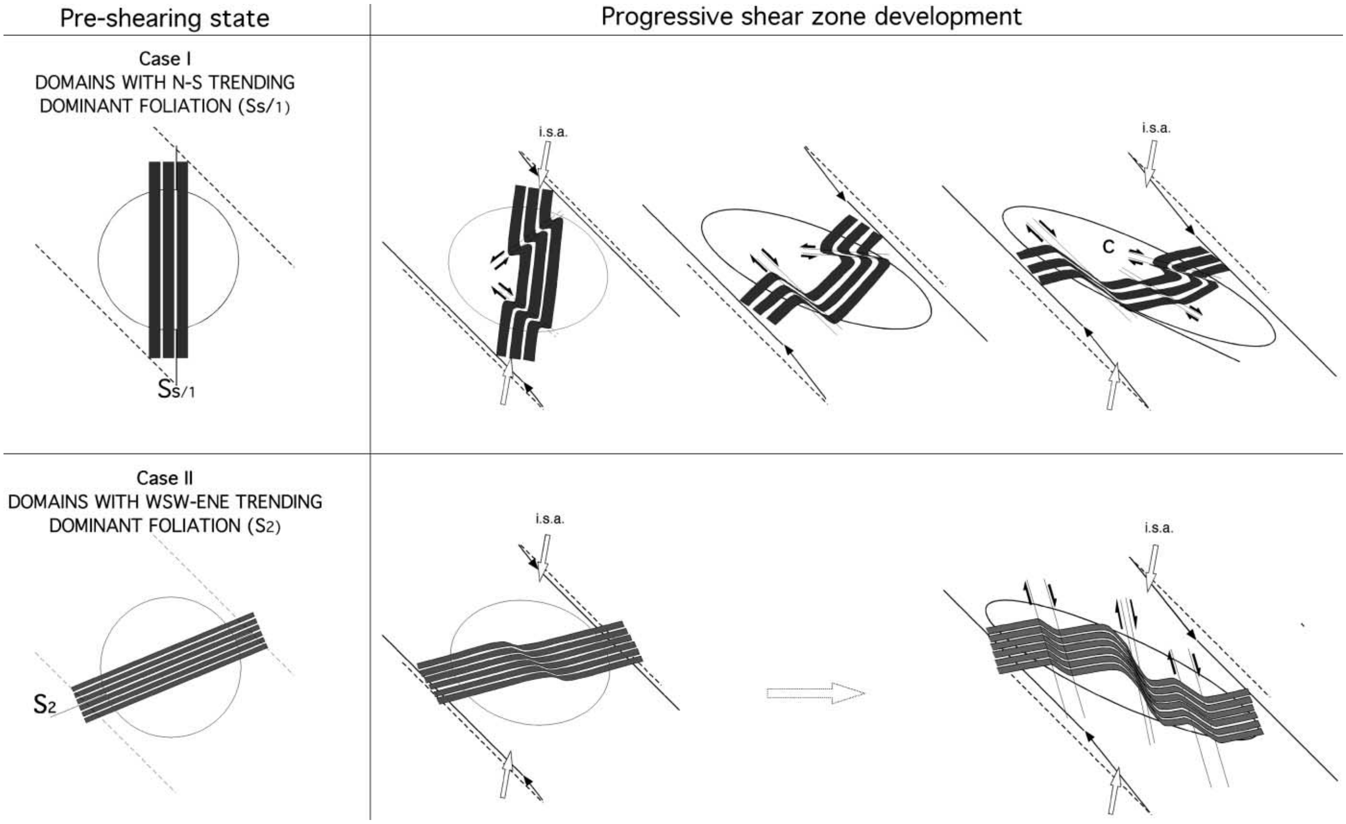


Fig. 18. Two-dimensional interpretation for the localisation and evolution of the northern Cap de Creus shear zones as a function of the nature and orientation of pre-existing foliations. It is assumed a NW–SE wrench dominated, slightly transpressive tectonic regime. In domains with conjugate shear zones, the antithetic set will be gradually less active, and even invert the sense of shear in late stages, giving rise to kinematically complex shear zones (labelled 'c'). Symbols: i.s.a. instantaneous shortening axis. Ellipses represent finite bulk strains. Case I represents the situation shown in Fig. 7, while Case II corresponds to structures shown in Fig. 6.

5.1. Towards the shear zone interior: minor structures, microstructures and microfabrics

These structures are most recognisable in low strain marginal domains, while towards the shear zone interiors they become transposed or obliterated as a result of high shear strains. Marginal folds are best developed in domains where shear zones cut across lesser crystalline, well-foliated schists. These folds initially form with axial planes and fold axes oblique to the shear plane and to the stretching lineation, respectively. Parallelism is approached with increasing strain (Figs. 12 and 19). It is expected that shear zones showing abundant marginal fold development are those which deviate more from ideal simple shear.

Extensional crenulations are common microstructures developed in the marginal low strain domains. They form at a low angle (25–32°) to the pre-existing foliation in agreement with the values given by Platt and Vissers (1980). When the pre-existing foliation lies at low angles to the *XY* plane of the incremental strain ellipsoid, two sets of extensional crenulation cleavages, ecc1 and ecc2, may form (Fig. 22). Towards the shear zone interior, the ecc2 set tends to disappear. The remaining set is the one expected to form by oblique stretching of a pre-existing foliation whose asymmetry is consistent with the shear sense (Fig. 23). The stable extensional crenulation cleavage (ecc1) at the shear zone margins is commonly closely parallel to the overall trend of the shear zone and could be considered a 'C' plane (terminology after Berthé et al., 1979). However, this parallelism is considered as coincidental, rather than a general rule. Towards the high-strain domains, rotated and newly formed crenulations and folds coexist. These latter structures arise from new instabilities. Folds are better displayed in sections normal to the stretching lineation, while extensional crenulations are most evident in parallel sections. It must be remarked that there is a continuous spectrum of extensional crenulation cleavages relative to their orientation to the shear zone and the nature of the affected foliation surface (varying from pre-existing foliation to a completely newly formed mylonitic foliation). As a consequence, the conventional S–C terminology should be avoided for such fabrics.

