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Hercynian structure of the Axial Zone of the Pyrenees: the Aran Valley cross-section (Spain–France)

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Abstract—In the Axial Zone of the Pyrenees Palaeozoic rocks crop out with superimposed Hercynian and Alpine deformation. The Hercynian structures present two domains: infrastructure and suprastructure. A complete cross-section of the central part of the Axial Zone of the Pyrenees is presented, in which Hercynian structures of both domains can be observed. In the Aran Valley cross-section, the infrastucture corresponds to the Garona Dome Domain, where the main Hercynian structures are north-vergent recumbent folds, with an associated, mainly flat-lying, foliation (S_2). In the Garona Dome Domain, the peak of the medium-to-high grade metamorphism is simultaneous with the main Hercynian structures. The Aran Valley and Noguera Ribagorça areas correspond to the suprastructure. In these domains the main Hercynian structures are upright folds, with an associated, mainly steep-lying, foliation (S_3). In these suprastructural domains, low grade syn-S_3 metamorphism is present. This cross-section relates the structures of the infra- and the suprastructure, suggesting a deformational sequence for the Axial Zone of the Pyrenees. This sequence is deduced from the fact that the north-vergent folds observed in the Garona Dome are earlier than the main structures of the suprastructure. The boundary of the Garona Dome and the suprastructure is a décollement level observed at the base of the Silurian. The deformational sequence proposed here provides some constraints for models of Hercynian evolution and the emplacement of the Alpine structures of the Pyrenees. Copyright © 1996 Elsevier Science Ltd

INTRODUCTION

The Axial Zone of the Pyrenees, which constitutes the core of the Pyrenean Alpine Orogen, contains Palaeozoic rocks with superimposed Hercynian and Alpine deformations (Fig. 1a). The importance of the Hercynian deformation of the Palaeozoic rocks of the Pyrenees varies from place to place. In the Axial Zone of the Pyrenees two structural domains have been distinguished. The first of these domains has undergone low grade metamorphism and its main Hercynian structures are upright folds and axial planar cleavage. In the other domain, metamorphism ranges from medium-to-high grade and the main Hercynian structures are recumbent folds with an associated flat-lying foliation. Relying on these structural differences, Zwart (1963b) called the former domain the suprastructure and the latter the infrastructure. Carreras & Capellá (1994) reviewed the different models explaining the Hercynian structure of the Pyrenees, which gave rise to an interesting Discussion and Reply (Aerden 1995, Carreras & Capellá 1995). Some of these models involve important crustal extension, either: (i) during the early stages of deformation as a consequence of rifting (Wickham & Oxburgh 1985,1986); or (ii) during the last stages of the Hercynian deformation, simultaneously with the metamorphic peak and as a consequence of gravitational collapse (van den Eeckhout 1986, van den Eeckhout & Zwart 1988, Kriegsman *et al.* 1989, Vissers 1992, Aerden 1994). A key argument in the latter models is that the flat-lying foliation of the infrastructure occurred later than the suprastructure's upright folds and their associated steep cleavage, and that it derives from gravitational collapse of a thickened orogen. Other authors propose that the main Hercynian structures resulted from migmatitic and granitic rock diapirism (Soula 1982, Soula *et al.* 1986a,b, Pouget 1991). Pouget (1991) considers the main Hercynian structures of both the supra- and the infrastructure to have been formed simultaneously in the Bossost Dome area (Fig. 2a), due to diapiric emplacement of migmatites and granites coinciding with the metamorphic peak. However, Matte (1969) claimed that the steep structures postdate the flat-lying foliation.

This paper focuses on the study of the Aran Valley traverse, located in the central part of the Axial Zone of the Pyrenees (Fig. 2). This section has often been used to illustrate the different models proposed for the Hercynian structure (Matte 1969, Zwart 1979, Pouget 1991, Aerden 1994). Knowledge of the structure of this sector will thus provide a deformation sequence of the Hercynian structures with which these different models must be compatible. Most of the structures in the Aran Valley traverse belong to the suprastructure, whereas the Garona Dome, located in the northern part, is part of the infrastructure (see Fig. 2). In this sector of the Pyrenees, the boundary between both domains is at a décollement level located in the Silurian (Matte 1969, García-Sansegundo 1990,1992).

The starting point of this work is a complete N–S cross-section of the Axial Zone Palaeozoic rocks, passing through Ariège, Arán Valley and Alta Ribagorça (Fig. 1). This cross-section shows the characteristics of the Hercynian and Alpine structures in areas affected by Hercynian medium-to-high grade metamorphism (Bossost Dome) and in regions where metamorphism is weak.

