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New insights in the structural and metamorphic history of the western Lys-Caillauas massif (Central Pyrenees, France)

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Abstract

We distinguish four deformation phases of major importance in the Cambro-Ordovician metasediments of the western Lys-Caillauas massif. During D1 a steep, regionally dominant, foliation developed. This foliation was folded during D2 into a large overturned isoclinal antiform. The third deformation phase D3 resulted in complex rotations of porphyroblasts, crenulation of the S1 foliation and development of a gently dipping crenulation cleavage S3. At a deep structural level, S1 was completely transposed into the plane of S3, which forms a dome-type D3 structure. D1, D2 and D3 are of Variscan age. D4 included Alpine thrusting along the Gavarnie thrust and accompanying folding and faulting. The Alpine folding led to a considerable amount of ductile shortening in the footwall rocks of the thrust.

The Cambro-Ordovician rocks show a Variscan plurifacial metamorphism during D3. Peak values were 600–660° C and 2.5–3.5 kbar. Two metamorphic domes of slightly different age with respect to D3 are distinguished. One of these metamorphic domes coincides with the D3 structural dome. The Alpine D4 occurred under low grade metamorphic conditions.

In comparison with current notions on the Lys-Caillauas massif the data presented a) reveal a more complex deformation history; b) fit the development of some controversial large scale structures within the deformation history; c) give a more complete picture of the metamorphic evolution of the rocks and d) show that Alpine folding was more widespread than previously thought.

Introduction

The Pyrenees are an Alpine mountain chain with a distinct central or 'axial' zone. This axial zone consists of Paleozoic sediments which have been deformed, metamorphosed and intruded by igneous rocks during the Variscan orogeny (Zwart 1979). Influences of the Alpine orogeny on the Variscan Pyrenees which have recently been reported are large scale thrusting (Parish 1984; Fischer 1984), development of shear zones (Lamouroux et al. 1980), minor folding (Majesté-Menjoulas 1979) and a low grade metamorphic imprint (McCaig 1983).

In recent years detailed remapping and microstructural analyses have led to a better understanding of the structural and metamorphic evolution of the axial zone (e.g. Verhoef et al. 1984; Van den Eeckhout 1984). In this paper results are presented of a study of the western part of the Variscan Lys-Caillauas massif. The location of the studied area is indicated in Fig. 1.

Lithology

The Lys-Caillauas massif consists of a megacrystal granite, situated in metamorphosed Paleozoic sediments. Details on the lithology of the western part of the massif are given by Clin (1959) and Trouiller (1976). The lithostratigraphic sequence is in good agreement with successions in other parts of the massif (cf. Wennekers 1968; André 1985).

The deepest part of the sequence consists of at least 1700 m of Cambro-Ordovician metapelitic rocks with an irregular intercalation of quartzite beds, the number of which decreases upwards. Halfway, there is a matrix-supported metaconglomerate, with quartz pebbles up to 25 cm. It is overlain by a series of lightcoloured quartzitic metapelites, meta-microconglomerates and a ('lower') alternation of marble and quartzite beds. Laterally the quartz pebble metaconglomerate and associated rocks vary strongly in character and thickness; the metaconglomerate and marble-quartzite alternation wedge out, roughly in a south- and a southwestward direction respectively. In the top of the Cambro-Ordovician part of the sequence another ('upper') alternation of marble and quartzite beds occurs.

The Cambro-Ordovician succession is tectonically separated from overlying Silurian black slates and Devonian limestones by the Alpine Gavarnie thrust (Roddaz 1977; Majesté-Menjoulas 1979). This thrust cuts down into the stratigraphy from west to east (see Fig. 1). Locally a fossiliferous Cretaceous (Majesté-Menjoulas 1979) limestone rests with an unconformity on the upper Cambro-Ordovician rocks in the footwall of the thrust.

Metamorphism has transformed the Cambro-Ordovician sediments into slates, phyllites and schists with various assemblages of metamorphic minerals. Isograds intersect the sedimentary layering, although some minerals, e.g. chloritoid, seem restricted to specific lithostratigraphic levels.

Stocks and sills of medium grained muscovite-biotite granite, with some accompanying pegmatite veins, are locally exposed in the la Pez valley. The granite bodies are intrusive in the metasediments; the contacts are sharp.

Deformation

Based on overprinting relations, we distinguished seven deformation phases, four of which are of major importance (see map, Fig. 1 and sections, Fig. 2).

Structures from the earliest part of the deformation history include a cleavage or isoclinally folded bedding. They have rarely been observed.

During D1 (or 'mainphase') an S1 foliation developed, which is the dominant planar structure in most of the area. S1 is formed by parallel arrangement of micas and elongate quartz grains (e.g. Fig. 6b). S1 is axial planar to open to isoclinal folds on all scales, in higher grade rocks often intrafolial. Generally, bedding-S1 relations and D1 folds are SW vergent. Foldaxes plunge gently WNW or ESE, S1 mostly dips steeply SSW or NNE.

