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Hercynian tectonics in the Pyrenees: a new view based on structural observations around the Bassiès granite pluton

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Abstract—The structural study of the metasedimentary rocks surrounding the Bassiès granite pluton allows definition of the succession and kinematics of the Hercynian tectonic events in this part of the Pyrenees, and supports a model of syntectonic emplacement of the pluton. The following sequence of events has been identified: (i) an early phase of N–S compression, essentially resulting in southwards thrusting; (ii) emplacement of the granite body along a NE–SW sinistral strike-slip shear zone which could be a late effect of the first phase; (iii) a major dextral and transpressive phase, responsible for the main fold structures trending, on average, N100°, and following a sigmoidal pattern around the granite body. This succession of events and their kinematics could be extended to the whole Hercynian basement of the Pyrenees. © 1997 Elsevier Science Ltd. All rights reserved

INTRODUCTION

The complex structure of the Hercynian basement of the Pyrenees results from a tectonic evolution which has been the subject of a number of different interpretations. It has been considered as essentially extensional by Wickham and Oxburgh (1985), but most workers agree that the Hercynian structures result either from a single compressional event (Zwart, 1979; Matte and Mattauer, 1987; Carreras and Capella, 1994), or from a compressional event followed by an extensional one (Verhoef *et al.*, 1984; Van den Eeckhout and Zwart, 1988; Vissers, 1992). The major WNW–ESE Hercynian structural trend of the Pyrenees, mostly defined in the field by a pervasive and generally subvertical cleavage or foliation, is called D_2 and considered by most workers as the main compressive phase.

The numerous granodioritic bodies that constitute a large part of the basement of the Pyrenees are still generally considered as post-tectonic. This is principally based on the Permian Rb/Sr ages obtained in most massifs (e.g. Vitrac-Michard and Allègre, 1975; Vitrac-Michard *et al.*, 1980; Majoor, 1988). Among them, the Bassiès granite massif has been dated at 276 Ma (Debon and Zimmerman, 1988). In a previous study focused solely on the internal fabric of the latter granite massif, Gleizes *et al.* (1991) proposed that the pluton was emplaced before the D_2 Hercynian tectonic event. This was in contradiction with the Permian age of the pluton since, according to the stratigraphic record (Nagtegaal, 1969), the end of the Hercynian orogeny occurred in the Pyrenees before the Westphalian *D*. Therefore, it is essential to investigate the structural evolution recorded in the wall rocks of the Bassiès massif in order to confirm its pre-D_2 character. In addition, the structural analysis of both the granite body and its aureole provides a basis for inferring the mode of emplacement of the pluton in relation to the regional deformational events.

The Bassiès area is one of the best places to attempt the unravelling of the Hercynian tectonic evolution of the Pyrenees since: (i) the country rocks have suffered a regional metamorphism of only low greenschist facies; therefore, the contact metamorphism is more clearly imprinted in the rocks, and the 'frozen-in' structural features pre-dating pluton emplacement and coeval with it, are easier to unravel; (ii) Alpine deformation seems to have had little influence in this area. In the present study, the detailed structural analysis of the wall rocks and neighbouring areas of the Bassiès granite confirms the early (pre-D_2) emplacement of the body. This breakthrough in the understanding of a segment of the Hercynian belt of the Pyrenees has a profound bearing in the interpretation of the tectonic evolution of the whole chain.

GEOLOGICAL SETTING OF THE BASSIES AREA

The Bassiès granite massif is located in the Axial Zone of the Central Pyrenees, mostly in France except for its southern fringe which extends into Spain. Its outcrop area covers about 90 km^2 with a parallelogram shape whose long axis trends approximately N60°E (Fig. 1). The northernmost part of the pluton was truncated by the North Pyrenean Fault during the Alpine orogeny. Further north, the North Pyrenean Zone consists of Mesozoic sediments deposited along narrow basins and metamorphosed during the Cretaceous (Choukroune, 1976; Golberg and Leyreloup, 1990).

To the east and south (Fig. 1), the pluton is surrounded by Cambro-Ordovician low grade metasediments (Zwart, 1979). They consist of grey-coloured quartzitic slates and pelitic hornfels with some interbedded coarse-

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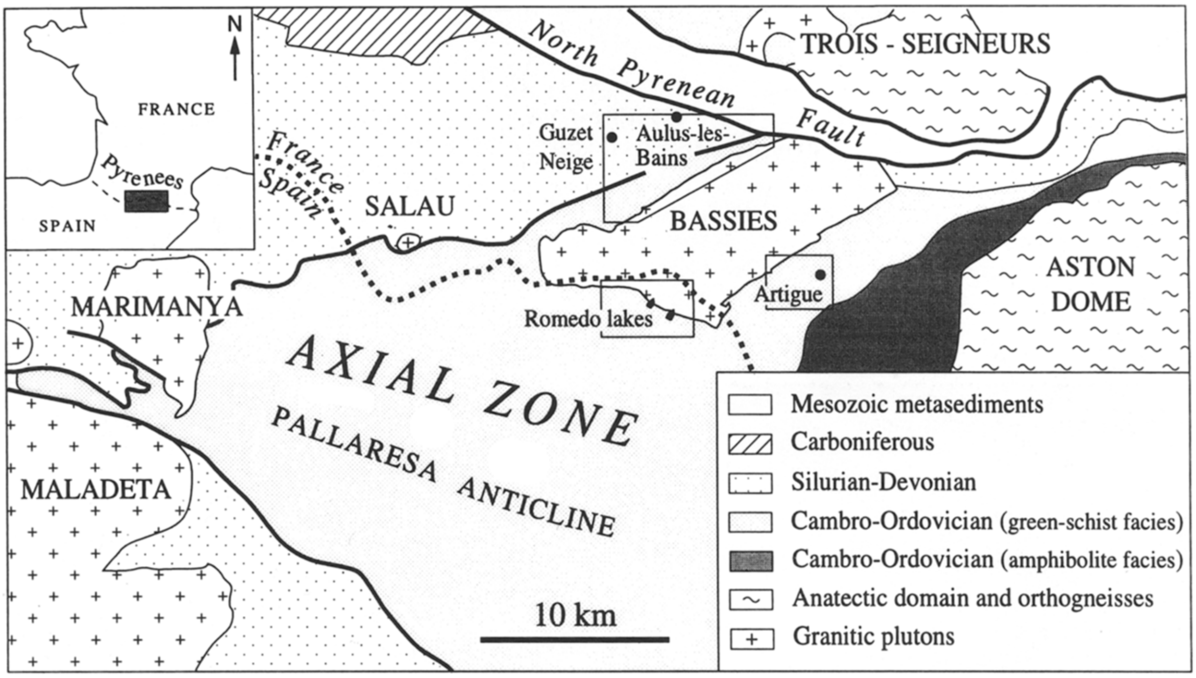


Fig. 1. Geological sketch map of the Central Pyrenees with location of the three study areas along the contacts of the Bassiès granite.

grained quartzite layers. On the eastern side of the pluton, which also constitutes the western termination of the Aston Dome, the series becomes older and rocks are of higher metamorphic grade, up to amphibolite facies conditions on approaching the dome (Vissers, 1992). The western side of the pluton (Fig. 2) is more complex. Cambro-Ordovician sediments outcrop alongside the western border, but the width of this series is reduced in map view to about 1 km, due to a ENE–WSW-trending fault on the northern side of which Devonian marbles and quarzitic slates outcrop (Bodin and Ledru, 1986; Palau and Sanz, 1989). Further northwest from the pluton, these formations abut against another fault beyond which black graphitic Silurian slates crop out. Still further to the northwest, the latter formations are overlain by Devonian calcareous slates alternating with thick and massive limestones. These two fault planes are originally low angle thrust planes that were subsequently folded and steepened by D_2 (Losantos *et al.*, 1986; Bodin and Ledru, 1986).

Around the pluton, the contact metamorphic aureole, approximately 2.5 km wide, can be subdivided into three zones, which, with increasing distance from the granite, are characterized by the presence of sillimanite, andalusite, and biotite $\pm$ muscovite. Along the western side of the pluton, this gradation of metamorphism is not disturbed by the thrust faults.

FIELD STRUCTURAL OBSERVATIONS

We have studied three areas in detail, one on each side of the pluton (Fig. 1): (i) the Guzet–Aulus area along the western contact; (ii) the Artigue area along the eastern contact, whose study was complemented by the Pla de Soulcem cross-section; and (iii) the Romedo area in Spain, at the southern contact.

Main structural features around the Bassiès pluton

In all of the lithological units, the S_0 (bedding) is generally prominent. However, a closer examination reveals that a cleavage is commonly parallel to S_0, hence forming a composite $S_{0\text{–}1}$ planar fabric, and that a layer-parallel stretching lineation L_1 is commonly observed. Many workers in the Pyrenees have attributed this particular fabric to the early D_1 event (Losantos *et al.*, 1986; Calvet and Charvet, 1988; Soliva *et al.*, 1989).

These early structures always display the imprint of a superimposed deformation. However, the nature of this imprint depends on the distance from the granite. Within about 1 km from the contact, the structures which deform $S_{0\text{–}1}$ have a characteristic style, described below, easily related to granite emplacement. These wall rock structures are called S_g, L_g and F_g, respectively for foliations, lineations and fold axes, and the corresponding event is called D_g. This neutral terminology has the advantage of not specifying any particular chronological relationship with respect to the other Hercynian tectonic phases. By contrast, farther from the granite, the structures which deform $S_{0\text{–}1}$ are similar to those which everywhere in the Pyrenees have been attributed to the D_2 'main phase' (Zwart, 1979; Vissers, 1992). The structures related to this phase of deformation are therefore designated by S_2, L_2 and F_2.