Using the data provided by the geological map and

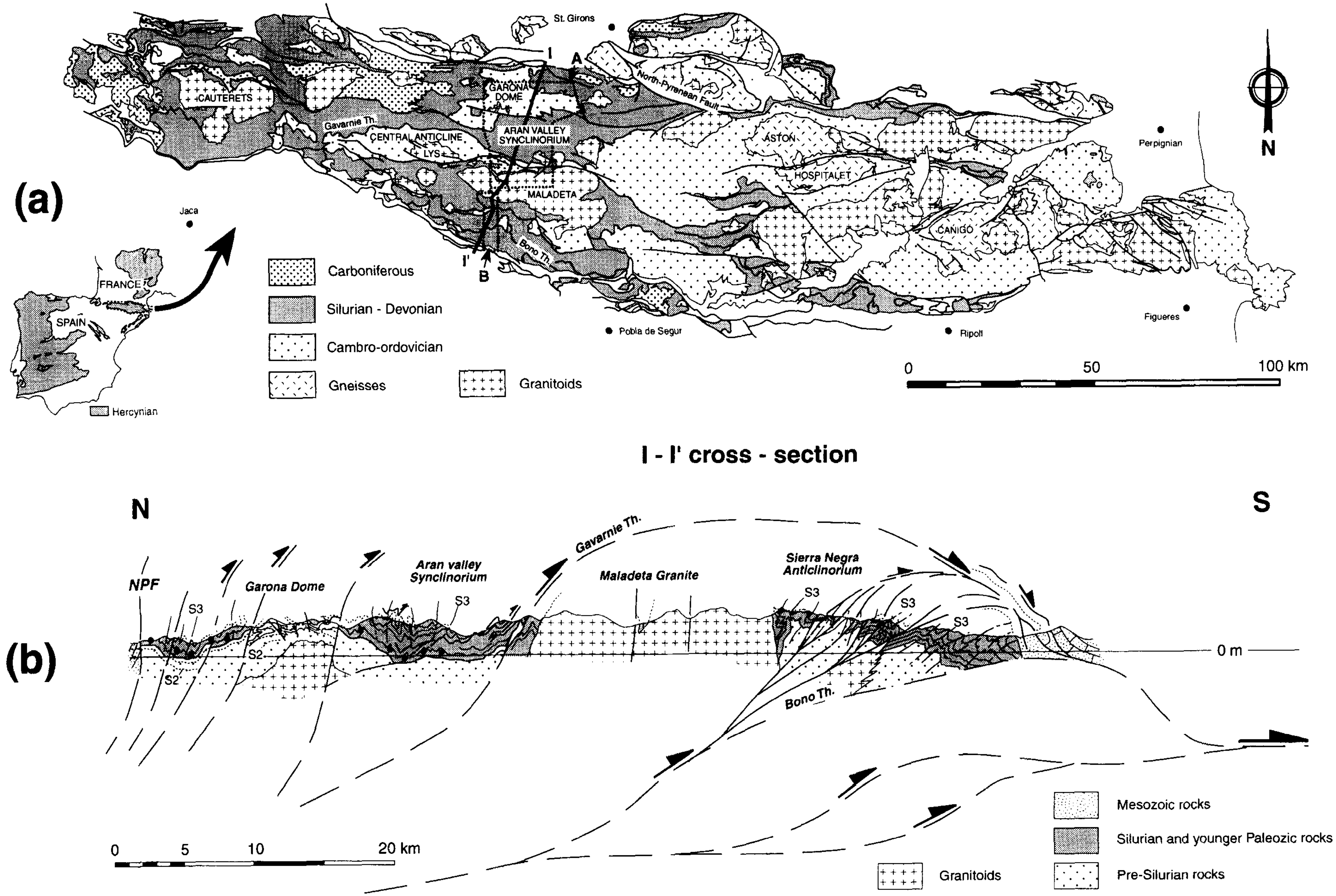

Fig. 1. (a) Geological map of the Axial Zone of the Pyrenees. A: Geological map of Fig. 2(a), B: Geological map of Fig. 2(b). I–I' cross-section of (b). Th: thrust. (b) Cross-section through the Axial Zone of the Pyrenees in the Aran Valley and Noguera Ribagorça areas; the deep structures have been interpreted from the ECORS profile. NPF: North Pyreneean fault. Location in (a).

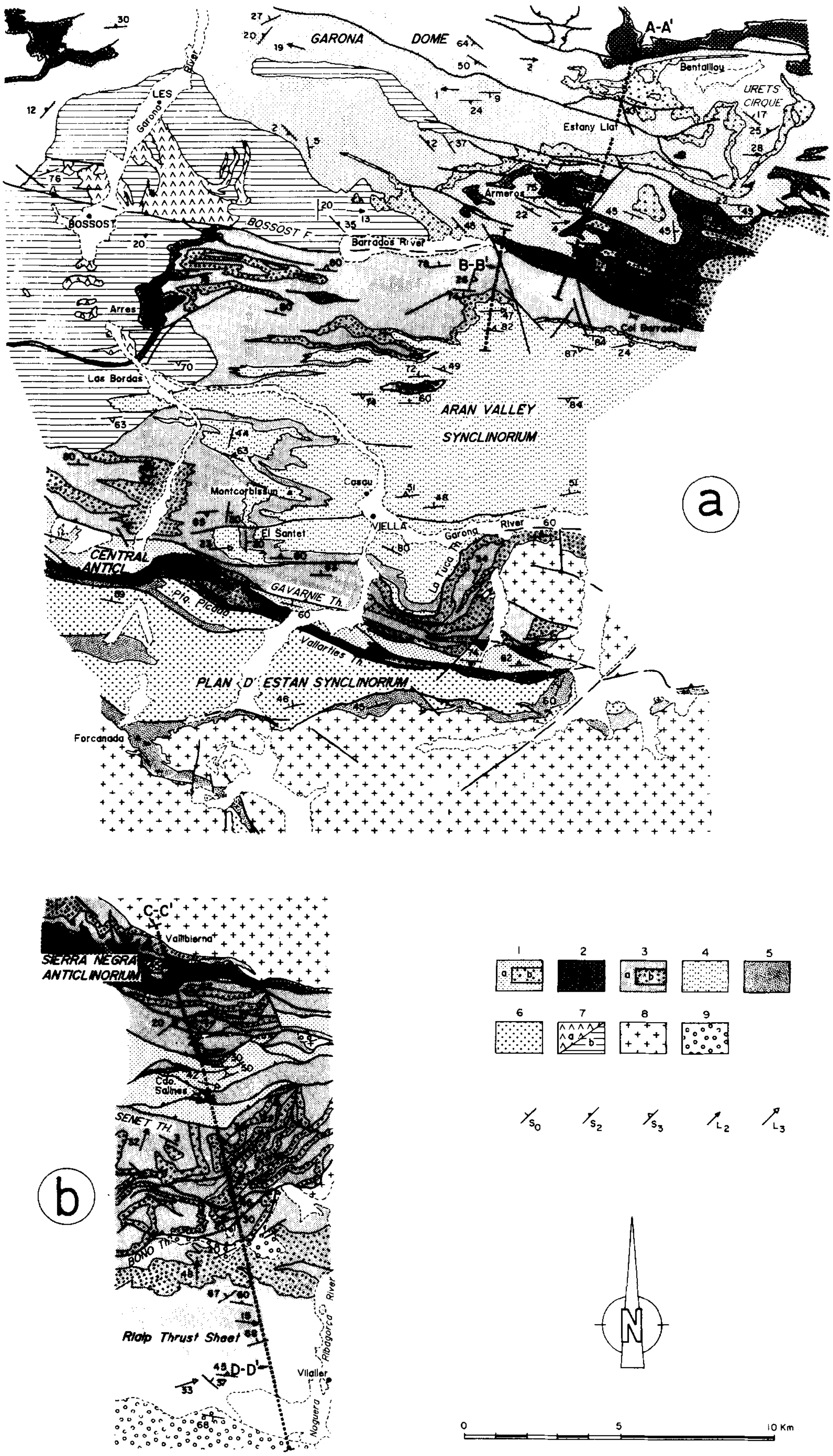

Fig. 2. (a) Geological map of the Aran Valley area. Northern part of study area: Garona Dome, Aran Valley domain and northern sector of Noguera Ribagorça domain. A–A′, cross-section of Fig. 4. B–B′, cross-section of Fig. 5(a). (b) Geological map of the Alta Ribagorça area. Southern part of the study area: southern sector of Noguera Ribagorça domain, C–C′, cross-section of Fig. 6, D–D′, cross-section of Fig. 7. Legend: (1a) Cambro–Ordovician rocks; (1b) Bentaillou limestone (Cambro–Ordovician age); (2) Silurian black slates; (3a) Devonian slates and limestones; (3b) Devonian limestones; (4) Devonian sandstones and slates; (5) Upper Devonian and Lower Carboniferous limestone; (6) Carboniferous greywackes and slates (Culm); 7a) Leucogranities, migmatites and pegmatites. (7b) Rocks with medium-to-high grade metamorphism. (8) Granodiorites; (9) Post-Hercynian rocks.

cross-section, the structures of the different domains in this traverse section are related and a deformation sequence is proposed (Figs. 1 and 2). Since the results of this study put constraints upon other models, these are discussed in a final section. In the same way, by studying the Hercynian structure of this cross-section, the position of the Alpine thrust sheets within Palaeozoic rocks can be established, which is essential for the adequate interpretation of the ECORS-Pyrenees deep seismic profile (ECORS Pyrenees team 1988, Choukroune *et al.* 1990, Mattauer 1990, Muñoz 1992).

