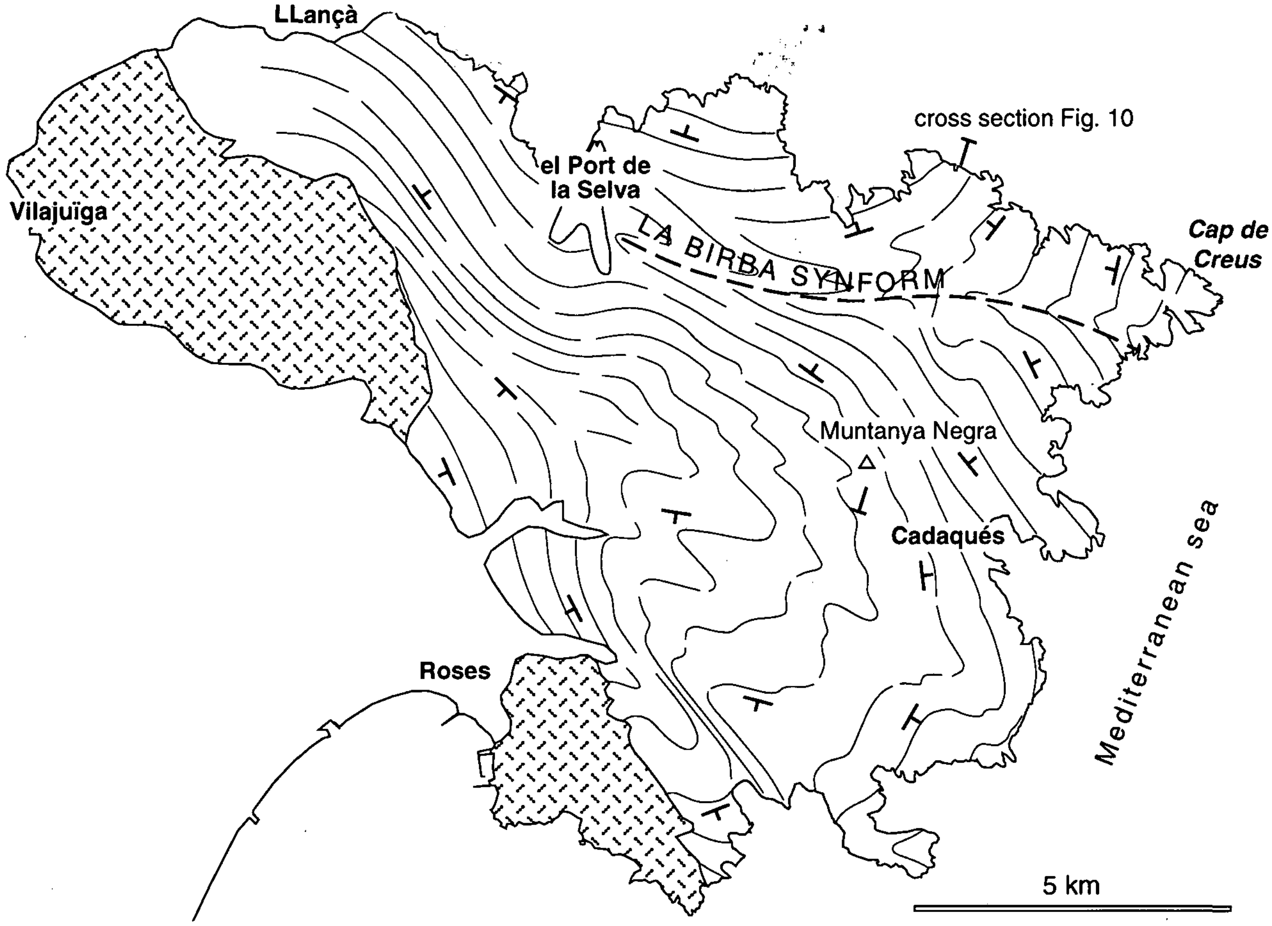


Fig. 9. Structural traces of dominant foliation in the Cap de Creus peninsula, according to Carreras & Losantos (unpublished).

