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Development of high thermal gradients by coeval transpression and magmatism during the Variscan orogeny: insights from the Cap de Creus (Eastern Pyrenees)

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Abstract

The Cap de Creus massif in the Eastern Pyrenees is a fragment of the Variscan mid-crust. Geological mapping and meso- and micro-scale analysis of the relationships between fold structures, metamorphic mineral assemblages and fabrics affecting igneous bodies suggest the interaction of three processes: deformation, metamorphism and magmatism. Low pressure prograde metamorphism grades laterally from greenschist to amphibolite facies over a distance of ~5 km. Zones of intense metamorphism coincide with transpressional deformation and with the intrusion of small sheets of diorite and granitoid. Sillimanite schists and small migmatite pods are distributed around the small intrusions of basic–intermediate composition. The inferred maximum thermal gradient is >80°C/km. This thermal pattern is similar to patterns observed in other Variscan massifs in the Pyrenees, where high metamorphic gradients are interpreted in different ways. In the NE Cap de Creus, the type and distribution of prograde metamorphism was caused, or at least enhanced, by the intrusion of diorites and granitoids. Alternatively, deformation developed at prograde conditions and synchronously with magmatism, displays a highly heterogeneous distribution, producing a sub-vertical zone of high strain that coincides with domains of highest grade metamorphism and igneous activity. This zone of high strain, developed in a broad transpressive regime, could have favoured the injection of magmatic sheets from deep to mid-crustal levels. Progressive folding from prograde to retrograde metamorphic conditions might account for the narrowing of metamorphic zones.

Keywords: Variscan structure; Pyrenees; LP/HT metamorphism; thermal gradients; magmatism

1. Introduction

The recurrent association of deformation, magmatism and metamorphism in mid to deep crustal levels in orogenic belts is well known. However, only over the last two decades has such an association been based on the detailed analysis of structures, which record an often complex evolution (e.g. Reavy, 1989; Gray, 1997). An approach based exclusively on the distinction of different tectonic phases and interkinematic periods is often inadequate, because progressive deformation is involved. Moreover, magmatic and metamorphic episodes appear to be strongly controlled by the overall deformation rather than being independent events with a sole spatial coincidence. In the study of processes in orogenic belts, multidisciplinary approaches are required and data is often presented in terms of three binary relationships: deformation–plutonism, plutonism–metamorphism and metamorphism–deformation (Karlstrom and Williams, 1995).

In this context, interpretation of the Variscan basement of the Pyrenees is contentious. Despite a huge accumulation of geological data, diverse

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interpretations have arisen, some of them highly contradictory. Most of these discrepancies derive from different interpretations of the geodynamic evolution, mainly in areas where mid to deep crustal levels are exposed. Moreover, precise dating of the age of emplacement of igneous rocks and of metamorphism remains to be completed.

Variscan metamorphism in the Pyrenees is a classic example of LP/HT metamorphism. The causes for the development of high thermal gradients related to this LP/HT metamorphism are challenging and still under discussion (Guitard et al., 1996). This question has been variably interpreted as related to a combination of: (i) thermal processes, such as the control effect of gneissic domes on the metamorphic distribution and the influence of granitoid intrusions in the development of the so called plutonometamorphism (Zwart, 1962; Soula, 1970; Liesa, 1994); and (ii) extensional tectonics-dominated geodynamic processes (van den Eeckhout, 1986; Wickham and Oxburgh, 1986, 1987; Vissers, 1992).

Another controversy concerning the Variscan of the Pyrenees is the identification of the geotectonic setting controlling the evolution of this crustal domain. A variety of very different models has been put forward depending on the deformation phase considered most important. These different models may involve: (i) crustal contraction without significant extension, (ii) early crustal extension followed by crustal contraction or (iii) early crustal contraction followed by crustal extension (see Carreras and Capellà, 1994 for review). Recent structural studies mainly conducted in granitoids have led to the recognition of the syntectonic character of most magmatic rocks and to their emplacement in a broadly transpressive regime (Gleizes et al., 1991; Leblanc et al., 1996; Evans et al., 1997; Druguet, 1997; Gleizes et al., 1998; Druguet and Hutton, 1998). Similar contractional and/or transpressional orogenic settings involving coeval magmatism and/or metamorphism have been invoked for other Variscan massifs (e.g. Melka et al., 1992; Matte et al., 1998) and for other orogens (e.g. McCaffrey, 1992; D'Lemos et al., 1992; Ingram and Hutton, 1994; Karlstrom and Williams, 1995; Davidson et al., 1996; Tikoff and Saint Blanquat, 1997; Brown and Solar, 1998a,b).

The main emphasis of this paper lies on the tectonothermal evolution of the Cap de Creus massif. In this easternmost outcrop of the Variscan basement in the Pyrenees, rocks are well exposed. A detailed knowledge of this mid-crustal domain will not only help in understanding the evolution of the Variscan basement of the Pyrenees, but also will provide constraints for geodynamics in orogenic belts. A likely explanation for the development of high thermal gradients is considered in light of coeval magmatism and transpression inferred from the observed relationships between structural and petrological features in the Cap de Creus.

2. Tectonothermal patterns in the Variscan basement of the Pyrenees

In the NE of the Iberian Peninsula, the Variscan basement outcrops along the Pyrenean Axial Zone and along the Catalonian Coastal Ranges (Fig. 1). In spite of the lack of continuity, these different massifs build up a section that belongs to the southern branch of the Variscan belt. However, no clear correlation with established zones in the Iberian Massif (Julivert et al., 1972) has been established. This segment of the Variscan belt (Iberian block) is separated from the North Pyrenean Massifs and the further north Monthoumet, Montagne Noire and French Massif Central (European plate) by the North Pyrenean Fault, an Eoalpine sinistral transtensional fault related to rotation of Iberia with respect to the European plate, mainly produced during the Cretaceous (Le Pichon and Sibuet, 1971; Choukroune and Mattauer, 1978). After the North Pyrenean Fault and the effect of Alpine contraction are restored, the present Pyrenean domain appears in the Late Palaeozoic as a segment of Variscan crust (Fig. 1). Within this piece of the Variscan, a NW–SE trending zonation is recognized from external shallow-seated zones in the southwest (corresponding to the present day Iberian Chain, Catalonian Coastal Ranges, Western and Central Pyrenees) to more internal and deep-seated rocks north-eastwards (the present day eastern Axial Zone of the Pyrenees and the North Pyrenean Massifs).

A widespread feature in different massifs of this more internal zone is the distribution of low pressure prograde metamorphic zones (Fig. 2), with large areas of low or very low grade metasediments bounding amphibolite to granulite facies metamorphic and