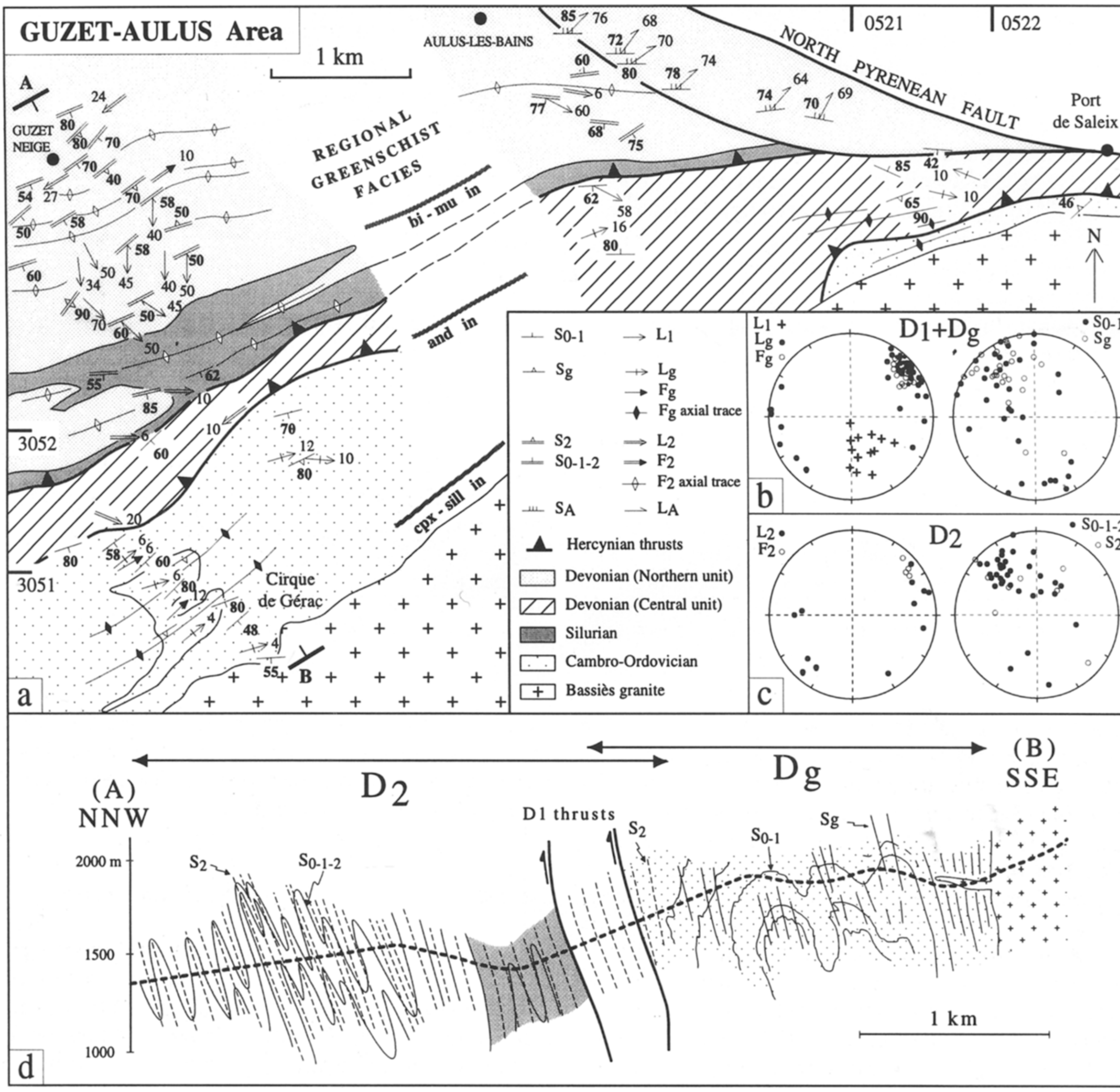


Fig. 2. Guzet–Aulus area. (a) Structural map with location of contact metamorphic isograds; (b) and (c) lower hemisphere orientation diagrams of linear and planar structures for D_l + D_g (b) and D_2 (c); (d) A–B cross section from Guzet-Neige to Cirque de Gérac underlining the shape differences between F_g and F_2 folds: the F_g folds, near to the granite, are open whilst, far from the granite, the F_2 folds are tight and isoclinal.

Guzet–Aulus area (Fig. 2)

$S_{0\text{–}1}$ is clearly visible everywhere in this area (Fig. 2a). In the hinges of the folds, it is possible to observe that $S_{0\text{–}1}$ carries a L_1 stretching lineation defined by the elongation of mineral aggregates, and by quartz ± chlorite pressure shadows around the pyrite cubes of the Devonian calcareous slates. These lineations are particularly visible near Guzet where they plunge 50° to the SE (Fig. 2b).

Within about 1 km from the granite contact, all the structures are attributed to D_g (Fig. 2a) and are nearly parallel to each other, and to the pluton border. Their average strike is N60° in the Cirque de Gérac area, and close to E–W near Port de Saleix. The D_g deformation has intensely folded $S_{0\text{–}1}$ into hectometric to decimetric open folds (Fig. 2d). The F_g folds are imprinted by an axial planar cleavage S_g, somewhat difficult to see in the field, which dips steeply to the SE (Fig. 2b). The fold limbs are crenulated and these crenulations are parallel to the F_g axes and also to the L_g stretching lineations defined by the elongation of mineral aggregates. All these linear structures are subhorizontal (Fig. 2b). The D_g deformation is characterized, in this area, by an intense NE–SW stretching which induced boudinage of the quartzitic layers and numerous sheath folds in the carbonates. In addition, mesoscopic extensional fractures perpendicular to L_g are common. All these stretching features are only observed along the pluton border, and therefore related to magma emplacement. They make it possible to distinguish the D_g structures from those of D_1 and D_2.

Farther than 1.5 km from the pluton, the evidence of D_g stretching disappears and the structures superimposed onto those of D_1 are attributed to D_2. The F_2 folds are isoclinal and particularly tight, in contrast with the open F_g folds (Fig. 2d). The F_2 fold axes are subhorizontal and trend N60° to E–W, as shown by their trajectories in map view (Fig. 2a) and their orientations in stereographic projection (Fig. 2c). Except in the hinges, the S_2 crenulation cleavage is generally parallel to S_{0-1}, and is therefore named S_{0-1-2}. The S_{0-1-2} planar fabric dips 60° on average to the SE or to the S. Its N60° to E–W strike is parallel to that of S_g (Fig. 2c), and this parallelism impedes the analysis of the overprint of S_g by S_2. In the slaty lithologies, the S_2 plane generally carries a L_2 crenulation parallel to the F_2 fold axes.

In the northern Devonian unit near Aulus-les-bains, peculiar structures are recorded in a zone elongated parallel to the North Pyrenean Fault. They probably represent an Alpine imprint, as argued by Lamouroux *et al.* (1980). The main planar fabric (S_A in Fig. 2) dips to the N, and a well developed lineation (L_A) plunges 70° to the NE.

Artigue area (Fig. 3)

Along the eastern side of the Bassiès pluton, the D_g structures are parallel to those of the Cirque de Gérac. The foliation S_g is more penetratively developed than along the western side, and is often parallel to S_{0-1}. Therefore, the large scale folds, which are clearly delineated in the Cambro-Ordovician interbedded lithologies, are difficult to recognize in the homogeneous slates. All the structural trends are N60° in azimuth, with S_g subvertical, parallel to the granite contact, and the linear structures L_g and F_g are subhorizontal (Fig. 3b).

The D_2 structures appear gradually at a distance of less than 1 km from the contact, and become more strongly imprinted further to the SE where all the planar structures rotate progressively from N60° to N90°. A spaced S_2 crenulation cleavage, slightly oblique to the S_{0-1}/S_g plane, carries a L_2 lineation but, unlike at Guzet Neige, no obvious L_1 stretching lineation was observed. Further east, F_2 isoclinal folds with subhorizontal axes can be found. Their limbs constitute a S_{0-1-2} planar fabric, striking E–W like all the structures in this area. There, L_2 plunges gently to the west (Fig. 3c).