GEOLOGICAL SETTING

The Pyrenees Mountain Range (Fig. 1) resulted from the collision between the Iberian and European plates during the Upper Cretaceous and the Neogene. Its hinterland consists mainly of Palaeozoic rocks, the Axial Zone, that were uplifted by crustal scale Alpine thrusts. The southern margin of this Axial Zone presents an antiformal stack of the thrust sheets involving Palaeozoic rocks with Hercynian deformation and Mesozoic rocks (Parish 1984, Williams & Fischer 1984, Deramond *et al.* 1985, Muñoz 1985, Williams 1985). The total estimated Alpine shortening of the Pyrenees Mountain Range is between 100 and 150 km (Roure *et al.* 1989, Muñoz 1992), which roughly represents between 35% and 50%.

Several authors have studied the stratigraphic succession in the Aran Valley cross-section (Kleinsmiede 1960, Mey 1967,1968, García-Sansegundo & Alonso 1989, García-López *et al.* 1990, 1991, García-Sansegundo 1992), which is shown in Fig. 3. The older rocks are Cambro–Ordovician slates, quartzites, limestones and conglomerates, over which Upper Ordovician rocks were laid down unconformably (Den Brok 1989, García-Sansegundo & Alonso 1989). The Silurian rocks consist of black slates, often serving as a décollement level. The Devonian rocks form a thick series with remarkable facies changes, ranging from siliciclastic series in the northern area, to slates and limestones in the middle zone and predominantly slaty series in the south (Fig. 3). The Carboniferous rocks exhibit limestones at the base, overlain by greywackes and slates in culm facies.

Three structural domains have been distinguished in this cross-section. The first one is the Garona Dome (De Sitter & Zwart 1962) and is part of the infrastructure. This domain is located in the northern part of the area under study. Its main Hercynian structures are recumbent folds with an associated subhorizontal foliation. The central part of this domain, in which the outcropping rocks show medium-to-high grade metamorphism, is called the Bossost Dome (De Sitter & Zwart 1962). The peak of the low-pressure metamorphism in this dome is synchronous with the main flat-lying Hercynian foliation of the Garona Dome. The metamorphic zones of the Bossost Dome are quite narrow and it is possible to shift from medium-to-high to low grade metamorphic areas within a few metres. This metamorphism is a good

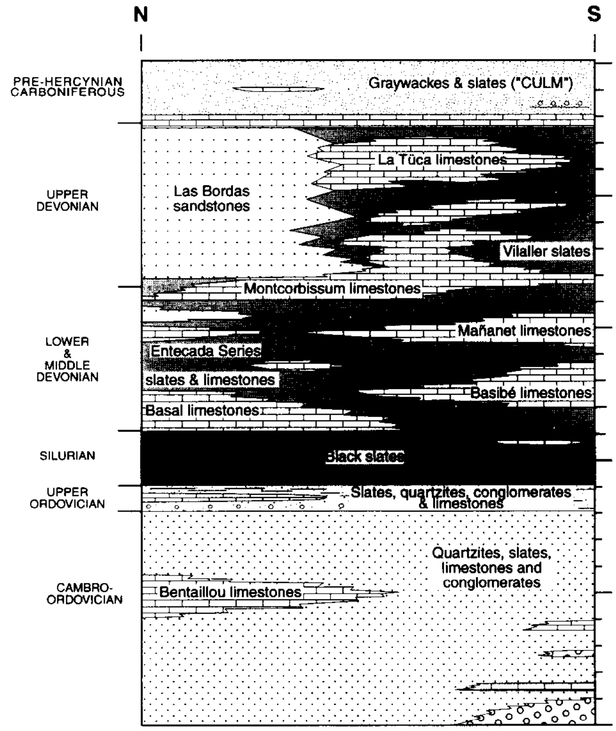

Fig. 3. Correlation chart showing the stratigraphic succession of the Palaeozoic of the Aran Valley traverse. The vertical scale is approximate.

example of low-pressure metamorphism (Zwart 1958, 1962, 1963a, 1979, 1986, Pouget 1991).

South of the Garona Dome, the Aran Valley Domain is found, where the main Hercynian structures are upright folds with an associated steep cleavage. The Aran Valley Domain is located within the suprastructure. In this domain metamorphism is weak. The Silurian décollement level forms the boundary between the Garona Dome and Aran Valley domains.

The third domain is the Alta Ribagorça Domain, whose Hercynian structures are essentially similar to those in the Aran Valley Domain. The difference with the previous domain is the important development of Alpine structures in this third domain. In order to analyse the complete cross-section, the structures of each domain will be first described and then compared.

STRUCTURE OF GARONA DOME DOMAIN

The Garona Dome is located in the northern part of the study area (Fig. 2a) and consists predominantly of pre-Silurian rocks. The first structures to be recorded in the Garona Dome (D_1) consist of a regional slaty cleavage (S_1), mainly visible in thin section. S_1 is defined by dimensional orientation of small muscovite, chlorite and quartz crystals, formed under low grade metamorphism. Provided the younging direction of the sequence is known, the relationship between bedding and S_1 has been used to deduce that these structures are south vergent (in Stille's sense; see Bell 1981, p. 197). No folds have been found associated with S_1. In the Garona Dome, where metamorphism is from medium to high grade, S_1 is difficult to recognise since it has been overprinted by later deformations.

D_2 structures in the Garona Dome are dominated by kilometre-scale, north-vergent recumbent or asymmetric, tight, type 1C or 2 (classification of Ramsay 1967) folds. Some of the main structures of this generation are the Urets, Liat and Armeros anticlines (Fig. 4). In sectors of this domain where metamorphism is weak, the main foliation (S_2) is a subhorizontal, axial planar crenulation cleavage. With medium to high grade metamorphism, S_2 is a well-developed schistosity. Often andalusite, staurolite and cordierite crystals grow synchronously with this schistosity. According to Zwart (1963a), the peak of the metamorphism developed after S_2. Other authors like Pouget (1991) claim that the development of both the foliation and the climax of the metamorphism occurred simultaneously.

In the Garona Dome domain, apart from D_1 and D_2 structures, two sets of upright folds can be recognized whose trends are oblique. These two fold systems deform the previously described structures (D_1 and D_2) and are responsible for the dome structure (Figs 2 and 4). The first set consists of local upright folds trending N-150/170-E. The trend of the second set of folds (D_4) varies from E–W to N-100-E. These folds have an associated penetrative crenulation that overprints S_1, S_2 and the décollement level of the Silurian, which is associated with the main folds of the Aran Valley domain (D_3). From the crenulation lineations, it can be inferred that the set closer to the E–W trend (D_4) is younger than the N-150/170-E one.

Apart from these folds, the Garona Dome also shows south-directed thrusts, with minor displacement cutting across the main foliation (S_2) and the N-150/170-E trending local upright folds. In addition, the thrusts are folded by the upright E–W trending folds (D_4). As will be shown below, these thrusts are associated with Hercynian folds (D_3) of the upper domain. The upright folds trending E–W (D_4) may be Alpine structures.

STRUCTURE OF THE ARAN VALLEY DOMAIN

This domain is located between the Garona Dome and the Alpine Gavarnie Thrust (Figs. 1 & 2). The rocks outcropping in the Aran Valley are mainly Devonian. They are folded, at cartographic scale, giving rise to the Aran Valley Synclinorium.