A large anticline occurs in the Rioumajou valley (see Fig. 1). Various interpretations exist on the precise nature of this fold. These will be discussed later. Our data indicate that both sedimentary layering and S1 foliation were folded during D2 into a large overturned anticline (section BB′, Fig. 2), with a gently WNW plunging foldaxis. No D2 meso- or microstructures have been proven. However, within the low grade rocks of the structurally higher level of the fold, a crenulation cleavage has locally been observed which might be axial planar to the anticline. Lack of unambiguous overprinting relations prevent definite conclusions.

The hinge of the isoclinal D2 fold has not been found in the eastern part of the Rioumajou valley (section CC′, Fig. 2). However, at the Col de la Pez (Fig. 1) upright tight folds have been observed which appear to fold both bedding and main foliation (cf. Gilman 1980). These folds probably are D2 structures representing the core of the antiform (see section EE′, Fig. 2).

D3 deformation started (D3a) with rotation of porphyroblasts of metamorphic minerals, around a WNW-ESE axis and a sinistral movement looking WNW. The dominant S1 foliation was crenulated subsequently and a (differentiated) crenulation cleavage S3 developed (D3b; Fig 6c + e). Usually, S3 dips gently southward. The intersection lineation with S1 is oriented WNW–ESE and S1–S3

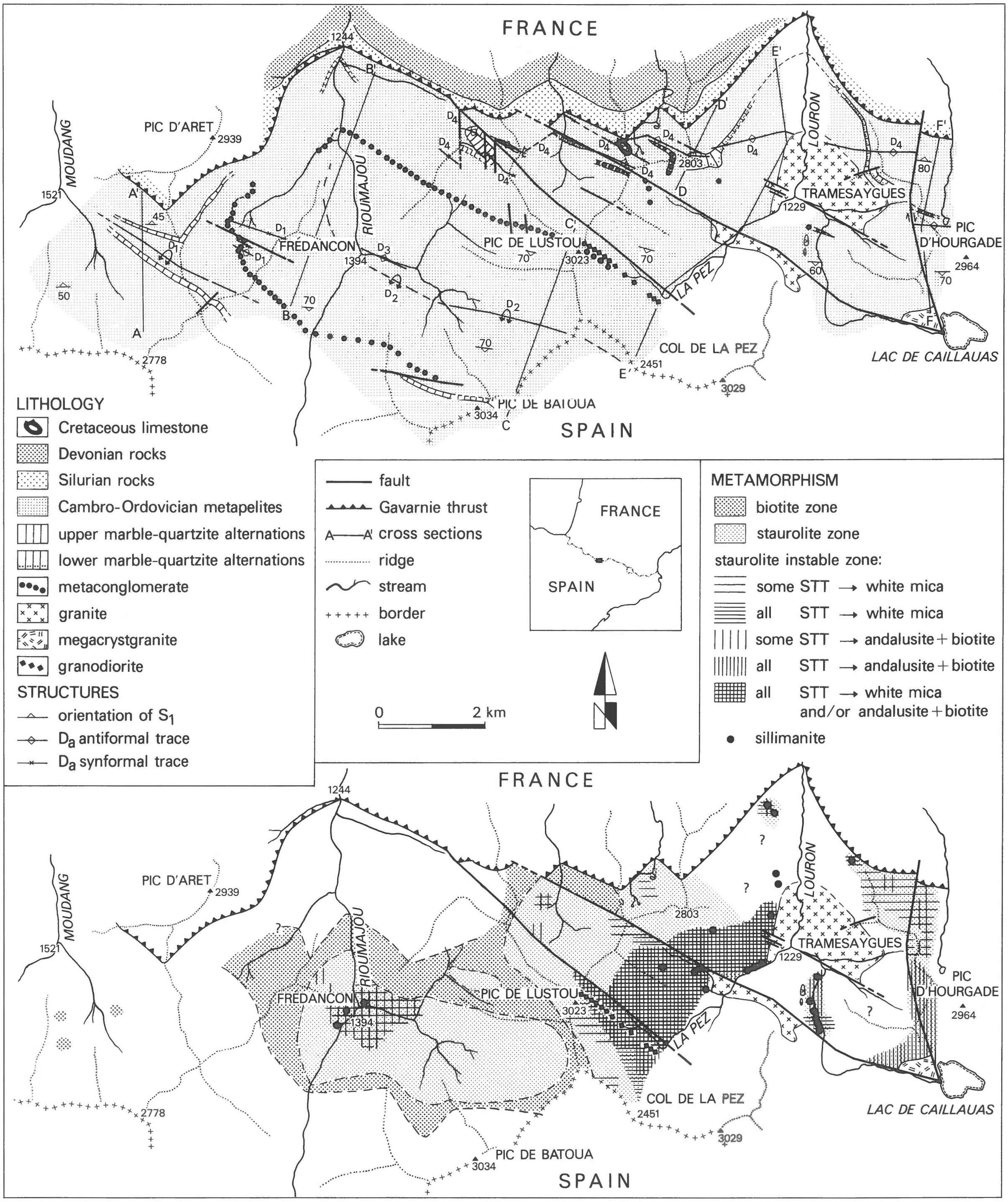


Fig. 1. Geological map of the western Lys-Caillauas massif. a) (top) Structural and lithological features. b) (bottom) Metamorphic zones. STT = staurolite.