The study of the Artigue area has been complemented,

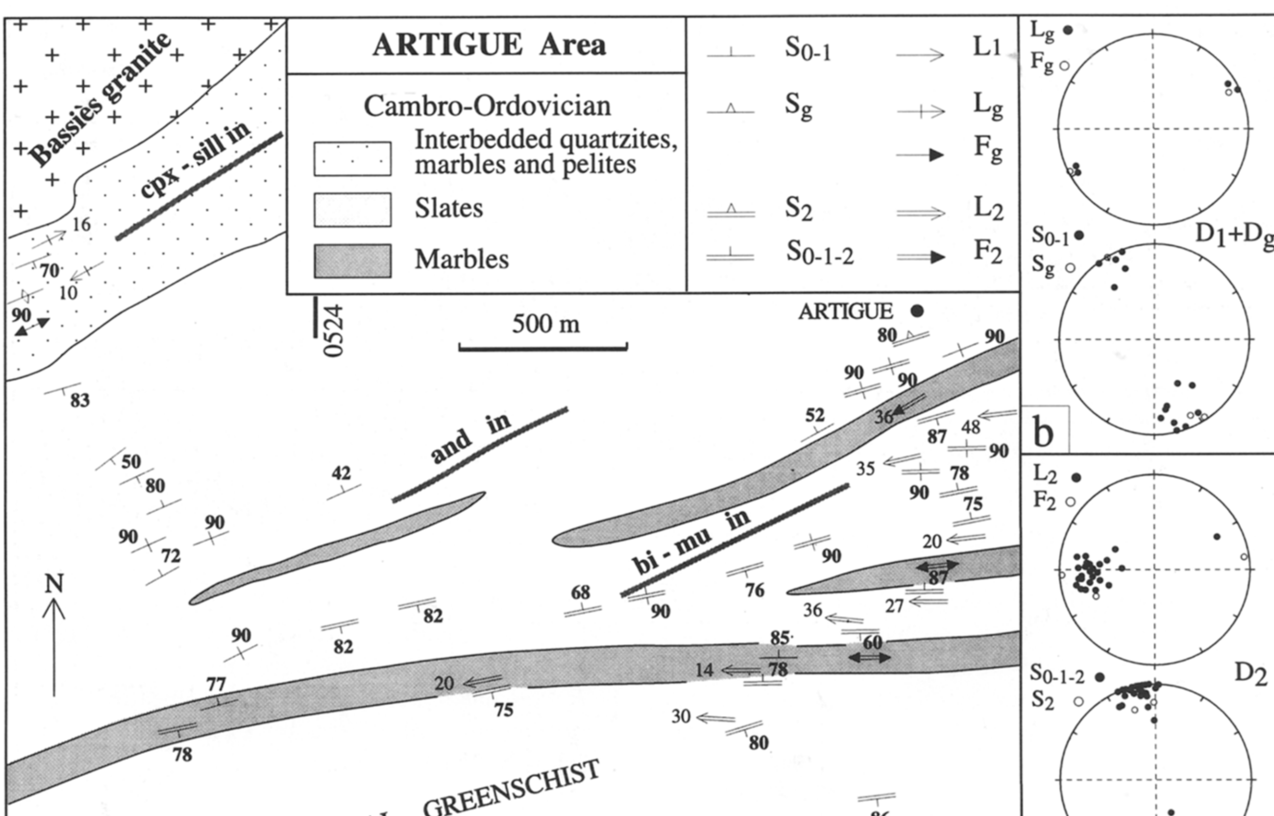


Fig. 3. Artigue area. (a) Structural map with location of contact metamorphic isograds; (b) and (c) orientation diagrams of linear and planar structures for D_1 + D_g (b) and D_2 (c).

further east, by a N–S section joining the NE corner of the Bassiès pluton to the Pla de Soulcem metamorphic dome, the W part of the Aston dome (Fig. 5). This section had been previously described by Verhoef *et al.* (1984) and Vissers (1992) who attributed the structures to three distinct tectonic events on the basis of the steep or flat-lying attitude of the foliation. We agree with the change of the foliation dip, but not with their interpretation which attributes the subhorizontal structures of the Pla de Soulcem area to a D_3 event. According to our observations, the overall structure results from only two tectonic events, like everywhere else: from the north to the south of the section, the S_2 subvertical foliation planes and the associated L_2 lineations rotate progressively, up to the Pla de Soulcem area where these D_2 structures becomes subhorizontal. In the northern part of the section, the L_2 lineations are mostly crenulations which are parallel to the F_2 fold axes. Further south, with increasing metamorphic grade, the L_2 crenulations fade progressively while stretching lineations appear, underlined by mineral orientation which is roughly parallel to the crenulations.

Romedo area (*Fig.* 4)

At the southern border of the pluton, the structures of the Romedo area strongly contrast with those of the two previous areas. Close to the granite, S_{0-1} remains clearly visible in the interbedded quartzites and slates dipping steeply to the north. The N80° average strike of S_{0-1} is oblique to the granite contact whose average trend is N110° (Fig. 4a). Within 1 km from the contact, and particularly in the rocks surrounding the Romedo lakes, cylindrical folds, with axes plunging steeply to the NE, deform the S_{0-1} plane. These folds, whose frequency decreases with distance from the pluton, are attributed to the emplacement of the granite, hence named F_g in Fig. 4. The associated L_g mineral lineations, although weakly defined, can be traced by the rough alignment of the andalusite porphyroblasts with an orientation comparable to the F_g fold axes (Fig. 4b).

S_2 appears in the pelite-dominated horizons at a distance of a few hundred metres from the contact. Since S_2 is generally oblique to S_{0-1}, itself defined by interbedded quartzite layers folded by F_g, it is more easily

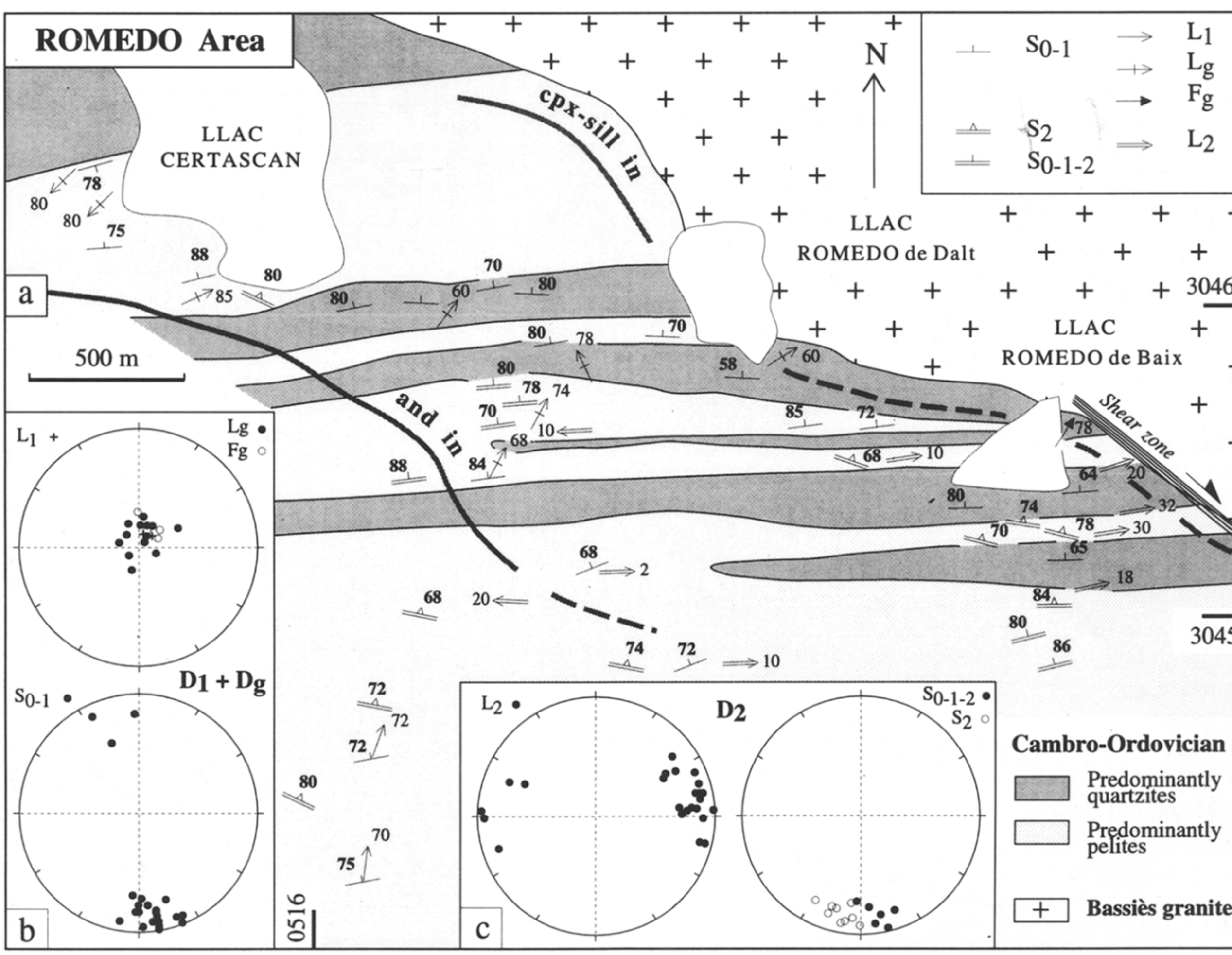


Fig. 4. Romedo area. (a) Structural map with location of contact metamorphic isograds; (b) and (c) orientation diagrams of linear and planar structures for D_l + D_g (b) and D_2 (c).

recognized than in the two previous areas. It dips steeply to the north, strikes N90°–N120° (Fig. 4c) and becomes more strongly developed towards the south. Farther than 1 km from the granite, S_2 strikes generally parallel to S_{0-1}, and is therefore called S_{0-1-2}. A subhorizontal L_2 lineation is systematically found on the S_2 foliation plane, but F_2 folds were not observed within a distance of 1.5 km from the granite.

MICROSTRUCTURAL AND KINEMATIC OBSERVATIONS

To understand better the relationships between D_1, D_g and D_2 and their associated kinematics, a detailed microstructural analysis has been undertaken to complement the previous mesoscopic geometrical descriptions. Numerous thin sections from oriented samples were cut parallel to the prominent lineation and perpendicular to the foliation.