D_1 structures of the Garona Dome have not been found in the Aran Valley Domain. The first structures recorded in this domain are E–W trending and northward vergent asymmetric folds, that can be correlated with the D_2 structures of the Garona Dome. These folds have an associated slaty cleavage (S_2). On the whole, they are poorly developed and have overturned limbs seldom greater than 100 m. S_2, in turn, occurs almost all over the domain. In thin section, it is marked by a predominant muscovite, chlorite and quartz preferred grain shape dimensional orientation.

The second set of structures in the Aran Valley Domain (D_3 structures) is the most important. It comprises N-110-E trending upright or south-vergent folds that are responsible for the current configuration of the Aran Valley Synclinorium (Figs. 1b & 5a). They span a wide range of sizes, are quite tight and belong to type 1C (classification of Ramsay 1967). In this cross-section, the fold axes plunge 30°–50° eastward. These structures have an associated foliation (S_3), which is mostly developed as a crenulation cleavage. D_2 and D_3 fold systems give rise to interference figures of type 3 in Ramsay's (Ramsay 1967) classification (Fig. 5a). This domain is affected by the medium grade metamorphism of the Bossost Dome. Some thin sections reveal that the metamorphic minerals pre-date the main cleavage (S_3) development.

Apart from these two generations of folds, the southern limb of the Aran Valley Synclinorium has thrusts deformed by D_3 folds (La Túca Thrust, Fig. 2a).

At the base of the Silurian black slates, a generalized décollement level is recorded throughout the area. This décollement level forms a sole for south-trending thrusts which converge towards the base and cut across the subhorizontal foliation S_2. The thrusts become D_3 folds in the upper part of the domain (Fig. 2a, southern part of Figs 4 and 5b). The Silurian décollement level forms the boundary between the Garona Dome and Aran Valley domains.

STRUCTURE OF THE ALTA RIBAGORÇA DOMAIN

The Alta Ribagorça domain is situated between the Alpine Gavarnie Thrust and the southern boundary of the Axial Zone. This domain is the hangingwall of the Gavarnie Thrust (cross-section of Fig. 1). Its Hercynian structure is similar to that of the Aran Valley Domain. However, the Alpine thrusting within the Alta Ribagorça Domain gave rise to an antiformal stack that can be observed at many different places in the southern part of the Axial Zone (Parish 1984, Williams & Fischer 1984, Deramond *et al.* 1985, Muñoz 1985, 1992, Williams 1985).

D_1 structures of the Garona Dome have not been found in this domain. The first set of structures observed in the Alta Ribagorça Domain consists of predominantly E–W trending, asymmetrical and northward vergent folds (D_2). Their E–W trend is evident on the southern limb of the Sierra Negra Anticlinorium, where some of the most prominent structures of this generation can be seen (Fig. 6). To the south, these earlier structures, and the later ones, show a N–S trend (Fig. 2b). These folds have an associated slaty cleavage, which, on the basis of correlation with other domains, is interpreted to be S_2. This foliation is best developed in the northern sectors. In fact, south of the traverse section, some sectors do not develop S_2 cleavage at all.

The most outstanding structures in this domain, as in the previous domain, are D_3 upright or south-vergent folds with E–W trend, and subhorizontal axes. The E–W

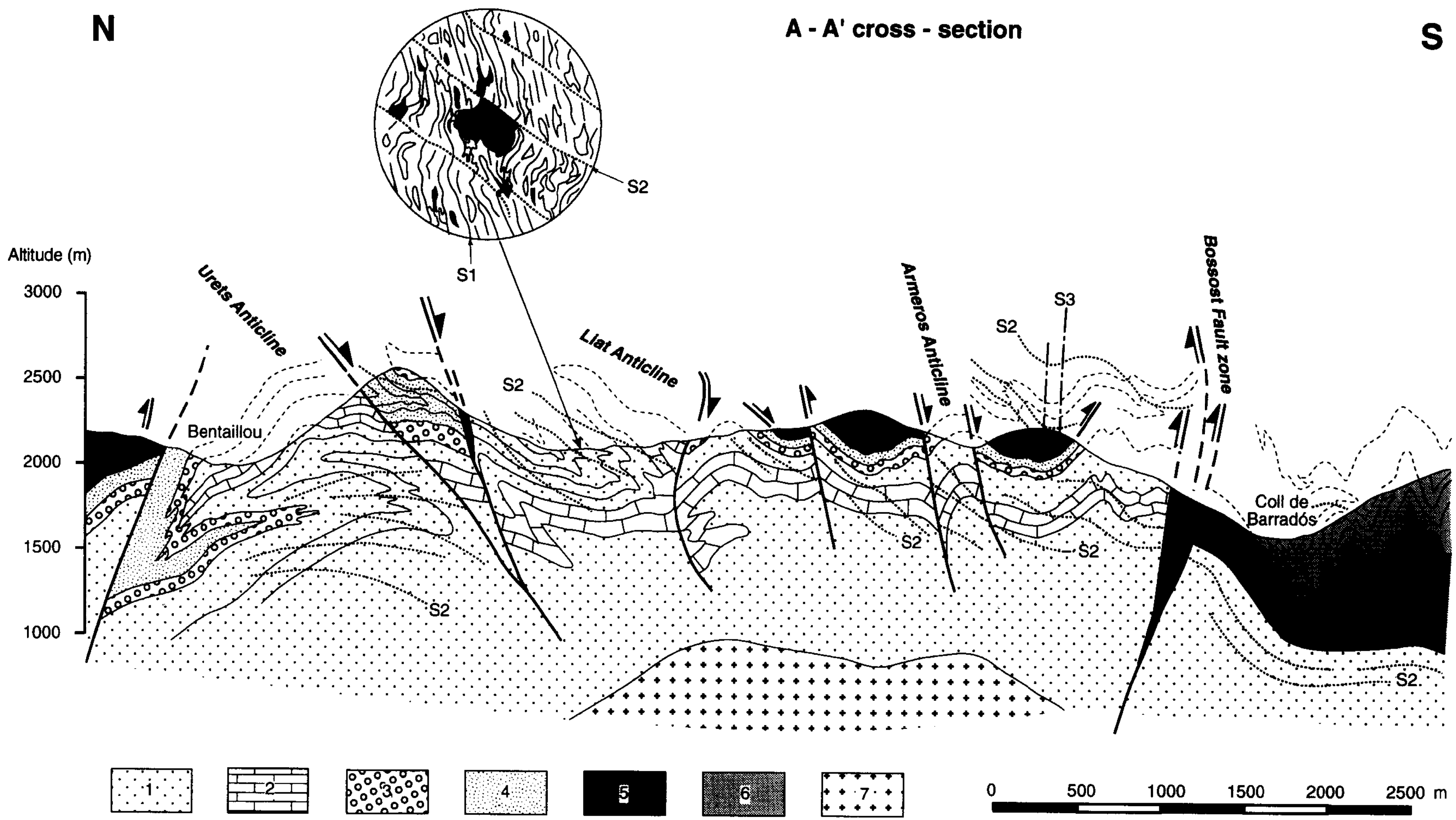

Fig. 4. Cross-section of the Garona dome. (1) Cambro–Ordovician quartzites and slates; (2) Bentaillou Limestone; (3) Upper Ordovician discordant conglomerates and sandstones; (4) Upper Ordovician slates, quartzites and limestones; (5) Silurian black slates; (6) Devonian slates and limestones; (7) Leucogranites, pegmatites and migmatites. Note the microscopic relationship between S_1 and S_2 cleavages. Location in Fig. 2(a).