When data on the quartz *c*-axis fabrics of the Cap de Creus shear zones were first published (Carreras et al., 1977), the observed rotation of the quartz girdles moving towards the shear zone boundary emerged as controversial (Fig. 24). Roermund et al. (1979) and Lister and Williams (1983) proposed a strain partitioning interpretation, which

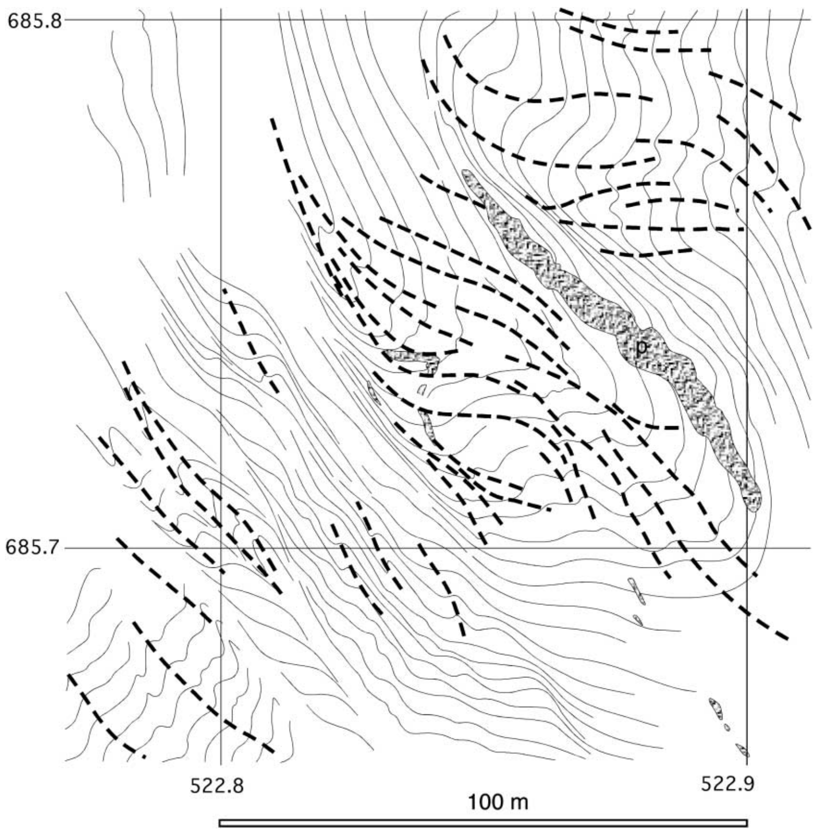


Fig. 19. Shear zones and high strain transposition bands developed within the transition zone. The S_1 foliation becomes transposed into a mylonitic foliation. Symbols: dashed lines: axial traces of shear-related folds, p: pegmatite.

apparently contradicted the idea that the fabric was controlled by the simple shear kinematics, and girdles should remain stable with respect to the shear plane. Although that strain partitioning interpretation would still receive large consensus, available evidence suggests that fabric rotation would occur in a more complex manner than that proposed by Lister and Williams (1983). This topic is further documented in Section 5.3.

5.2. Internal structures

A main characteristic of mylonitic rocks is the development of penetrative foliated and/or well-banded fabrics. Distinctive, more competent bands are formed by thin quartz–feldspathic or quartzitic layers derived from the deformation of pegmatites and quartz nodules, respectively. Banding is the result of mechanical stretching without any chemical differentiation (Myers, 1978; Carreras and Druguet, 1994; Carreras et al., 1997b). Because of the high strains achieved, these quartz–feldspathic rocks become transformed into thin layers or lenses, commonly with a corrugated surface, corrugations being parallel to the stretching lineation and sometimes defined by morphologic alignment of minerals.

The presence of rocks with variable strengths caught in the shear zones causes the development of lozenges and pinch-and-swell structures. These structures are asymmetric and their shapes reflect the sense of shear in sections parallel to the shear direction. In these structures there is a strong strain gradient from the centre towards the edges and tails. Knots of pegmatites display nearly isotropic cores and penetratively foliated boundaries. The banding and the mylonite foliation are commonly folded (intramylonitic folds), and/or extended by an extensional crenulation cleavage (referred to as the ecc1 set; Figs. 11 and 23). Intramylonitic folds and extensional crenulation cleavage, although widespread, do not require a particularly high amount of shear strain to form (White et al., 1980). While they may be absent in shear zones recording deformation with strain ratios exceeding 500/1, they may appear in most low-strain domains. However, these intramylonitic structures are always present in domains where there was a disturbance of the shear flow. Such a disturbance is reflected by complex mylonitic foliation patterns in domains where the shear zones coalesce. Flow disturbances also occur in shear zones recording changes in shear direction or shear sense, as indicated by oblique stretching lineations, or by the coexistence of opposite shear sense indicators. Thus, these internal structures can be considered to reflect a new pulse of shearing, involving the reactivation of an already formed mylonitic foliation. The development and evolution of intramylonitic folds has been treated by Carreras and Santanach (1973) and by Carreras et al. (1977).

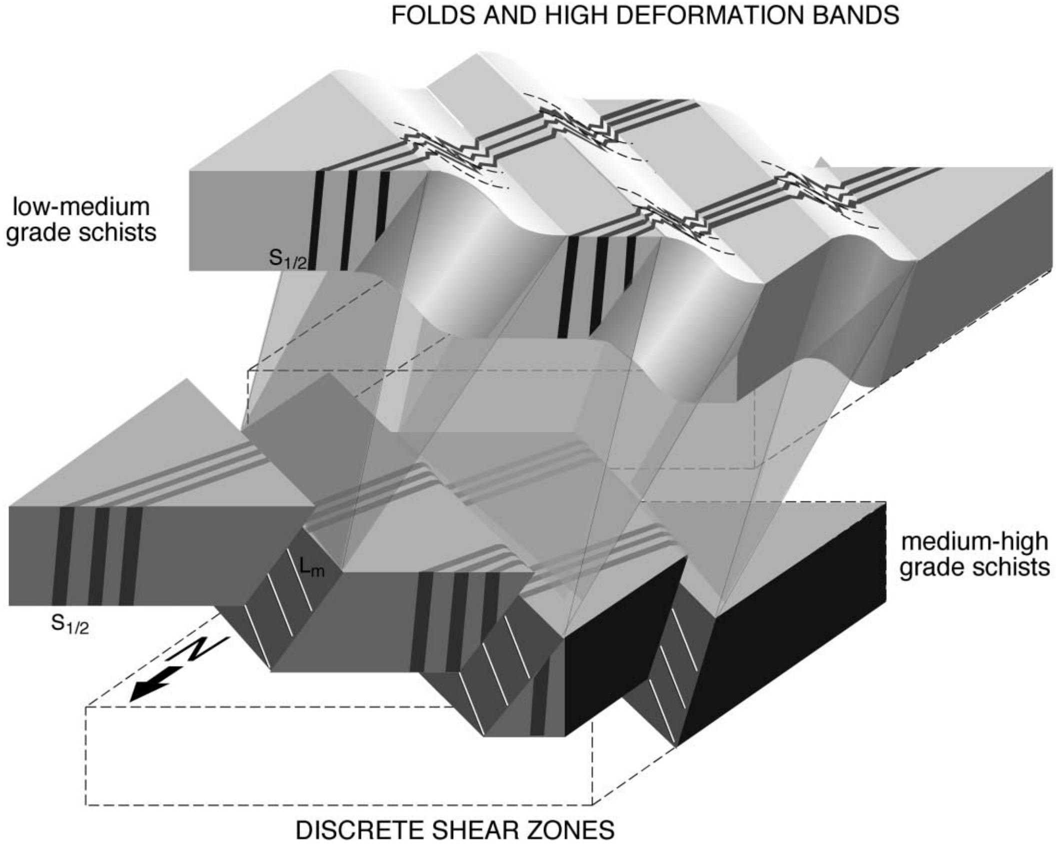


Fig. 20. Model showing the change from domains where dominant wrenching is accommodated by discrete shear zones, towards domains where strain is accommodated by broader deformation bands involving non-coaxial folding. At Cap de Creus this transition occurs southwards.