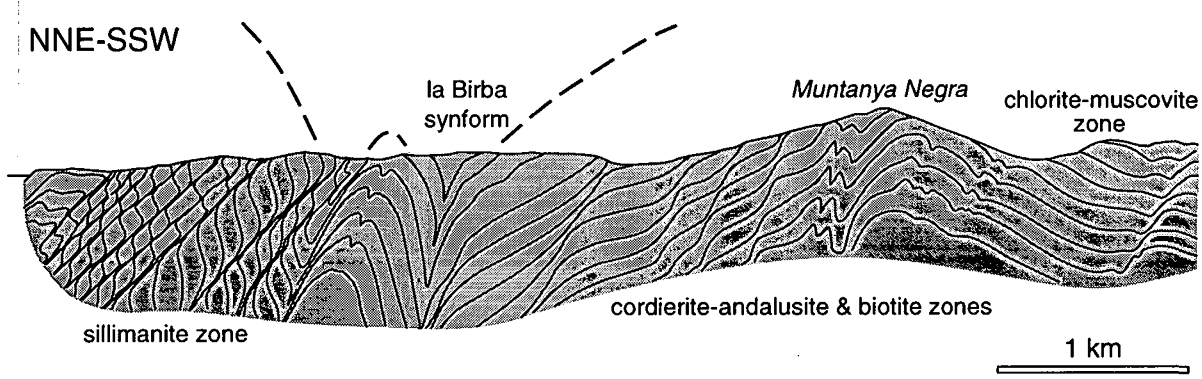


Fig. 10. Structural section across the northern part of the Cap de Creus peninsula (after Carreras et al. 1980). See location in Fig. 9.

The described setting, where medium to high grade deep-seated rocks are located on the northern areas and low grade, upper stratigraphic sequences and granodiorites on the south reflects, in analogy to other areas of the Hercynian basement, the spatial dissociation between high grade domains and granitoid batholiths and stocks. Only south of Llancà granodiorites are found adjacent to medium grade metasediments.

This setting has analogies with other areas in the Pyrenees, where deep-seated structural levels are exposed. However some singularities make difficult the correlation of structures between this massif and other massifs in the Eastern and Central Pyrenean Axial Zone. A main singularity of the Cap de Creus peninsula is the absence of gneiss domes as those of the nearby Alberes, Roc de Frausa or Canigó Massifs. The only dome-shaped structure is the gently antiform located in the centre of the Peninsula, separating the north dipping domains from the south dipping domains, but the core of this structure is made of low or very low grade metasediments. On the other hand, medium to high grade rocks appear in a domain with steep attitudes of prevalent foliations, and also with moderate to steep fold axes. This setting prevents to assimilate this structural domain to the "infrastructure" of Sitter & Zwart (1960), characterized by a flat-lying attitude of all structures. Possibility that this domain corresponds to a subsequently steepened original flat-lying structure will be later discussed. Furthermore, this steep domain is not unique in the Pyrenees, as similar steep foliations are observed in the nearby Albera massif, where medium grade schists form a sub-vertical sheet pinched between granodiorites in the south and the orthogneisses in the north.

As in other deep-seated Hercynian domains, the tectonic history of the northern Cap de Creus peninsula is complex and different tectonic events are difficult to separate. Often, a well developed structural element like a prevalent foliation is difficult to correlate with major structures, and is even more difficult to associate its formation to a specific tectonic regime. Early works on the Hercynian structure of the Pyrenees tried to characterize the tectonic evolution by distinguishing different tectonic phases of widespread distribution, each of them characterized by a specific tectonic style and orientation. This lead to distinguish in the Central Pyrenees four main deformation events (Boschma 1963, Oele 1966, Hartevelt 1970). In the Eastern Pyrenees, Laumonier & Guitard (1978) and Laumonier et al. (1984) arrive to individualize 8 different tectonic phases. In an attempt to simplify the tectonic evolution, Carreras & Santanach (1983), using a criteria previously used by Guitard (1960), grouped tectonic phases in three main events: an early pre-schistose, a syn-schistose and a late post-schistose. However, Carreras & Capellà (1994) recently challenged the validity of this criteria, arguing that phases ascribed to one of this main events may correlate to phases from another event across different tectonic levels. Furthermore, these authors consider that structures arise as the result of a progressive deformation, rather than being the consequence of distinct deformation episodes. This postulate is justified with examples of the late deformation event from different areas (e.g. in the eastern Pallaresa (Carreras & Cirés 1986) and in the Cap de Creus (Carreras & Casas 1987)).

Under such circumstances, a problem arises when attempting to conceive a description of structures arose from a progressive deformation, and when structures interpreted as contemporaneous may differ in style from place to place, even in a reduced area such as the here studied. However, as a recount requirement, structures have been grouped in different deformation episodes (D_1 to D_3), although in some instances structures which are differently labelled might have been created during a unique deformation episode. In addition to the structure overprinting criteria, used in classical analysis for phases separation, two other criteria have been subsequently used to group different structures. Firstly, the progression of deformation during metamorphism has enabled to separate, in medium and high grade domains, structures preceding, developed close to and succeeding the metamorphic peak. Secondly, the fold axial planes have rather constant attitudes for each deformation event. The combination of these two criteria enables the distinction of three deformation events.

1) The oldest deformation recorded in the area (D_1), led to the development of a widespread first penetrative schistosity (S_1), formed prior to the metamorphic climax. Its main characteristic is the sub-parallelism with bedding (S_s) almost throughout the entire area. No clear D_1 macrostructures have been recognized, although some repetitions of quartzite beds could be due to tight F_1 folds. In domains where the S_1 foliation dominates without significant overprinting of later structures, it displays a dominantly N-S trend, with a moderate to steep easterly dip.