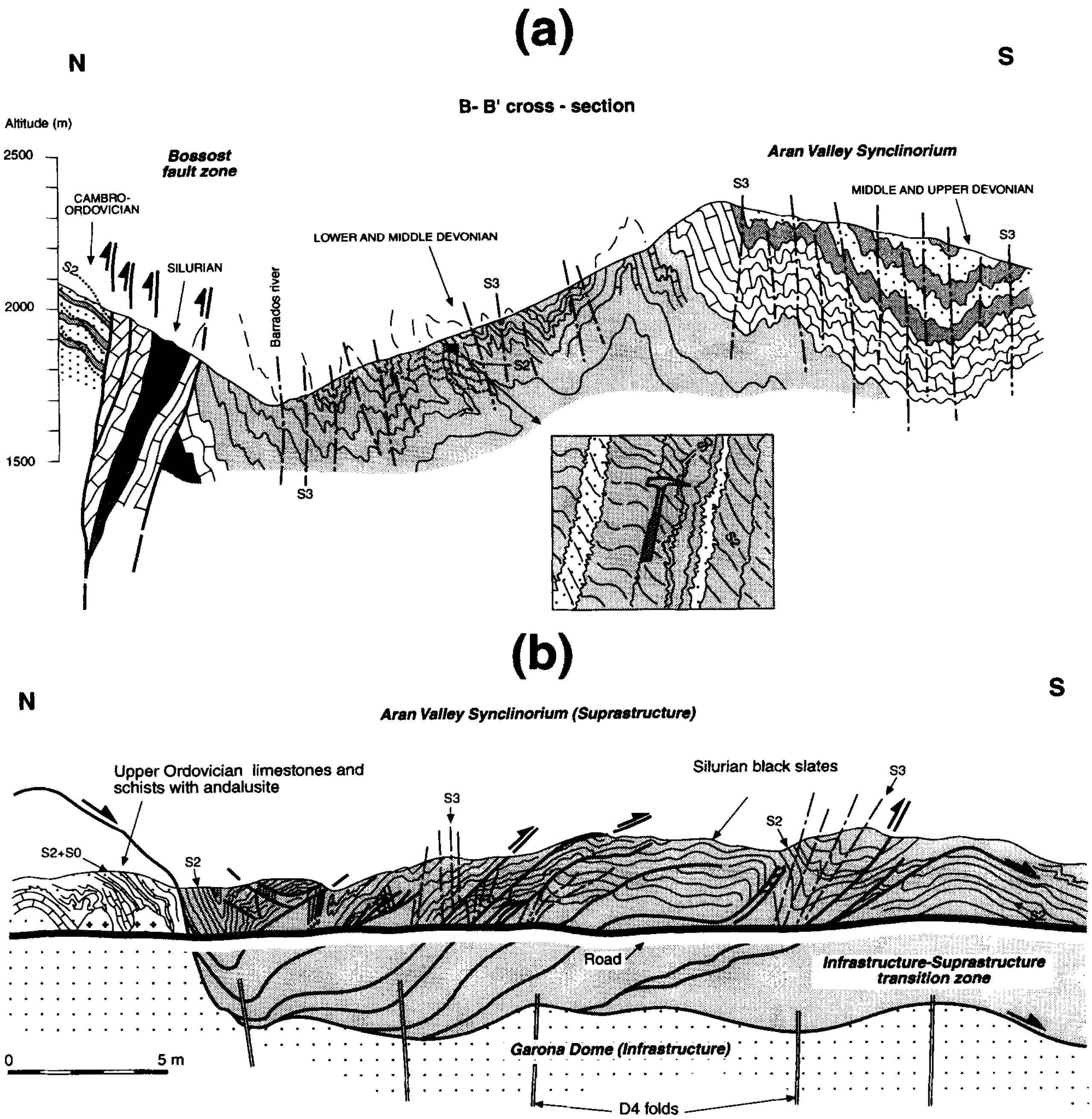

Fig. 5. (a) Cross-section of the northern limb of Aran Valley synclinorium. Notice the north-vergent folds deformed by the upright structures of the main phase of this domain. In the outcrop indicated, an overturned D_2 limb with vertical S_0 is observed. S_2 is gently folded by D_3 folds and S_3 is not developed. Location in Fig. 2(a). (b) Décollement level located at the base of the Silurian black slates in Arres. This décollement level is the boundary between the Garona Dome and Aran Valley domains and it deforms the S_2 of the Garona Dome. S_3 foliation is associated with the décollement level. The metamorphism in this area is middle to high grade. Location in Fig. 2(a).

trend of these structures is evident practically everywhere. However, in the south, all the folds in the Bono and Rialp thrust sheets, have almost N–S trends. Similar structures have been described in nearby areas to the east of the traverse section (Speksnijder 1987). The D_3 folds are tight type 1C folds and account for the the main cartographic structures of the traverse section (Central Anticline, Plan d'Estan Synclinorium and Sierra Negra Anticlinorium. See Figs. and 2). The main foliation in this domain (S_3) is associated with them. In this domain, D_2 and D_3 structures produce good examples of Ramsay's type 3 fold interference patterns (Figs. 6 and 7). The foliation associated with D_3 folds is a well-developed crenulation cleavage, especially in the northern sectors where it completely overprints the S_2 and bedding. In the south, it may occur as a primary cleavage.

Apart from the folds, between the Central Anticline and the Plan d'Estan Synclinorium (Fig. 2a), a thrust places Silurian rocks on top of Carboniferous rocks (Puerto Picada–Valarties Thrust). A sector of the Puerto Picada–Valarties Thrust is deformed by the main D_3 folds of this domain. Nevertheless, in other sectors of this domain, this thrust cuts across the D_3 folds. Hence, it is interpreted as a Hercynian thrust which was later reactivated during the Alpine deformation (García-Sansegundo 1992).