Criteria used to determine the sense of shear

Contrary to rocks plastically deformed at relatively high temperatures, and for which unambiguous shear criteria are available, in weakly metamorphosed sedimentary rocks such as those of the Bassiès area, the sense of shear is not particularly easy to determine. However, because of the variety of lithologies, mostly made up of quartzites and pelites but also of some marble levels, several criteria could be used. They include some well established shear sense criteria but also, using the finely interbedded lithologies, sets of criteria based upon the precise geometries of microfolds, kink bands, offsets, and thickness variations of layers.

The classical criteria which were used are the following: (i) obliquity of quartz *c*-axis fabric (Bouchez, 1977; Plate 1a & b); (ii) rotation of rigid markers during growth of parallel-to-stretch pressure fringes (Etchecopar and Malavielle, 1987; Plate 1c); (iii) boudinaged and offset quartz veins (Hanmer and Passchier, 1991; Plate 1d); (iv) S/C structures (Berthé *et al.*, 1979) which are often associated with antithetic minor shear structures (Bell and Johnson, 1992); and (v) asymmetric microfolds related to buckling (Ramberg, 1963).

The ubiquitous kink bands within the finely layered sediments were used to determine the sense of shear but, to ensure their reliability, they were systematically associated with the more classical criteria; for example: (i) antithetic kink bands with S/C shear bands (Plate 1e); (ii) sigmoidal synthetic kink bands associated with shear bands (Plate 2a); (iii) antithetic and synthetic kink bands associated with asymmetric microfolds and S/C shear bands (Plate 2b); and (iv) synthetic kink bands associated with asymmetric micro-drag folds.

Microstructures and kinematics of D_1

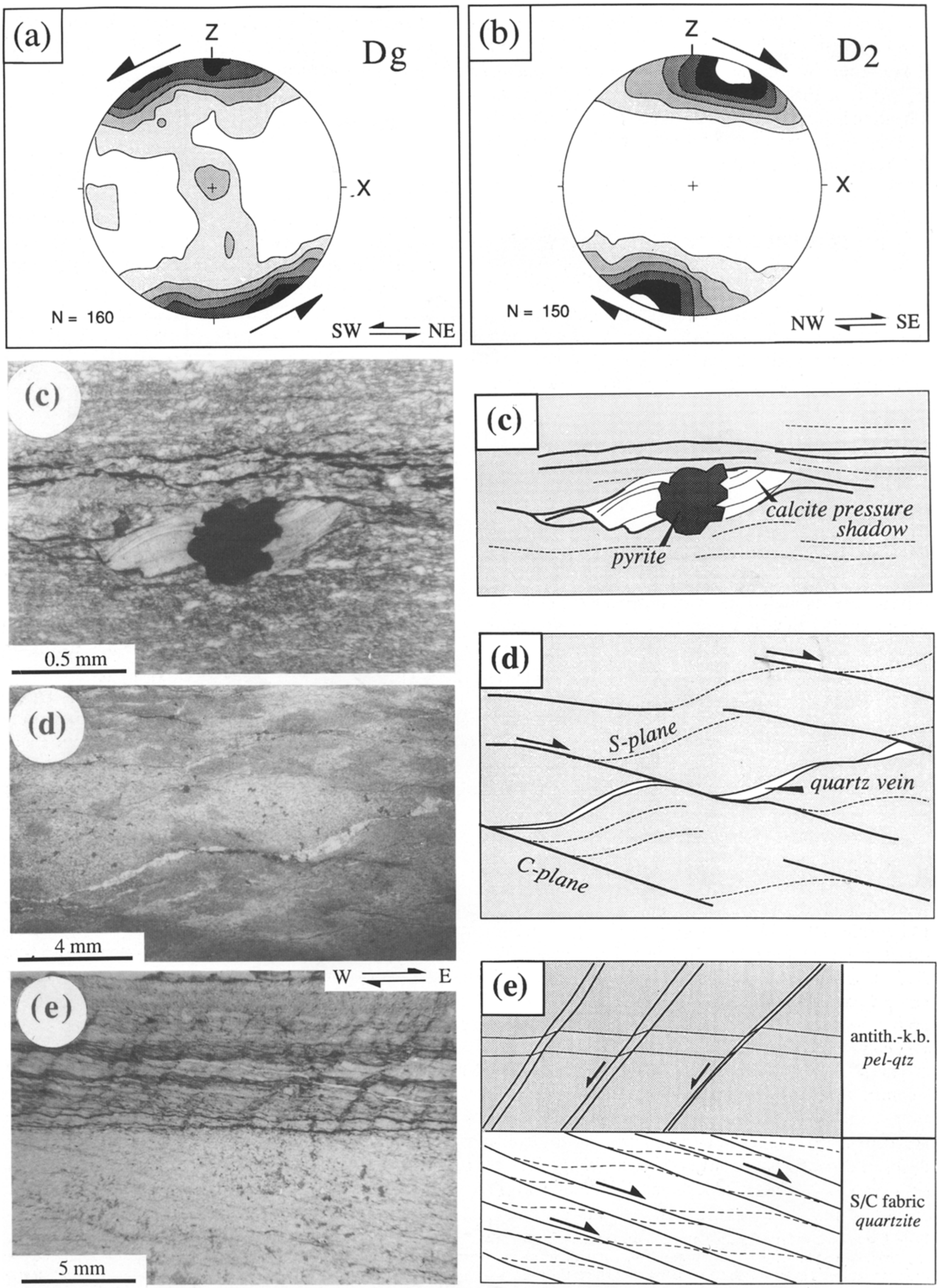
The trace of the S_{0-1} plane is always clearly observed in thin sections whatever the distance from the granite. However, L_1 is rather difficult to see in both the quartzitic and pelitic lithologies. By contrast, L_1 is easy to define in the Devonian carbonates, marked by elongate pressure shadows attached to pyrite cubes which provide unambiguous shear-senses near Guzet-Neige. In this area, L_1 plunges moderately to the southeast, and the XZ thin sections in this D_1 frame systematically show a top-to-the-southeast sense of shear (Fig. 5 and Plate 1c).

Microstructures and kinematics of D_g

Along the contacts of the pluton in the *Guzet and Artigue areas*, where S_g is vertical and L_g horizontal, X–Z sections are horizontal. In such sections, the microstructures are characteristic of intensely stretched rocks and asymmetries clearly defining the sense of shear are hardly observed. In addition, the fine-grained carbonate and pelitic lithologies contain few objects which could be used as kinematic markers, and the contact metamorphic minerals overgrow the previous structures.

Shear sense criteria being generally unclear, a quartz *c*-axis fabric was analysed in an impure quartzite sampled 200 m from the granite contact in the Cirque de Gérac area. The quartz grains have a prolate elliptical shape and are often subdivided into subgrains. The subgrain pattern, essentially made of two sets of subperpendicular boundaries, resembles the chessboard pattern typical of high temperature plastic strain in quartz. Accordingly, the *c*-axis diagram (Plate 1a) shows a girdle, with maximum near Z, and submaxima at Y and in-between Z and Y, indicating the activation of the high temperature prismatic and pyramidal *a* slip systems in addition to the basal *a* system (Bouchez and Pêcher, 1981). Also, the weak submaximum in proximity to the lineation (X), may indicate the contribution of the {prism} [*c*] slip system, supporting the high temperature straining of this rock. Such a high temperature demonstrates that D_g is necessarily related to magma emplacement. The overall asymmetry of the diagram demonstrates the sinistral sense of shear. This sense is confirmed in the field in subhorizontally exposed sections such as large glacially polished surfaces providing easy to examine X–Z planes.

Plate 1. Examples of shear sense criteria in the metasedimentary rocks (photographs: plane polarized light). Samples are located in Fig. 5. (a) and (b) Quartz *c*-axis fabric diagrams in quartzitic samples. Schmidt lower hemisphere, Kamb's contours with contour interval = 2σ. (a) D_g diagram characterizes high temperature deformation and sinistral sense of shear. (b) D_2 diagram displays typical low temperature fabric and dextral sense. (c) Dextrally rotated pyritic grain with asymmetric development of pressure fringes, in Devonian calcareous slates. Subvertical X–Z section parallel to L_1. (d) Sigmoidal quartz vein boudinaged and offset by C-planes during dextral shear in spotted quartzitic slates. Subhorizontal section parallel to the L_2 crenulation. (e) S/C fabric in a quartzitic layer gives evidence of dextral shear. Therefore, the kink bands (k.b.) in the more pelitic layers (pel-qtz) must be regarded as antithetic. Subhorizontal section parallel to the L_2 crenulation.