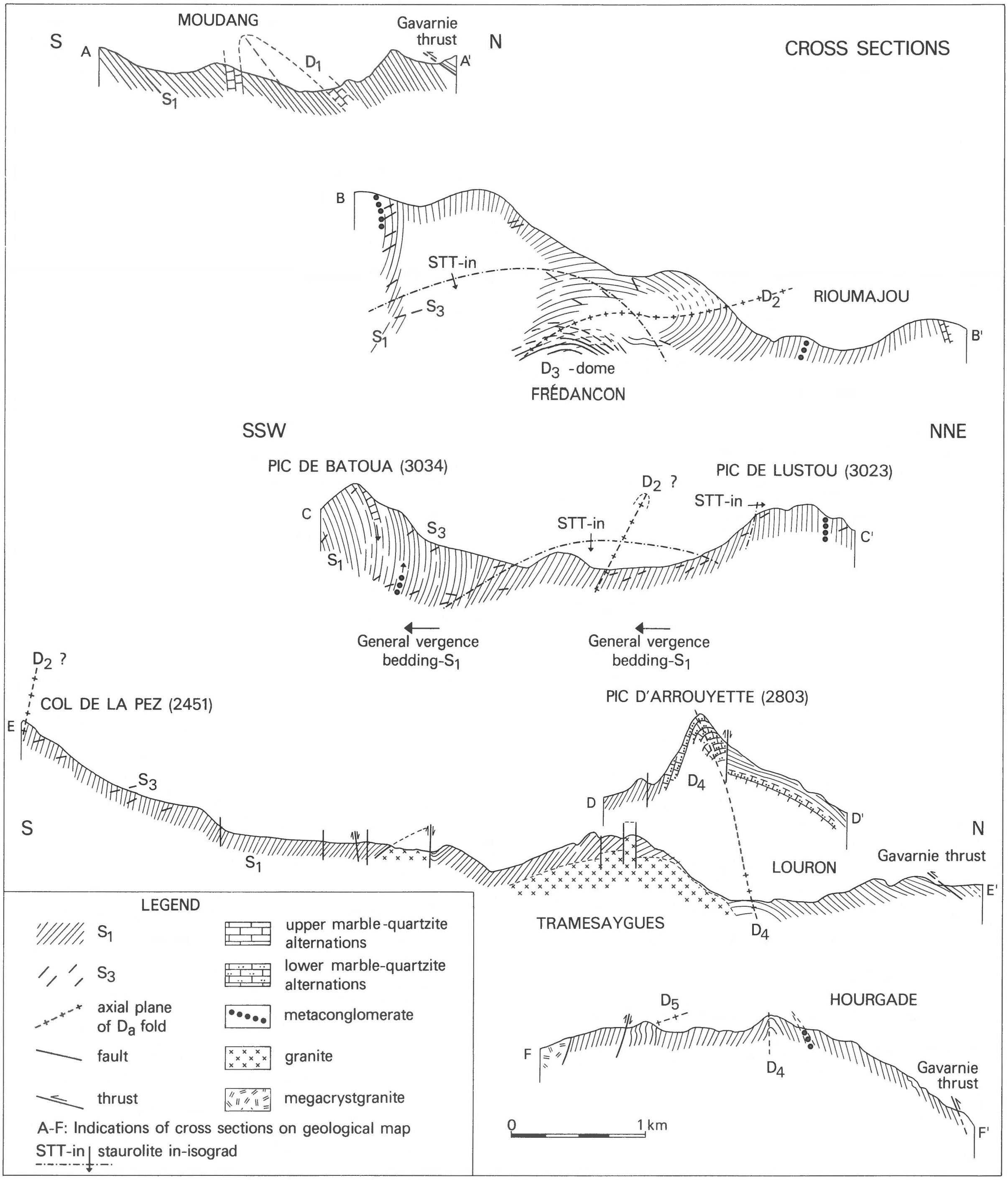


Fig. 2. Cross sections (for locations see Fig. 1). Bedding is generally subparallel to S1.

vergence is in general SSW. Consequently, the sense of asymmetry is similar to that of D3a. This suggests that D3a rotation and D3b crenulation resulted from one progressive phase of south directed flow (top over base).