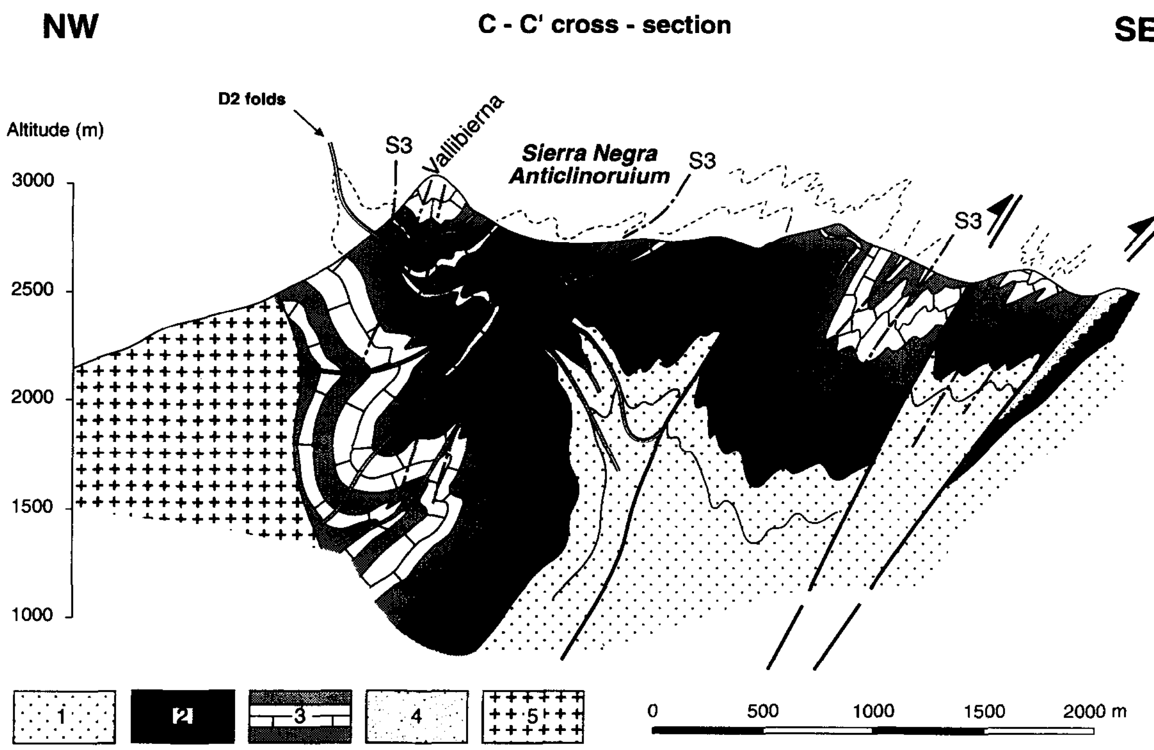

Fig. 6. Cross-section of the Sierra Negra anticlinorium. Note the Ramsay's type 3 fold interference pattern (north-vergent D_2 and south-vergent D_3 folds). (1) Cambro–Ordovician rocks; (2) Silurian black slates and limestones; (3) Devonian limestones and slates; (4) Permo–Triassic rocks; (5) Granodiorites. Location in Fig. 2(b).

As mentioned above, Alpine structures are dominant in the Alta Ribagorça Domain. Here, thrusts often deform Mesozoic rocks with folds and Alpine rough cleavage, and Palaeozoic rocks that preserve their Hercynian structure. One of the most important effects of Alpine structures on Palaeozoic rocks with Hercynian deformation is the subvertical attitude of the main Hercynian cleavage (S_3) in the northern part of the domain. Towards the south, however, S_3 cleavage dips some 30° northwards. This fan pattern of S_3 cleavage is caused by the Alpine antiformal stack (Fig. 7).

DEFORMATIONAL SEQUENCE

Considering the data provided by this cross-section of the Axial Zone of the Pyrenees, a deformational sequence is proposed with which the structures of both infrastructure (Garona Dome) and suprastructure domains (Aran Valley and Alta Ribagorça domains) are related. The deformational sequence of the Hercynian structures in the study area is as follows:

(D_1) *S_1 slaty cleavage*: This only found in the Garona Dome. This foliation may be associated with presumed south-vergent structures. The south-vergent attitude of S_1 is inferred using S_0 and S_1 relations when the younging direction is well known.

(D_2) *North-vergent, asymmetrical or recumbent folds with crenulation cleavage or schistosity (S_2).* This is the main foliation in the Garona Dome. In the rest of the study area, these north-vergent folds and associated cleavage are the earliest structures recognised. In the Garona Dome these structures are synchronous with the peak of the metamorphism (Pouget 1991).

(D_3) *Upright folds trending E–W.* The associated crenulation or slaty cleavage (S_3) is the main foliation in the Aran Valley and Alta Ribagorça domains. In the Aran Valley domain, these structures are related to a décollement level located in the Silurian black slates, which is also considered as a third generation structure. Moreover, the Puerto de la Picada–Valarties and La Túca thrusts (Fig. 2a), which can be related to the décollement level located in the Silurian black slates, are also considered to be D_3 structures.

(D_4) *Upright folds trending E–W, which are only found in the Garona Dome.* These structures deform the décollement level of the Silurian rocks.

Other minor local structures are not included in this deformational sequence. The N-150/170-E trending upright folds of the Garona Dome and the N–S trending folds described by Speksnijder (1987), close to the Alta Ribagorça Domain would be among these.

CONCLUSIONS AND DISCUSSION

In the traverse section of the Axial Zone of the Pyrenees, the deformation sequence suggested is valid for both medium-to-high grade and low-to-very-low