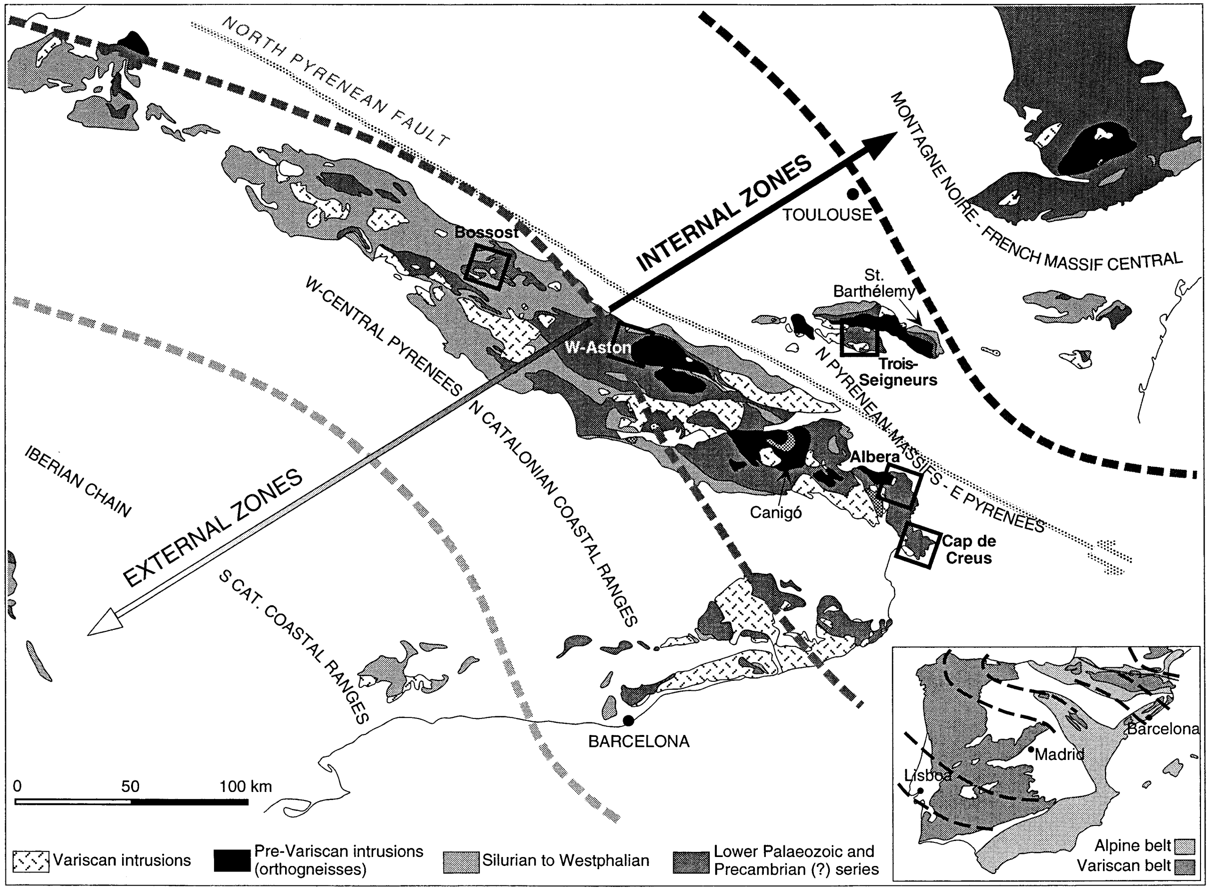


Fig. 1. Sketch of the Variscan zonation in NE Iberia and S France approximately at the end of the Palaeozoic. Restorations of the North Pyrenean Fault and the Alpine contractional effect are based on models in Barnolas and Chiron (1996).

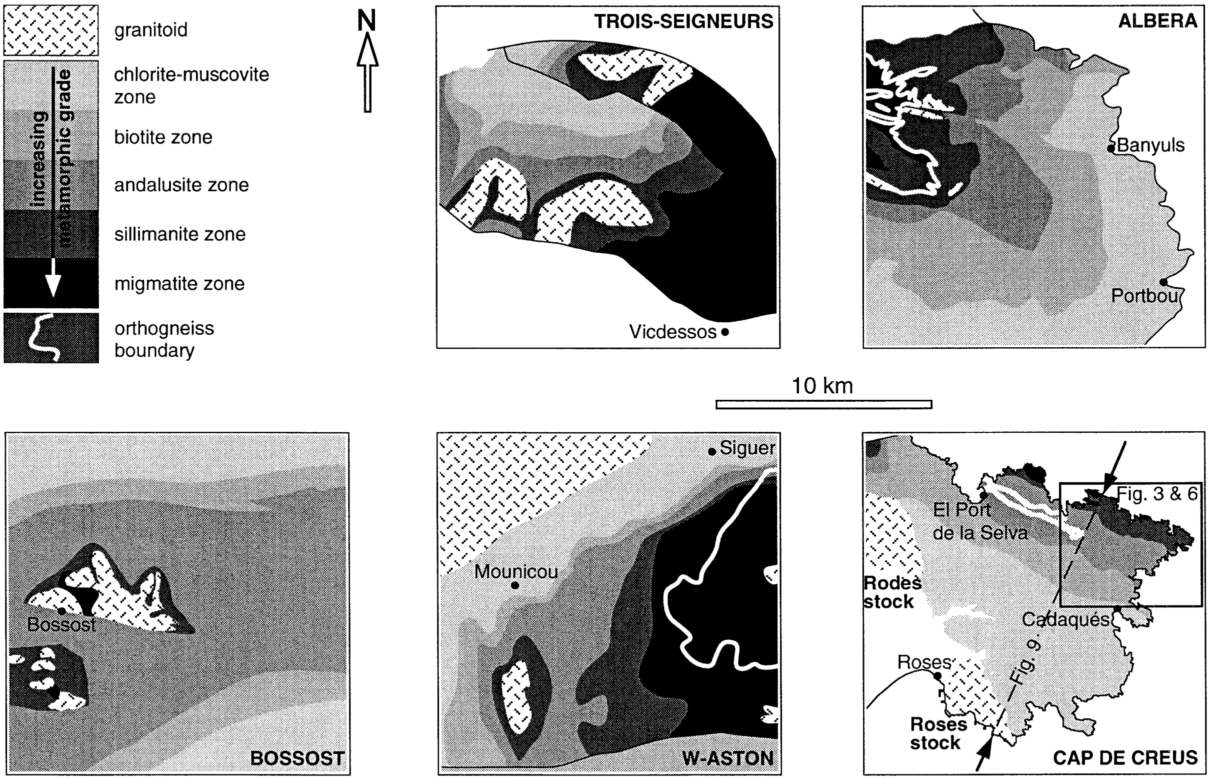
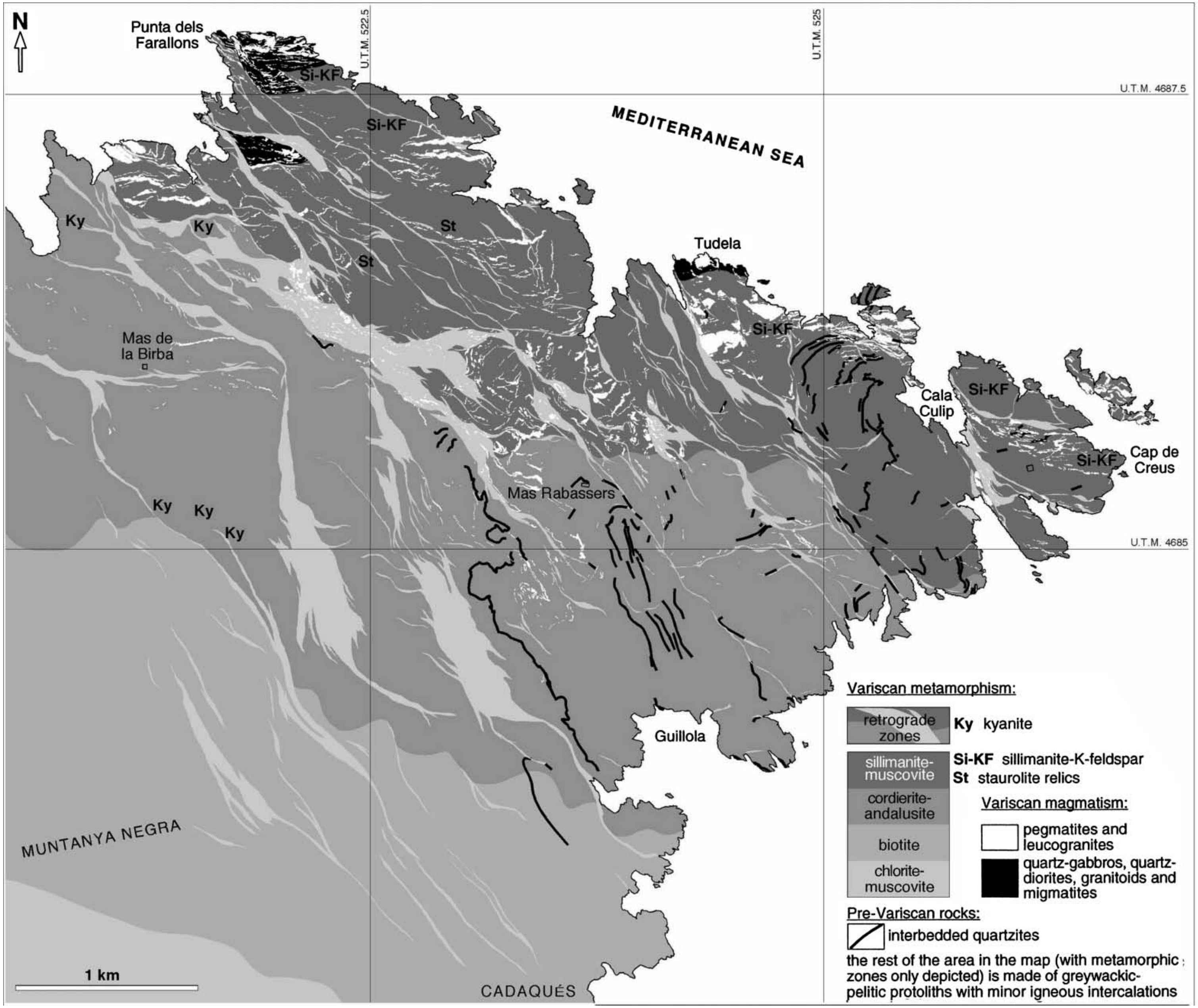


Fig. 2. Metamorphic gradients in some Variscan massifs in the Pyrenees (location shown in Fig. 1), based on original drafts in Barnolas and Chiron (1996).

migmatitic cores. Zones of highest metamorphic grade recorded pressures <8 kbar (Vielzeuf et al., 1990). Most of these zones are concentric and dome-shaped, and therefore called thermal domes. The following metamorphic zones can be identified in most massifs: chlorite–muscovite, biotite, cordierite–andalusite, sillimanite–muscovite, sillimanite–K-feldspar and migmatite (Guitard et al., 1996). Thus, a high thermal gradient metamorphism is found in the interval between the beginning of the amphibolite facies until the beginning of anatexis, i.e. at middle crustal depths, inferred from the present day distribution of isograds. Geothermal gradients are not easy to constrain because the thermal structure may has been modified by late tectonics and because of the difficulty to establish the spatial attitude of isograds. However, an average metamorphic gradient for the five massifs in Fig. 2 is ~75°C/km (qualitatively estimated from data in Barnolas and Chiron, 1996), which is close to the 70 and 65°C/km assessed by Zwart (1986) and Guitard et al. (1996), respectively for the entire Variscan massifs in the Pyrenees. The Trois-Seigneurs massif is one of these massifs with a higher thermal gradient (over 100°C/km, Wickham and Oxburgh, 1986). The most moderate values obtained are of about 50°C/km (for instance, the Albera massif). The Cap de Creus study case depicts an average thermal gradient >80°C/km.