The observations suggest that differently named structures (shear bands, s–c cleavages, extensional crenulations), as well as normal crenulations and folds in shear zones, are all the result of instabilities formed during re-shearing of a pre-existing planar anisotropy, which can be either a pre-existing, a partially transposed, or a newly formed foliation during the shearing event.

5.3. A model for shear zones of the Cap de Creus type

The Cap de Creus shear zones bear marked longitudinal strain gradients. Such shear zones must bear other strain components other than a unique shear, using the strain compatibility criterion of Ramsay and Graham (1970) in the opposite sense. Volume loss can be ruled out, considering the lack of evidence for material transfer in highly crystalline schists, which deformed predominantly by dislocation creep, and without the action of solution transfer as a deformation mechanism. In addition, volume loss would not explain the observed changes in shear-zone parallel thickness.

The zigzag isogon pattern indicates that the pre-existing foliation was not passively sheared. Instead, shearing developed from instabilities in ways likely similar to those noted in experiments on deformed anisotropic materials (Cobbold et al., 1971; Cosgrove, 1976; 1989; Williams and Price, 1990). This indicates that shear zone formation involves, at least in the initial stages, transverse shortening, accommodated by means of buckling and involving a considerable amount of strain partitioning.

The proposed behaviour is somewhat analogous to the model postulated by Lister and Williams (1983), although with a significant difference. There is no need to develop a coaxial stretching in competent domains to keep a bulk simple shear strain regime. The layer-parallel coaxial stretching that Lister and Williams (1983) proposed in order to maintain a bulk simple shear situation seems unlikely to occur in the analysed setting, where the competent layer might experience initial shortening and later extension in two dimensions, with deformation being internally accommodated by a complex non-coaxial strain. Structures like those in Fig. 12, cannot be explained by layer parallel coaxial stretching of competent layers during shear zone development.

The preferred interpretation for shear zone development is schematically summarised in Fig. 25. It is assumed that deflection at the shear zone margins starts as a buckling instability, controlled not only by a mechanical anisotropy represented by the pre-existing foliation, but also by the variably transposed metagreywacke beds which act as competent layers.

The low mica content and massive character of the metagreywackes enables us to assume that the internal

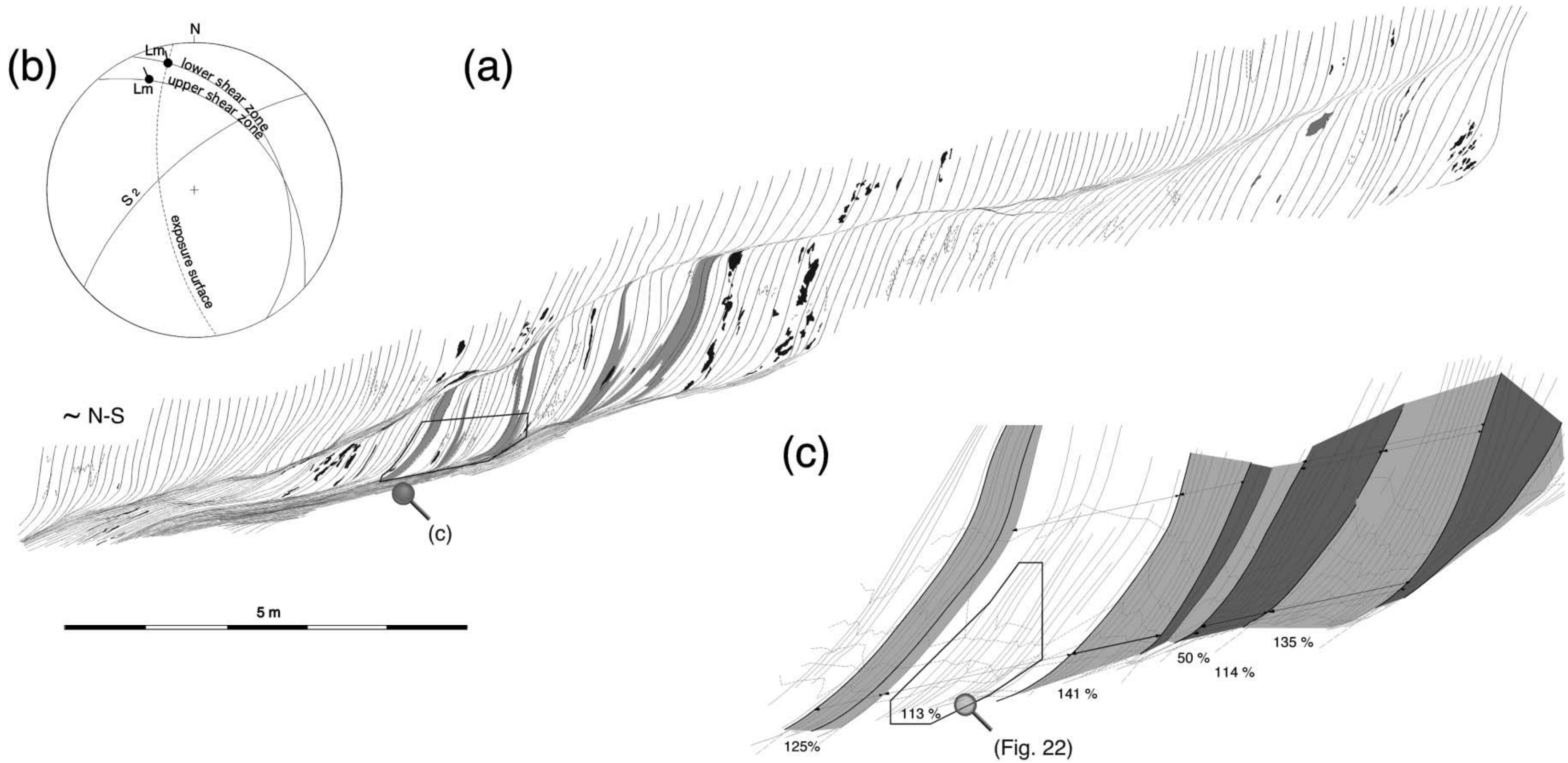


Fig. 21. Geometry of the Cala Prona shear zones. (a) Sketch showing traces of the sheared S_2-foliation. Some folded layers have been drawn (in grey) representing mica-rich bands. Black objects represent segregated quartz nodules. Some markers (L1 to L8) have been drawn to visualise the lateral variation of shear. (b) Lower hemisphere, equal area stereogram showing the arrangement of surfaces and lineations in (a). Note that the sub-vertical exposure surface corresponds to a shear parallel section. (c) Zigzag pattern of the S_2-isogons, reflecting almost parallel-fold geometry of some metagreywacke layers (light grey) in low strain margins of the shear zone. Dark grey bands correspond to predominately pelitic layers, while white domains correspond to intermediate lithologies. Percentages indicate the increase or decrease of parallel thickness of some layers.