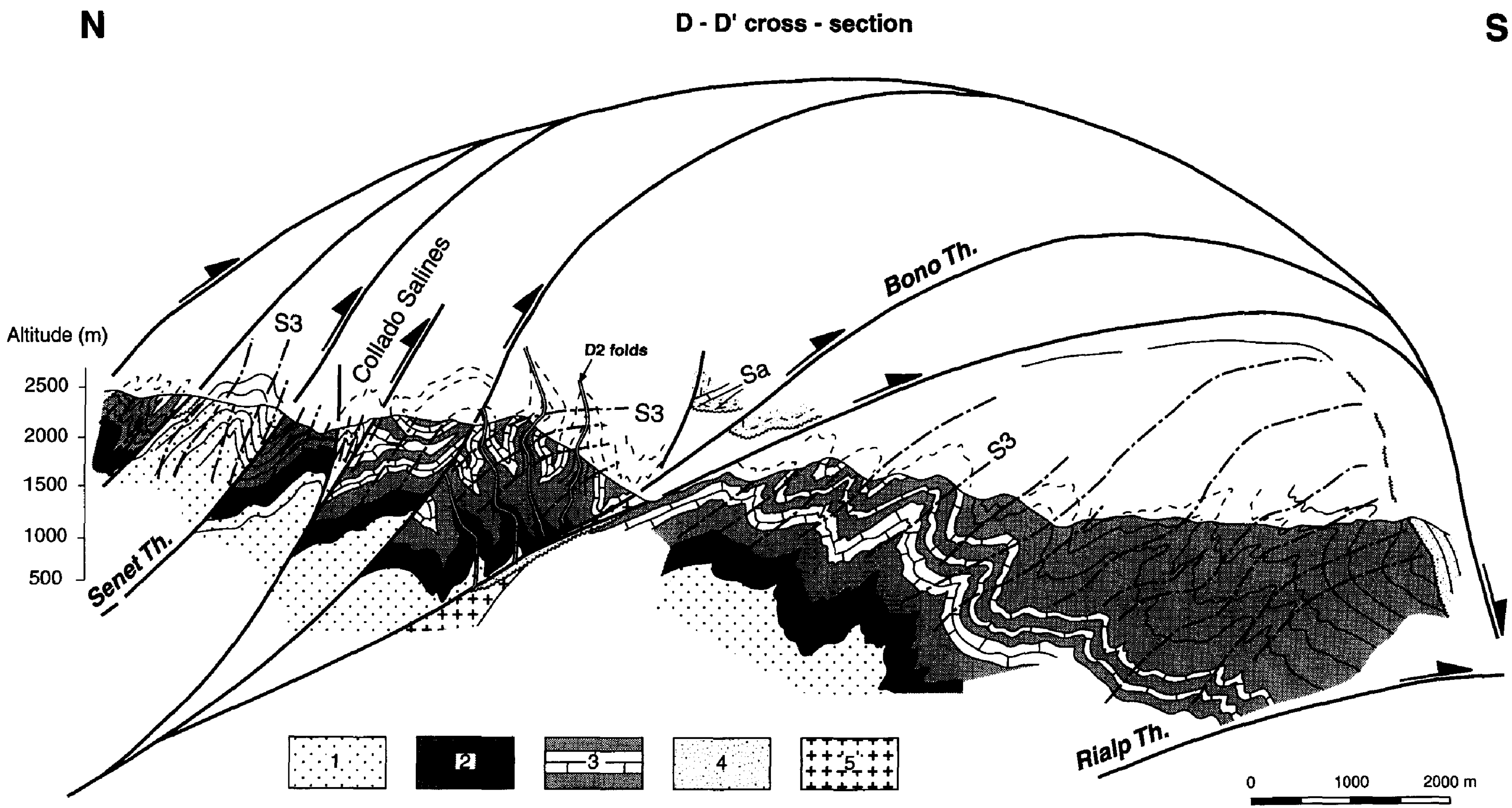

Fig. 7. Cross-section of southern part of Alta Ribagorca domain. Note in the Bono thrust sheet the Ramsay's type 3 fold interference pattern (north-vergent D_2 and south-vergent D_3 folds). The Rialp thrust sheet shows a gently Alpine folding of S_3. (1) Cambro–Ordovician rocks; (2) Silurian black slates; (3) Devonian limestones and slates; (4) Permo–Triassic rocks; (5) Granitic rocks. Location in Fig. 2(b).

grade metamorphic areas. Considering the data presented in this paper, the flat-lying foliation (S_2) of the infrastructure developed before the steep cleavage (S_3) of the suprastructure. Using superimposed structures criteria, it is possible to infer that the recumbent structures (D_2 folds and foliation) in the infrastructure domain, with medium-to-high grade metamorphism, originated before the upright folds and associated steep S_3 cleavage of the suprastructure domain, where metamorphism is from low to very low grade. Owing to this, the distinction between synchronous suprastructure and infrastructure, in the sense stated by several authors (Zwart 1963b, Pouget 1991) must be revised. As stated in the introduction to this paper, Pouget (1991) claims that the formation of the Garona Dome is associated with the main structures (D_2), and would have resulted from the diapiric uplift of migmatic rocks. This interpretation must be revised, since according to the information provided in the present paper, the structures that gave rise to the current configuration of the Garona Dome (D_4) formed after the main structures (D_2 and D_3) had developed.

The most pronounced of all the structures found in the traverse of the Axial Zone of the Pyrenees are the second generation structures (D_2). Although, on the whole, the Hercynian structures of this zone are southward-vergent, the second generation structures, both in the infrastructure and suprastructure, are northwards vergent. The difference could lie in the fact that, at a certain stage during the Hercynian deformation, some difficulties could have arisen which would have hampered general shortening of the mountain range. This, in turn, could have caused north-vergent structures to occur. No evidence has been found, however, of any such difficulties. Nevertheless, correlation of the north-vergent structures of the infrastructure and the suprastructure seems reasonable.

Regarding the theories on the extensional nature of the main Hercynian structures of the infrastructure (D_2 structures of the Garona Dome), which mainly rely on the fact that pressure conditions during the metamorphism was low, it should be pointed out that after the hypothetical extensional deformation (D_2) there was another compressive episode, characterized by the main structures of the suprastructure. After recent data recorded by Gibson & Bickle (1994) from the Canigó massif to the east of the study area (Fig. 1), which the authors applied to the rest of the Hercynian structures of the Pyrenees, the temperature conditions of the peak of the metamorphism were $725 \pm 25°C$ and the pressure was in the range of 4.5 ± 0.5 kbar (equivalent to 16 ± 2 km crust thickness). Such conditions correspond to a geothermal gradient of 45–50 °C km^{-1}, remarkably lower than the one previously suggested by other authors (Zwart 1963a, Pouget 1991). These conditions are quite similar to those obtained at many other locations of the Hercynian Mountain Range.

The Axial Zone of the Pyrenees may have undergone extension after the compressive deformation, consistent with the igneous activity observed at different points, which could have developed a contact metamorphism after deformation D_3 had taken place (south Bossost Dome, Maladeta Granodiorite, etc.). However, the main Hercynian structures (D_2 and D_3 structures) do not seem to be associated with this extensional episode. On the other hand, flattening of foliation S_2 of the Bossost Dome, as a result of late-Hercynian extension cannot be rejected by the present data.

As a consequence of the Alpine deformation, the Hercynian structures of the southern part of the study area dip strongly southwards. Considering the angle between the bedding of the Mesozoic rocks and the main Hercynian cleavage (S_3) of the suprastructure, as well as the current position of some Hercynian structures (e.g. décollements at the base of the Silurian), it is possible to conclude that in the northernmost parts of this zone, the Alpine thrust sheets have not rotated much. This conclusion has been independently reached by McClelland & McCaig (1989) on the basis of palaeomagnetic data.

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REFERENCES

Aerden, D. G. A. M. 1994. Kinematics of orogenic collapse in the Variscan Pyrenees deduced from microstructures in porphyroblastic rocks from the Lys-Caillaouas massif. *Tectonophysics* **238**, 139–160.

Aerden, D. G. A. M. 1995. Tectonic levels in the Paleozoic basement of the Pyrenees—a review and a new interpretation: Discussion. *J. Struct. Geol.* **17**, 1489–1491.

Bell, A. M. 1981. Vergence: an evaluation. *J. Struct. Geol.* **3**, 197–202.

Carreras, J. & Capellá, I. 1994. Tectonic levels in the Paleozoic basement of the Pyrenees: a review and a new interpretation. *J. Struct. Geol.* **16**, 1509–1524.

Carreras, J. & Capellá, I. 1995. Tectonic levels in the Paleozoic basement of the Pyrenees—a review and a new interpretation: Reply. *J. Struct. Geol.* **17**, 1493–1495.

Choukroune, P., Pinet, B., Roure, F. & Cazes, M. 1990. Major Hercynian thrusts along the ECORS Pyrenees and Biscay lines. *Bull. Soc. géol. Fr., 8 Ser* **6**, 313–320.

De Sitter, L. U. & Zwart, H. J. 1962. Geological map of the Paleozoic of the Central Pyrenees, 1:50 000; sheet 1 Garonne, sheet 2 Salat. *Leid. geol. Meded.* **27**, 191–236.

Den Brok, S. M. J. 1989. Evidence for pre-Variscan deformation in the Lys-Caillouas area, Central Pyrenees, France. *Geologie Minjb.* **68**, 377–380.

Deramond, J., Graham, R. M., Hossack, J. R., Baby, P. & Crouzet, G. 1985. Nouveau modèle de la Chaîne des Pyrénées. *C. r. Acad. Sci., Paris* **301**, 1213–1216.

