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Strain analysis in the Axial Zone of the Variscan basement of the Pyrenees

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Abstract

The structural zonation in the Axial Zone of the Pyrenees, in which an infrastructure and a suprastructure are distinguished, can be correlated with a variation in the orientation of the principal strain axes. The deformation associated with the main foliation in the infrastructure is accommodated heterogeneously. Computed strain ellipsoids in gneisses show an apparent flattening strain (k = 0.29), which is in contrast to the apparent constrictional forms of pebbles in adjacent layers. Strain values in the suprastructure indicate an apparent flattening strain (k = 0.36) of low magnitude (R_{xy} = 1.36 and R_{yz} = 2.01), which can be taken as representative of the bulk finite strain throughout that domain. The trend of the X-axes is roughly E–W, parallel to the fold axes in the infrastructure, whereas it is NW to N-plunging and at a wide angle to the ESE–WNW trending folds in the suprastructure. The Z-axes are steeply plunging in the infrastructure and almost horizontal in the suprastructure. Strain patterns in each of the two structural domains are suggested to be associated with different tectonic events. Between the infrastructure and the suprastructure, a transition zone exists where both domains grade into one another.

Introduction

The Pyrenean mountain chain is of Alpine age and involves a Variscan basement and a Mesozoic and Paleogene cover. The main body of basement rocks occurs in the approximately E–W trending centre of the chain, which is called the Axial Zone. The Variscan basement of the Axial Zone and north-Pyrenean massifs constitutes an isolated segment of the Variscan belt of western Europe, although its structural and metamorphic events and igneous activity are analogous to those elsewhere in the belt.

The Axial Zone consists of Paleozoic rocks which were affected by polyphase tectonics, and metamorphosed and intruded during the Variscan orogeny in late Carboniferous times. Paleogene compressive events affecting the Variscan basement have led to the development of conspicuous thrust structures (Williams & Fischer 1984, Déramond et al. 1985, Muñoz 1985, Roure et al. 1989). In addition, some relatively narrow shear zones have been considered to be associated with Alpine tectonics (Lamouroux et al. 1980, McCaig 1986, Soula et al. 1986a).

Within the Variscan basement of the central Pyrenees, there is a variation in structural style and metamorphic grade. De Sitter & Zwart (1960) and Zwart (1963) distinguished two domains, termed infrastructure and suprastructure. The infrastructure is characterized by metamorphic rocks above the biotite isograd, and has a gently dipping foliation as its dominant structure. The rocks, commonly micaschists and augengneisses, occur in large-scale dome structures with an E–W to ESE–WNW trend, some of them with a gneissic core (e.g. the Aston-Hospitalet massif). The suprastructure is characterized by low-grade metamorphism and a steep axial plane foliation, often a slaty cleavage. This fabric is related to ESE–WNW trending folds up to kilometre-scale. Between these two structural domains both main foliations coexist; this zone is designated as the transition zone (Oele, 1966).

Autran & Guitard (1969) and Guitard (1970) did not recognize the subdivision of infrastructure and suprastructure in the eastern Pyrenees (Roc de Frausa and Canigó-Carancà massifs). These authors inferred that the main foliation was flat-lying across the entire tectonic edifice and that later deformation modified this initial geometry. However, in the southwest part of the Canigó-Carancà massif, where an upper front of the main flat-lying foliation is inferred (Santanach 1972, Capellà 1991), recent work (Capellà 1991) has revealed a vertical variation in the structural style similar to that described earlier for the central Pyrenees. It is important to note that the main foliation in the suprastructure, a spaced cleavage and usually a lower strain fabric than in the central Pyrenees, is gently dipping. In this area an infrastructure, a suprastructure and even a transition zone have also been established (Capellà 1995).

There are some problems concerning the relationship between the two structural domains distinguished. On the one hand, several authors (Oele 1966, Zwart 1979, Soula 1982, Soula et al. 1986b) suggest that the main foliation in rocks with relatively high-grade, as well as that in rocks with low-grade metamorphism are the same and developed simultaneously. Verhoef et al. (1984), Van den Eeckhout (1986), De Bresser et al. (1986), Kriegsman et al. (1989), García-Sansegundo (1991) and Capellà (1991), on the other hand, claim that there are two different foliations which are not contemporaneous. Another major unsolved question for the last authors is which foliation is the oldest. Verhoef et al. (1984), Van den Eeckhout (1986), De Bresser et al. (1986) and Kriegsman et al. (1989) have shown that the steep axial plane foliation is overprinted by the gently dipping structures. In contrast, new data from the Garona dome (García-Sansegundo 1991), the Pallaresa anticlinorium and the Canigó-Carancà massif (Capellá 1995) allow the opposite to be interpreted.

Problems as to the geometry and structural correlation have provoked different tectonic models. The models proposed for the Variscan basement of the Pyrenees can be subdivided into three types; structural successions involving compression (Zwart 1963, 1979, Guitard 1970, Matte & Xu Zhi 1988, Carreras & Capellà 1994), extensional models (Wickham & Oxburgh, 1985), and models incorporating extension and compression (Soula et al. 1986b, Van den Eeckhout 1986, Van den Eeckhout & Zwart 1988).

There are, however, significant differences between authors who invoke the same model. In the compression model, two main discrepancies are apparent regarding the succession of the structures, i) the timing between the main structures in the relatively high-grade and low-grade rocks, and ii) the initial disposition of the structures in the relatively high-grade metamorphic rocks. According to Zwart (1963, 1979), one main foliation simultaneously developed throughout the entire tectonic edifice. On the other hand, Carreras & Capellà (1994) concluded that there are two main foliations, one in the high-grade, and the other in the low-grade rocks, which mostly developed at different times. Zwart (1963, 1979) deduced that the main foliation in the high-grade metamorphic domains had an initial dome shape, whereas Guitard (1970) inferred that the main foliation was more or less flat-lying and that it was overprinted by late folding. With regard to the models that involve extension and compression, the disagreement on the time relationship between the metamorphic peak and the main structural events has led to two completely different views. The school of Toulouse (Soula et al. 1986b) holds that two episodes can be clearly distinguished in the Variscan evolution of the Pyrenees. Crustal extension allowed the intrusion into the lower crust of basic rocks originating from partial melting of the upper mantle and producing the onset of the metamorphism. The extension was succeeded by shortening with southward moving thrusts controlling the diapiric rise of plutonic magmas as well as the metamorphic peak. The interference of the crustal shortening and the diapiric emplacement of magmas caused the geometry of the folds and the regional foliation: flat-lying above the plutonic rocks and steeply dipping at the margins. The assumption that the two main foliations were not coeval and that the metamorphic peak postdated the main structural events, led the Leiden-Utrecht group (Van den Eeckhout 1986, Van den Eeckhout & Zwart 1988) to develop a model also including both an extension and a compression episode, but with a time relationship opposite to that outlined above. A N–S crustal shortening gave rise to mostly E–W trending structures such as thrusts and steeply dipping folds and foliations in both shallow and deep structural levels. The overprinting of these steep structures by flat-lying foliations observed in metamorphic rocks (Verhoef et al. 1984), and the fact that the peak of a high-temperature and low-pressure metamorphism postdated all main structures, led to the supposition that a horizontal crustal extension with vertical thinning succeeded the earlier shortening event. Bulk extension was accommodated through a horizontal, large-scale shear zone.