2) Later, intense and markedly inhomogeneous D_2 deformation led to folding of bedding, S_1 surfaces and early segregated quartz veins. F_2 folds are characterized by vertical or steeply inclined axial surfaces and associated crenulation cleavage, trending approximately NE-SW in less deformed domains and about ENE-WSW in domains of intense D_2 deformation. In highly deformed domains, the crenulation cleavages leads to the transposition of the previous S_1 foliation. F_2 folds

often have steeply plunging axes. D_2 deformation took place about the peak metamorphic interval, and lasted till the emplacement of the pegmatites, shortly after the peak of metamorphism.

3) Following this, in retrograde metamorphic conditions, E-W to NW-SE trending folds (F_3) developed. This folds are the so-called late folds by Carreras (1975). E-W to NW-SE trending ductile shear zones, associated to F_3 folds, were also developed under retrograde metamorphic conditions (Carreras and Casas 1987). Mylonitic bands in the Cap de Creus area are related to this anastomosed network of late shear zones, with predominantly dextral strike-slip or oblique-slip movements.

D_2 and D_3 are characterized by the progressive character of deformation and, in consequence, they are difficult of separate. In addition, structures evidence a high deformation inhomogeneity and thus, domains containing interference structures belonging to both deformation episodes coexist with domains where only one (either D_2 and D_3) is manifest. In such situations, ascribing structures to one or the other event is complex. The steady state metamorphic conditions existing in low grade domains makes inappropriate to differentiate different deformation events on the basis of the metamorphism.

2.3. SETTING OF THE METAMORPHISM AND PLUTONISM

The studied area is characterized by the presence of a metamorphic gradient which increases from south to north. This gradient reflects the effects of a prograde low pressure regional metamorphism, initiated during the early Hercynian deformational events and affecting all the pre-Hercynian lithologies in the area. Over a horizontal distance of approximately 5 km, the pelitic metasediments show a gradient from the chlorite-muscovite zone in the south to the sillimanite - K-feldspar zone in the north (Fig. 11). A retrograde metamorphism, uniform in grade but strongly inhomogeneous in distribution, is preferentially developed along the mylonitic zones, and superimposed on the prograde metamorphic pattern.

Although prograde zoning reflects low-P mineral assemblages developed about the metamorphic peak, medium pressures must have been reached according to the following features: (i) the sporadic presence of relics of staurolite inside andalusite or cordierite, which would indicate an early stage with medium pressure conditions and (ii) the localized existence of kyanite pseudomorphosing andalusite and partially reemplaced by muscovite produced during a late metamorphic stage. The presence of kyanite appearing in late stages of metamorphic evolution was first reported by Autran & Guitard (1970) in the nearby Llançà area.

In a more extended setting, the described zonation extends all along the northern part of the Cap de Creus Peninsula, with medium grade zones appearing north of an imaginary line linking Cadaqués and Port de la Selva. South of this line, all rocks show a low or very low metamorphic grade, and no marked differences in metamorphic conditions occurred throughout the entire Hercynian evolution. Outside the northern metamorphic belt, medium grade schists are restricted to the area South of Llançà, (Montoto 1968, Morales 1975). Although metasediments with analogue metamorphic zonation appear in the neighbouring Albera massif, the medium grade zones do not show continuity, since schists from the south Llançà area are truncated by the La Valleta fault, which brings them in contact with the very low grade metasediments of the western Albera Massif (Fig. 3).

This metamorphic distribution, where vast domains of low or very low grade metasediments bound metamorphic cores with high gradient, is a common feature in the Hercynian basement of the Pyrenees. In most settings, these medium to high grade cores broadly coincide with structural domes which frequently contain orthogneisses. It was suggested that these cores acted as heat channels which caused and controlled the metamorphic isograds. The control effect of gneissic domes on the metamorphic distribution was named "effect de socle" (Guitard 1960, 1970, 1976, 1989; Fonteilles & Guitard 1977). However, in some metamorphic zones an orthogneissic core does not outcrop. This is the case in the Bossost area (Zwart 1962) and also in the Cap de Creus.

In the Cap de Creus, the Hercynian igneous activity is represented by two groups, on the basis of the level of emplacement and the volume of the intrusions: (i) small intrusives located in the northern metamorphic belt and (ii) the Roses and Rodes granodiorite stocks.