D3 folds on dm to m scale have been observed locally, S3 being axial planar. The overall change in orientation of S1 from steeply south to steeply north dipping in the structurally higher levels (Fig. 2) can possibly also be attributed to D3 deformation.

At Frédancon, there is a gradual transition from a weakly developed crenulation cleavage S3 to complete transposition of S1 into the plane of S3, forming a new schistosity (see also Zwart 1979). Microstructures show various effects of strong flattening, with a WNW–ESE directed extension lineation which is indicated by alignment of reoriented minerals. S3 planar structures form a dome about 1 km wide and 1.5 km long (see Fig. 2, section BB'). Porphyroblasts witness a second event of rotation, now with a dextral movement looking NNE. The movement parallels the extension lineation. Comparable eastward rotated porphyroblasts have been found locally throughout the area. Because of the gradual character of the transition from D3b deformed rocks into the flattened rocks within the dome, and the parallelism between extension lineation and rotation direction, both transposition and rotation are considered as prolonged D3 deformation (D3c). The transition then marks a change in deformation pattern from south directed flow to strong flattening with east directed flow.

D4 deformation combined thrusting along the Gavarnie thrust with folding and faulting, as will be argued later. The thrust brought Silurian and younger rocks upon the Cambro-Ordovician rocks and their Mesozoic cover. The thrust is marked by extensive development of shear bands within the Silurian slates, which indicate a southward displacement (cf. Simpson & Schmid 1983). Intercalated limestones in the Silurian rocks locally show a mylonitic foliation. D4 folds are upright and tight, with 0–30° WNW plunging foldaxes and roughly E–W striking axial planes. These folds have been found on all scales and they occur mainly in the northern part of the area. Locally an S4 crenulation cleavage is developed which, unlike S3, shows almost no differentation. D4 deformed micas are not polygonized, which is in contrast with the polygonal arcs frequently observed in D3 crenulations. D4 folds affect the Mesozoic limestone and the Gavarnie thrust (Fig. 3). One of these folds is traceable to the east where it shows complexities near the granite body at Tramesaygues and is displaced dextrally along a N–S trending subvertical fault (see Fig. 4 and discusssion).

Post-D4 deformation resulted in asymmetric, often kink-like, folds up to 100 m scale, with gently WNW plunging foldaxes and 10–40° north or south dipping axial planes. A locally differentiated crenulation cleavage is frequently developed.

Two steeply dipping fault systems were found, striking N–S and WNW–ESE respectively. Some of the N–S faults prove to be dextral. At least one of them seems related to D4 (see discussion). Most of the WNW–ESE faults have a downthrown N-block. They displace a D4 fold and the Gavarnie thrust (Fig. 1). The WNW–ESE Caillauas fault (Clin 1959) has a minimum vertical displacement of 300 m and probably a dextral component, as indicated by displaced isograds.

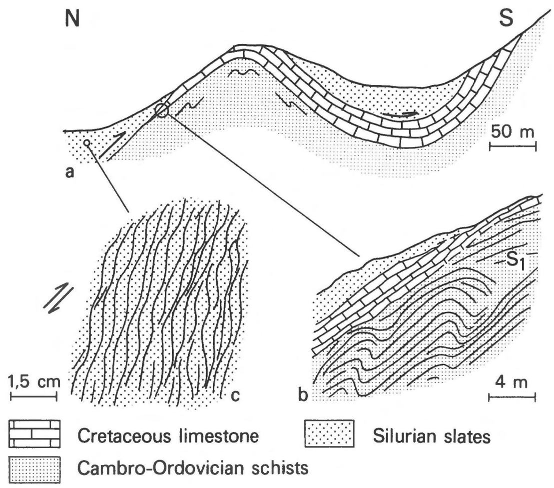


Fig. 3. a,b) D4 folds and c) shearbands near the Gavarnie thrust NNE of the Pic de Lustou (Fig. 1).

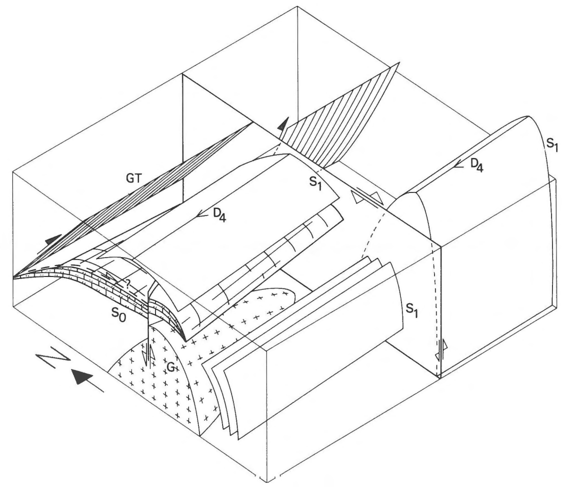


Fig. 4. Simplified diagram of the large scale D4 antiform in the Hourgade area (lithology symbols as in Fig. 1). The S1 foliation shows an open antiform. The lower marble-quartzite alternation does not show the same D4 fold hinge, but is faulted on top of the Tramesaygues granite (G). The upper surface of the alternation may be a décollement horizon. The antiform is displaced about 1200 m by a dextral N-S striking fault. In the eastern block the interlimb angle of the fold is about 30° smaller than in the western block. The Gavarnie thrust (GT) is displaced over a smaller distance and dips more steeply in the eastern block.