Strain data in different areas and structural levels of the Variscan edifice can be used to evaluate the proposed tectonic regimes, and in fact should be taken into account in the tectonic interpretations. These data would allow the subdivision into an infrastructure and a suprastructure from a distinct viewpoint, important in the understanding of the configuration of these domains. To this purpose, strain measurements have been carried out in some areas of the central and eastern Pyrenees.

Strain analysis

Geological setting

Two pre-Variscan lithological units can be distinguished in the Axial Zone: augengneisses derived from either Cadomian or Ordovician granites (see Zwart 1979 and references therein), and a Paleozoic metasedimentary sequence. Samples for strain measurements in the infrastructure were collected from gneisses in the Canigó-Carancà and Hospitalet massifs. Strain estimates in the suprastructure were carried out in coarse-grained deposits of the Paleozoic, mainly thin microconglomerate layers interbedded in the Cambro-Ordovician, and conglomerates at the base of the Upper Ordovician. The samples come from the southern part of the Pallaresa anticlinorium, the Orri and Rabassa domes, and from the southern border of the Canigó-Carancà massif. Strain analysis of the transition zone was based on samples from the northern area of the Pallaresa anticlinorium. The location of the samples studied is shown in Figure 1.

Methodology

The R_f/Φ method (Ramsay 1967) of has been selected for strain analysis due to the elliptical geometry of the strain markers: K-feldspar megacrysts and quartzo-feldspathic aggregates in gneisses, and slate, quartz and quartzite pebbles in coarse-grained sediments. Microconglomerates and conglomerates are usually clast and matrix-supported, respectively.

2D analysis. Two different versions of the R_f/Φ method are applied, the algebraic version of Matthews et al. (1974) and the theta-curves method of Lisle (1977). Both methods have certain inherent assumptions, including i) homogeneous deformation of the markers without a ductility contrast with the matrix, and ii) an initially random fabric (or, if not random, a symmetric one in the algebraic version). Both versions (adopted for computer use) allow the strain ratio (R_s) to be calculated. Computational procedures used to test the initial distribution of the particles and to estimate statistically the error in the strain determination, are included in Matthews et al. (1974). In Lisle's theta-curves method, the value of χ^2 at the best-fit for the R_s value, allowing assessment of the distribution of the markers on an R_f/Φ plot, is comparable with that of Matthews et al. (1974). These additional procedures suggest that the 2D strain measurements are sufficiently accurate.

The strain ratios calculated according to both versions are very similar. The difference in 91% of the sections, is smaller than 10% of the lower R_s value. Higher or lower R_s values obtained through the two versions of the R_f/Φ method do not appear to be related systematically to one of the specified versions. Quartz pebbles in several samples of Upper Ordovician conglomerates have been analysed separately, to avoid the effects of ductility contrasts. R_s values obtained from quartz grains of usually low eccentricity are between 60 and 16% lower than the R_s value of the bulk strain of the section.

The Fry method (Fry 1979) has been applied to some microconglomerates with a high concentration of quartz grains, but with limited success, possibly due to an original non-anticlustered distribution. Strain ratios by this centre-to-centre technique vary from 70 to 125% of the value obtained through the R_f/Φ method.

3D analysis. For calculating three-dimensional strain ratios, the procedure used in Ramsay & Huber (1983: 188–190) is selected. In that procedure, the cleavage plane is assumed to be the XY plane. The procedure is based on finding values of the extensions in the three principal planes of the strain ellipsoid, through the strain ratios determined in the pertinent sections of the sample. As in most of the samples the X and Y-directions are not known, three sections of each sample are required by this procedure. Arbitrarily-oriented sections were cut at high angles to the XY plane and to each other.

Because the three sectional ellipses are expected to represent sections of the same ellipsoid, they need to be compatible. As the absolute principal extensions are not known (only strain ratios were computed) the sectional ellipses had to be scaled through their intersection lines in order to become compatible. After scaling these ellipses, the elongation along the intersection lines between the XY plane and three sections must be computed in order to determine three values of extension on the XY plane. The algebraic solution of De Paor (1988) or a Mohr circle construction (Ramsay 1967) can be applied to calculate the orientations and lengths of the X and Y principal axes, from any three known stretches on the plane. The Z-axis extension