(a)
Z
Dg
X
N = 160
SW NE
(b)
Z
D2
X
N = 150
NW SE
(c)
0.5 mm
(c)
calcite pressure shadow
pyrite
(d)
4 mm
(d)
S-plane
quartz vein
C-plane
W E
(e)
5 mm
(e)
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pel-qtz
S/C fabric
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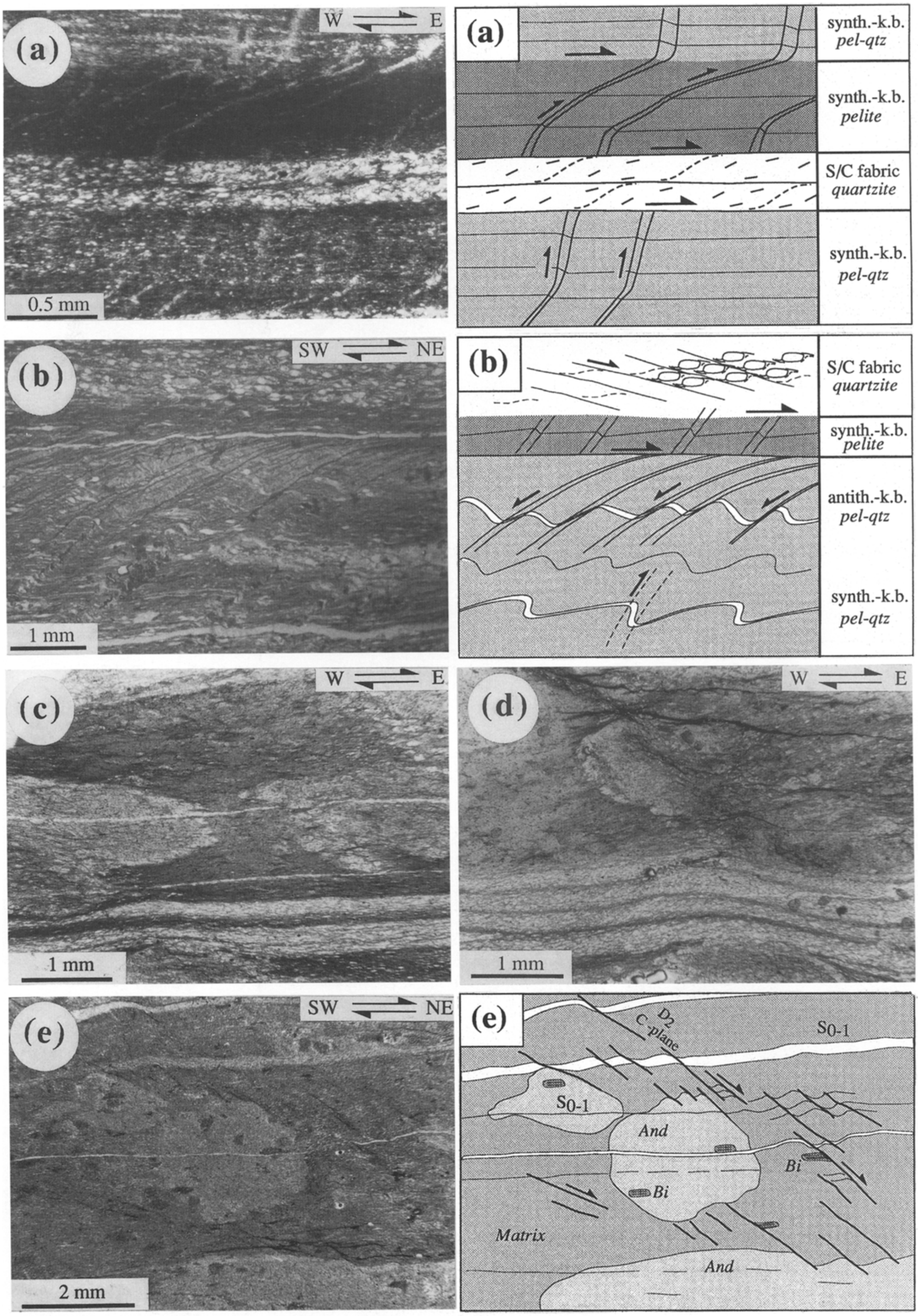
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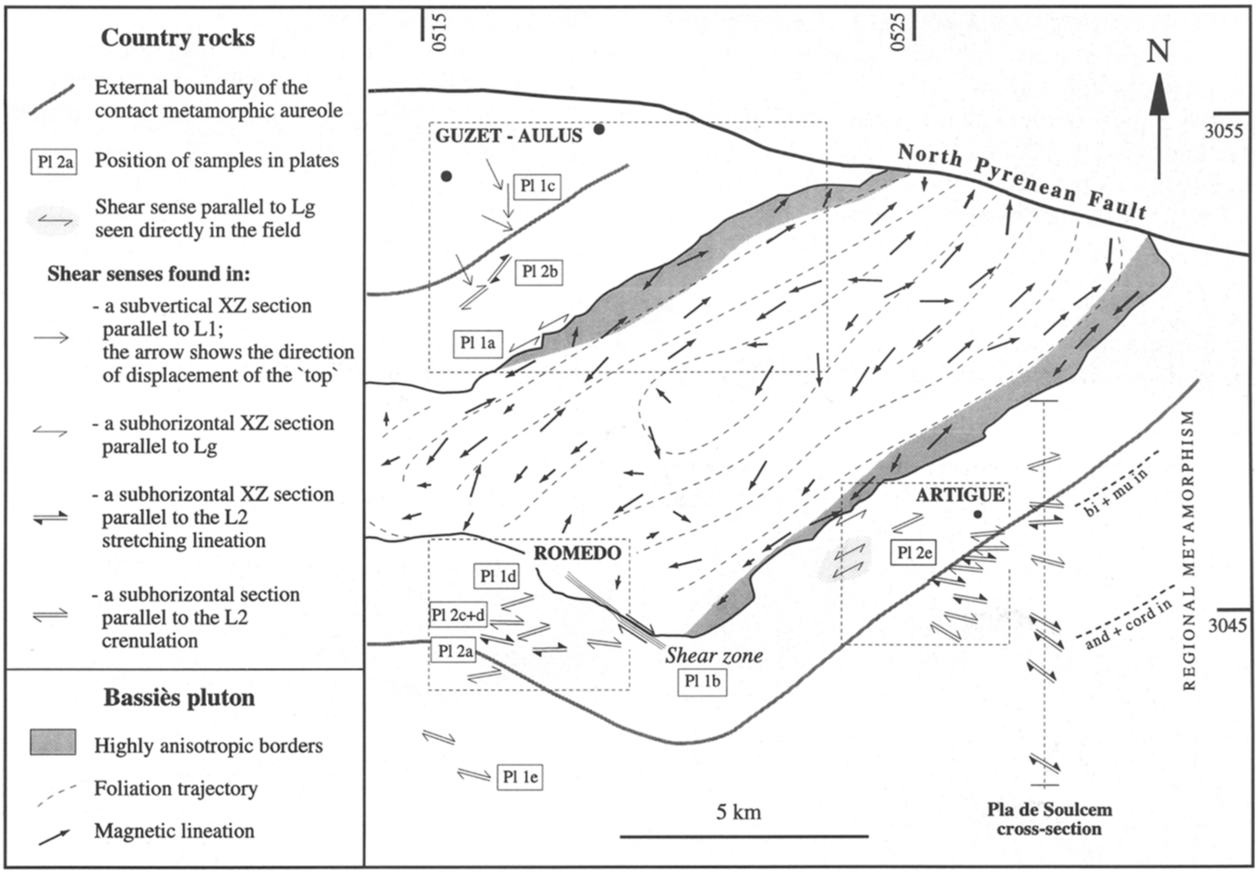

Fig. 5. Shear senses observed in the Bassiès country rocks. The structures within the Bassiès pluton are after Gleizes *et al.* (1991).

In the Artigue area, extensional fractures and boudinaged quartz veins clearly evidence the sinistral strike-slip parallel to the pluton border (Fig. 5).

Along the granite contact *in the Romedo area* where the foliation strikes E–W and dips steeply to the north, the fabric is essentially planar. However a subvertical stretching component is locally apparent, marked by a weakly defined subvertical L_g lineation. Observed in a vertical and N–S-trending section plane (i.e. X–Z section), the microstructures show that D_g deformation was not intense. This interpretation is strengthened by the observation of quartz grains in microconglomerates which are only slighly deformed, with rare subgrains and shape ratios not exceeding 2:1. In these weakly deformed rocks, no sense of shear could be determined and it may be suggested that the deformation was essentially coaxial.

Microstructures and kinematics of D_2

The D_2 linear structure visible in the field is generally only a crenulation in low-grade metasediments. However, in the amphibolite facies rocks of the Pla de Soulcem cross-section to the east of Artigue, a well-defined stretching lineation is apparent and is always subparallel to the crenulation. A similar observation of a crenulation in low-grade rocks parallel to a mineral streching lineation in higher grade rocks, has been described in the Trois-Seigneurs massif (Leblanc *et al.*, 1996). In fact, even in the low-grade samples, a detailed microscopic investigation of the S_2 planes often reveals the existence of a stretching lineation (L_2), marked by the alignment of mineral aggregates subparallel to the crenulation. Therefore, the thin sections that we have

Plate 2. Finely interbedded slates. Samples are located in Fig. 5. All sections are subhorizontal and parallel to L_2 (plane polarized light, except (a): cross polars). (a) Synthetic kink bands (k.b.) developed in slates. They are more stretched and dextrally rotated in the pelitic layers where deformation is concentrated. In the quartz-rich band, the sigmoidal arrangement of the elongated quartz grains and their obliquity relative to the lithological contacts (S_{0-1-2}) also indicate dextral shear. Section parallel to the L_2 stretching lineation. (b) As a function of lithological contrasts, various convergent criteria indicate an overall dextral sense of shear. The lower part of the section displays synthetic kink bands. On approaching a pelitic level where the dextral deformation is concentrated, the dextral deflection of fractures resulting from the evolution of other kinks indicates their antithetic character. Section parallel to the L_2 stretching lineation. (c) Quartzitic levels, which define the S_{0-1} trace, cross the contact andalusite minerals without deflection. Therefore the porphyroblasts have grown statically on D_1 structures. (d) A contact andalusite mineral is offset by a D_2 synthetic shear plane: it has grown before D_2. (e) In the matrix, S_{0-1} is overprinted by D_2 C-planes which offset the borders of the andalusite porphyroblasts (And) and the biotite contact minerals (Bi). By contrast, the S_{0-1} trace and the biotites are not deflected within the andalusites.

studied and which were cut parallel to the crenulation and perpendicular to the S_2 plane, (i.e. subhorizontally), are mostly true X–Z sections.