Metamorphism and its relation with deformation

The Cambro-Ordovician sediments were metamorphosed during the Variscan event (Zwart 1979). Two spatially separate metamorphic domes could be distinguished: at Frédancon and at Tramesaygues (see Fig. 1). The domes show an increase in metamorphic grade from very low to medium, locally even high (terminology after Winkler 1979). This increase is indicated by a progressive sequence in both space and time of biotite-in, staurolite-in, staurolite-instable and sillimanite-in.

These isograds outline five metamorphic zones in which in metapelites white mica, quartz and the following main metamorphic assemblages are observed:

I chlorite zone: + chlorite ± andalusite ± cordierite ± feldspar ± chloritoid;

II biotite zone: + biotite ± chlorite ± andalusite ± cordierite ± feldspar ± chloritoid ± spessartine-rich garnet;

III staurolite zone: + biotite + staurolite ± andalusite ± cordierite ± feldspar ± chloritoid;

IV staurolite-instable zone: + biotite + andalusite + staurolite-relics ± cordierite ± feldspar ± almandine-rich garnet;

V sillimanite zone: + feldspar + biotite + sillimanite ± andalusite ± cordierite ± staurolite-relics ± almandine-rich garnet.

Relations between development of deformation structures and growth of specific metamorphic minerals for both metamorphic domes are indicated in Fig. 5.

The metapelites outside the biotite-in isograd show chlorite crystals parallel to the S1 foliation. This indicates that D1 took place under low grade metamorphic conditions. In higher grade metapelites biotite crystals contribute to S1 in the Tramesaygues metamorphic dome.

Porphyroblasts of biotite, staurolite, andalusite and cordierite commonly show inclusion trails continuous with the S1 foliation, indicating post-D1 growth. In the Rioumajou valley isograds of biotite-in and staurolite-in cut through the D2 isoclinal fold (section BB′, Fig. 2). This means that porphyroblastesis occured after D2.

Biotite appeared before staurolite because it is often included in staurolite. This is in accordance with the larger spreading of biotite in the area. A

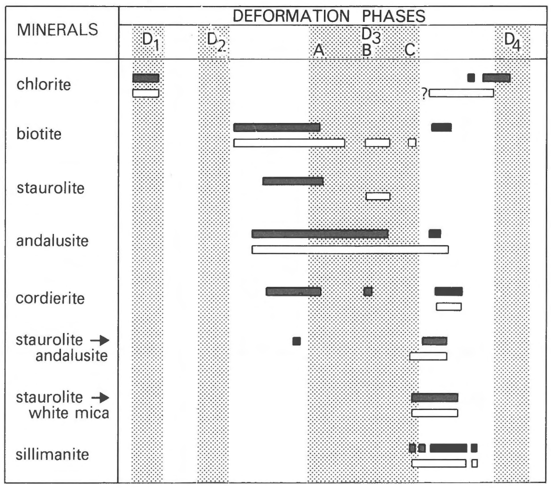


Fig. 5. Relations between deformation structures and growth of metamorphic minerals. Closed bars: Tramesaygues metamorphic dome; open bars: Frédancon metamorphic dome.

reverse inclusion relation has been observed in one sample from the westernmost part of the Tramesaygues staurolite zone. Here, unlike the staurolite, biotite may be interpreted as belonging to the Frédancon metamorphic dome. This implies later growth of biotite around Frédancon than in the Tramesaygues metamorphic dome.

Inclusion patterns in biotite and staurolite indicate different relations with D3 for the two metamorphic domes. In the Tramesaygues metamorphic dome biotites prevail that were grown pre- to synkinematically with D3a. Directly near the Tramesaygues granite bodies, in the sillimanite zone, rocks occur with locally abundant biotite, which probably has grown post-D3. Staurolite commonly has been rotated during D3a, in most cases postcrystalline (Fig. 6b) but in the west of the dome also partly syncrystalline. Throughout the Frédancon metamorphic dome pre- to syn-D3a biotites have been found (Fig. 6a). Within the staurolite zone also many biotites occur which included D3b crenulations before the D3c transposition took place. The staurolite in this zone is exclusively post-D3b (Fig. 6c). The observations indicate that part of the biotite and all staurolite in the Frédancon metamorphic dome are younger with respect to D3 than the common biotite and the staurolite in the Tramesaygues metamorphic dome. This, together with the mutual growth relation of biotite and staurolite, indicates that the metamorphism at Frédancon may have reached it climax somewhat later than at Tramesaygues.