(i) In the sillimanite-muscovite or sillimanite-K-feldspar zones of the northern metamorphic belt, there are small intrusions of quartz diorites, tonalites, leucogranites and voluminous pegmatite dykes, usually surrounded by small migmatite pods, so the whole set is called migmatite complex. Three migmatite complexes has been recognized in the Cap de Creus peninsula: the Cap Gros (Ramírez 1983a, b), The Punta dels Farallons (Druguet 1992, Druguet et al. 1995) and the Tudela complexes. Occupying a wider area, from the cordierite-andalusite zone to the north, a swarm of anatectic pegmatite dykes forms the so called pegmatite area. In spite of this marked compositional heterogeneity, all rocks appear in medium to high grade metamorphic perianatectic domains with intense D2 deformation, and were all emplaced close in time. Thus, these rocks not only show close space and time links with metamorphism and tectonics, but also a genetic relationship, as will be later explained.

(ii) Hercynian magmatism in the Cap de Creus peninsula also consists of major granodiorite stocks (the Roses and Rodes granodiorites) which were emplaced in the low grade metasediments, south of the study area (Fig. 3), recording Hercynian structures. No conspicuous relationship between the regional metamorphism and the emplacement of these stocks has been recognized. 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. According to the mentioned features, they are similar to other calcalcaline granitoid stocks and batholiths of the Hercynian basement of the Pyrenees.

Similarly to the Cap de Creus, migmatites are also present in other metamorphic zones in the Pyrenees. In these migmatite domains the amount of granitoids is also restricted. Considering their structural setting, the rare granitoids located in deep-seated structural levels were named intermediate granites by Autran et al. (1970). In contrast, the large batholiths and stocks, as described earlier, are usually emplaced at higher structural levels and in low grade metasediments. These constitute the shallow granites of Autran et al. (op. cit.). The term deep-seated granites is uniquely applied to a charnockitic granite in the Agly Massif.

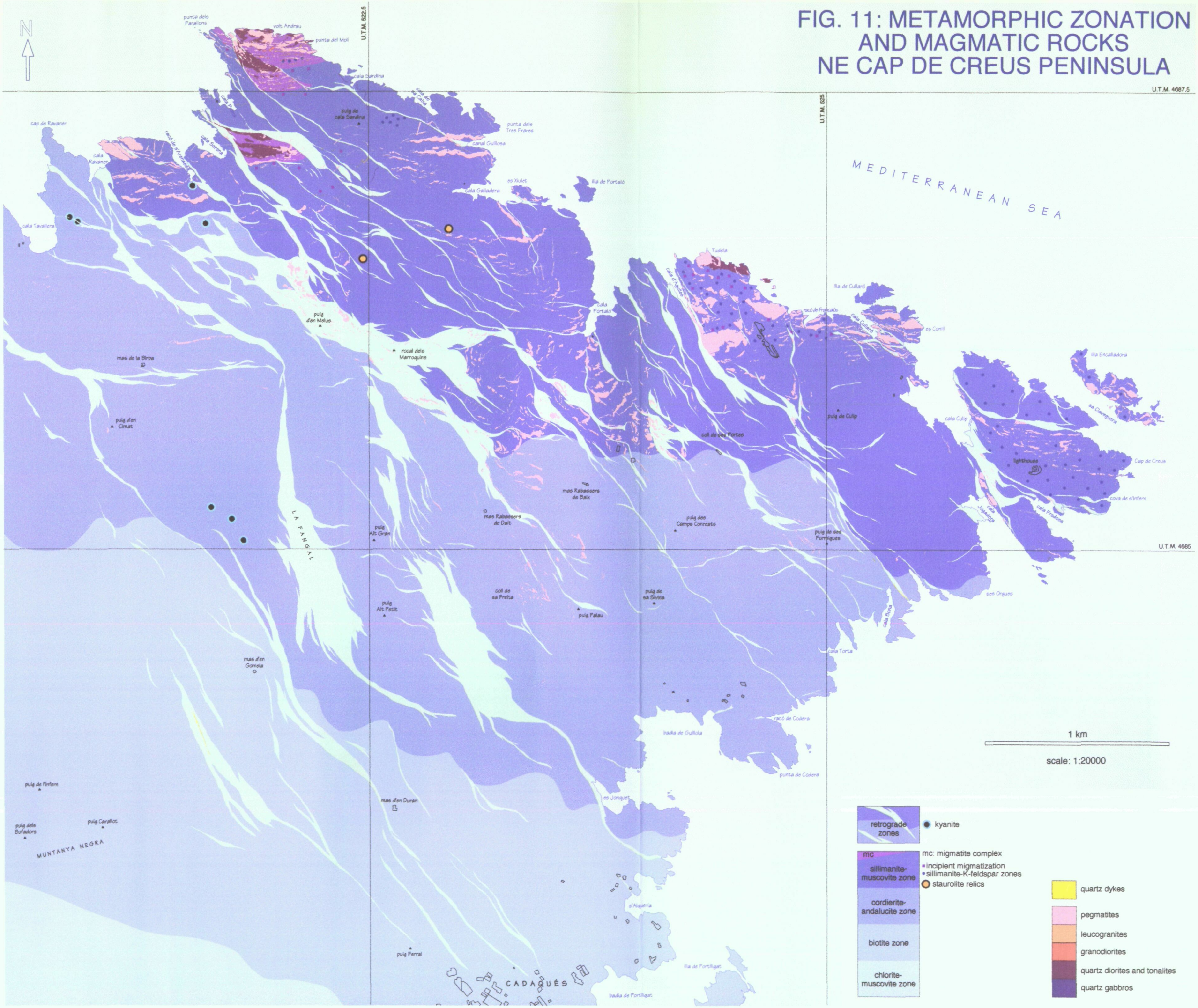
Granitoids in metamorphic cores affecting the metasediments (i.e. the intermediate ones) exhibit a similar range in composition that those in Cap de Creus, with quartz diorites and tonalites being the most abundant types. Similar metamorphic and magmatic settings to Cap de Creus can be observed in the Western Aston (Zwart 1965) or in the Trois Seigneurs (Allart 1959, Wickham 1987) massifs. It is significant that in the metamorphic-migmatitic cores located in the orthogneisses (e.g. Albera massif), leucogranites prevail over other types of granitoids. Such links between the type of migmatitic country rock and related intrusives was remarked by Zwart (1968), who in addition suggested a genetic connection between deep seated quartz diorites and tonalites and shallower tonalite-granodiorite batholiths.