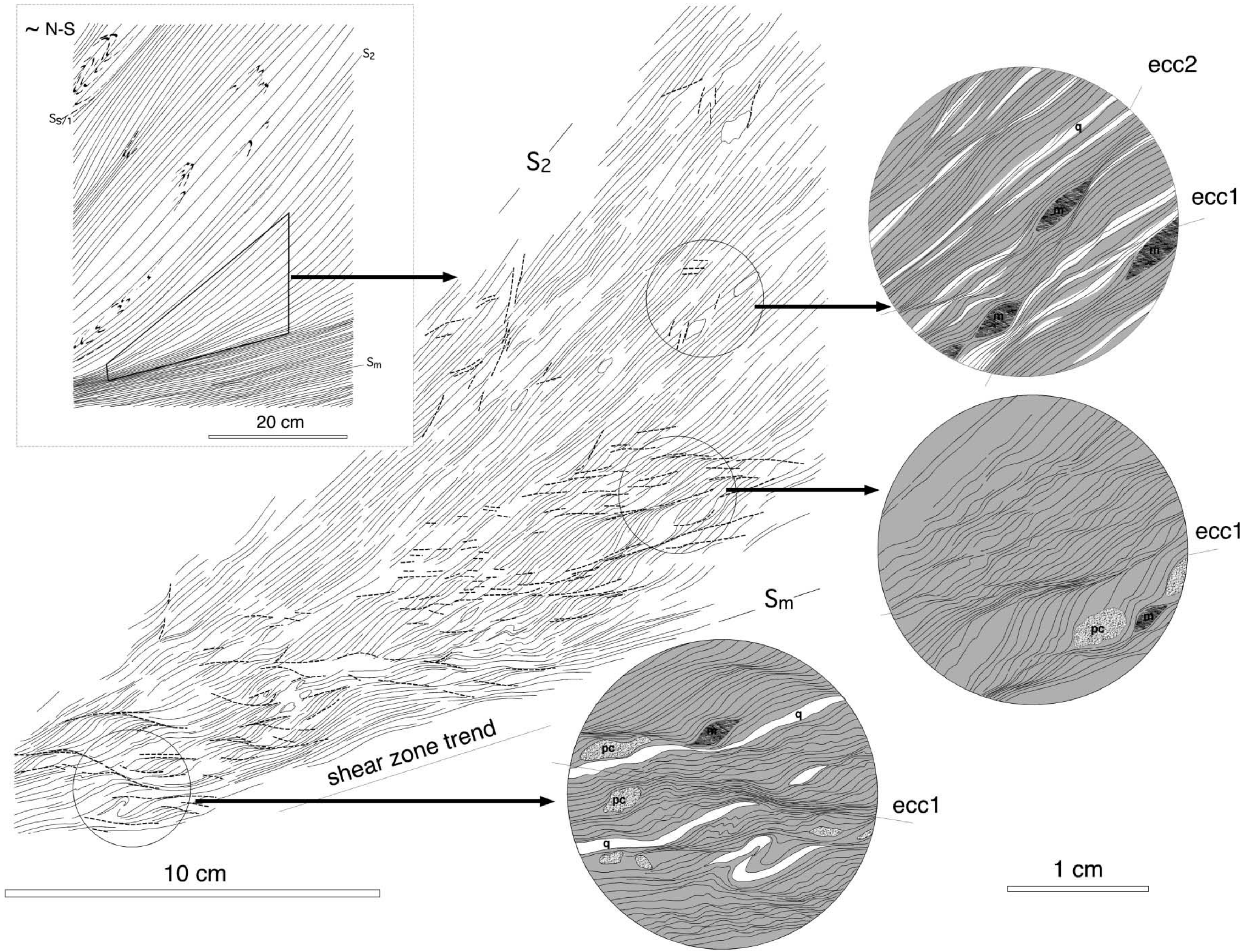


Fig. 22. Development of extensional crenulations sets ecc1 and ecc2 in the margin of the Cala Prona shear zone shown in Figs. 9 and 21 (m: mica fishes; pc: porphyroclasts, mainly feldspars; q: quartz-ribbons). $S_{s/1}$: layer parallel foliation; S_2: transposition foliation; Sm: mylonitic foliation.