The D_2 microstructures characterize a weak deformation at low temperature, i.e. almost without recrystallization in quartz. They are predominantly represented by kinks in the pelitic layers, probably accommodating the movement in-between the rigid quartzitic layers where strain is accommodated by oblique shear planes. The kinematics of D_2 (Fig. 5) is unambiguously established since the shear sense is found to be always dextral in X–Z sections (Plate 2a & b). The other sections, those cut parallel to the crenulation (Plate 1d & e), also systematically show a dextral component of shear in the three study areas. This again confirms that the crenulation and stretching directions are almost parallel, in accordance with the observations on the S_2 planes in samples where both types of lineations are visible. In conclusion, the D_2 deformation appears to be characterized by a dextral shear component in the whole Bassiès area. The opposite senses of shear evidenced for D_2 and D_g strengthen our conviction that these fabrics cannot have developed simultaneously.

Some shear zones which cross-cut the Bassiès granite body (Blès and Gros, 1980; Gleizes *et al.*, 1991) are also attributed to D_2. They strike N120° and dip steeply to the north with stretching lineations moderately plunging to the NW. One of these shear zones, 300 m wide, may be observed to the east of the Romedo de Baix lake; it cross-cuts the granite contact and also the D_g structures of the aureole (Fig. 4a). There, a quartz c-axis fabric was analysed in a sample from a highly stretched quartzitic layer. The diagram (Plate 1b) is characteristic of low temperature deformation and its asymmetry indicates dextral shearing.

Deformation of the contact metamorphic minerals

The contact aureole around the pluton may be easily subdivided into metamorphic zones (Figs 2–4) based on the presence of index minerals. Starting from the intrusive contact, the texture is clearly hornfelsic in a 300 m wide zone, marked by the presence of sillimanite $\pm$ clinopyroxene in the quartz-rich slates, and clinopyroxene $\pm$ amphibole $\pm$ epidote in the impure carbonates. Then, over a distance of about 1.5 km away from this zone, contact metamorphism is defined by andalusite $\pm$ biotite in the quartzo-pelitic rocks, and biotite $\pm$ muscovite $\pm$ epidote in the impure carbonates. The boundary of the outermost zone, approximately 2.5 km from the granite, is represented by the transformation of chlorite and sericite into biotite and muscovite.

Relationships between the contact metamorphic porphyroblasts and the deformations related to the different tectonic events, further constrain the relative timing of these events. In the three study areas of the aureole, the contact metamorphic minerals have everywhere grown statically over the D_g fabric. This is particularly obvious in the quartz-rich slates where the andalusite porphyroblasts, although locally aligned parallel to L_g, show no deflection of their internal fabric with respect to the fabric of their matrix. The relationships of the contact metamorphic porphyroblasts with the fabrics of D_1 and D_2 have been studied in thin sections of rocks sampled approximately 1 km from the granite contact, i.e. where D_g deformation is weak. In the Romedo area, the andalusites have grown statically upon the D_1 fabric since S_{0-1} is not deflected (Plate 2c). However, in the same thin section, andalusite porphyroblasts sometimes appear to be offset by D_2 shear planes (Plate 2d). Similar features were observed in many other places: for example, a phyllite rich sample from the Artigue area (Plate 2e) exhibits D_2 C-planes which are imprinted in the matrix and which deform both the contact metamorphic biotites and the boundaries of the andalusites; by contrast, within the andalusites, the biotite inclusions and the trace of S_{0-1} were protected from D_2 and remain undeflected. The synthesis of these microstructural observations shows that the contact metamorphic minerals grew after D_1 and D_g but were deformed by D_2.

INTERPRETATION

Model of pluton emplacement

According to a previous detailed structural study inside the pluton (Gleizes *et al.*, 1991), the magma was intruded into a strike-slip shear zone parallel to the long sides of the pluton. The new structural data from the wall rocks strengthen this conclusion since the D_g structures of the wall rocks present striking similarities with the internal structures of the pluton.

Along the eastern and western sides of the pluton, the wall rocks are characterized by open folds and microfolds with subhorizontal F_g axes, and by a discrete axial planar cleavage S_g. More characteristic are the features, such as sheath folds, indicative of a strong streching parallel to the sides of the pluton. These structures in the wall rocks fit with the strong anisotropic fabric of the nearby granite, which results from magmatic to incipient solid-state deformation. This fabric in the granite has a foliation everywhere vertical, and a lineation mostly subhorizontal (Fig. 5) whose N60° orientation is the same as that of D_g in the wall rocks. Along the southern contact, a weak deformation is characterized (i) in the wall rocks, by the slight deformation by cylindrical F_g folds, and (ii) in the granite, by low anisotropy magnitudes. In addition, the discrete subvertical L_g mineral lineations of the wall rocks are equated with the steeply plunging lineations in the granite. All these similarities confirm that the specific D_g structure, which is imprinted only in the vicinity of the granite, is related to the emplacement of the magma.

We are therefore allowed to propose the model of emplacement of Fig. 6(a). The D_g structure characterizes

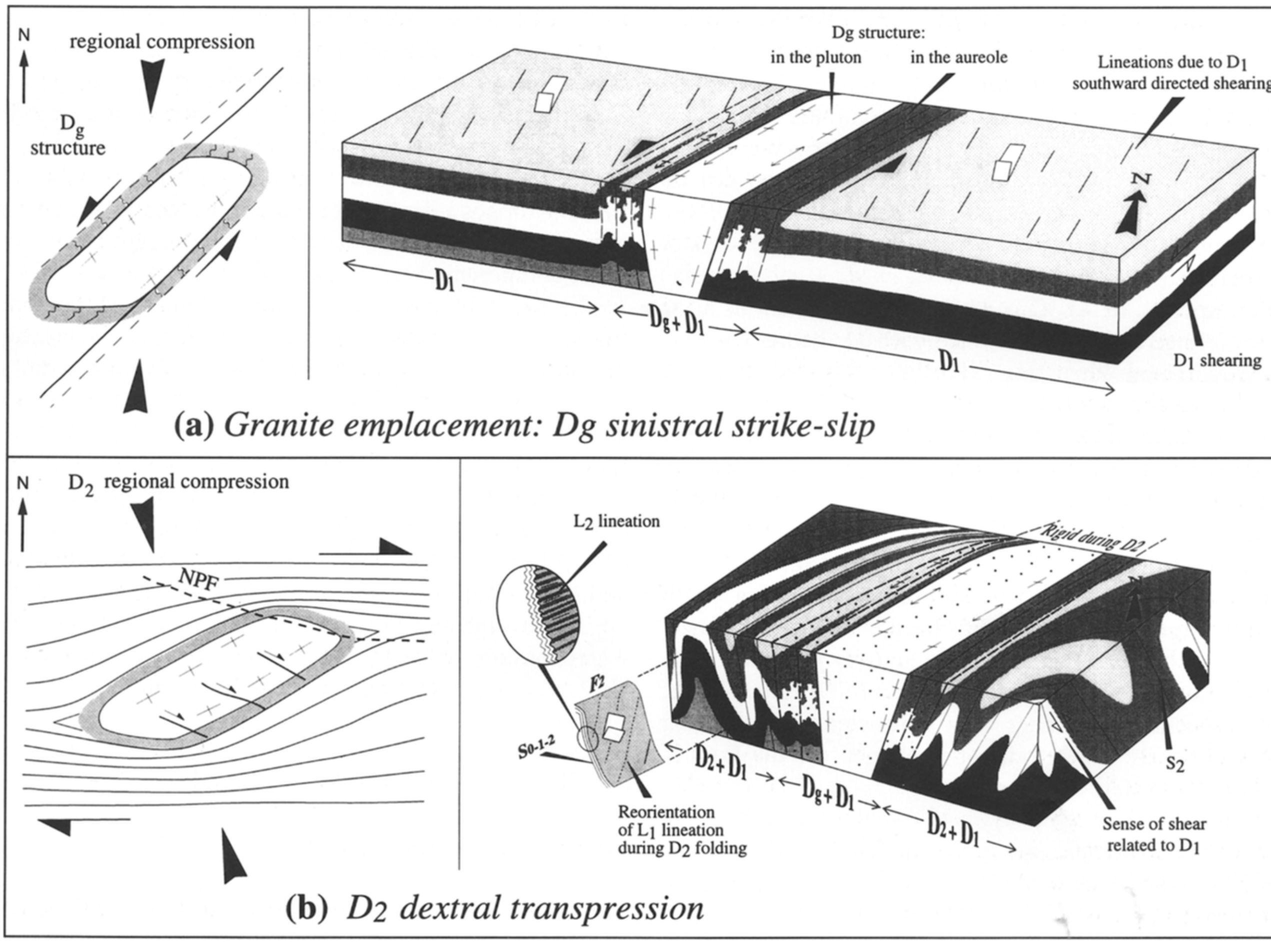

Fig. 6. Tectonic evolution of the Bassiès pluton and its country rocks in plan section (on the left) and block diagram (on the right). (a) Emplacement of the pluton into a sinistral strike-slip pull-apart at the end of the D_1 event and generation of the D_g deformation of the wall rocks. (b) After full crystallization, the pluton and its immediate wall rocks acted as a rigid porphyroclast undergoing a 'clockwise' rotation during D_2 dextral transpression; the D_2 structures of the country rocks envelop the pluton; NPF is the North Pyrenean Fault which truncated the pluton during Alpine orogenesis.

a NE–SW strike-slip. In the upper crust, where the pluton was emplaced, some strain rate variation during shear would have been accomodated by a tear fault which created an opening able to trap the magma that finally pooled to form the Bassiès massif. The sinistral sense of this tear fault is deduced from the quartz fabric observed near the pluton (Plate 1a), and from local field observations (Fig. 5). In this model of emplacement, the development of intensely stretched D_g structures, such as sheath folds along the NE–SW-striking sides of the pluton, is attributed to the shearing. The vertical lineations observed in the southern part of the granite underline the 'flow' fabric of the magma during pluton feeding which, in this area, has not been overprinted by horizontal shearing. Here, the subvertical F_g fold axes can be attributed to the flattening of the wall rocks during emplacement.