In the migmatite settings, there are swarms of leucogranite and pegmatite dykes which invade the envelopping non-migmatitic domains. These widespread dyke swarms have been referred to as perianatectic pegmatites by Autran et al. (1970).

Radiometric data from Hercynian intrusive rocks forming batholiths in the eastern Pyrenees yield ages ranging between 282±5 Ma (Cocherie 1985) and 275±12 Ma (Vitrac-Michard & Allègre 1975b). Intrusions from the Catalonian coastal ranges give ages of 284±7 Ma (Del Moro & Enrique 1996). The Canigó granite, located together with migmatites in a deep seated structural level, has yielded ages of about 335±15 Ma (Vitrac-Michard & Allègre 1975b). In Cap de Creus, K-Ar analyses in muscovites from a pegmatite gave an age ≥ to 267 Ma (Enrique et al. 1995, 1997). However, younger ages have also been obtained from different muscovites, biotites and K-feldspars (about 240 Ma, and even an age of 85 Ma). According to the authors, these results show that this zone was affected by at least two heating events during post-Hercynian times.

3 METAMORPHISM AND MAGMATISM

FIG. 11: METAMORPHIC ZONATION AND MAGMATIC ROCKS NE CAP DE CREUS PENINSULA



3.1. ZONATIONS OF METAMORPHISM AND MAGMATISM

The distribution of metamorphic and magmatic rocks in the NE of the Cap de Creus peninsula is shown in Fig. 11. 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, presented in the figure, 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, two observable features characterizes this spatial distribution: (i): a sub-parallelism between the enveloping boundaries of the zones of isometamorphism, the pegmatite dyke swarm and the migmatite complexes, with progressively higher grade metamorphic zones northwards, (ii): the described pattern appears heterogeneously cross-cut and overprinted by narrow zones of low grade metamorphism, related to late folding and mylonitization. The present metamorphic pattern appears to be the result of assembling both prograde and retrograde traits.

The establishment of a rather accurate structural pattern has allowed to distinguish in each case whether the primary isograd surfaces (metamorphic origin), or the late shear zones were responsible for the geometry of the metamorphic zones. This present day metamorphic pattern is formed by approximately WNW-ESE trending bands of isometamorphism. Assuming that the zones of isometamorphism display a similar spatial attitude than the pegmatite dykes and the intrusions of the migmatite complexes, isograds might have steeply north dipping boundaries. In consequence, the outcrop surface is cutting metamorphic zones obliquely or at a high angle. Whereas in the the study area prevalent foliations and isograds display steep monoclinal dispositions, to the south, large-scale late folding might had lead to a more complex pattern of folded, dome-shaped isograds (Carreras et al. 1980, Fig. 10).

In the study area, **the chlorite-muscovite zone** outcrops only in the southwestern part. This zone continues southwards of the study area, to become the dominant metamorphic zone of the Cap de Creus peninsula.

The **biotite zone** extends over 1.5 to 2 km from Cadaqués and Muntanya Negra to the vicinities of Mas d'en Gomeia and Es Jonquet (Fig. 11). The first appearance of biotite is irregular and difficult to detect, so that the southern margin of this zone represent the boundary by which phyllites grade into micaschists with the presence of macroscopically visible biotite. Genesis of biotite could be related to the next equilibrium reaction:

$$Ms + Chl + Qtz = Ms + Chl + Bt$$

North of the biotite zone, biotite and muscovite are present as several generations through the metamorphic belt.