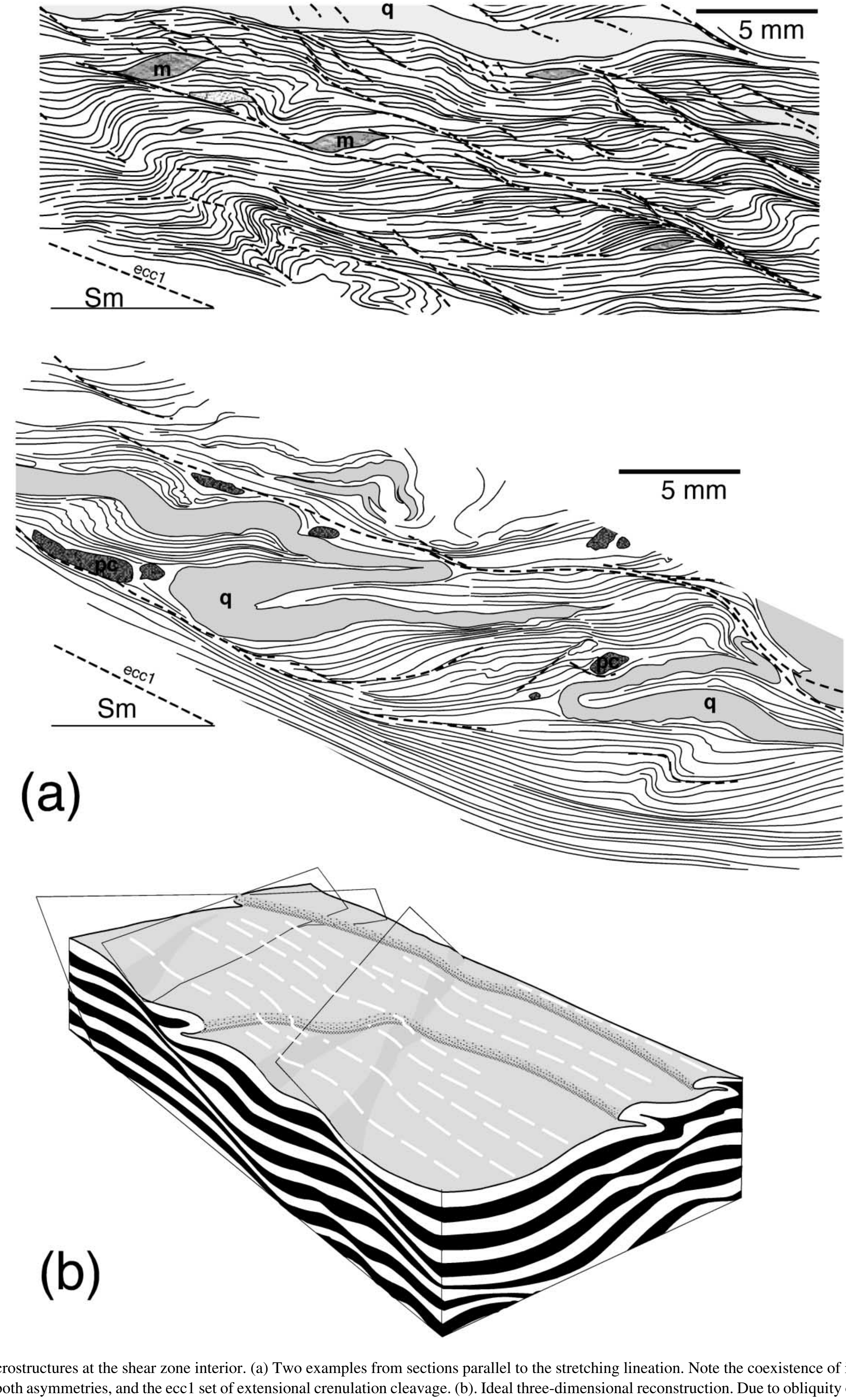
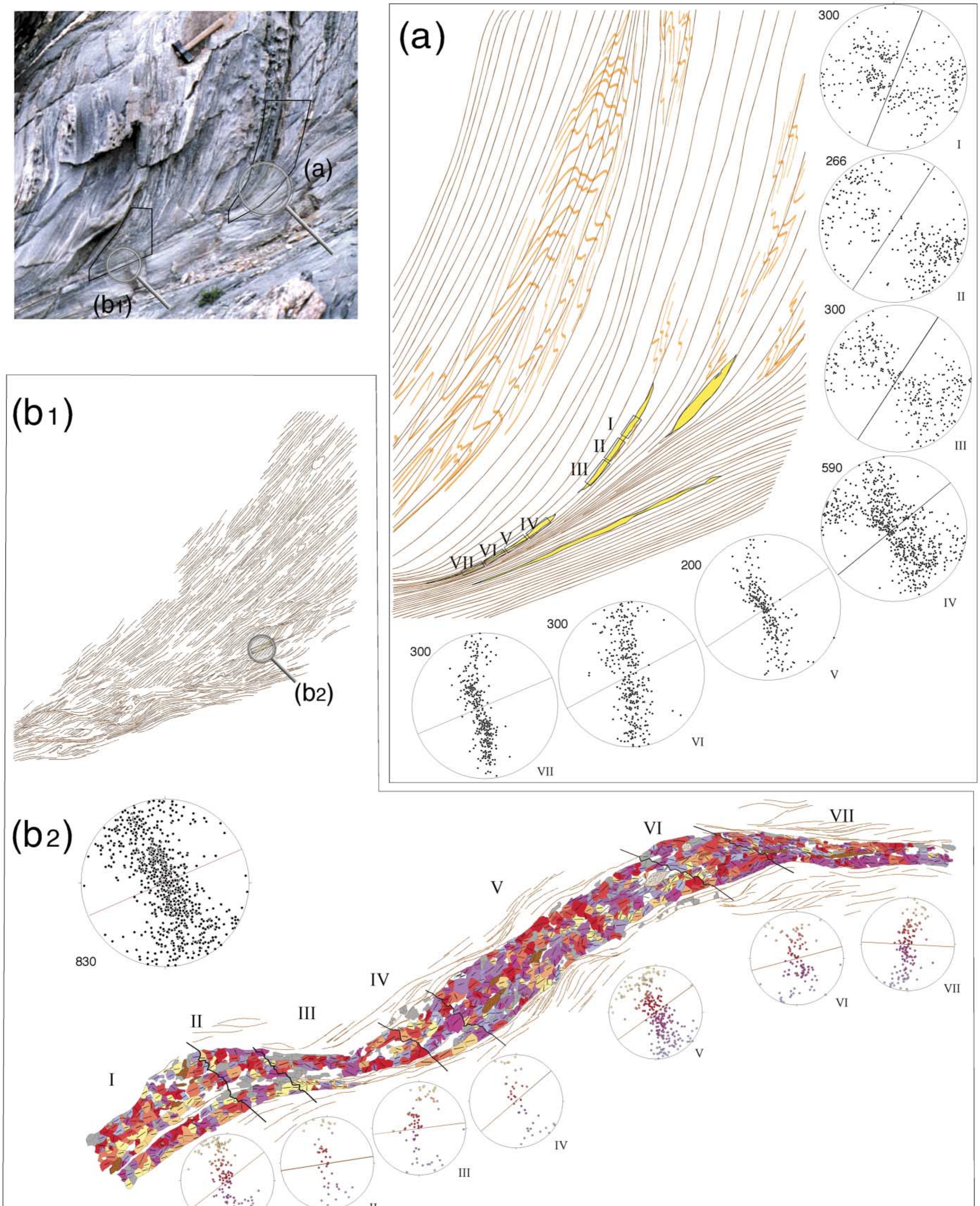


Fig. 23. Microstructures at the shear zone interior. (a) Two examples from sections parallel to the stretching lineation. Note the coexistence of intramylonitic folds, with both asymmetries, and the ecc1 set of extensional crenulation cleavage. (b). Ideal three-dimensional reconstruction. Due to obliquity of ecc1 planes with respect to the stretching lineation, on sections normal to the lineations, the ecc1 set might appear as forming a conjugate sets framework.


(a)
(b1)
(b2)
300
I
266
II
III
590
IV
200
V
VI
VII
830

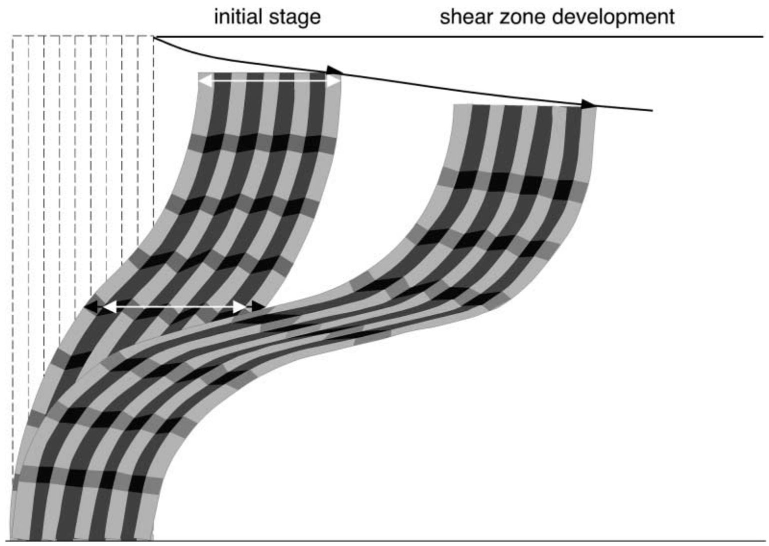


Fig. 25. A two dimensional model for the development and evolution of Cap de Creus shear zones. At the initial stage, pre-existing foliation and layering are deflected as a consequence of a buckling instability. It is assumed that competent layers (light grey) accommodate strain by tangential longitudinal strain, while in the mica-rich accommodative layers (dark grey) internal deformation might involve flexural-slip or flow. At initial stages, bulk shearing is not accompanied by coaxial stretching of competent layers; as a consequence, marked departure from a simple shear exists. This deviation from the heterogeneous simple shear model is also outlined by the increase in the parallel thickness (black arrow). It is suggested that as strain increases, accommodation gradually approaches a simple shear strain regime, with strain localisation in the central deflected domains.