ECORS Pyrenees team. 1988. The ECORS deep reflection seismic survey across the Pyrenees. *Nature* **331**, 508-511

Eeckhout, van den B. 1986. A case study of a mantled gneiss antiform, the Hospitalet massif Pyrenees (Andorra, France). *Geologica Ultraiectina* **45**, 1–193.

Eeckhout, van den B. & Zwart, H. J. 1988. Hercynian crustal scale extensional shear zone in the Pyrenees. *Geology* **16**, 135–138.

García-López, S., García-Sansegundo, J. & Arbizu, M. 1990. Datos estratigráficos y paleontológicos de la sucesión devónica del área del Río Baliera (Zona Axial, Pirineos centrales españoles). *Geogaceta* **7**, 33–35.

García-López, S., García-Sansegundo, J. & Arbizu, M. 1991. Devonian of the Aran Valley Synclinorium, Central Pyrenees, Spain: Stratigraphical and Paleontological data. *Acta geol. Hisp.* **26**, 55–66.

García-Sansegundo, J. 1990. Structure of the Paleozoic in the Aran valley, Axial zone, central Pyrenees. *Bull. Soc. géol. Fr., 8 Ser* **6**, 229–239.

García-Sansegundo, J. 1992. Estratigrafía y Estructura de la Zona Axial Pirenaica en la Transversal del Valle de Aran y de la Alta Ribagorça. *Spec. Publs. Bol. geol. Min.*, 1–167.

García-Sansegundo, J. & Alonso, J. L. 1989. Stratigraphy and structure of the southeastern Garona Dome. *Geodinamica Acta* **3**, 127–134.

Gibson, R. L. & Bickle, M. J. 1994. Thermobarometric constraints on the conditions of metamorphism in the Canigou massif, Pyrenees: implications for Hercynian geothermal gradients. *J. geol. Soc. Lond.* **151**, 987–997.

Kleinsmiede, W. F. J. 1960. Geology of the Valle de Arán (Central Pyrenees). *Leid. geol. Meded.* **25**, 129–245.

Kriegsman, L. M., Aerden, D. G. A. M., Bakker, R. J., den Brok, S. W. J. & Schutjens, P. M. T. M. 1989. Variscan tectonometamorphic evolution of the eastern Lys-Caillaouas massif, Central Pyrenees—evidence for a late orogenic extension prior to peak metamorphism. *Geologie Minjb.* **68**, 323–333.

Mattauer, M. 1990. Une autre interprétation du profil ECORS Pyrénées. *Bull. Soc. géol. Fr., 8 Ser* **6**, 307–311.

Matte, Ph. 1969. Le problème du passage de la schistosité horizontale à la schistosité verticale dans le dôme de Garonne (Paléozoïque des Pyrénées Centrales). *C. r. Acad. Sci., Paris* **268**, 1841–1844.

McClelland, E. & McCaig, A. M. 1989. Palaeomagnetic estimates of rotations in compressional regimes and potential discrimination between thin-skinned and deep crustal deformation. In *Palaeomagnetic Rotations and Continental Deformation* (edited by Kissel, C and Laj, C.), 365–379.

Mey, P. H. W. 1967. The Geology of the Upper Ribagorzana and Baliera valleys, Central Pyrenees, Spain. *Leid. geol. Meded.* **41**, 153–220.

Mey, P. H. W. 1968. Geology of the Upper Ribagorzana and Tor valleys, Central Pyrenees, Spain. *Leid. geol. Meded.* **41**, 229–292.

Muñoz, J. A. 1985. Estructura Alpina i Herciniana a la vora sud de la Zona Axial del Pirineu Oriental. Unpublished Ph.D. Thesis, Universidad de Barcelona, 1–305.

Muñoz, J. A. 1992. Evolution of a Continental Collision Belt: ECORS-Pyrenees Crustal Balanced Cross-section. In: *Thrust Tectonics* (edited by McClay, K. R.). Chapman and Hall, 235–246.

Parish, M. 1984. A structural interpretation of a section of the Gavarnie nappe and its implications for Pyrenean Geology. *J. Struct. Geol.* **6**, 247–255.

Pouget, P. 1991. Hercynian tectonometamorphic evolution of the Bosost dome (French–Spanish central Pyrenees). *J. geol. Soc. Lond.* **148**, 299–314.

Ramsay, J. G. 1967. Folding and fracturing of rocks. McGraw-Hill, New York.

Roure, F., Choukroune, P., Berastegui, X., Muñoz, J. A., Villien, P., Matheron, P., Bareyt, M., Seguret, M., Cámara, P. & Deramond, J. 1989. ECORS Deep Seismic data and balanced cross sections: Geometric constraints on the evolution of the Pyrenees. *Tectonics* **8**, 41–50.

Soula, J. C. 1982. Characteristics and mode of emplacement of gneiss domes and plutonic domes in central-eastern Pyrenees. *J. Struct. Geol.* **3**, 313–342.

Soula, J. C., Debat, P., Deramond, J., Gucherreau, J. Y., Lamouroux, Ch., Pouget, P. & Roux, L. 1986. Evolution structurale des emsembles métamorphiques des gneiss et des granitoïdes dans les Pyrénées centrales. *Bull. Soc. géol. Fr.* **8**, 79–93.

Soula, J. C., Debat, P., Deramond, J. & Pouget, P. 1986. A dynamic model of the structural evolution of the Hercynian Pyrenees. *Tectonophysics* **129**, 115–147.

Speksnijder, A. 1987. The detection and significance of early deformation in the southern Variscan Pyrenees, Spain; implications for regional Paleozoic structural evolution. *Geol. Rdsch.* **76**, 451–476.

Vissers, R. L. M. 1992. Variscan extension in the Pyrenees. *Tectonics* **6**, 1369–1384.

Wickham, S. M. & Oxburgh, E. R. 1985. Continental rifts as a setting for regional metamorphism. *Nature* **318**, 330–333.

Wickham, S. M. & Oxburgh, E. R. 1986. A rifted tectonic setting for Hercynian high-thermal gradient metamorphism in the Pyrenees. *Tectonophysics* **129**, 53–69.

Williams, G. D. 1985. Thrust tectonics in the south central Pyrenees. *J. Struct. Geol.* **7**, 11–17.

Williams, G. D. & Fischer, M. W. 1984. A balanced section across the Pyrenean orogenic belt. *Tectonics* **3**, 773–780.

Zwart, H. J. 1958. Regional metamorphism and related granitization in the Valle de Arán (Central Pyrenees). *Geologie Minjb.* **20**, 18–30.

Zwart, H. J. 1962. On the determination of polymetamorphic mineral associations, and its application to the Bosost area (Central Pyrenees). *Geol. Rdsch.* **52**, 38–65.

Zwart, H. J. 1963a. Metamorphic history of the Central Pyrenees, Part II. Valle de Arán, sheet 4. *Leid. geol. Meded.* **28**, 321–376.

Zwart, H. J. 1963b. The structural evolution of the Paleozoic of the Pyrenees. *Geol. Rdsch.* **53**, 170–205.

Zwart, H. J. 1979. The Geology of the Central Pyrenees. *Leid. geol. Meded.* **50**, 1–74.

Zwart, H. J. 1986. The Variscan Geology of the Pyrenees. *Tectonophysics* **129**, 9–27.