The **cordierite-andalusite zone** occupies another band 1.5 to 2 Km wide, extending to an approximate line between Cala Ravaner and Ses Orgues (Fig. 11). The passage to this zone is characterized by a progressive increase in grain size and the sudden appearance of mainly cordierite porphyroblasts (appearing as poikiloblasts with many inclusions) subsequently followed by both cordierite and andalusite crystals, probably depending on the chemical composition of rocks. Genesis of cordierite and andalusite would be related to next equilibrium reactions:

$$Bt + Chl + Ms + Qtz = Bt + Chl + Crd + H_2O$$

$$Crd + Bt + Ms = Crd + Bt + And + Qtz$$

Staurolite is not frequent and, when found, it appears as relics inside andalusite crystals, so that the following equilibrium might have been attained:

$$St + Bt + Ms + Qtz = And + Bt$$

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 andalusite coexist with sillimanite at least in the southern margins of the zone, and cordierite still grows. In some places, relics of staurolite in cordierite have also been found.

In the westerly domains, the cordierite-andalusite and the sillimanite-muscovite zones are separated by a complex mylonite band (Cala Ravaner-Puig d'en Melus-Rocal dels Marroquins shear zones).

The sillimanite-K-feldspar isograd is difficult to determine mainly due to the common occurrence of retrograde muscovite. The presence of the association **sillimanite-K-feldspar** is distinctive in the easternmost area between Cala Culip and Cala Fredosa, and in the southern margin of the Tudela migmatite complex. The following reactions would have been responsible for the formation of K-feldspar:

Ms + Ab + Qtz = Kfs + Sil + H2O
Ms + Qtz = Sil + Kfs + H2O

Other small spots of sillimanite-K-feldspar schists have been detected in the eastern slopes of Puig de Cala Sardina and within the Punta dels Farallons migmatite complex. 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. However, the effective partial melting of the metasediments with development of migmatite schists might have taken 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 best developed Punta dels Farallons migmatite complex is located in contact with the schists of the sillimanite-muscovite zone, and relatively distant from the sillimanite-K-feldspar schists from the lighthouse area. Inferences from these features will be later discussed.

The first pegmatite dykes are found within the cordierite-andalusite zone and they extend as a 2.4 km wide irregular swarm towards the northern coast. They are most abundant in the sillimanite and migmatite areas.

Zones of retrograde metamorphism in greenschist facies conditions are heterogeneously developed, and appear intimately related to the development of late structures (folds and shear zones). In Fig. 11, 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. Sporadically, in westerly domains affected by late folding, kyanite is found as pseudomorphs of andalusite and/or sillimanite.

In what follows, description of the petrological features of the different metamorphic and magmatic rocks will be presented, while the space-time evolution of the thermal history in the area will be discussed later.

3.2. METAMORPHIC ROCKS

The sedimentary sequence and related pre-Hercynian igneous rocks described in section 2.1 were affected by the Hercynian LP-HT prograde metamorphism, transforming each rock into its metamorphic equivalent. The mineralogy and textural features of each particular lithology vary with the metamorphic grade, the mineral changes being most notable in the metapelites, and less evident in other lithologies.

3.2.1. METAPELITES AND METAGREYWACKES

Although these two types of metasediments will be described together in this section, those features specific of one group will be considered separately.

The majority of the new metamorphic minerals (micas and aluminium silicates) develop selectively within the pelitic layers, whereas the most greywacky types are dominated by recrystalized relic crystals of detritic origin, such as quartz and feldspars. The degree of grain growth by secondary recrystallization varies depending on the achieved temperature, which induced a manifest coarsening. Thus, greywackes have quartz and feldspar grain sizes of around 200μm in the chlorite-muscovite zone and of 400 μm in the sillimanite zone.

Metapelites show a penetrative planar fabric (cleavage and schistosity) defined by preferred orientation of micas. In metapsammites, mainly metagreywackes, granoblastic textures are dominant and fabrics are less anisotropic. In millimetric to decimetric alternances, sequences of foliated and granoblastic textures can be observed. A mimetic bedding may have developed by lithologically controlled mica and porphyroblast growth. Successive generations of micas formed, parallel to different fabric elements, either by synkinematic growth or by mimetic recrystallization. In the schists that posses only a bedding parallel foliation, micas are commonly parallel to the bedding-foliation planes, while porphyroblasts grow over them. In schists bearing crenulation cleavages, micas arrange parallel to them, and might define a tectonic banding. Microstructures are very complex in mica-rich metapelites, due to the existence of disharmonic folding induced by the high mechanical anisotropy caused by both the micas and the rigid porphyroblasts. In the schists made of metasedimentary alternances, a characteristic discordance arises between the schistosity in the metapelites (usually a bedding-oblique crenulation cleavage) and that in the metagreywackes (a bedding-parallel foliation) (Fig. 12).