deformation of the metagreywacke layers at the very initial stages of layer deflection is accommodated by an internal deformation closer to tangential longitudinal strain, than to passive shearing or flexural flow. This assumption, in agreement with Hudleston et al. (1996), is corroborated by the observation of crenulations that developed exclusively at the internal interface of banded metagreywackes at shear zone boundaries. It is assumed that metapelites behave as accommodation layers undergoing a complex deformation, most likely involving flexural flow, with occasional evidence of layer-parallel slip along some mica-rich lithological boundaries. Some domains with no marked compositional contrast, like the domain shown in Fig. 22, seem to closely approach the simple shear model. However, as a whole, strain distribution is far from homogeneous in longitudinal profiles. The existence of selective layer-parallel shearing seems the most likely explanation for the depicted quartz *c*-axis fabrics, which display a congruent girdle rotation (e.g. see Fig. 24; Carreras et al., 1970; Roermund et al., 1979; Carreras and García-Celma, 1982). Quartz *c*-axis fabric analysis reveals that the whole fabric pattern, as well as the girdle elements, are very sensitive to strain partitioning. In general, grains contributing to the girdle asymmetry pattern, whose *c*-axes lie at high angles to the foliation plane, appear to be controlled by the presence of ecc planes, (Fig. 24b; Carreras and García-Celma, 1982). This observation is in accordance with descriptions by García-Celma (1982, 1983) on Cap de Creus mylonites, and by Casas (1982) in other mylonites of the Eastern Pyrenees.

The variation in shear-zone parallel thickness at the margins of the shear zones has been stressed; it must also be noted that towards their interior, the geometry of the shear zones approaches that expected for heterogeneous simple shear. It is likely that, once the instability has formed, simple shearing takes over in the way the deformation is accommodated (Fig. 24). In the proposed interpretation, shear-like structures with transverse shortening can form without stretching along the intermediate finite strain axis, nor considering volume loss. Thus, in analogy to the millipede models of Bell (1981), it is proposed that some shear zones develop in a complex heterogeneous deformation process, which includes shearing components but also transverse shortening. It has often been suggested that this situation is unlikely, because of space problems due to incompatibility (e.g. Simpson and De Paor, 1997), but most arguments have neglected that the problems of generating a simple shear zone or a sub-simple shear zone within a

Fig. 24. Quartz *c*-axis variations at the margins of the Cala Prona shear zone shown in Figs. 9 and 21. (a) Quartz *c*-axis variation in a quartz vein affected by a shear zone. The congruent rotation of the girdles is attributed to strain partitioning, while the variable shape of the girdles is related to the strain intensity. The asymmetry also depends on the presence and orientation of extensional crenulation cleavage sets. (b) A.V.A. from a quartz-ribbon affected by the ecc1 dominant set. The cross-girdle fabric is a composite *c*-axis including different microstructural domains, where *c*-axes at a high angle to the mylonitic foliation are located in domains affected by the extensional crenulation cleavage. Grey scale represents the variation of the angle between the *c*-axis and the foliation plane (dark grey: low angle; light grey: high angle). Lines in each grain represent the trace of the basal plane. Modified from Carreras and García-Celma (1982). Numbers in the stereograms indicate the number of grains measured.

confined domain are of the same order. It is easier to generate a heterogeneous simple shear by moving undeformed rock domains in unconfined media, but as one considers boundary conditions, and as shear zones terminate, thus necessitating longitudinal strain gradients, one must not discount the possibility of plane strain or almost plane strain shear zones with transverse shortening. It should be noted that it is confusing to use the terminology of transpression for this proposed situation, as well as for other settings that include shortening transverse to the zone. The transpression is less confusing when it is restricted to the original use of Harland (1971), where geographical reference is most relevant.

6. Concluding remarks

The Cap de Creus area offers one of the best examples of shear zones developed in foliated rocks. The existing types of ductile shear zones show how the mid-crustal rocks accommodate deformation through different structures ranging from discrete ductile shear zones in isotropic rocks to shear zones associated with buckling structures. The main characteristics of these shear zones are the complex geometric–kinematic relationships and the control of the pre-existing foliations on the localisation and evolution of the shear zones. Depending on the orientation of the pre-existing foliation, one or two sets of shear zones may develop. During shearing, the pre-existing foliation does not behave passively, but leads to the development of instabilities, which commonly leads to the development of a self-similar pattern of shear zones in a wide range of sizes. In such shear zones, transverse shortening is likely and strain compatibility is achieved through a bulk anastomosed pattern and by remarkable longitudinal shear strain gradients. In spite of the existence of transversal shortening components, the use of the term ‘transpressive’ does not seem appropriate to describe these shear zones. The general kinematic arrangement of the Cap de Creus shear belt suggests that the shear belt developed in the framework of a bulk transpressive tectonic regime. This is corroborated by the dominant dextral-reverse shear zones, which implies that the two sides of the belt converge with a dextral component of movement.

The existence of self-similar geometries and kinematic patterns over a wide range of scales suggests that all the investigated structures arose as instabilities governed by the same mechanical controls, where processes other than ductile simple shear are involved in the development of shear zones in the middle crust. New shear zone models should explain satisfactorily, and in a unified way, the strain distribution in space and time of banded zones deviating from simple shear, to produce structures that range from the crenulation size to the crustal scale.

Acknowledgements

The author wishes to thank Peter Hudleston, whose comments on geology and English allowed a great improvement of the manuscript. Elena Druguet is also thanked for her careful review and constructive comments. The Department of Geology and Geophysics of the University of Minnesota, where this paper was written, is also thanked for the facilities provided. The stay at this institution was financially supported by the DGESIC (Ministerio de Educación y Ciencia) in Spain. This contribution benefited from the use of data obtained from different studies performed in the Pyrenees and Cap de Creus in collaboration with M.L. Arboleya, G. Bossière, J.M. Casas, J. Cirés, E. Druguet, A. García-Celma, M. Julivert, M. Liesa, M. Losantos, V. Morales, J. Palau, P: Santanach, J. Soldevila and S. White; and from stimulating discussions held with many other geologists who visited the area. The author wishes to thank Enrico Tavarnelli and an anonymous reviewer for detailed and helpful comments on the manuscript.