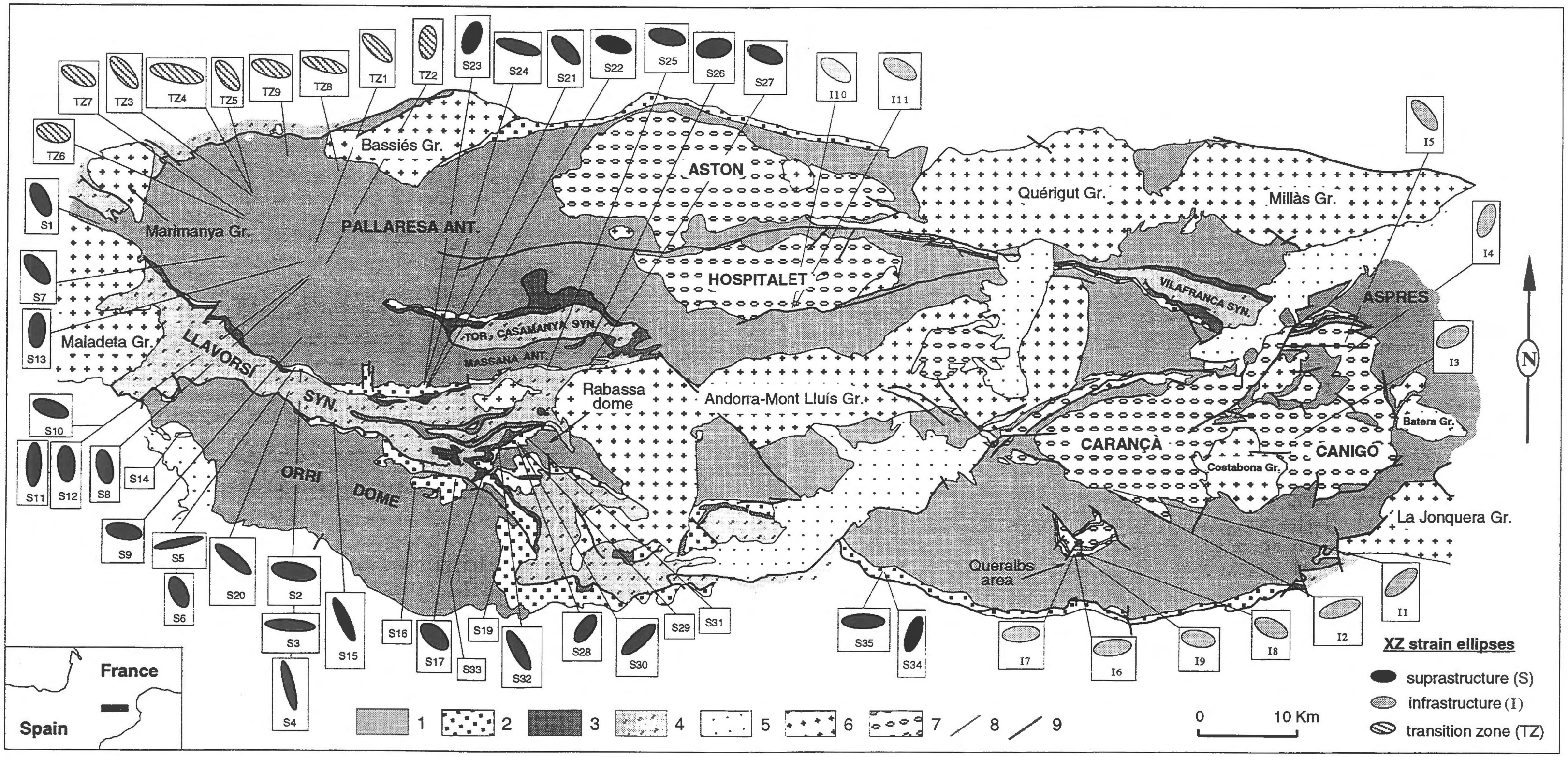

Figure 1. Geological sketch-map of the central and eastern Axial Zone of the Pyrenees (after Losantos et al. 1989, modified) showing the location of the samples used in strain analysis. Oriented ellipses correspond to the XZ principal planes of the finite strain ellipsoids. No volume change was assumed in the determination of the length of principal axes. Letter-number couples indicate structural domain and sample number, respectively. Samples used only to calculate the strain magnitude are not indicated with an XZ strain ellipse. Key: 1 Cambro-Ordovician; 2 Upper Ordovician; 3 Silurian; 4 Devonian and pre-Variscan Carboniferous; 5 post-Variscan cover; 6 gneisses; 7 Variscan batholiths; 8 lithological contact; 9 fault. Gr = granite.

is derived in a similar way. A computer program was written to perform these calculations for the present 3D analysis.

Strain distribution. Results

More than a hundred samples were collected, but only the strain data of those treated in this paper have been taken into account. More than half of the collected samples have been rejected because i) very often the sectional ellipses cannot be determined, or the sectional strain ratios are not reliable when the Fry method is applied; ii) statistical assessment procedures associated with the R_f/Φ method give values indicating that the assumptions on which the method is based have not been met; and iii) the three sectional ellipses do not fit into an ellipsoid, as the three intersection lines between the three sections and the XY plane do not unambiguously determine the Mohr circle. Three-dimensional strain was computed making use of strain ellipses derived from the algebraic version of Matthews et al. (1974). The number of measured markers for a section averages 80, exceeding the minimum sample sizes required by Matthews et al. (1974) and Lisle (1985), i.e. 20 and 10, respectively.

Only the augengneisses of the Canigó-Carancà and Hospitalet massifs allow strain magnitude calculations in the infrastructure domain. Finite strain ellipsoids reveal an apparent flattening strain (mean k = 0.29) of low magnitude (mean R_{xy} = 1.22, R_{yz} = 1.76 and R_{xz} = 2.16) in Figure 2a. Although clustered in the Flinn plot, the computed strain ellipsoids are not very representative of the apparent bulk strain related to the development of the main foliation in the infrastructure. Thus near the Carancà-Canigó gneisses (south flank), pebbles of conglomerate horizons usually show cigar-like forms, sometimes with a notable elongation when observed along the pervasive foliation planes. Also, the gneissic foliation displays an irregularly-developed stretching lineation, defined by the alignment of mainly K-feldspar megacrysts. The strain magnitude associated with the main foliation in the infrastructure seems to be accommodated inhomogeneously. Generally, the strain intensity in the competent felsic orthogneisses is lower than that recorded in the surrounding sediments.

With regard to the orientation of the X-axes, two areas in the Canigó-Carancà massif, the gneissic core and the Queralbs area, are distinguishable. As plotted in Figure 2b, two main orientations of the X-axes of the computed ellipsoids in the Canigó-Carancà massif can be identified. The ellipsoids from the gneissic core are usually arranged around the NE–SW stretching lineations (Figures 2b, c). In the Queralbs area, however, where the gneisses do not show a stretching lineation, the X-axes are roughly W-plunging (Figure 2b). In contrast to most of the axes in the gneissic core of the Canigó-Carancà massif, the X-axes of the ellipsoids in the Hospitalet gneisses differ from the measured E-plunging stretching lineations. Slight inaccuracy in the analytical procedure due to the low strain magnitude may have caused the inconsistency between both directions. Unlike the suprastructure, stretching lineations (although variably oriented) are parallel to minor fold axes in the sedimentary rocks in the infrastructure of the Canigó-Carancà massif (Santanach 1972, Capellá 1991). In the gneissic core, the axes of minor folds are mostly NE-trending (Casas 1984, Soliva et al. 1989). According to Santanach (1972), the bedding/main foliation intersection lineations in the Queralbs area plunge W to NW and thus match the orientation of the X-axes. A varying orientation of the strain ellipsoid occurs on the scale of the massif. However, on a broader scale, the main foliation planes in the infrastructural areas commonly show a W–E directed extension parallel to the fold axes (Zwart 1979, Alonso 1979, Van den Eeckhout 1986, De Bresser et al. 1986, Kriegsman et al. 1989).