An assessment of the context for the described metamorphic areas is essential for unravelling the causes for the development of a LP/HT metamorphism with high thermal gradients. In fact, most workers in the Pyrenees use models based on the spatial occurrence of metamorphism in a specific massif. Metamorphic massifs

Fig. 3. Metamorphic zonation and magmatism in the NE Cap de Creus peninsula (location shown in Fig. 2).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Variscan metamorphism:
retrograde
zones
Ky kyanite
sillimanite-
muscovite
Si-KF sillimanite-K-feldspar
St staurolite relics
cordierite-
andalusite
Variscan magmatism:
biotite
pegmatites and
leucogranites
chlorite-
muscovite
quartz-gabbros, quartz-
diorites, granitoids and
migmatites
Pre-Variscan rocks:
interbedded quartzites
the rest of the area in the map (with metamorphic
zones only depicted) is made of greywackic-
pelitic protoliths with minor igneous intercalations

in the Pyrenees may be the response to a complex spectrum of situations that prevent the establishment of a unique integral pattern for the space distribution of metamorphism. However, it is likely that there is some common first-order cause for the development of the widespread high metamorphic gradient.

The Variscan structure in the Pyrenees is due to the superposition of structures arising from polyphase tectonics that resulted in a dominant WNW–ESE structural trend, with the presence of large domes or broad antiforms with a flat-lying dominant foliation. These domes, which often include an orthogneiss core, are bounded by domains of folds affecting metasediments with steep to moderate dipping foliations. Many workers presently agree that the main Variscan phase in the Pyrenees was contractional. This is based on the prevalence in the metasediments of upright, isoclinal tight folds with a well developed axial planar schistosity (Matte and Mattauer, 1987; Carreras and Capellà, 1994; Carreras et al., 1996).

Medium to high grade metamorphism is often developed around structural domes with an orthogneissic core, as in the Aston, Saint Barthélemy, Canigó and Albera massifs (Figs. 1 and 2, see Guitard et al., 1996 for review). It was suggested that these gneissic cores acted as heat channels which caused and controlled the metamorphic isograds. This controlling effect of gneissic domes on the metamorphic distribution was named 'effet de socle' (Guitard, 1960; Fonteilles and Guitard, 1977; basement effect of Ayerton, 1980). However, this case appears as a common end-member in a complex mosaic of situations. In some other metamorphic massifs, an orthogneissic core does not outcrop. This is the case in the Bossost area (Zwart, 1962; Pouget et al., 1988; García-Sansegundo, 1996), in the Trois-Seigneurs massif (Allaart, 1959; Wickham, 1987; Mercier, 1996) and also in the Cap de Creus.

In the Cap de Creus example (Fig. 2), medium to high grade deeper stratigraphic rocks are located on the northern areas and low grade, upper stratigraphic sequences and granodiorites on the south, reflecting in analogy to other areas of the Variscan basement, the spatial dissociation between high grade metamorphic domains and granitoid batholiths and stocks. However, in many cases, metamorphic and migmatitic formations are usually associated with small funnel shaped intrusions of basic, mantle-derived, magmas emplaced in mid-crustal levels (Driouch et al., 1989). The Trois-Seigneurs, Western Aston and Cap de Creus massifs are among these occurrences. This observation has guided some authors to the belief that high thermal gradients were generated by heat transfer from the mantle into the crust in the form of basic intrusions (Soula et al., 1986; Wickham and Oxburgh, 1987), which also contributed to development of granulite facies and migmatites (Vielzeuf et al., 1990). Furthermore, Zwart (1968) suggested a genetic link between deeper quartz–diorites and tonalites and shallower tonalite–granodiorite batholiths.

Extensional tectonics was also invoked to explain LP/HT metamorphism in the Pyrenees. According to Wickham and Oxburgh (1986), crustal extension in a rifted tectonic setting, accompanied by intrusion of mantle-derived magmas, caused the extremely high thermal gradients. However, such a continental extensional model does not apply at the Cap de Creus massif, where, as will be shown, the low-P metamorphism is coeval with progressive crustal thickening. Moreover, this model also conflicts with the growing belief that main Variscan deformation in the Pyrenees was transpressional (Carreras and Capellà, 1994; Gleizes et al., 1998).

3. Metamorphism and magmatism in the NE Cap de Creus

The distribution of metamorphic and magmatic rocks in the NE mid-crustal domains of the Cap de Creus peninsula is shown in Fig. 3. Metamorphism affects all the pre-Variscan lithologies (late Pre-Cambrian and lower Palaeozoic) in the area, which mainly correspond to a metasedimentary sequence (metagreywackes, metapelites and few quartzite beds) with minor interlayered meta-igneous rocks (metabasites and orthogneisses). Different zones of metamorphism have been interpreted on the basis of changes in the mineral assemblages of pelitic metasediments. The general spatial pattern for the metamorphism does not show significant differences in regard to that proposed by Carreras (1973), but incorporates more precisely traced isograds as a result of compiling old and new metamorphic data.

As a first approximation, four observable features characterize this spatial distribution: (i) the presence