Timing of pluton emplacement

Emplacement of the Bassiès pluton was previously considered as post-tectonic on the basis of the Permian Rb–Sr 276 Ma age obtained by Debon and Zimmerman (1988). However, a recent study of the internal fabric of the Bassiès pluton (Gleizes *et al.*, 1991) suggested that it was emplaced relatively early in the Hercynian orogeny. Our new data from the country rocks agree with the latter proposal and specifically show the emplacement as being pre-D_2.

The pre-D_2 emplacement of the pluton can be deduced from the following results: (i) D_g and D_2 are kinematically incompatible events, and therefore they must be separated in time, and (ii) the contact metamorphic minerals are static over D_g but systematically deformed by D_2. Therefore, D_g must pre-date D_2. The fact that the contact metamorphic porphyroblasts have grown everywhere statically over the D_g fabric is attributed to the transfer of heat to the wall rocks being slower than the local strike-slip deformation and emplacement of the magma. Hence, contact metamorphic minerals grew after the straining of rocks within which the magma was intruded.

D_g pluton emplacement being pre-D_2 and linked to a strike-slip tectonics, it remains to clarify the precise timing of this early Hercynian event within the tectonic and chronological frame of the Pyrenees. Is the strike-slip deformation D_g kinematically compatible with the D_1 tectonics? It is widely admitted that the early Hercynian D_1 event represents a S-verging thrust event characterized by nappes in the upper crustal section and ductile shearing at deeper levels, due to N–S convergence resulting in probably moderate crustal thickening (Lagarde, 1978; Soliva *et al.*, 1989; Carreras and Capella, 1994). Our kinematic observations near Guzet-Neige are in accordance with D_1 being a southward directed thrusting. The D_g NE–SW-trending sinistral strike-slip is compatible with the general N–S shortening that characterizes D_1. Therefore, it is envisaged that the emplacement of the pluton took place at the end of D_1.

This timing of emplacement deduced from the structures in both the pluton and its aureole contradicts the Permian Rb–Sr age. To solve this contradiction, a U–Pb zircon dating of the pluton has been prompted. The new age at 312 ± 2 Ma (Paquette *et al.*, in press) reconciles the geochronological data with our structural results. Hence, the different approaches now consistently indicate a relatively early syntectonic emplacement of the Bassiès pluton in the Hercynian orogeny

D_2: *a dextral transpressive event*

Ubiquitous E–W-trending isoclinal folds in the Hercynian basement of the Pyrenees are the most conspicuous regional-scale structures. They provide evidence of considerable N–S shortening attributed to N–S compression during the main phase, termed D_2 (Zwart, 1979; Vissers, 1992). The D_2 structures here-above described in the Bassiès country rocks are assigned to this major phase. However, our microstructural observations show that D_2 implies a dextral component of shear. This event, apparently dominated by compression, but with a dextral shear component, was therefore transpressive.

The rocks of the inner aureole became rigid due to thermal metamorphism, and were therefore weakly imprinted by D_2. In the granite, which was fully crystallized before D_2, the imprint of this event is localized into dextral shear zones which are scarce and have a low-temperature plastic to brittle nature. Therefore, the pluton and its inner aureole were more rigid than the outer country rocks, and behaved just like a porphyroclast during D_2 (Fig. 6b). This interpretation explains the triple point of planar fabric trajectories which appears on the south-western termination of the pluton where the country rocks are not deformed by D_2 (Gleizes, 1992). Figure 6(b) suggests that another triple point should have been found at the northeastern corner, but it has been truncated by the Alpine North Pyrenean Fault. This interpretation also explains why F_2 axial traces envelop the pluton, passing from a general E–W trend far from the pluton into a N60° orientation along its sides. Such a deflection of 30° of the F_2 folds during D_2 allows the inferred pre-D_2 N–S orientation of the L_1 lineations in the Guzet area (Fig. 6a).

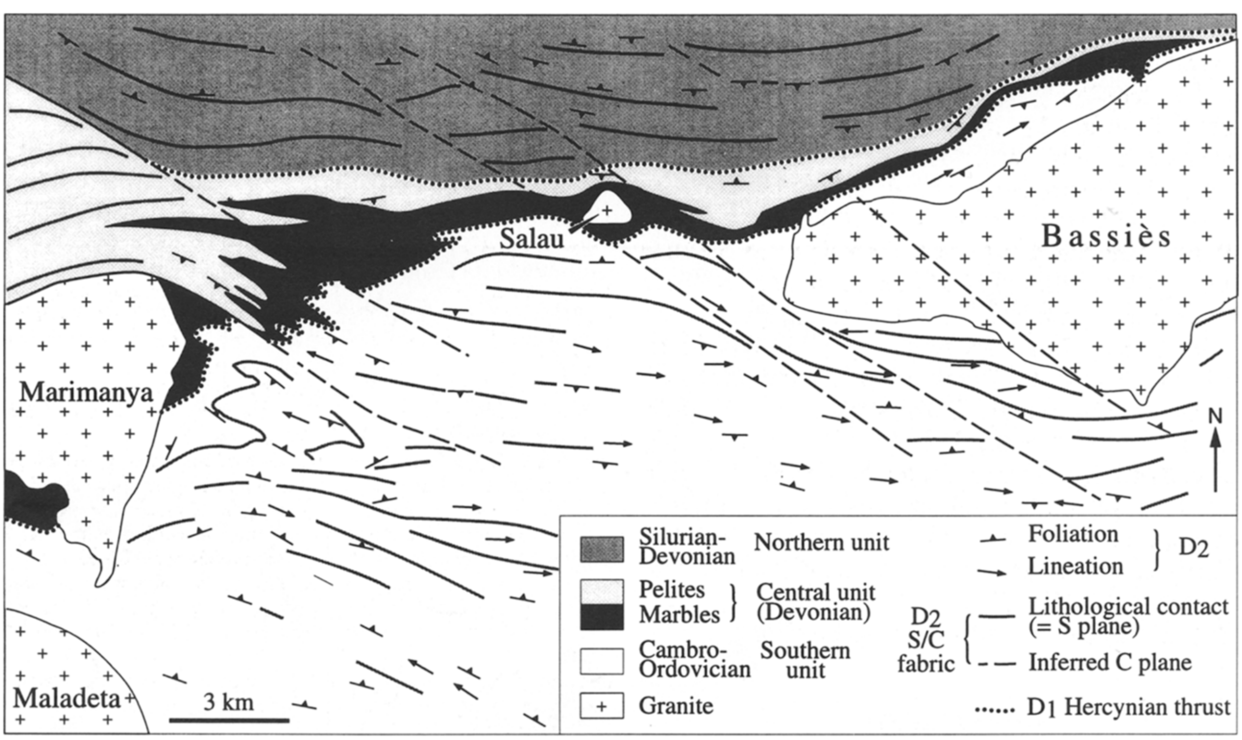

Fig. 7. Structural sketch map of the Pallaresa area (Central Pyrenees) regarded as a X–Z section at a regional scale. The regional dextral sense of shear for D_2 is indicated by the sigmoidal pattern, which is compared to a S–C fabric, and by the asymmetrical geometry of the structures around the Marimanya and Salau plutons.

Tectonic implications to the Hercynian of the Pyrenees

The Hercynian structure of the Central Pyrenees may be reinterpreted in the light of our observations in the Bassiès area. The sketch map (Fig. 7) has been drawn by reproducing the structural features given by the geological map of the Pallaresa area located between the Bassiès and Marimanya plutons (de Sitter and Zwart, 1959). Traces of the D_1 thrust planes (Fig. 7: dots), mapped by Losantos *et al.* (1986) and Bodin and Ledru (1986), and inferred shear corridors (dashes) have been added. All the structures of this map have evolved during the main phase D_2, since (i) the subhorizontal lineations, which trend about N100°, can be equated with the L_2 of the Romedo area; (ii) the foliations, which strike N100° on average and mostly dip steeply, are, in some places (e.g. NE of Marimanya), axial-planar of the F_2 folds which are clearly oblique to the trace of S_{0-1} (more commonly, these foliations are parallel to S_{0-1} but interpreted as S_{0-1-2} like in the Bassiès area), and (iii) the D_1 thrust planes are themselves subparallel to the bedding and reworked by tight F_2 folds, particularly in-between the Salau and Marimanya granite bodies.