The strain in the suprastructure corresponds to an apparent flattening type. It can be roughly defined by an ellipsoid with k = 0.36 and the mean principal strain ratios R_{xy} = 1.36, R_{yz} = 2.01 and R_{xz} = 2.75. Although the vast majority of the constructed finite strain ellipsoids for the suprastructure lie within the flattening field in the Flinn diagram (Figure 3a), it should also be noted that there is some dispersion. There is a variation in principal strain ratio values, mainly in the R_{yz}, yielding k-values varying from 0.05 to 0.95, but for the greater part ranging from 0.2 to 0.4.

In order to test the strain distribution in the suprastructure, four areas have been selected in the central Pyrenees. These are the north flank of the Orri dome, the southern area of the Pallaresa anticlinorium, the Massana anticline and the Rabassa dome. Although the principal strain ratios of the ellipsoids show some variability within each area (Figures 3b–e), the strain magnitude displays only a slight variation on a larger scale. Thus the means of R_{xy} and R_{yz} values range only from 1.26 and 1.84 in the Massana anticline, to 1.47 and 2.04 in the Rabassa dome and the Pallaresa anticlinorium respectively. Flattening strain also appeared in the eastern Pyrenees in two ellipsoids which fit very well with those computed for the central Pyrenees. Strain

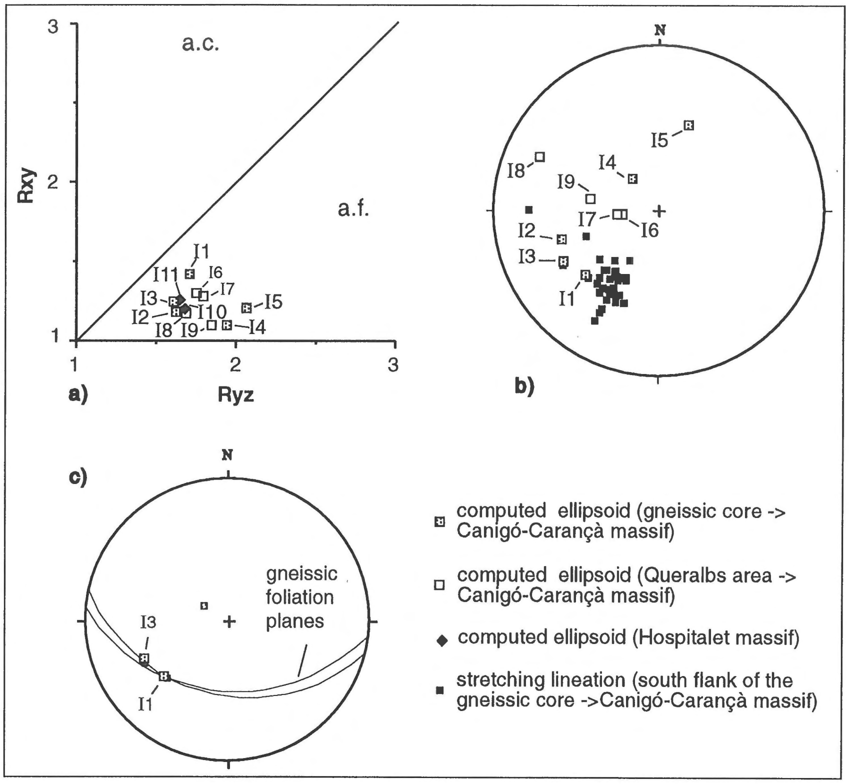

Figure 2. a) Flinn diagram of the calculated ellipsoids in the infrastructure domain (a.c.: apparent constriction field, a.f.: apparent flattening field). b) Lower hemisphere stereographic projection of the X-axes of ellipsoids and stretching lineations in the Canigó-Carancà massif. Ellipsoids obtained from the gneissic core of this massif (samples I1–I5, see Figure 1) are mainly close to stretching lineations which were also measured in the gneissic core. In the Queralbs area (samples I6–I9), to the south, the ellipsoids appear to be west-X-directed, matching the bedding/main foliation intersection lineations obtained by Santanach (1972) in that area. c) Plot of well-defined stretching lineations in samples I1 and I3 and their relation to the X-axes and the XY planes.

gradients have not been established in the suprastructure along or across the central Pyrenees.

Principal extension directions were derived by means of the three-dimensional strain computation procedure, and of the elongation of sulphide aggregates and associated pressure shadows in the cleavage planes. In the stereonets of Figure 4, the X-axes of the ellipsoids (limited to samples whose cleavage plane does not deviate from the regional cleavage attitude) and stretching lineations for the above-mentioned areas and adjacent zones have been plotted. Stereonets show on the one hand a relative scattering of the X-axes with regard to the mostly clustered stretching lineations, and on the other that there are no notable changes in the orientation of these principal extensions in the different areas. In spite of the dispersion shown by the X-axes, a prevalent orientation of the major extension is still discernible, plunging NW to N (Figure 5). That direction is at a wide angle to the ESE–WNW trending and relatively steep folds associated with the foliation of the suprastructure (the pitch ranges from 45–65° in the Massana anticline, to 70–85° in the Pallaresa anticlinorium). Other relationships between the fold and X-axes are observed locally. They may result from variably-oriented minor folds axes, which were probably produced by the superposition of the foliation on previous structures.