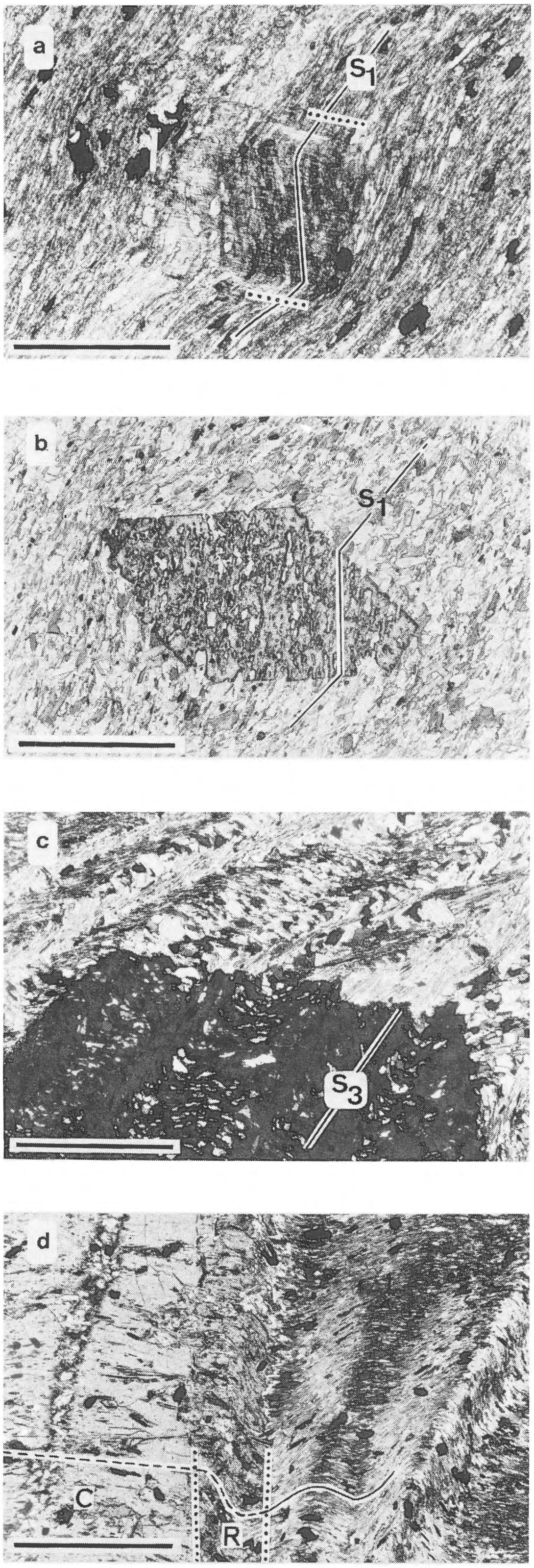
Andalusite is usually abundant. Locally it is present outside, in other places it is absent inside the biotite-in isograd. In the low grade metapelites andalusite grew at the cost of white mica, quartz and either chlorite or biotite. In the higher grade metapelites andalusite may partly or completely replace staurolite (see further below). It follows that it is hardly useful to delineate a single andalusite zone, since andalusites of different origin would be grouped together. The pattern is the more complicated as single andalusite crystals may show different growth stages, especially in some rocks from the area between the two metamorphic domes (Fig. 6d). This multi-stage growth has possibly been caused by the superposition of the Frédancon metamorphic dome on the Tramesaygues metamorphic dome.

Observations on few cordierite crystals indicate growth of these crystals before, during and after the third deformation phase in the la Pez area and only post-D3 around Frédancon (Fig. 5).

In lower grade rocks uncorroded staurolite may be included in andalusite, whereas in the higher grade metapelites it is commonly partly or completely transformed into andalusite together with small anhedric biotite crystals. This prograde reaction (Hoschek 1969) occurred after staurolite rotated during D3a in the Tramesaygues metamorphic dome and after staurolite overgrew D3 crenulations in the Frédancon metamorphic dome. In both cases the original inclusion pattern of the transformed staurolite is usually preserved in andalusite. In the eastern part of the area part of the transformation probably occurred somewhat earlier. Here, unrotated andalusite has included unrotated staurolite whereas staurolite crystals outside the andalusite blasts show D3a rotation. Rotation of the andalusite crystals might have been prohibited by their elongate shape and unfavourable position with respect to the D3a rotation-axis.

In the higher grade metapelites staurolite is also frequently transformed into an aggregate of muscovite with biotite. The distribution of these mica pseudomorphs (see Fig. 1) suggests that the transformation can best be considered a progradation phenomenon as suggested for other areas in the Pyrenees (e.g. Zwart 1979; Verhoef et al. 1984). Results of electron microprobe analyses of the transformations support this assumption (Majoor & De Bresser 1984). Throughout the area the transformation of staurolite to mica occurred predominantly after the third deformation phase (Fig. 6e). In the Rioumajou valley micas have replaced staurolite crystals after their boudinage due to D3c deformation. Few examples in the Hourgade area indicate that D4 folding affected mica pseudomorphs.