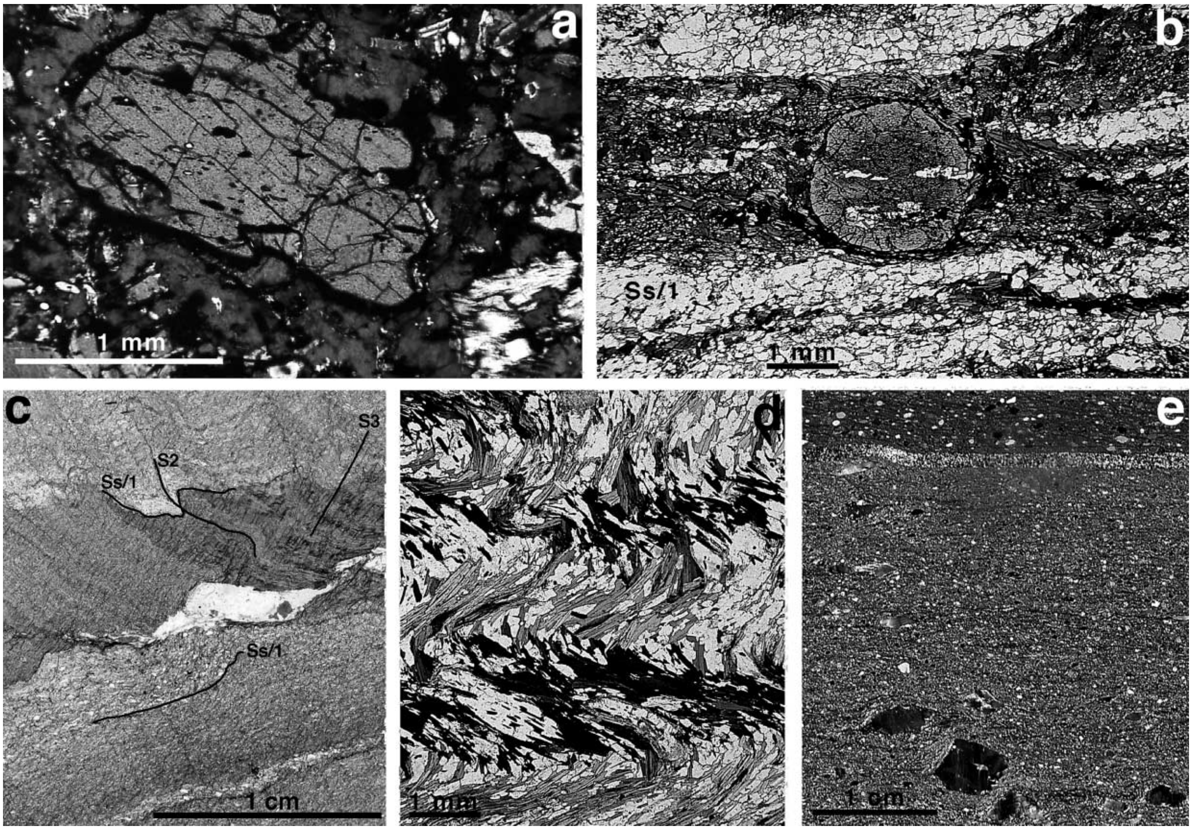


Fig. 4. Microphotographs of metamorphic rocks. (a) Staurolite crystal preserved in a poikiloblast of cordierite, in cordierite–sillimanite schist. Crossed polarised light. (b) Almandine garnet grown over Ss/1 in psammitic schist from the sillimanite–muscovite zone. Garnet shows inclusion trails of quartz and biotite parallel to Ss/1, and a weak zonation (prograde). Plane polarised light. (c) Overprinting foliations in a ritmite from the biotite zone. S2 crenulation cleavage overprints Ss/1 and is most marked in metapelitic layer (centre of the photograph) where it forms a tectonic banding defined by the alignment of early and syn-D2 biotite. Late retrograding S3 crenulation cleavage is also selectively developed in metapelite. Plane polarised light. (d) High grade sillimanite–K-feldspar micaschist showing biotite crystals that grew at least partially after crenulation related to the E–W folding stage in the north. Folded surfaces correspond to S2 foliation. Plane polarised light. (e) Mylonitic schist derived from a high grade metasediment. The mylonitic foliation (S3) is defined by alternating quartz ribbons and philosilicate-rich layers (chlorite, muscovite ± biotite). Mica fish and porphyroclasts of plagioclase are also displayed. Crossed polarised light.

of a metamorphic gradient which increases from SW to NE; (ii) a high angle cross-cutting relationship between WNW–ESE trending isograde surfaces and pre-Variscan lithologies; (iii) a sub-parallelism between the isograds, the pegmatite dyke swarm and the migmatite complexes, with progressively higher grade metamorphic zones northwards, coinciding with the appearance of voluminous pegmatite dykes; and (iv) the described pattern appears heterogeneously cross-cut and overprinted by narrow zones of low grade metamorphism, related to late folding and mylonitization. Assuming that the zones of prograde metamorphism display a similar spatial attitude as the pegmatite dykes and the intrusions of the migmatite complexes, isograds may have steeply north-dipping boundaries. In consequence, the outcrop surface cuts metamorphic zones obliquely or at a high angle. Whereas prevalent foliations and isograds display steep monoclinal dispositions in the study area, to the south, large-scale late stage folding resulted in a more complex pattern of folded, dome-shaped isograds (Carreras et al., 1980, Fig. 10).

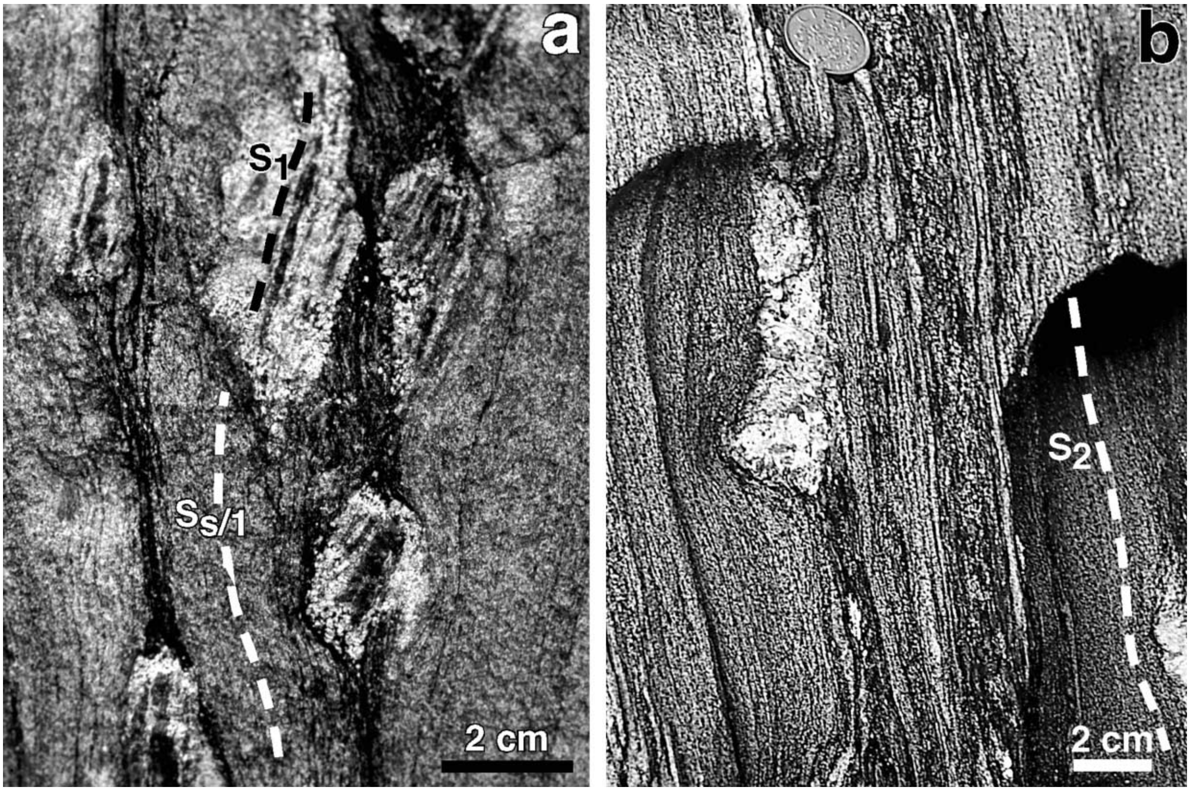


Fig. 5. Field photographs (cross-section views). (a) Partial sillimanite pseudomorphs of andalusite porphyroblasts from the sillimanite–muscovite zone. Sillimanite forms reaction rims along the boundaries of andalusite. S1-related inclusion trails display an oblique-Si geometry. This suggests that andalusite is post-tectonic with respect to S1 but pre-tectonic with respect to S2, and that overgrowths of sillimanite are syntectonic with respect to relative rotation during S2. (b) Migmatite schist with lens-shaped granitic leucosomes parallel to S2 transposition foliation. The high temperature fabric is also characterized by the presence of biotite, sillimanite and K-feldspar in the schist.

3.1. Metamorphism

The chlorite–muscovite zone outcrops only in the southwestern part of the mapped area (Fig. 3), although it continues southwards (Fig. 2), where all rocks show a low or very low metamorphic grade. The biotite zone extends over 1.5–2 km (Fig. 3). The southern margin of this zone represents, more or less, the boundary in which phyllites grade into micaschists with the presence of macroscopically visible biotite. Small garnets appear sporadically in psammitic layers, and chloritoid in dark pelites.

North of the biotite zone, several generations of biotite and muscovite form through the whole metamorphic area. The cordierite–andalusite zone occupies a band 1.5–2 km wide. The passage to this zone is characterized by a progressive increase in grain size and the sudden appearance of cordierite porphyroblasts (appearing as poikiloblasts with many inclusions) subsequently accompanied by andalusite crystals. The typical mineral assemblages in this zone are biotite–cordierite, biotite–cordierite–andalusite, biotite–andalusite and biotite–garnet. Staurolite is scarce and, when present, it appears as relics inside andalusite crystals. *P–T* conditions estimated by Reche et al. (1996) from a petrogenetic grid yielded 600°C and 4.3 kbar for the upper part of the cordierite–andalusite zone.

Towards the sillimanite–muscovite zone, a gradual increase in grain size is accompanied by sillimanite growth, often as fibrolite growing epitaxially on biotite. Porphyroblasts of almandine garnet with inclusions of biotite and quartz appear sporadically in psammitic layers (Fig. 4b) and, locally, relics of staurolite in cordierite have also been found (Fig. 4a). Porphyroblasts of andalusite coexist with sillimanite (Fig. 5a) at least in the southern margins of the zone, in association with cordierite. In the westerly domains, a complex mylonite band related to late shear zones separates the cordierite–andalusite and the sillimanite–muscovite zones. The achieved $P–T$ conditions for the sillimanite zone vary from 670°C and 4.7 kbar (values corresponding to the core of garnets with retrograde zoning) to 560°C and 2.4 kbar (obtained from garnet rims). The estimation of both temperature and pressure has been made using thermobarometric calculations developed by Reche and Martínez (1996). These data agree with previous estimations made by Reche et al. (1996) using petrogenetic grids.