The regular pattern of the ellipsoids in the Flinn plot and of the major axes of these ellipsoids in a stereonet in different areas of the suprastructure, may indicate a rather homogeneous deformation on the scale of that domain, although some local heterogeneities are also present. In the fine-grained rocks (such as in the Cambro-Ordovician and Devonian), where stretching lineations were observed, the strain is higher than in coarse-grained ones. Thus, the ellipsoids, mostly from samples of the Upper Ordovician conglomerates, show rather low strain ratios, which might cause the

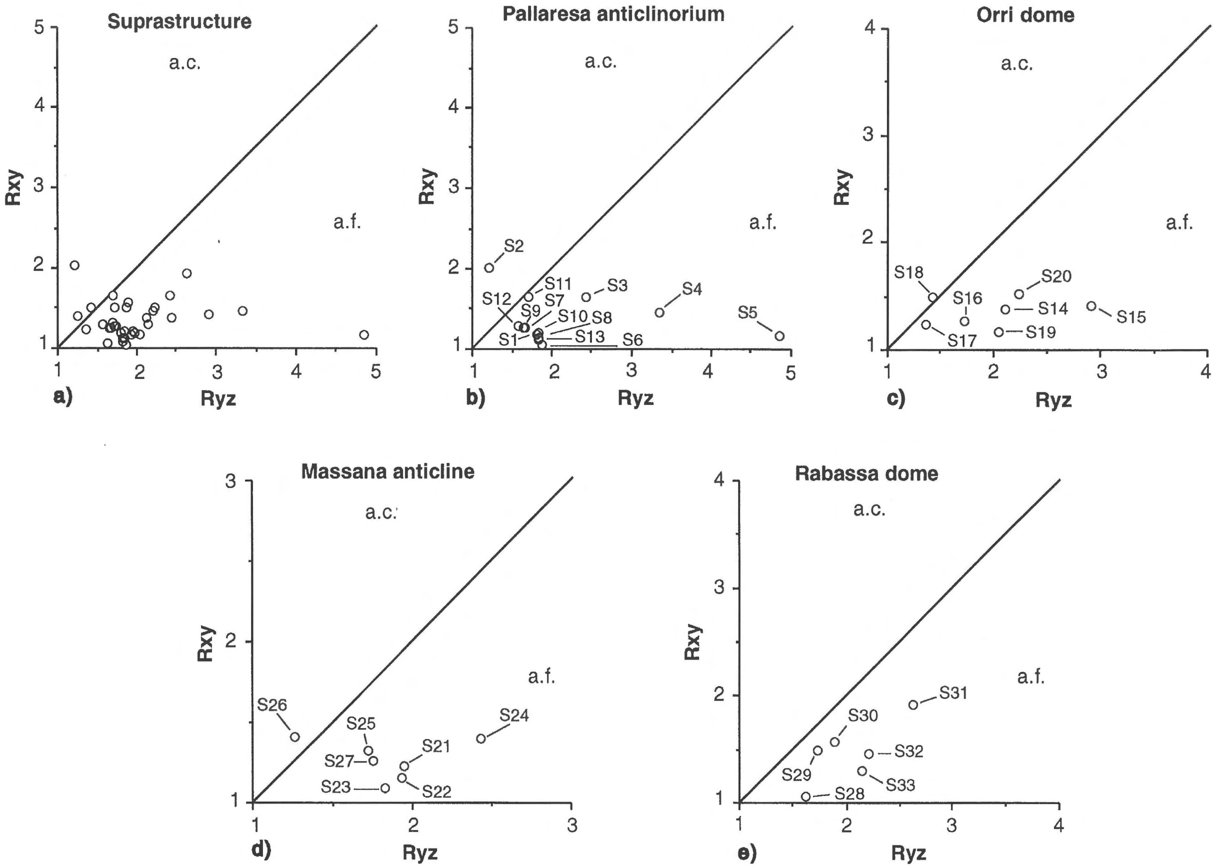

Figure 3. Flinn diagrams of the computed finite strain ellipsoids for the suprastructure (a). The same ellipsoids are shown for different structural units of the central Axial Zone (b–e). Sample S20 is located in the south flank of the Llavorsí syncline. (a.c. and a.f.: see Figure 2.)

spread observed of the X-axes. The low magnitude of the determined ellipsoids, which is probably a minimal one, contrasts with the well-developed fabric in the surrounding rocks. The fact that the assumptions of the method might have been not completely satisfied and that a possible loss in accuracy may occur due to low strain ratios, seems to result only in a slight variation in the ellipsoid form and orientation on a local scale.

Strain evaluation in the transition zone was carried out in the northern area of the Pallaresa anticlinorium. The gently-dipping foliation in the infrastructure in that area is intensively affected by one, and locally two steep pervasive crenulation cleavages, of which the oldest one corresponds to the foliation of the suprastructure (Capellà 1991). A low-magnitude apparent flattening strain (mean values, k = 0.43, R_{xy} = 1.33, R_{yz} = 1.76 and R_{xz} = 2.37) is evidenced by the plotted ellipsoids in the Flinn diagram of Figure 6a. These values indicate the bulk finite strain. Because of the superposition of the deformations it is impossible, however, to assess the strain associated with each of the successive structures.

The X-axes of the ellipsoids show an anticlustered distribution in the stereonet of Figure 6b. This probably results from the superposition of the different deformation structures, and precludes any conclusion with regard to the relationships between the strain X-axes and the fold axes.

A certain consistency is apparent between the low strain ratios, and the shape and orientation of the ellipsoids of the transition zone and the suprastructure. It is suggested that the structural configuration of the analysed levels of the transition zone is related to the main structural event in the suprastructure.