Sillimanite has been formed as fibrolite by decomposition of biotite without clear involvement of other reactors or reactants. Timing with respect to deformation phases could only be demonstrated locally: in the east D4 folding affected the mineral



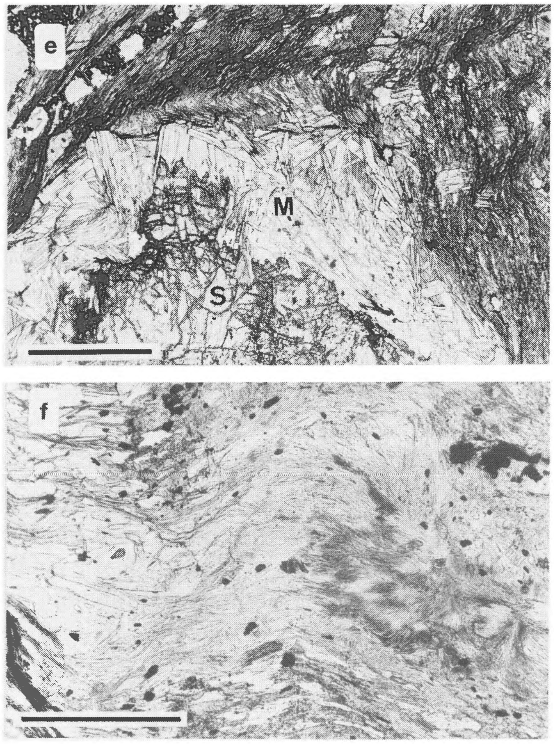


Fig. 6. Microstructures. Scale bars: 0,5 mm except e: 1 mm; c and d crossed nicols. a) Biotite porphyroblast grown pre- to syn-D3a, indicated by the transition from a straight inclusion pattern (S1) in the core to a curved pattern in the rim (Rioumajou valley). b) D3a rotated staurolite (Tramesaygues metamorphic dome). The angle between the internal and external foliation (S1) indicates pre-D3a growth. c) D3b crenulations included in staurolite (Frédancon metamorphic dome). Note the flattening (D3c) of the S3 crenulation cleavage outside the porphyroblast. d) Multi-stage growth of andalusite (pre- and post-D3b) shown by different inclusion patterns of S1 in core (C) and rim (R) of the crystal. The rim is undeformed (sample W of Col de la Pez). e) Coarse grained mica crystals (M) have pseudomorphically transformed a staurolite (S) but, unlike the S1 in the matrix, are unaffected by the D3b microfolding and therefore postdate D3b (la Pez valley). f) D4 deformed aggregate of fibrolitic sillimanite (Hourgade area).

(Fig. 6f): around Frédancon aggregates of fibrolite extending from biotites overgrew a recrystallized quartz fabric that formed during the D3c transposition. Formation of sillimanite post-D3 and pre-D4 is therefore most likely.

The sillimanite occurs predominantly near the granite bodies and can be interpreted as a contact metamorphic mineral. The granite thus probably intruded in the last stages of the Variscan history.

The above observations indicate a single progressive plurifacial metamorphism: the higher grade metamorphic zones developed later with respect to deformation than the lower grade metamorphic zones. The maximum P-T conditions were at 2.5–3.5 kbar and 600–660° C, as concluded from a) sillimanite stability conditions (after Greenwood 1976); b) the transformation from staurolite to andalusite + biotite (Hoschek 1969); and c) the generally stable association of muscovite + quartz within the sillimanite zone (Chatterjee & Froese 1975), except for a small area southwest of the granite body at Tramesaygues where, however, no migmatites have been found. The peak conditions of metamorphism are in good agreement with the results for other areas in the Pyrenees (Zwart 1962; Verhoef et al. 1984).

The metamorphic minerals, except for chlorite, do not include D4 crenulations and may be deformed due to D4. Therefore, the main porphyroblastesis clearly predates the fourth deformation phase (see Fig. 5). Chlorite is locally parallel to the S4 foliation. Some crystals show slightly curved inclusion trails probably indicating syn-D4 growth. This implies that low grade metamorphic conditions reigned during the (Alpine) D4.

In the Siluro-Devonian rocks north of the Alpine Gavarnie thrust distinct porphyroblasts of biotite, andalusite and very small staurolite occur only locally. Silurian slates thrusted upon Cambro-Ordovician medium grade schists confirm the post-metamorphic nature of the thrust.

The D4 affected Cretaceous limestone, unconformably overlying the Cambro-Ordovician schists, contains tiny chlorite-calcite fringes.