Fig. 12. Micrograph of a ritmite with a metapelitic layer (centre of the photograph) showing a different structural pattern than the adjacent metagreywackes. $S_{s/1}$: bedding-parallel early foliation; S_2: crenulation cleavage; S3: late crenulation cleavage.

The **chlorite-muscovite zone** contains phyllites and fine grainsized metagreywackes. The main minerals in these rocks are quartz, white mica, chlorite, relic feldspars and opaques. Minor amounts of relic tourmaline can also be present. The major element composition of a phyllite from this zone is given in Table 1 (left column).

Micaschists from the **biotite zone** differ from the phyllites of the previous zone by their grainsize and by the presence of macroscopic visible biotite, which progressively replaces chlorite. Biotites are slightly magnesian, with an average Fe/Fe+Mg = 0.55 (Fig. 13). Small garnets appear sporadically in psammitic layers, and chloritoid in dark pelites.

Rock	Phy	Sil-s		
SiO_2	63	68,26	69,06	69,21
TiO_2	0,74	0,76	0,73	0,62
Al_2O_3	16,4	13,53	13,46	13,22
Fe_2O_3	6,03	6,54	5,69	5,22
MnO	0,07	0,08		0,07
MgO	2,55	2,45	2,62	2,45
CaO	0,83	2,31	2,08	2,04
Na_2O	2,63	2,67	2,8	3,22
K_2O	3,67	2,68	2,61	3,33
P_2O_5	0,24	0,18	0,6	0,18
H_2O	2,25	0,56	0,68	0,55
Total	98,41	100	100,3	100,1

Table 1. Chemical composition of metasediments from the Cap de Creus. Phy: phyllite; Sil-s: sillimanite schists. After Druguet (1992) and Druguet et al. (1995).

With some increase in metamorphic grade, **the cordierite-andalusite zone** is reached, characterized by the generalized presence of porphyroblasts of aluminium silicates. Inclusion-rich cordierite poikiloblasts appear first, followed by the growth of tiny porphyroblasts of andalusite. As will be explained in a next chapter, porphyroblasts grow over the S_1 foliation and most of them predate crenulation cleavages. From a few tens of meters north of the cordierite-andalusite isograd and until the sillimanite zone, massive porphyroblastesis of cordierite and andalusite becomes a common feature in the pelitic layers (Fig. 14a). In some cases, the abundance and coarse grainsize (up to 7 cm) of disoriented porphyroblasts cause a partial obliteration of the preexisting schistosity, with the development of a new rough fabric.

The typical mineral assemblages detected in this zone are: Bt-Cd, Bt-Cd-And, Bt-And and Bt-Gt (Fig. 15a), being the last one relatively less abundant and restricted to psammitic layers. Biotites show little deviations in composition from those corresponding to the biotite zone, with a small decrease of the Fe/Fe+Mg ratio, which is consistent with an increase in the metamorphic grade (Fig. 13).