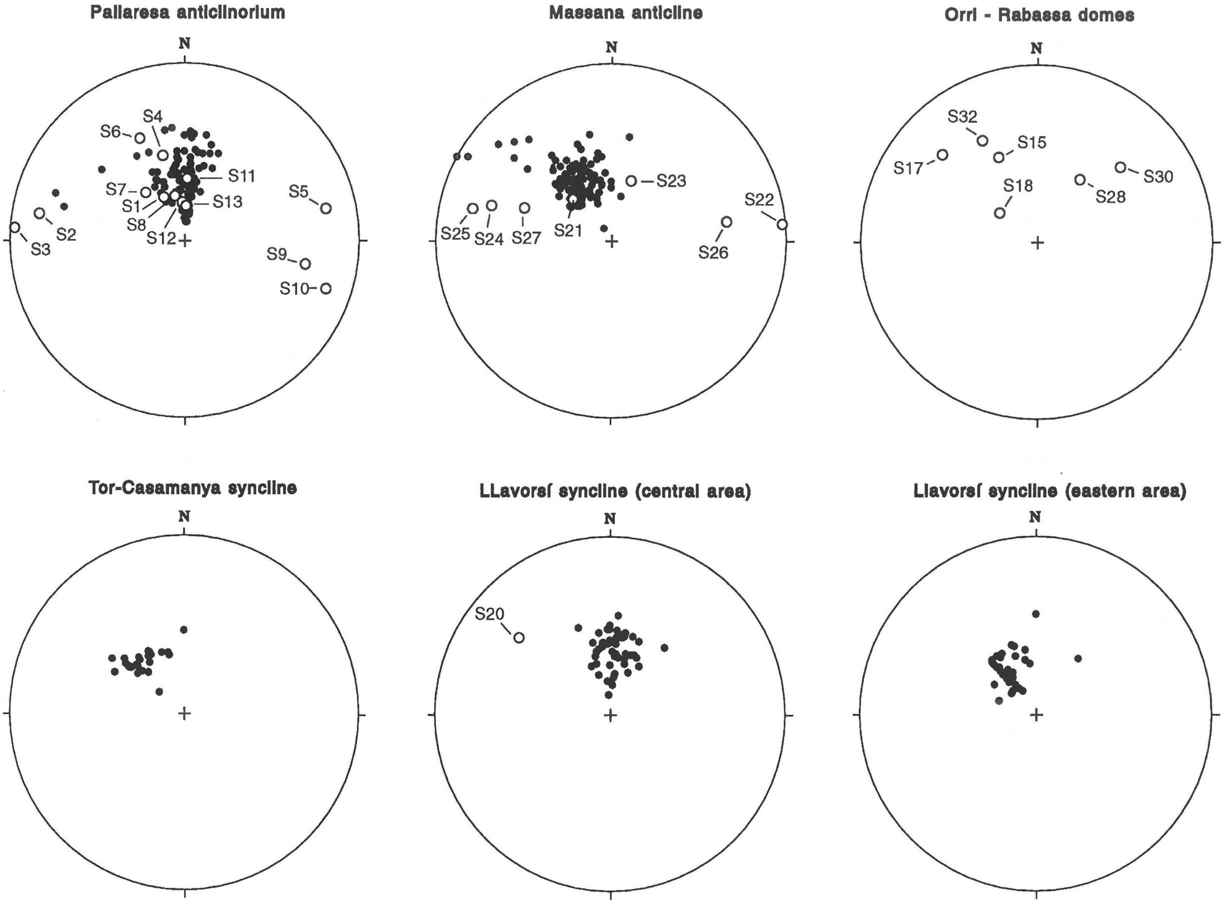

Figure 4. Lower hemisphere orientation diagrams of the X-axes (open circles) and stretching lineations (black circles) for several structural units in the suprastructure of the central Pyrenees.

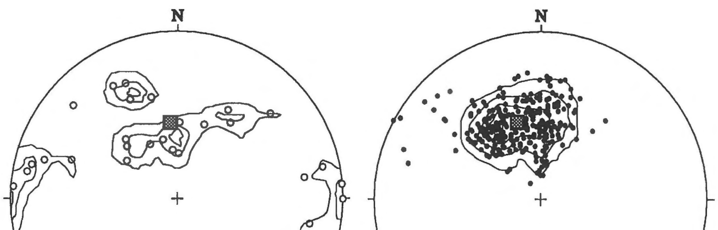

Figure 5. Lower hemisphere stereoplot of the computed ellipsoids' X-axes (a) and stretching lineations (b) of the suprastructure. a) 28 measurements, contours at 3, 6, 9% per 1% area. b) 301 measurements, contours at 3, 6, 12, 15, 18% per 1% area. The mean orientations, filled squares, of the X-axes and stretching lineations are 356/53 and 343/51, respectively.

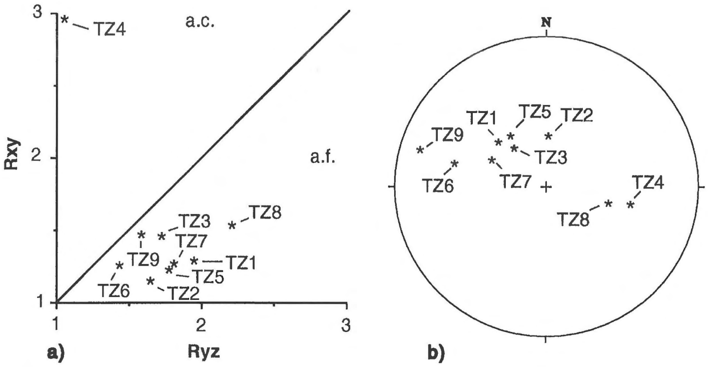

Figure 6. Flinn diagram (a) and the X-axes lower hemisphere stereographic projection (b) of the constructed ellipsoids in the transition zone. (a.c. and a.f.: see Figure 2.)

Discussion and conclusions

The subdivision in the Axial Zone of the Variscan basement of the Pyrenees on the basis of the structural style into two different domains, an infrastructure and a suprastructure, is consistent with the strain variation through a vertical section.

Microstructural analysis in the Canigó-Carançà massif (Capellà 1991) suggests that the main foliation in the infrastructure formed originally with a flat-lying attitude. Carreras & Capellà (1994) attributed the main flat-lying to gently dipping structures in the deep-seated levels to non-coaxial horizontal shear related to crustal shortening. Most authors conclude that N–S directed shearing was associated with the development of the main structures in the metamorphic levels (De Bresser et al. 1986, Kriegsman et al. 1989, Soliva et al. 1989, Van den Eeckhout 1990), in agreement with that proposed by Matte (1986) for the Variscan fold belt of Europe. Both in the Canigó-Carançà massif and the Pallaresa anticlinorium, the main foliation in the infrastructure is overprinted by one and very often two inclined to upright fold systems. In the latter case, the most important one, with ESE–WNW trend, is aligned according to the dome-shaped structures in the infrastructure. The interference of two fold systems postdating the development of the main foliation in the metamorphic rocks is considered to be responsible for the dome-shaped foliation (Guitard 1970, Santanach 1972, Liesa & Carreras 1989, García-Sansegundo & Alonso 1989).

The orientation of the strain ellipsoids and the relationship of the principal X- and Y-axes with regard to fold axes constitute two essential differences between the infrastructure and the suprastructure. Thus, the NE–SW and the more common E–W stretching lineations in the flat-lying foliation planes in the infrastructure are parallel to the fold axes. This variability of the structural trends, through space and time, is interpreted as a consequence of a changing tectonic regime (Carreras & Capellà 1994). The infrastructural foliation was achieved by a deep-seated inhomogeneous horizontal shearing during an earlier compressive event, and the rather steep foliation and folds of rocks in shallow structural levels developed mainly as a result of significant subhorizontal shortening during a late transpressive event.