References

Bell, T.H., 1981. Foliation development: the contribution, geometry and significance of progressive bulk inhomogeneous shortening. Tectonophysics 75, 273–296.

Berthé, D., Choukroune, P., Jegouzo, P., 1979. Orthogneiss, mylonite and non coaxial deformation of granites: the example of the South Armorican shear zone. Journal of Structural Geology 1, 31–42.

Carreras, J., 1974. Progressive mylonitization in quartzitic and quartz–feldspathic rocks in the shear belts of the Cabo de Creus (E-Pyrenees, Spain). M.Sc. thesis, Imperial College, London.

Carreras, J., 1975. Las deformaciones tardi-hercínicas en el litoral septentrional de la península del Cap de Creus (prov. Gerona, España): la génesis de las bandas miloníticas. Acta Geológica Hispánica 10, 109–115.

Carreras, J., 1992. Zonas de cizalla en rocas foliadas: Geometría y cinemática. Simposios II Congreso Geológico de España Salamanca 2, 355–364.

Carreras, J., 1997. Shear zones in foliated rocks: geometry and kinematics. In: Sengupta, S. (Ed.). Evolution of Geologic Structutres in Micro- to Macro-scales. Chapman and Hall, London, pp. 185–201.

Carreras, J., Capellà, I., 1994. Tectonic levels in the Palaeozoic basement of the Pyrenees: a review and a new interpretation. Journal of Structural Geology 16, 1509–1524.

Carreras, J., Casas, J.M., 1987. On folding and shear zone development: a mesoscale structural study on the transition between two different tectonic styles. Tectonophysics 135, 87–98.

Carreras, J., Druguet, E., 1994. El papel de las zonas de cizalla en la configuración estructural del complejo migmatítico del sector septentrional de la península del Cap de Creus (Girona). Revista de la Sociedad Geológica de España 7, 21–29.

Carreras, J., García-Celma, A., 1982. Quartz *c*-axis fabric variation at the margins of a shear zone developed in schists from Cap de Creus (Spain). Acta Geológica Hispánica 17, 137–149.

Carreras, J., Losantos, M., 1982. Geological setting of the Roses granodiorite (E-Pyrenees, Spain). Acta Geológica Hispánica 17, 219–225.

Carreras, J., Santanach, P., 1973. Micropliegues y movimiento en los cizallamientos profundos del cabo de Creus (prov. Gerona). Estudios Geológicos 29, 439–450.

Carreras, J., Estrada, A., White, S., 1977. The effects of folding on the c-axis fabrics of a quartz–mylonite. Tectonophysics 39, 3–24.

Carreras, J., Julivert, M., Santanach, P., 1980. Hercynian Mylonite Belts in the Eastern Pyrenees: an example of shear zones associated with late folding. Journal of Structural Geology 2, 5–9.

Carreras, J., et al., 1996. Debat., P. (Coord.), Tectonique hercynienne. In: Barnolas, A., Chiron, J.C. (Eds.), Synthese geologique et geophysique des Pyrénées. BRGM-ITGE, pp. 585–677.

Carreras, J., Casas, J.M., Losantos, M., 1997a. Some mesoscopic aspects of mylonitic shear zones. In: Snoke, A.W., Tullis, J., Todd, V.R. (Eds.). Fault-related Rocks—A Photographic Atlas. Princeton University Press, Princeton, NJ, pp. 234–235.

Carreras, J., Casas, J.M., Losantos, M., 1997b. Banding and foliation in mylonites. In: Snoke, A.W., Tullis, J., Todd, V.R. (Eds.). Fault-related Rocks—A photographic Atlas. Princeton University Press, Princeton, NJ, pp. 296–297.

Casas, J.M., 1982. Pseudo-two girdles *c*-axis fabric patterns in a quartz–feldspar mylonite (Costabone granodiorite, Canigó massif). Acta Geológica Hispánica 17, 151–157.

Casas, J.M., 1986. Shear bands and related extensional structures in a mylonitized quartz dyke. Journal of Structural Geology 8, 693–699.

Choukroune, R., Mattauer, M., 1978. Tectonique des plaques et Pyrénées: sur le fonctionnement de la faille transformante Nord-Pyrénéenne: comparaison avec les modèles actuels. Bulletin Société Géologique de France 7 20 (5), 689–700.

Cobbold, P.R., Cosgrove, J.W., Summers, M., 1971. The development of internal structures in deformed anisotropic rocks. Tectonophysics 12, 23–53.

Corsini, M., Vauchez, A., Caby, R., 1996. Ductile duplexing at a bend of a continental-scale strike-slip shear zone: example from NE Brazil. Journal of Structural Geology 18, 385–394.

Cosgrove, J.W., 1976. The formation of crenulation cleavage. Journal geological Society London 132, 155–178.

Cosgrove, J.W., 1989. Cleavage, folding and the finite strain ellipsoid. Proceedings Geological Association 100, 461–479.

Coward, M.P., 1976. Strain within ductile shear zones. Tectonophysics 34, 181–197.

Davidson, I., McCarthy, M., Powel, D., Torres, H.H.F., Santos, C.A., 1995. Laminar flow in shear zones: the Pernambuco Shear Zone, NE-Brazil. Journal of Structural Geology 17, 149–161.

Delaperrière, E., Saint Blanquat, M., Brunel, M., Lancelot, J., 1994. Géochronoloige U–Pb sur zircons et monazites dans le massif du Saint Barthélémy (Pyrénées, France): discussion des âges des événements varisques et pré-varisques. Bulletin Société Géologique de France 165, 101–112.

Druguet, E., 1997. The structure of the NE Cap de Creus peninsula. Relationships with metamorphism and magmatism. Ph.D. thesis, Universitat Autònoma de Barcelona.

Druguet, E., Hutton, D.H.W., 1998. Syntectonic anatexis and magmatism in a mid-crustal transpressional shear zone: an example from the Hercynian rocks of the eastern Pyrenees. Journal of Structural Geology 20, 905–916.

Druguet, E., Passchier, C.W., Carreras, J., Victor, P., den Brok, S., 1997. Analysis of a complex high-strain zone at Cap de Creus, Spain. Tectonophysics 280, 31–45.

Eeckhout, B.v.d., Zwart, H.J., 1988. Hercynian crustal-scale extensional shear zone in the Pyrenees. Geology 16, 135–138.

Evans, N.G., Gleizes, G., Leblanc, D., Bouchez, J.L., 1997. Hercynian tectonics in the Pyrenees: a new view based on structural observations around the Bassiès granite pluton. Journal of Structural Geology 19, 195–208.

García-Celma, A., 1982. Domainal and fabric heterogeneities in the Cap de Creus Quartz mylonites. Journal of Structural Geology 4, 443–456.

García-Celma, A., 1983 *C*-axis and shape fabrics in quartz mylonites of Cap de Creus (Spain); their properties and development. Ph.D. thesis, Institute of Earth Sciences, University of Utrecht.