The presence of the association sillimanite–K-feldspar is evidenced by the appearance of conspicuous clusters made of fibrolite ± quartz ± K-feldspar (microcline) in the easternmost area and in the margins of the migmatite complexes. The sillimanite–K-feldspar isograde is difficult to determine mainly due to the common occurrence of coarse flakes of secondary muscovite, especially related to areas with extensive intrusion of pegmatite dykes. As a result of these observations, small areas or local spots within the sillimanite–muscovite zone have been mapped, instead of a distinctive metamorphic zone. In a similar way, zones of incipient migmatization have been observed within the sillimanite zone (Fig. 5b). However, the effective partial melting of the metasediments with development of migmatite schists took place in the areas where intrusive granitoids and quartz–diorites had been emplaced. The spatial association of granitoids and migmatites form the so called migmatite complexes.

The prograde metamorphic assemblages reflect the effects of a low pressure regional metamorphism affecting all the pre-Variscan lithologies in the area. Over a horizontal distance of approximately 5 km, the pelitic metasediments show an average metamorphic gradient >80°C/km.

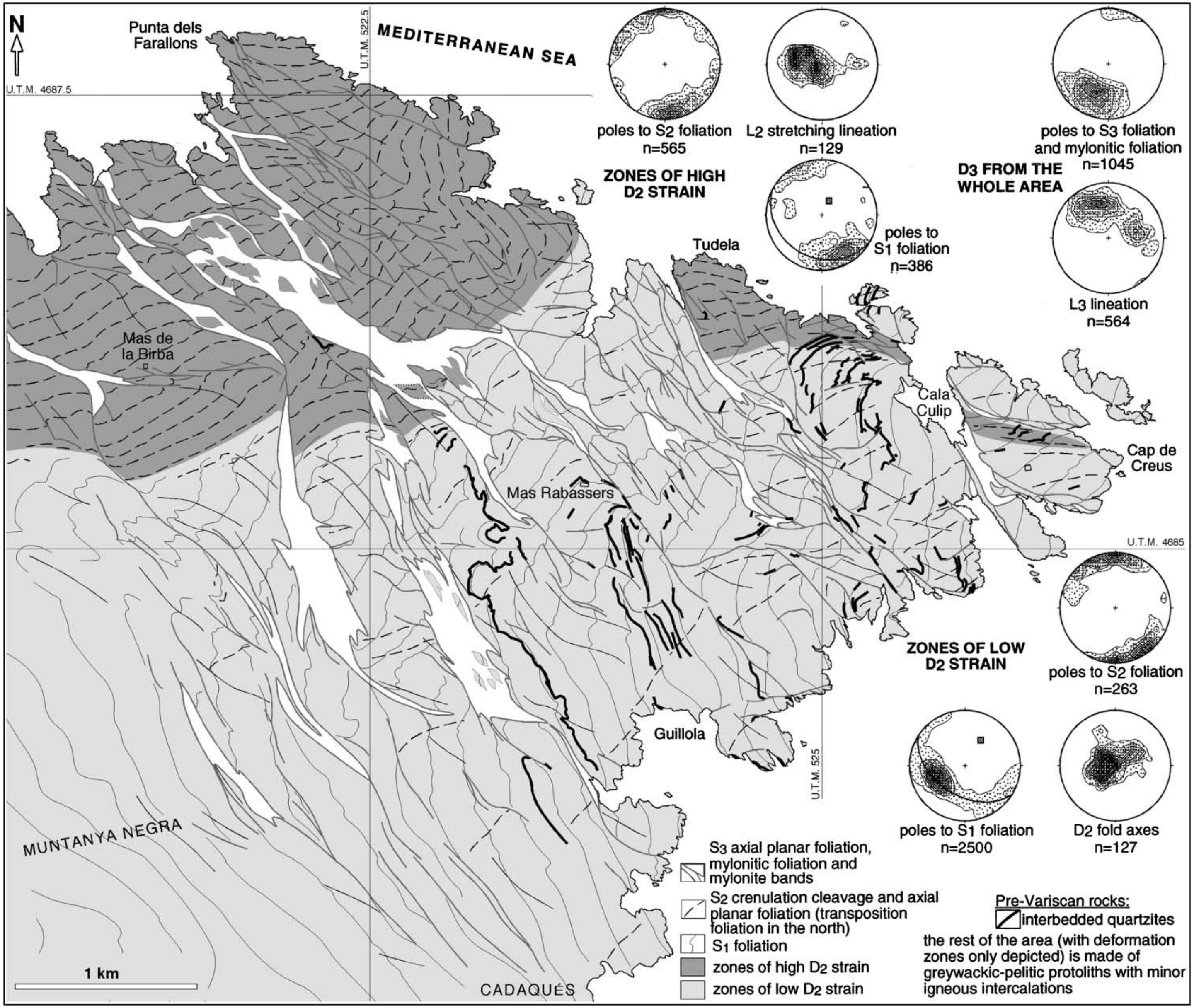
Zones of retrograde metamorphism and mylonitization in greenschist facies conditions (Fig. 4e) are heterogeneously developed, and appear intimately related to the development of late structures (folds and shear zones). In Fig. 3, only those bands of intense folding and mylonitization are represented, although more diffuse zones of retrograde metamorphism do exist throughout the area, which involve, for instance, the alteration of porphyroblasts into pinite and sericite. Sporadically, in westerly domains affected by late folding, kyanite is found as pseudomorphs of andalusite and/or sillimanite.

Although prograde zoning reflects low-P mineral assemblages developed about the metamorphic peak, medium pressures may have been reached, as recorded by the following features: (i) the sporadic presence of relics of staurolite inside andalusite or cordierite, which may indicate an earlier stage with higher pressure conditions; and (ii) the localized existence along pinched synforms of kyanite pseudomorphing andalusite and partially replaced by muscovite produced during a late metamorphic stage. Staurolite relics are relatively common in many other Variscan massifs of the Pyrenees and of NW-Iberia, and they have been interpreted as corresponding to a medium pressure metamorphic stage coeval with an early compressional event (Guitard et al., 1996; Reche et al., 1998). Autran and Guitard (1970) first reported the presence of kyanite appearing in late stages of the metamorphic evolution in the Cap de Creus and Albera massifs, and interpreted it as an increase in pressure during late folding.

3.2. Magmatism

In the Cap de Creus, two groups of Variscan igneous rocks are represented, on the basis of the level of emplacement and the volume of the intrusions: small intrusives located in the northern metamorphic zone, and the Roses and Rodes granodiorite stocks (Fig. 2).