Calculated strain values in the suprastructure show an apparent flattening strain type (k = 0.36) of low magnitude (R_{xy} = 1.36, R_{yz} = 2.01) which, although possibly a minimum estimate, can be taken as being representative for the entire domain. In contrast to the computed ellipsoids, Corstanje et al. (1989) assessed constrictional strain in the Pallaresa anticlinorium. The samples collected by the Dutch colleagues are considered to belong to the transition zone, where two and three deformation phases can be distinguished (Capellà 1991). Although a similar relationship with regard to the fold and X-axes is provided in both analyses, the multiple deformation history may explain such differences in strain type, even inside the transition zone.

Unlike the suprastructure, apparent flattening strains (k = 0.29) calculated in the infrastructure seem representative for some upper levels in the Canigó-Carançà and Hospitalet gneisses. Apart from the contrast in ductility between different lithologies, these strains seem to reflect the heterogeneous character of the deformation responsible for the main foliation. Shear zones and sheath folds related to the development of the main foliation in the infrastructure in rocks deeper than those analysed in this study, as reported from the eastern Pyrenees (Lagarde 1978, Soliva et al. 1989, Ramírez 1983, J. Carreras pers. comm.), suggest an increasing shear magnitude downwards. Also, since a detachment horizon in relation to the upper front of the main foliation has not been observed, the strain magnitude must progressively decrease upwards.

The deformation associated with the main foliation in the suprastructure is revealed by varying structural styles predominating at different structural levels. This deformation gives rise to a slaty cleavage or spaced cleavage in the suprastructure, and to a still pervasive fabric (crenulation cleavage) in the transition zone. At higher levels in the transition zone, the bulk strain magnitude is approaching that in the suprastructure, as is reflected by the computed ellipsoids in the northern area of the Pallaresa anticlinorium (k = 0.43, R_{xy} = 1.33 and R_{yz} = 1.76). However, in the infrastructure this deformation is taken up by some flexure-type folds and related shear zones, which affect the main foliation (Carreras et al. 1980). These structures are broadly WNW–ESE and steeply dipping, parallel to the major folds in the suprastructure as observed on the geological map.

An overall strain distribution and structural zonation similar to that in the Pyrenean Variscan basement is present in other orogenic belts. In the Variscan fold belt of southwest England and the east-central Cariboo mountains of British Columbia, Sanderson (1979) and Murphy (1987), respectively, have attributed the transition from upright to recumbent folds at depth to a near-horizontal shortening in the upper crust to an increased component of non-coaxial strain downwards, usually related to major thrusts. In the Variscan basement of the Pyrenees, some authors also consider the main foliation in the infrastructure to be the result of compressive events, such as major nappes or thrusts in the Canigó-Carancà massif (Guitard 1970, Soliva et al. 1989) or a mid-crustal decollement in the Garona dome (Matte & Xu Zhi 1988). On the other hand, structure and metamorphism have led Van den Eeckhout (1986) and Van den Eeckhout & Zwart (1988) to relate the flat-lying foliation development to crustal extension. Microstructural analysis carried out across the northern Pallaresa anticlinorium (Capellà 1991) reveals that the infrastructural foliation corresponds, in shallower structural levels, to the pre-main cleavage folds. This interpretation is in contradiction with the view that the infrastructure would be associated with an extensional event.

Although finite strain analysis cannot be used by itself as a criterion to discern the relationship between the infrastructure and the suprastructure, it provides information on the strain pattern of the structural domains. Furthermore, strain analysis may contribute to the determination of the type of tectonic regime when additional criteria are also considered.

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References

Alonso, J.L. 1979 Deformaciones sucesivas en el área comprendida entre Liat y el Puerto de Orlá. Control estructural de los depósitos de sulfuros (Valle de Arán Pirineos centrales) – Tesis de Licenciatura, Univ. Oviedo, 26 pp

Autran, A. & G. Guitard 1969 Mise en évidence de nappes hercyniennes de style pennique dans la série métamorphique du Massif du Roc de France (Pyrénées orientales: liaison avec la Nappe de Canigou) – C.R. Acad. Sci. Paris 269: 2497–2499

Capellà, I. 1991 Variació de l'estil estructural a l'Hercinià del Pirineu. Infrastructura-Zona de Transició-Suprastructura – Tesi Doctoral, Univ. Autónoma de Barcelona, 281 pp

Capellà, I. 1995 El estilo estructural del hercínico en el sector suroccidental del Macizo del Canigó-Carancà (Pirineos orientales) – Rev. Soc. Geol. España 8: 7–20

Carreras, J. & I. Capellà 1994 Tectonic levels in the Paleozoic basement of the Pyrenees: a review and a new interpretation – J. Struct. Geol. 11: 1509–1524

Carreras, J., M. Julivert & P. Santanach 1980 Hercynian mylonite belts in the eastern Pyrenees: an example of shear zones associated with late folding – J. Struct. Geol. 2: 5–9

Casas, J.M. 1984 Estudi de la deformació dels gneissos del Massís del Canigó – Tesi Doctoral, Univ. Barcelona, 284 pp

Corstanje, R., Ch. Klepper, B. Rutgers, I. Van den Wal & B. Van den Eeckhout 1989 Quantification of finite strain in the Pyrenean slate belt: a first assessment using R_f/Φ method – Geol. Mijnbouw 68: 313–321

De Bresser, J.H.P., F.J.M. Majoor & M. Ploegsma 1986 New insights in structural and metamorphic history of the Western Lys-Caillaouas massif (Central Pyrenees, France) – Geol. Mijnbouw 65: 177–187