Discussion and conclusions

Clin (1959) described a repetition of stratigraphy in the Rioumajou valley (Fig. 1) and correlated it with an open antiform in the schistosity at Frédancon. Trouiller (1976) argued that the fold in bedding is an inclined tight structure to which S1 (our terminology) is axial planar. This fold is refolded at Frédancon into the open antiform. However, throughout the Rioumajou valley we did not observe a systematic change in bedding-S1 vergence, which rules out the possible existence of a large D1 structure.

From microstructural data Zwart (1979) inferred that the flatlying schistosity at Frédancon is a younger foliation superimposed on S1. Outside the Frédancon area this younger foliation is a crenulation cleavage. Combining data of Trouiller (1976) and Groen (1978), Zwart interpreted the Frédancon structure as a fourth phase (F4) antiform. The F4 folds of Zwart (1979) in general can be correlated with our D4 folds. Gilman (1980) showed that the Frédancon antiform does not continue eastward. Consequently, it can not be responsible for the repetition in stratigraphy in the area (cf. sections BB′ and CC′, Fig. 2).

In our interpretation, in which both bedding and S1 are folded, crosscutting of isograds through the antiform (section BB′, Fig. 2) indicates large scale folding before the main metamorphism, hence before D3. The apparent lack of the D2 fold hinge in the eastern part and of small scale D2 structures throughout the Rioumajou valley is possibly due to obliteration during D3. Within the Central Pyrenees no comparable pre- main-metamorphic D2 folds have as yet been described.

The small dome at Frédancon is superimposed on the D2 isoclinal fold. There is, however, no reason to consider the Frédancon antiform as a D4 structure, as its geometry is not comparable with known D4 folds, and no axial plane foliation S4 has been found. The spatial coincidence of dome and D3 structures indicates a mutual relation. Moreover, the structural dome matches the central part of the Frédancon metamorphic dome. Metamorphism took place during D3, and granite bodies probably intruded during a late stage of D3 or shortly thereafter. An intrusion of a granite body below Frédancon led possibly to the development of the structural and coincident metamorphic dome. It is, however, unclear why such a combined structural and metamorphic dome does not occur around the granite body at Tramesaygues.

The transition in the Rioumajou valley from a steep dominant S1 to a shallower dipping younger dominant foliation, together with increasing metamorphic grade, is strikingly comparable with the transition from low grade suprastructure to high grade infrastructure elsewhere in the Pyrenees, for example in the western Aston massif (Verhoef et al., 1984) and the Hospitalet massif (Van den Eeckhout 1984).

D4 folding affected the Gavarnie thrust and the Mesozoic limestone, implying an Alpine age for D4. Moreover, the restriction of D4 folds to the vicinity of the thrust suggests a causal relation between thrusting and folding. A similar relation exists between the thrust, the large scale antiform and the dextral fault in the eastern part of the area (Fig. 4). The thrust and the fold show a different amount of displacement along the fault and a change in orientation and shape respectively on either side of the fault. This suggests that all three structures were actively being formed at the same time. We propose that the thrusting caused the folding. Whereas the lower thick marble-quartzite alternation is tightly folded in the la Pez area (section DD′, Fig. 2), the presence of the Tramesaygues granite locally hampered folding of this unit in the Hourgade area (Fig. 4). Deformation has led to faulting, possibly décollement along the upper surface of the marble-quartzite alternation and induced the N-S dextral fault allowing the fold to become tight east of the granite.

In the Gavarnie area to the west, Parish (1984) described basement thrusts which deform the Gavarnie nappe, leading to monoclinal folds. He interpreted the structures in terms of a piggy-back thrusting sequence. In the present area no thrust structures nor duplications of stratigraphy have been found in the Cambro-Ordovician footwall rocks and their Mesozoic cover. The observed tightening of the major D4 antiform east of the N-S fault and the general tight geometry of D4 folds throughout the area cannot be attributed to the existence of deeper level thrusts, but must be due to a considerable amount of ductile shortening within the footwall rocks of the Gavarnie thrust. This ductile behaviour is in agreement with the presence of limestone mylonites in the thrust zone and the supposed low grade metamorphism associated with D4.

To the east, the large scale antiform can be traced towards the Lac d'Oo antiform, proposed to be Variscan by Zwart (1979). It means that the Lac d'Oo antiform is Alpine, as well. This may indicate that large scale Alpine deformation is more widespread in the Lys-Caillauas massif than previously thought.

In conclusion detailed field and thin-section study revealed important aspects of the structural and metamorphic history of the western Lys-Caillauas massif: a) large scale folding affected both sedimentary layering and the dominant foliation before the peak of the Variscan metamorphism; b) the Variscan structural-metamorphic evolution of the massif comprised a transition from low grade rocks with a steeply dipping foliation (suprastructure) into higher grade rocks with a flat lying foliation (infrastructure); c) considerable ductile shortening occurred in the footwall rocks of the Alpine Gavarnie thrust.

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