(1) In the sillimanite–muscovite and the sillimanite–K-feldspar zones, there are, in order of emplacement, small intrusions of hornblende quartz–gabbro, quartz–diorite, tonalite, granodiorite and granite (a calc-alkaline association) and dykes of leucogranite and pegmatite (a peraluminous association). These intrusions are surrounded by small migmatite pods, so the whole set is called migmatite complex. Three migmatite complexes have been recognized in the Cap de Creus peninsula: the Cap Gros (Ramírez, 1983), the Punta dels Farallons (Druguet et al., 1995; Druguet and Hutton, 1998) and the Tudela (Druguet, 1997) complexes (Fig. 3). The pegmatites form a 2.4 km wide irregular dyke swarm from the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ZONES OF HIGH D2 STRAIN
poles to S1 foliation n=386
poles to S3 foliation and mylonitic foliation n=1045
D3 FROM THE WHOLE AREA
L3 lineation n=564
Tudela
Mas de la Birba
Cala Culip
Cap de Creus
Mas Rabassers
U.T.M. 4685
ZONES OF LOW D2 STRAIN
poles to S2 foliation n=263
Guillola
U.T.M. 525
poles to S1 foliation n=2500
D2 fold axes n=127
MUNTANYA NEGRA
S3 axial planar foliation, mylonitic foliation and mylonite bands
S2 crenulation cleavage and axial planar foliation (transposition foliation in the north)
S1 foliation
zones of high D2 strain
zones of low D2 strain
Pre-Variscan rocks:
interbedded quartzites
the rest of the area (with deformation zones only depicted) is made of greywackic-pelitic protoliths with minor igneous intercalations
1 km
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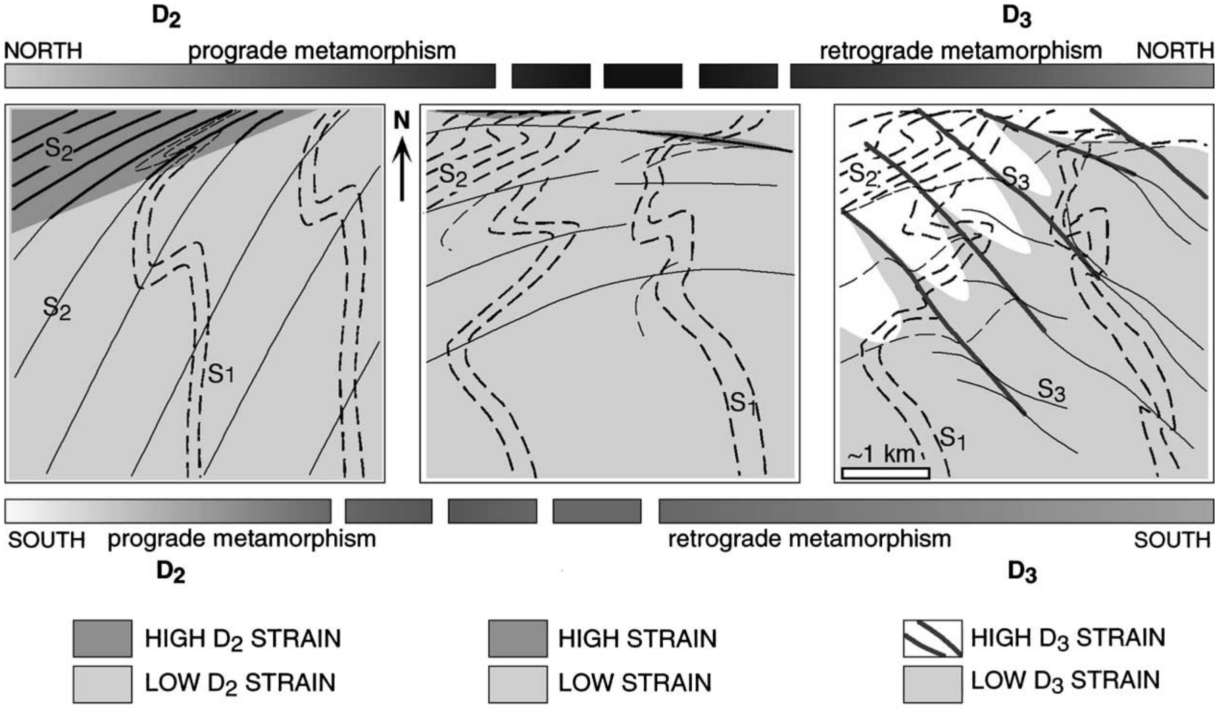

Fig. 7. Sketches of map views to show three stages in the structural evolution of the area. Thick lines represent the dominant foliations at each stage. Solid lines represent developing foliations and dashed lines are foliations being deformed. D2 deformation takes place during the prograde metamorphism and D3 deformations occur at retrograde conditions. An intermediate stage between D2 and D3 has different spatial correlations to metamorphism: while in the north it develops just before and after peak metamorphism conditions, in central and southern domains it develops at more clear retrograde conditions.

cordierite–andalusite zone to the sillimanite zone and migmatite complexes, sub-parallel to the trend of the metamorphic zones. The rocks of the peraluminous association (leucogranites and pegmatites) could derive from partial melting of the pelitic metasediments, by analogy with the perianatectic pegmatites defined by Autran et al. (1970) in other Variscan mid-crustal domains of the Pyrenees, and consistent with data obtained by Damm et al. (1992). In contrast, the rocks of the calc-alkaline association may represent mantle-derived magmas with different degrees of crustal contamination. In spite of the lack of trace element and isotopic data, major element data and the calcic composition of plagioclase (bytownite to labradorite) in quartz–gabbros, likely suggest a mantle origin for the less differentiated magmas (Druguet, 1997). The increase in temperature associated to ascent of these intermediate–basic magmas induced local anatexis, which may be the reason for migmatites being restricted to a few small areas around the granitoid intrusions. All these rocks display magmatic to solid state deformational fabrics, being mainly contemporaneous with high temperature deformation, as will be explained later.

(2) Variscan magmatism in the Cap de Creus peninsula also includes granodiorite stocks (the Roses and Rodes granodiorites) which were emplaced in the low grade metasediments, south of the study area (Fig. 2),

Fig. 6. Structural map of the NE Cap de Creus peninsula (location shown in Fig. 2), with special emphasis in D2-related structural zonation. Lower-hemisphere equal-area plots of lineations and poles to foliations (contour interval 1%). Location of poles of cylindrical best fits (squares) is also shown for plots referred to S1 foliation.

recording both synmagmatic fabrics and low-temperature mylonitization. They have a rather homogeneous composition, varying between granodiorite and tonalite. They clearly correspond to shallow intrusions, since they are sheet-shaped and emplaced in the low grade, upper series metasediments, producing a narrow aureole of contact metamorphism (Carreras and Losantos, 1982). According to the mentioned features, they are similar to other calc-alkaline granitoid batholiths emplaced at upper crustal levels of the Variscan basement of the Pyrenees.

4. Tectonothermal evolution of the NE Cap de Creus

4.1. The structure

In the NE Cap de Creus, a complex structural pattern results from progressive, heterogeneous non-coaxial deformation (Figs. 6 and 7). In such a situation, different tectonic phases or events are difficult to distinguish. However, as a requirement, structures have been grouped in three different deformation episodes according to the structure overprinting criteria and the attitudes of the fold axial planes for each deformation event. In addition, the progression of deformation during metamorphism enables separation, in medium and high grade domains, of structures preceding (D1), developed close to (D2) and succeeding (D3) the metamorphic peak.

(1) The oldest deformation recorded in the area (D1), led to the development of a widespread first penetrative schistosity (S1), formed prior to the metamorphic climax and well defined in rocks of the metasedimentary sequence by the alignment of phyllosilicates (Fig. 4c). Its main characteristic is the sub-parallelism with bedding (Ss) almost throughout the entire area, thus named Ss/1. In domains where there is no significant overprinting of S1 by later structures, it displays a dominantly N–S trend, with a moderate to steep easterly dip. There is little evidence on the regional significance of the D1 early deformation. However, it seems that this early event was governed by tangential tectonics, producing recumbent folds and/or thrusts with an associated gently dipping foliation. Although the original orientation of these structures cannot be established, attitudes of intersection lineations point to a near N–S trend. This tectonic event would be consistent with crustal thickening, involving a first metamorphic stage.

Before the onset of D2 deformation, the S1 foliation had a variable attitude, reflecting the existence of a NNW–SSE trending major dome-shaped structure (Druguet, 1997). In zones of dominant S1 foliation, this macrostructure is inferred from changes in orientation of the S1 foliation and from the dispersion of F2 and F3 fold axes, with S1 and fold axes being steeper towards the east. D2 and D3 events do not explain the flexure of early foliation, so that a pre-D2 macrostructure is required. However, neither penetrative foliations nor minor structures related to such a macrostructure have been recognized.

(2) The D2 event, defined as a prograde folding event, is characterized by the folding of bedding and the S1 first schistosity in prograde metamorphic conditions up to the peak of metamorphism. Crenulations and tectonic bandings in metapelites are mainly defined by the preferred orientation of biotite growth (Fig. 4c), and by biotite + sillimanite in the medium to high grade metamorphic zones (Fig. 5b). D2 non-coaxial progressive deformation gave rise to the development of upright folds. The prevalent original trend of these structures is close to NE–SW and about ENE–WSW in domains of intense D2 deformation. The existence of fold-related strain gradients across the area enables one to distinguish between structural domains with relatively high and low strain. In general, two main structurally homogeneous domains related to D2 folding event can be differentiated, a southern low strain domain with a dominating S1 foliation and a northern high strain domain where S2 dominates, giving rise to a structural zonation (Figs. 6 and 7).