Gleizes, G., Leblanc, D., Bouchez, J.L., 1997. Variscan granites of the Pyrenees revisited: their role as syntectonic markers of the orogen. Terra Nova 9, 38–41.

Graham, R., 1978. Wrench faults, arcuate fold patterns and deformation in the southren French Alps. Proceedings Geological Association 89, 125–142.

Guitard, G., Geyssant, J., Laumonier, B., 1980. Analyse structurale des gneiss oeillés blastomylonitiques du versant nord du Canigou (Pyrénées Orientales); présence de calcite dans ces roches. Comptes Rendues Academie Sciences, Paris 290, 1103–1106.

Harland, W.B., 1971. Tectonic transpression in Caledonian Spitsbergen. Geological Magazine 108, 27–42.

Hartevelt, J.J.A., 1970. Geology of the Upper Segre, Valira Valleys, Central Pyrenees. Andorra/Spain. Leidse Geologische Mededelingen 45, 167–236.

Hippertt, J., 1999. Are *S–C* structures, duplexes and conjugate shear zones different manifestations of the same scale-invariant phenomenon?. Journal of Structural Geology 21, 975–984.

Hudleston, P.E., 1999. Strain compatibility and shear zones: is there a problem? Journal of Structural Geology 21, 923–932.

Hudleston, P.E., Treagus, S.E., Lan, L., 1996. Flexural flow folding: does it occur in nature? Geology 24, 203–206.

Iglesias, M., Choukroune, P., 1980. Shear zones in the Iberian Arc. Journal of Structural Geology 2, 63–68.

Julià, R., Santanach, P., 1997. Banded structures in gouge. In: Snoke, A.W., Tullis, J., Todd, V.R. (Eds.). Fault-related Rocks—A photographic Atlas. Princeton University Press, Princeton, NJ, pp. 56–57.

Lister, G.S., Williams, P.F., 1983. The partitioning of deformation in flowing rock masses. Tectonophysics 92, 1–33.

Losantos, M., Palau, J., Sanz, J., 1986. Considerations about Hercynian Thrusting in the Marimanya Massif (Central Pyrenees). Tectonophysics 129, 71–79.

McCaig, A.M., Miller, J.A., 1986. $^{40}Ar–^{39}Ar$ Age of mylonites along the Mérens Fault, Central Pyrenees. Tectonophysics 129, 149–172.

Myers, J.S., 1978. Formation of banded gneisses by deformation of igneous rocks. Precambrian Research 6, 43–64.

Norton, M., 1982. The kinematic and microstructural development of some shear zones. Ph.D. thesis. Imperial College, London.

Passchier, C.W., 1982. Pseudotachylite and the development of ultramylonite bands in the Saint-Barthélémy Massif French Pyrenees. Journal of Structural Geology 4, 69–79.

Passchier, C.W., Williams, P.F., 1996. Conflicting shear sense indicators in shear zones; the problem of non-ideal sections. Journal of Structural Geology 18, 1281–1284.

Platt, J.P., Vissers, R.L.M., 1980. Extensional structures in anisotropic rocks. Journal of Structural Geology 2, 397–410.

Ramsay, J.G., 1980. Shear zone geometry: a review. Journal of Structural Geology 2, 83–99.

Ramsay, J.G., Allison, I., 1979. Structural analysis of shear zones in an Alpinised Hercynian granite, Schweizerische. Mineralogische und Petrographische. Mitteilungen 59, 251–279.

Ramsay, J.G., Graham, R.D., 1970. Strain variations in shear belts. Canadian Journal of Earth Sciences 7, 786–813.

Ramsay, J.G., Huber, M.I., 1987. The Techniques of Modern Structural Geology. Volume 2: Folds and Fractures. Academic Press, London.

Raymond, D., 1986. Tectonique tangentielle varisque dans le Paléozoïque supérieur de l'Est des Pyrénées françaises: l'example du pays de Sault (Nord du granite du Querigut), Aude et Ariège) et des régions voisines. Bulletin Société géologique de France 8 3, 479–485.

Roermund, H.L.M.v., Lister, G.S., Williams, P.F.W., 1979. Progressive development of quartz fabrics in a shear-zone from Monte Mucrone, Sesia–Lanzo Zone, Italian Alps. Journal of Structural Geology 1, 43–52.

Saillant, J.P., 1982. La faille de Mérens (Pyrénées Orientales) microstructures et mylonites. 3eme cycle thesis Université Paris VII.

Saint Blanquat, M., 1993. La faille normale ductile du massif du Saint Barthélémy. Evolution hercynienne des massifs nord-pyrénées catazonaux considérée du point de vue de leur histoire thermique. Geodinamica Acta 6, 59–77.

Simpson, C., 1983. Displacement and strain patterns from naturally occurring shear zone terminations. Journal of Structural Geology 5, 497–506.

Simpson, C., De Paor, D.G., 1997. Practical analysis of general shear zones using the porphyroclast hyperbolic distribution method: an example from the Scandinavian Caledonides. In: Sengupta, S. (Ed.). Evolution of Geologic Structures in Micro- to Macro-scales. Chapman and Hall, London, pp. 169–184.

Simpson, C., Carreras, J., Losantos, M., 1982. Inhomogeneous deformation in Roses granodiorite. Acta Geológica Hispánica 17, 219–226.

Skjernaa, L., 1980. Rotation and deformation of randomly oriented planar and linear structures in progressive simple shear. Journal of Structural Geology 2, 101–109.

Soler, D., Teixell, A., García-Sansegundo, J., 1998. Amortissement lateral du chevauchement de Gavarnie et sa relation avec les unites sud-pyreneennes. Comptes Rendues Academie Sciences, Paris 327, 699–704.

Soula, J.C., Debat, P., Déramond, J., Pouget, P., 1986a. A dynamic model of the structural evolution of the Hercynian Pyrenees. Tectonophysics 129, 29–51.

Soula, J.C., Lamouroux, C., Viallard, D., Bessière, G., Debat, P., Ferret, B., 1986b. The mylonite zones in the Pyrenees, their place in the alpine tectonic evolution. Tectonophysics 129, 115–147.

Teixell, A., 1996. The Anso transect of the Pyrenees: basement and cover thrust geometries. Journal of the Geological Society London 153, 301–310.

Wheeler, J., 1987. The determination of true shear sense from the deflection of passive markers in shear zones. Journal of the Geological Society London 144, 73–77.

White, S., Burrows, S.E., Carreras, J., Shaw, N., Humphreys, J., 1980. On mylonites in ductile shear zones. Journal of Structural Geology 2, 175–187.

Williams, P.F., Price, G.P., 1990. Origin of kinkbands and shear-band cleavage in shear zones: an experimental study. Journal of Structural Geology 12, 145–164.

Zwart, H.J., 1958. La faille de Mérens dans les Pyrénées Arièges. Bulletin Société géologique de France 6 8, 783–796.