Since S_2 or S_{0-1-2} are subvertical, and L_2 subhorizontal, the geological sketch map of (Fig. 7) provides a 'regional X–Z section' which can be analysed for its D_2 kinematics. A regional dextral sense of shear can therefore be derived by analysing the asymmetrical sigmoids that develop, within the Devonian marbles, on sides of the Marimanya and Salau granite bodies. The asymmetrical triangular shaped domain to the east of the Marimanya body, with its bedding affected by open folds oblique to the overall D_2 structure, is similar to the area around the southwest corner of the Bassiès pluton (Fig. 6b, left-hand side). They look like domains protected from the D_2 straining, just like neutral zones found behind rigid porphyroclasts observed in thin sections. The dextral sense of regional shearing is also attested by the dextral sigmoids and offsets affecting the bedding and the foliation plane trajectories, passing from regional N100° strikes to N120° strikes in narrow shear corridors (Fig. 7), just like C–S structures in thin sections.

In conclusion, the whole area of (Fig. 7) appears to have been affected by D_2, a pervasive and probably compressive regional-scale dextral shear event. No other significant tectonic event has occurred later in the Hercynian orogeny. This model for the Central Pyrenees should also apply to the whole chain where similar fabric patterns seem to be present. It is therefore proposed that the Hercynian structure of the Pyrenees resulted predominantly from the effect of dextral wrenching (D_2) during a crustal convergence that began with D_1.

CONCLUSIONS

This structural study emphasizes the early syntectonic emplacement character of the Bassiès pluton, and brings into question the significance of the previous isotopic data. Emplacement before the major D_2 event is demonstrated by the field structures at different scales and by the deformation features of contact metamorphic minerals during D_2. Along the western and eastern borders of the pluton, specific structures characterize a sinistral zone of shear deformation, coeval with the intrusion of the magma in the upper brittle crust by the end of the D_1 thrust event. We also demonstrate that the D_2 structures, which result from N–S compression, systematically display the imprint of a non-coaxial dextral component. The Hercynian basement of the Pyrenees is therefore viewed as a dominantly transpressive chain.

From a rheological point of view, our study underlines that the contact metamorphism hardened the rocks of the inner aureole preventing them from being overprinted by a subsequent deformation event. Hence, the absence of regional structures alongside a granite pluton, generally interpreted as due to an overprint during late magma emplacement, may indicate, on the contrary, an early emplacement. This study of the Bassiès area exemplifies that the structural knowledge of both the plutons and their aureoles significantly improves the geodynamical models of orogens based only on structural information derived from the country rocks.

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REFERENCES

Bell, T. H. and Johnson, S. E. (1992) Shear sense: a new approach that resolves conflicts between criteria in metamorphic rocks. *Journal of metamorphic Geology* **10**, 99–124.

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

Blès, J. L. and Gros, Y. (1980) La fracturation du granite de Bassiès (Pyrénées ariégeoises, France): chronologie des phases tectoniques, évolution des diaclases. *Bulletin de la Société de géologique de France* **22**, 377–390.

Bodin, J. and Ledru, P. (1986) Nappes hercyniennes précoces à matériel dévonien hétéropique dans les Pyrénées ariégeoises. *C. R. de l'Académie des Sciences Paris* **302**, 969–974.

Bouchez, J. L. (1977) Plastic deformation of quartzites at low temperature in an area of natural strain gradient. *Tectonophysics* **39**, 25–50.

Bouchez, J. L. and Pêcher, A. (1981) The Himalayan main central thrust pile and its quartz-rich tectonites in central Nepal. In *The Effect of Deformation on Rocks. Tectonophysics* **78**, 23–50.

Calvet, P. and Charvet, J. (1988) Mise en évidence microtectonique d'une phase hercynienne précoce dans la région de Pierrefitte (Hautes-Pyrénées). *C. R. de l'Académie des Sciences Paris* **307**, 997–1003.

Carreras, J. and Capella, I. (1994) Tectonic levels in the Paleozoic basement of the Pyrenees: a review and a new interpretation. *Journal of Structural Geology* **11**, 1509–1524.

Choukroune, P. (1976). Structure et évolution tectonique de la zone Nord-Pyrénéenne. Analyse de la déformation dans une portion de chaîne à schistosité subverticale. *Mémoires de la Société géologique de France* **127**.

Debon, F. and Zimmerman, J. L. (1988) Le pluton hercynien de Bassiès (Pyrénées, Zone Axiale): typologie chimique, âge et remaniements isotopiques. *C. R. de l'Académie des Sciences Paris* **306**, 897–902.

Etchecopar, A. and Malavielle, J. (1987) Computer models of pressure shadows: a method for strain measurement and shear-sense determination. *Journal of Structural Geology* **9**, 667–677.

Gleizes, G. (1992). Structure des granites Hercyniens des Pyrénées de Mont Louis-Andorre à la Maladeta. Unpublished thesis, Université de Toulouse.

Gleizes, G., Leblanc, D. and Bouchez, J. L. (1991) Le pluton granitique de Bassiès (Pyrénées ariégoises): zonation, structure et mise en place. *C. R. de l'Académie des Sciences Paris* **312**, 755–762.

Golberg, J. M. and Leyreloup, A. F. (1990) High temperature–low pressure Cretaceous metamorphism related to crustal thinning (Eastern North Pyrenean Zone, France). *Contributions to Mineralogy and Petrology* **104**, 194–207.

Hanmer, S. and Passchier, C. (1991). Shear sense indicators: a review. *Geological Survey of Canada* 90-17.

Lagarde, J. L. (1978) Analyse du déversement d'une nappe gneissique: application à la nappe hercynienne du Canigou (Pyrénées orientales). *C. R. de l'Académie des Sciences Paris* **286**, 937–940.

Lamouroux, C., Soula, J. C., Deramond, J. and Debat, P. (1980) Shear zones in the granodiorite massifs of the Central Pyrenees and the behaviour of these massifs during the Alpine orogenesis. *Journal of Structural Geology* **2**, 49–53.

Leblanc, D., Gleizes, G., Roux, L. and Bouchez, J. L. (1996) Variscan dextral transpression in the French Pyrenees: new data from the Pic des Trois-Seigneurs granodiorite and its country rocks. *Tectonophysics* **345**.

Losantos, M., Palau, J. and Sanz, J. (1986) Considerations about Hercynian thrusting in the Marimanya massif (Central Pyrenees). *Tectonophysics* **129**, 71–79.

Majoor, F. J. M. (1988). A geochronological study of the axial zone of the central Pyrenees, with emphasis on Variscan events and Alpine resetting. Ph.D. thesis, Institute of Earth Sciences, Utrecht.

Matte, Ph. and Mattauer, M. (1987) Hercynian orogeny in the Pyrenees was not a rifting event. *Nature* **325**, 739–740.

Nagtegaal, P. J. C. (1969) Sedimentology, paleoclimatology and diagenesis of post-hercynian continental deposits in the south-central Pyrenees, Spain. *Leidse Geol. Meded.* **42**, 143–238.

Palau, J. and Sanz, J. (1989) The Devonian units of the Marimanya massif and their relationship with the Pyrenean Devonian facies areas. *Geodinamica Acta* **3**, 171–182.

Paquette, J. L., Gleizes, G., Leblanc, D. and Bouchez, J. L. (in press) Le granite de Bassiès (Pyrénées): un pluton syntectonique d'âge westphalien. Géochronologie U–Pb sur zircons. *C. R. l'Académie des Sciences Paris.*

Ramberg, H. (1963) The evolution of drag folds. *Geological Magazine* **100**, 97–106.

de Sitter, L. U. and Zwart, H. J. (1959) Geological map of the Central Pyrenees, 1:50000, Sheet 1. Ariège. *Geol. Inst. Leiden Univ. and Leidse Geol. Meded.* **22**, 191–236.

Soliva, J., Salel, J. L. and Brunel, M. (1989) Shear deformation and emplacement of the gneissic Canigou thrust nappe, Eastern Pyrenees. *Geologie Mijnbouw* **68**, 357–366.

Van den Eeckhout, B. and Zwart, H. J. (1988) Hercynian crustal-scale extensional shear zone in the Pyrenees. *Geology* **16**, 135–138.

Verhoef, P. N. W., Vissers, R. L. M. and Zwart, H. J. (1984) A new interpretation of the structural and metamorphic history of the western Aston massif (central Pyrenees, France). *Geologie Mijnbouw* **63**, 399–410.

Vissers, R. L. M. (1992) Variscan extension in the Pyrenees. *Tectonics* **11**, 1369–1384.

Vitrac-Michard, A., Albarede, F., Dupuis, C. and Taylor, H. P. (1980) The genesis of Variscan (Hercynian) plutonic rocks: inferences from Sr, Pb and O studies on the Maladeta igneous complex, Central Pyrenees (Spain). *Contributions to Mineralogy and Petrology* **72**, 57–72.

Vitrac-Michard, A. and Allègre, C. (1975) A study of the formation and history of a piece of continental crust by ^{87}Rb–^{86}Sr method: the case of the French Oriental Pyrenees. *Contributions to Mineralogy and Petrology* **50**, 257–285.

Wickham, S. M. and Oxburgh, E. R. (1985) Continental rifts as a setting for regional metamorphism. *Nature* **318**, 330–338.

Zwart, H. J. (1979) The geology of the Central Pyrenees. *Leidse Geol. Meded.* **50**, 1–74.