The D2 folding structures evolved to folds and crenulation cleavages with an E–W trend (Fig. 7, central sketch). In the northern area, it is difficult to distinguish from the D2 event, because of the progressive character of deformation and the relative high temperature conditions at which both structures were developed (Fig. 4d). On the other hand, in the central and southern domains, these E–W folds were developed at retrograde conditions, being difficult to distinguish from the following D3 structures. Consequently, the E–W trending structures may be assembled with either D2 or D3 depending on

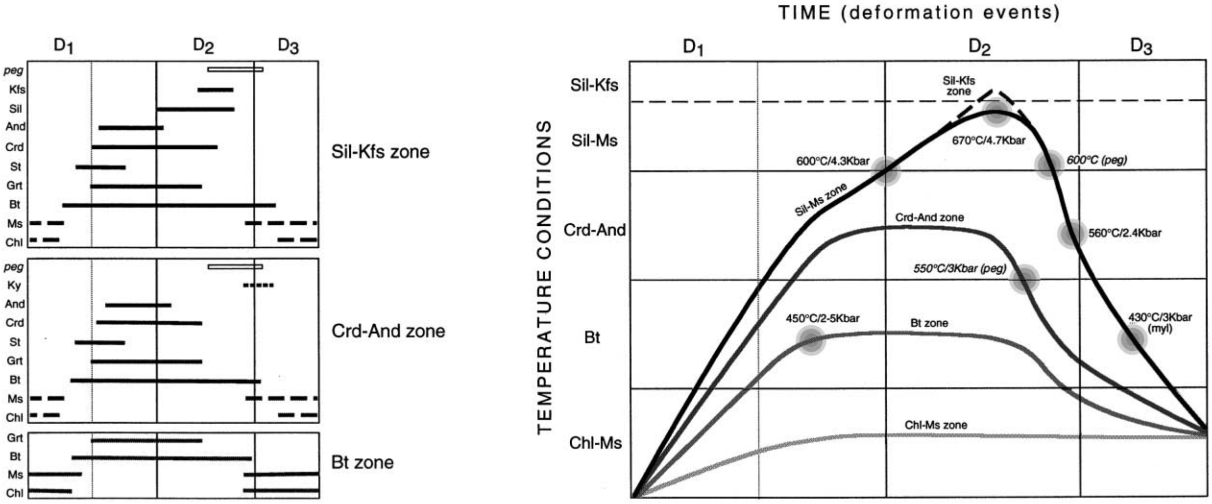

Fig. 8. Sketches showing relations between successive metamorphic zones and deformation events. The plot in the left side shows the relative time of blastesis of different minerals plotted for three different metamorphic zones. In the right side, an interpretation of the evolution of mapped metamorphic zones is shown. *P–T* estimates are roughly plotted as graded circles; those referred to pegmatite emplacement (peg) come from Damm et al. (1992) and Alfonso (1995), and that referred to mylonitization (myl) is from Bossière et al. (1996).

the metamorphic conditions at which they were developed.

F2 folds are characterized by vertical or steeply inclined axial surfaces and associated crenulation cleavage. Folds have moderate plunging (in the westernmost areas) to down-dip fold axes (eastwards), which are closely parallel to the L2 stretching lineations. These fold axes and L2 lineations usually plot near to the fold axes corresponding to the deduced cylindrical folds affecting the S1 foliation (Fig. 6).

Towards the centre of the high strain zone, an increase in strain is manifested by the increasing tightness of D2 folds and the clockwise rotation of both Ss/1 and S2 towards a steeply north dipping ENE–WSW trend, with coeval development of a S2 transposition foliation. The change in orientation of both foliations, from the low strain into the high strain zone, defines a sigmoidal folding structure, with steeply NE plunging axis (Fig. 6). High strain zone, mostly developed in the north, could represent the stretched and sheared limb of this large sub-vertical fold. Strain analysis performed in medium strain domains by the use of deformed quartz and pegmatite veins reveals an average horizontal shortening greater than 50%.

Previous works (Carreras and Druguet, 1994a, Druguet, 1997; Druguet et al., 1997) point towards a kinematic model where the whole structure is interpreted as a complex transpressive shear zone involving vertical extension, NNW–SSE sub-horizontal bulk shortening with reverse and minor dextral components, and bedding-S1 parallel sinistral flexural flow. This complex structural pattern, highly influenced by the presence of a previous anisotropy (Ss/1), is kinematically consistent with a broadly transpressive D2 strain regime.

(3) Progressive deformation under clear retrograde metamorphic conditions gave rise to D3 folds and shear zones. This deformational event was responsible for heterogeneous folding and shearing of D1 and D2 structures. The prevalent orientation of the D3 late folds and shear zones is close to NW–SE (Figs. 6 and 7), although the attitude of planar and linear structures depends on strain intensity, with axial planes and stretching lineations rotating, respectively, towards the NW–SE and to the NNW with increasing strain. These directions correspond to the main trend of the shear zones and of the associated stretching lineation. The E–W trending folds described in (2) would represent early stages of D3 progressive deformation, while the NW–SE folds and shear zones probably represent the latest stages of this event. This is specially manifest in lower

grade domains where, in addition, there is no significant change in metamorphic conditions during these events. Gradual spatial changes in the structural style during this late deformation produced a second structural zonation, evidenced by the disappearance in higher grade domains of late folds and the preponderance of shear zones. Mylonitic bands in the Cap de Creus area are related to this anastomosed network of shear zones, with predominantly dextral-reverse movements (Carreras, 1975; Carreras and Casas, 1987).

4.2. Relationships between deformation, metamorphism and magmatism

The metamorphic conditions at which rocks were deformed have been used in the previous section as a tool to separate different deformational events. In what follows, an outline is given on the relations between metamorphism, magmatism and deformation.

Regional prograde metamorphism approximately started during D1 deformation event (Fig. 8). Metamorphism probably did not exceed the greenschist facies conditions (or the lower amphibolite facies at the most) during D1. The possibility of staurolite growing during D1 deformation is open, since, where observed, andalusite or cordierite always replaces it. These first metamorphic stages are also inferred from the abundance of pre- to syn-D1 quartz segregation veins, with metamorphism involving dehydration reactions for water-saturated sedimentary rocks.

The observed relationships between mineral growth and deformation indicate that the blastesis of most minerals postdates D1 deformation. Most porphyroblasts (cordierite, andalusite, garnet) grew when S1 foliation was already well developed (Figs. 4b and 5a). The dome shaped macrostructure that formed in this intermediate stage may be associated to this metamorphic episode. However, the lack of meso and microstructures related to the macrostructure hinders attempts to unravel the question.

For rocks located in the biotite and cordierite–andalusite zones, the maximum metamorphic conditions were reached during this intermediate episode between D1 and D2 (Fig. 8). For rocks located in the sillimanite and sillimanite–K-feldspar zones, peak metamorphic conditions occurred later, since sillimanite (Fig. 5a) and K-feldspar grew synkinematically during D2. Therefore, it is assumed that the higher grade the metamorphic zone the later was achieved the metamorphic climax.

The prevalent succession of mineral associations in the entire area is a typical indicator for a metamorphism prograding from low to high temperature at low pressure and high thermal gradient conditions. D2-related structural zonation would take place simultaneously, together with migmatization and emplacement of calc-alkaline granitoids in domains with intense D2 deformation and upper amphibolite facies metamorphism. A sequence of intrusions from the more basic (quartz–gabbros and quartz–diorites of the migmatite complexes) to acid types (leucogranites and pegmatites) is established on the basis of overprinting relationships and intensity of deformation recorded by each magmatic type, being all syntectonic with respect to progressive D2 deformation. The intermediate–basic bodies registered part of D2 deformation, as they already carry xenoliths with F2 folds, but have fold-like shapes and record magmatic to solid state fabrics (consisting of an alignment of hornblende, biotite and plagioclase) which are correlated with the S2 transposition foliation in the metasediments. A predominantly steeply plunging stretching lineation is sometimes associated with the granitoid fabric, and it is sub-parallel to the lineation in the enclosing metasediments. Also the structures displayed by the migmatitic schists (Fig. 5b) provide good evidence for the contemporaneity of deformation and migmatization (Druguet and Hutton, 1998).

Retrogression of metamorphism approximately started at the latest stages of D2 deformation (Fig. 8). Pegmatites were emplaced just following peak metamorphism and, in relation with them, coarse muscovitization took place in the surrounding metasediments. Metamorphism, at least at this stage, seems to be intimately related in both space and time to the emplacement of such intrusive rocks. The latest stages of D2 were therefore contemporaneous with high temperature retrogression in the northern area. The large sub-vertical folding structure and associated sub-vertical extension, initiated in the D2 event, continued during this stage and enabled decompression of the rocks located at the northern high strain zones, accompanied by a slow temperature drop. In some specific localities, metamorphic rocks would