De Paor, D.G. 1988 Strain from three known stretches – an exact solution – J. Struct. Geol. 10: 639–642
Deramond, J., R.M. Graham, J.R. Hossack, P. Baby & G. Crouzet 1985 Nouveau modèle de la Chaîne des Pyrénées – C.R. Acad. Sci. Paris 301: 1213–1216
De Sitter, L.U. & H.J. Zwart 1960 Tectonic development in supra and infra-structures of a mountain chain. Proc. 21st Int. Geol. Congr. – Copenhagen 18: 248–256
Fry, N. 1979 Random point distributions and strain measurements in rocks – Tectonophysics 60: 80–105
García-Sansegundo, J. 1991 Estratigrafía y estructura de la Zona Axial Pirenaica en la Transversal del Valle de Aran y de la Alta Ribagorca – Tesi Doctoral, Univ. Oviedo, 339 pp
García-Sansegundo, J. & J.L. Alonso 1989 Stratigraphy and structure of the southeastern Garona dome – Geodinamica Acta 3: 127–134
Guitard, G. 1970 Le métamorphisme hercinien mésozonal et les gneiss oeillés du Massif du Canigou (Pyrénées orientales) – Bur. Rech. Géol. Minières Mem. 63, 353 pp
Kriegsman, L.M., D.G.A.M. Aerden, R.J. Bakker, S.W.J. den Brok & P.M.T.M. Schutjens 1989 Variscan tectonometamorphic evolution of the Eastern Lys-Caillaouas massif, Central Pyrenees – evidence for late orogenic extension prior to peak metamorphism – Geol. Mijnbouw 68: 323–333
Lagarde, J.L. 1978 Analyse du sens de déversement d'une nappe gneissique. Aplication à la nappe hercynienne du Canigou (Pyrénées orientales) – C.R. Acad. Sci. Paris 268: 937–940
Lamouroux, C., J.C. Soula, J. Déramond & P. Debat 1980 Shear zones in the granodioritic massifs of the Central Pyrenees and the behaviour of these massifs during the Alpine orogenesis – J. Struct. Geol. 2: 49–53
Liesa, M. & J. Carreras 1989 On the structure and metamorphism of the Roc de Frausa massif (eastern Pyrenees) – Geodinamica Acta 3: 149–161
Lisle, R.J. 1977 Clastic grain shape and orientation in relation to cleavage from the Aberystwyth Grits. Wales – Tectonophysics 39: 381–395
Lisle, R.J. 1985 Geological strain analysis. A manual for the R_f/Φ technique – Pergamon Press, Oxford, 99 pp
Losantos, M., E. Aragonès, X. Beràstegui & C. Puidefàbregues 1989 Mapa Geològic de Catalunya 1 : 250.000. (Servei Geològic de Catalunya). Institut Cartogràfic de Catalunya, Barcelona
Matte, Ph. 1986 Tectonics and plate tectonics model for the Variscan belt of Europe – Tectonophysics 126: 329–374
Matte, Ph. & Q. Xu Zhi 1988 Decollements in slate belts, examples from the European variscides and Qin Ling Belt of Central China – Geol. Rundschau 77: 227–238
Matthews, P.E., R.A.B. Bond & J.J. Van den Berg 1974 An algebraic method of strain analysis using elliptical markers – Tectonophysics 24: 31–67
McCaig, A. 1986 Thick and thin-skinned tectonics in the Pyrenees – Tectonophysics 129: 319–342
Muñoz, J.A. 1985 Estructura Alpina i Herciniana a la vora sud de la Zona Axial del Pirienu Oriental – Tesi doctoral, Univ. Barcelona, 305 pp
Murphy, D.C. 1987 Suprastructure/infrastructure transition, east-central Cariboo Mountains, British Columbia: geometry, kinematics and tectonic implications – J. Struct. Geol. 9: 13–29
Oele, J.A. 1966 The structural history of the Vall Ferrera Area, the transition zone between the Aston Massif and the Salat-Pallaresa anticlinorium (Central Pyrenees, France, Spain) – Leid. Geol. Meded. 38: 129–164
Ramírez, J. 1983 Els gneiss de Port de la Selva: Significació petrològica i relacions amb l'encaixant – Tesi Llicen., Univ. Barcelona, 210 pp
Ramsay, J.G. 1967 Folding and fracturing of rocks – McGraw-Hill, New York, 568 pp
Ramsay, J.G. & M.I. Huber 1983 Techniques of Modern structural Geology, Vol. 1: Strain Analysis – Academic Press, London, 307 pp
Roure, F., P. Choukroune, X. Berastegui, J.A. Muñoz, P. Villien, P. Matheron, M. Bareyt, M. Seguret, P. Cámara & J. Deramond 1989 ECORS Deep Seismic data and balanced cross sections: Geometric constraints on the evolution of the Pyrenees – Tectonics 8: 41–50
Sanderson, D.J. 1979 The transition from upright to recumbent folding in the Variscan fold belt of southwest England: a model based on the kinematics of simple shear – J. Struct. Geol. 1: 171–180
Santanach, P. 1972 Estudio tectónico del Paleozoico inferior del Pirineo entre la Cerdaña i el Río Ter – Tesis Doctoral, Univ. Barcelona, 238 pp
Soliva, J., J.F. Salel & M. Brunel 1989 Shear deformation and emplacement of the gneissic Canigou thrust nappe (Eastern Pyrenees) – Geol. Mijnbouw 68: 357–366
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., C. Lamouroux, P. Viallard, G. Bessière, P. Debat & B. Ferret 1986a The mylonite zones in the Pyrenees and their place in the alpine tectonic evolution – Tectonophysics 129: 115–147
Soula, J.C., P. Debat, J. Deramond & P. Pouget 1986b A dynamic model of the structural evolution of the Hercynian Pyrenees – Tectonophysics 129: 115–147
Van den Eeckhout, B. 1986 A case study of a mantled gneiss antiform, the Hospitalet massif, Pyrenees (Andorre, France) – Geol. Ultraiectina 45: 1–193
Van den Eeckhout, B. 1990 Evidence for large-scale recumbent folding during infrastructure formation in the Pyrenees: the structural geology of part of the eastern Hospitalet massif – Bull. Soc. Géol. France 8: 331–338
Van den Eeckhout, B. & H.J. Zwart 1988 Hercynian crustal scale extensional shear zone in the Pyrenees – Geology 16: 135–138
Verhoef, P.N.W., R.L.M. Vissers & H.J. Zwart 1984 A new interpretation of the structural and metamorphic history of the western Aston massif (Central Pyrenees, France) – Geol. Mijnbouw 63: 399–410
Wickham, S.M. & E.R. Oxburgh 1985 Continental rifts as a setting for regional metamorphism – Nature 318: 330–333
Williams, G.D. & M.W. Fischer 1984 A balanced section across the Pyrenean orogenic belt – Tectonics 3: 773–780
Zwart, H.J. 1963 The structural evolution of the Paleozoic of the Pyrenees – Geol. Rundschau 53: 170–205
Zwart, H.J. 1979 The Geology of the Central Pyrenees – Leid. Geol. Meded. 50: 1–74