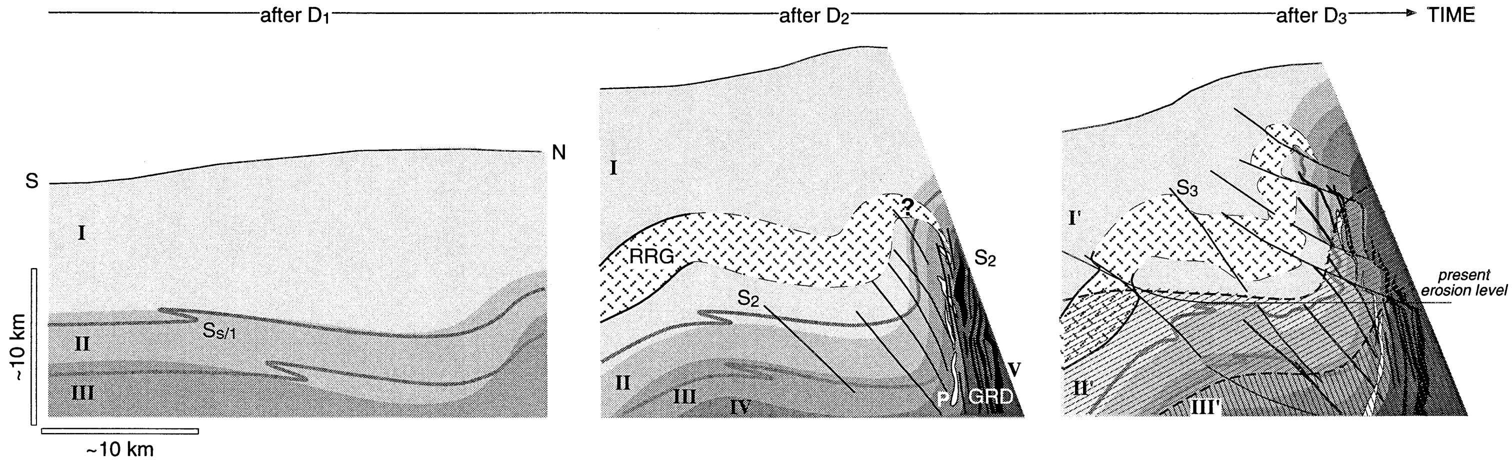

Fig. 9. Schematic cross-sections of a model for the tectonothermal evolution of the Cap de Creus area, with special reference to the NE zone (location shown in Fig. 2). Three ideal stages have been considered: after D1, after D2 and after D3 deformations. Zones of prograde metamorphism: (I) non-metamorphic to chlorite–muscovite zone; (II) biotite zone; (III) cordierite–andalusite zone; (IV) sillimanite zone; (V) migmatite areas. Intrusive rocks: GRD, granitoids and diorites from the migmatite complexes; RRG, Roses and Rodes granodiorite stocks; P, pegmatites. Zones of retrograde metamorphism: (I′) non- or very low metamorphic; (II′) lower greenschist facies; (III′) upper greenschist facies. After D1: early deformations produce a gently east dipping schistosity and onset of regional metamorphism. After D2: folding and development of high strain zones and a lateral high thermal gradient in the north, with granitoid sheets and pegmatites intruded vertically. The Roses and Rodes granodiorites are suggested to intrude as sheet-shaped and flat-floored bodies in shallower levels, having their root zone in the migmatite areas. Regional transpression with a south vergence produces vertical extension and horizontal shortening (crustal thickening). After D3: rapid erosion and uplift causes exhumation of metamorphic and igneous rocks, while resetting the normal, regional ambient, thermal conditions.

have been relatively buried along the hinge zones of pinched synforms, so that they would have passed through the kyanite stability field. In NW Spain, medium pressure kyanite bearing assemblages also occur in pinched synforms (Reche et al., 1998). In a general model for the whole Variscan massifs in the Pyrenees, Guitard et al. (1996) propose two hypothesis for the appearance of late kyanite: (i) a slight increase in pressure associated to late folding; and (ii) isobaric cooling during late stages of LP/HT metamorphism. The first is the most suitable explanation for the local presence of kyanite in the Cap de Creus.

Retrogression was more significant during D3, i.e. specially developed around late folds and shear zones, where low metamorphic grade conditions were newly reached. At this stage, the set of intrusive granitoids, included the pegmatite dyke swarm, would be completely crystallized, being able to deform at a low temperature solid state.

Horizontal shortening related to progressive D2 to D3 deformation would also account for the narrowing of metamorphic zones. Although it is difficult to determine average strain values in heterogeneously deformed rocks (only performed in D2 medium strain domains), a minimum horizontal shortening of 50% normal to isograds boundaries may be inferred for the whole D2 to D3 evolution. Assuming this percentage, and after restoring the shortening effect, a considerably lower thermal gradient of about 40°C/km could have preceded.

5. Discussion and conclusions

Polyphasic structures in the NE Cap de Creus peninsula are interpreted as part of a progressive deformational history. A continuity of events from prograde to retrograde metamorphic conditions is evidenced. Metamorphism started to develop high gradients at the time that deformation also displayed high strain gradients. Intensification of deformation and metamorphism coincides in space and time with magmatism (Fig. 9). The observed relationships between deformation, magmatism and metamorphism provide evidence that all three processes operated simultaneously in this segment of the Variscan belt.

The interaction between deformation, plutonism and metamorphism would explain the high thermal gradients observed in the study area (Figs. 2 and 3). The rocks now located in the north would have experienced a temperature rise due to the vicinity of hot, deep-level material. Magmas would ascend towards upper levels and along high strain domains, possibly favoured by vertical extension. The presence of sub-vertical deformation zones with down dip maximum elongation direction is thought to be crucial in determining melt transfer upward through the crust (Hutton, 1997; Brown and Solar, 1998a). Mantle derived magmas, preferentially channelled along the sub-vertical high strain zone, induced lateral heat flow in such a way that metamorphism increased and metamorphic zones became rather steep and thin towards the core of the coeval 'transpressive intrusion zone' (Fig. 9). Even the anatectic limit could have been superimposed over lower grade zones, such as the telescopic effect model (den Tex, 1963). As deformation proceeded, just after the peak of metamorphism (when pegmatites intruded), horizontal shortening would also account for the narrowing of metamorphic zones.

The overall tectonic evolution from D2 to D3 is characterized by: (i) a clockwise rotation of structures, from upright NE–SW to NW–SE with southern vergence; (ii) a broadly transpressive regional regime, starting with a contraction or pure shear-dominated transpression and later evolving to wrench-dominated transpression; and (iii) progressive strain localization along narrow transposition bands and shear zones (mylonite belt) at low temperature conditions. Therefore, the overall Variscan tectonometamorphic evolution involved a sequence of events which are difficult to separate in distinct phases. Consequently, the late folds and shear zones are interpreted to be Variscan.

The Cap de Creus massif offers insights into the interaction between deformation, metamorphism and magmatism in an orogenic belt governed by a regional transpressive regime. The high thermal gradients, evidenced here and in many other similar settings, can be explained on the basis of strong horizontal strain gradients superimposed on already magmatically-induced thin metamorphic zones. The continuity of some stratigraphic markers (e.g. quartzites) from low to high grade metamorphic zones (Figs. 3 and 6), and the absence of significant truncation of the structures and zones of metamorphism, preclude the existence in this area of any detachment or major fault responsible for the strong metamorphic gradients. Late shear zones, although locally affecting the isograds and the boundaries of the migmatite complexes, involve offsets of only a few hundred meters of displacement (Carreras and Druguet, 1994b). Progression of transpressional deformation during cooling is responsible for narrowing of the metamorphic zones and for uplifting deeper rocks on the north side of the high strain zone. The evolution from high to low temperature deformation during decompression and the subsequent exhumation require active denudation of the crustal domains overlying the high strain metamorphic belt, in accordance with kinematic and thermomechanical models proposed for similar settings (Thompson et al., 1997; Fossen and Tikoff, 1998; Stüwe and Barr, 1998).

The adjacent Roses and Rodes granodiorite massifs, although separated from the migmatite domains, are most likely connected to the proposed magmatic and tectonometamorphic evolution recorded in the northern part of the Cap de Creus peninsula. Spatial association of migmatite domains and deepest parts of batholiths is common in the Variscan of the Pyrenees, for instance in the Albera, Roc de Frausa (Liesa and Carreras, 1989) and in the Andorra-Mont-Lluis massif. Furthermore, these batholiths, mainly emplaced in shallow levels (Fig. 9), recorded structural features which have been interpreted as a result of progressive deformation during cooling (Simpson et al., 1982; Gleizes, 1992). Synmagmatic fabrics in the granodiorites can be correlated with D2 structures in the study area, and late granodiorite shearing with D3 folds and shear zones. Taking into account the bulk structure of the Cap de Creus Peninsula and the geometry drawn for other Eastern Pyrenean batholiths, it is conceivable that the Roses and Rodes granodiorites were linked to the migmatite domains but lying at a higher structural level. However, this possibility remains a hypothesis until geochemical evidence for the association between both magmatic rock types is established.

Concerning the bulk tectonic setting of the Variscan of the Pyrenees, evidence presented from the Cap de Creus structure is in agreement with the transpressive regime proposed for the main Variscan event in the Pyrenees by Carreras and Capellà (1994), Leblanc et al. (1996), Evans et al. (1997) and Gleizes et al. (1998). According to the first authors, the structural history is interpreted as a gradual evolution from an earlier compressive to a late transpressive and finally transcurrent regime, implying a continuous horizontal crustal shortening. Analysis of structures in the Cap de Creus area supports this hypothesis. No evidence has been found suggesting a regional extensional event or any tectonic excision of lithological units.

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