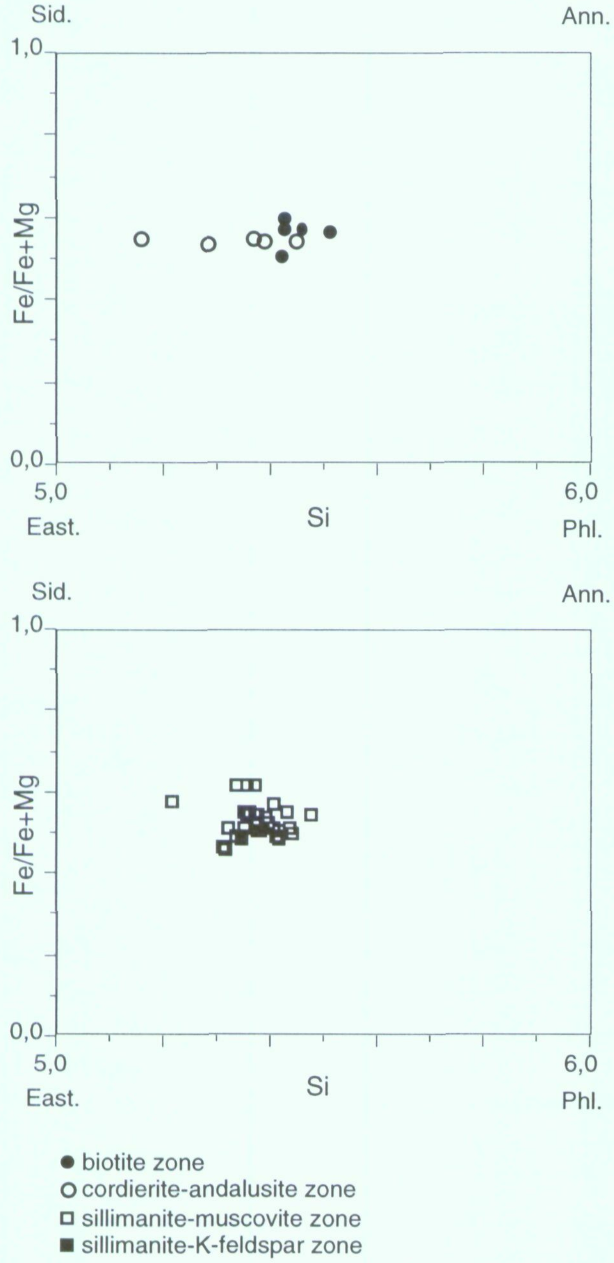


Fig. 13. Fe/Fe+Mg versus Si compositions of biotites from metapelites of the study zone.

Schists from the **sillimanite-muscovite zone** are first characterized by clusters of fibrolitic sillimanite associated to biotite and quartz. These minerals coexist with andalusite relics (Fig. 14b) in the southern half of the zone (i.e. Cala Portaló). Towards the north, andalusite becomes unstable and is altered to cordierite, fibrolite or muscovite, so that an assemblage Qtz-Pl-Bt±Ms±Cd-Sil appears (Fig. 15b). In the coarser grain schists from this zone, the schistosity tends to disappear at the microscopic scale, turning into decussate-like and granoblastic textures.

Major element compositions of three sillimanite schist samples are given in Table 1 (right columns).

Biotite compositions are shown in Fig. 13, displaying once more a trend marked by small decreases in the Fe/Fe+Mg ratio. A similar trend is recorded by cordierites from the cordierite-andalusite zone up to the sillimanite zone (BRGM-ITGE 1997).

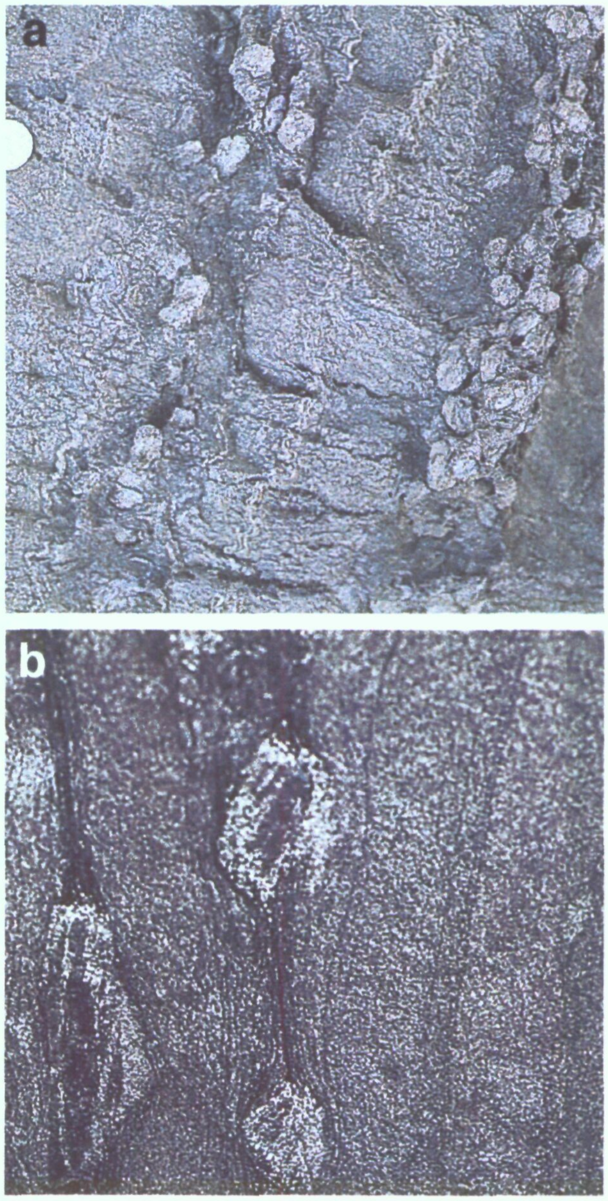


Fig. 14. Field photographs from schists of the Cd-And and sil-ms zones. (a): Porphyroblasts grown preferentially in metapelites. (b): Sillimanite growth forming reaction rims along the boundaries of andalusite porphyroblasts.

Garnet porphyroblasts appear sporadically in the sillimanite-muscovite zone. Some of the analyzed samples (Fig. 15c and Fig. 16) show prograde zonations (with Ca and Mn-rich cores and Fe and Mg-rich rims, together with a decrease in the Fe/Fe+Mg ratio), whereas others show retrograde zonations.

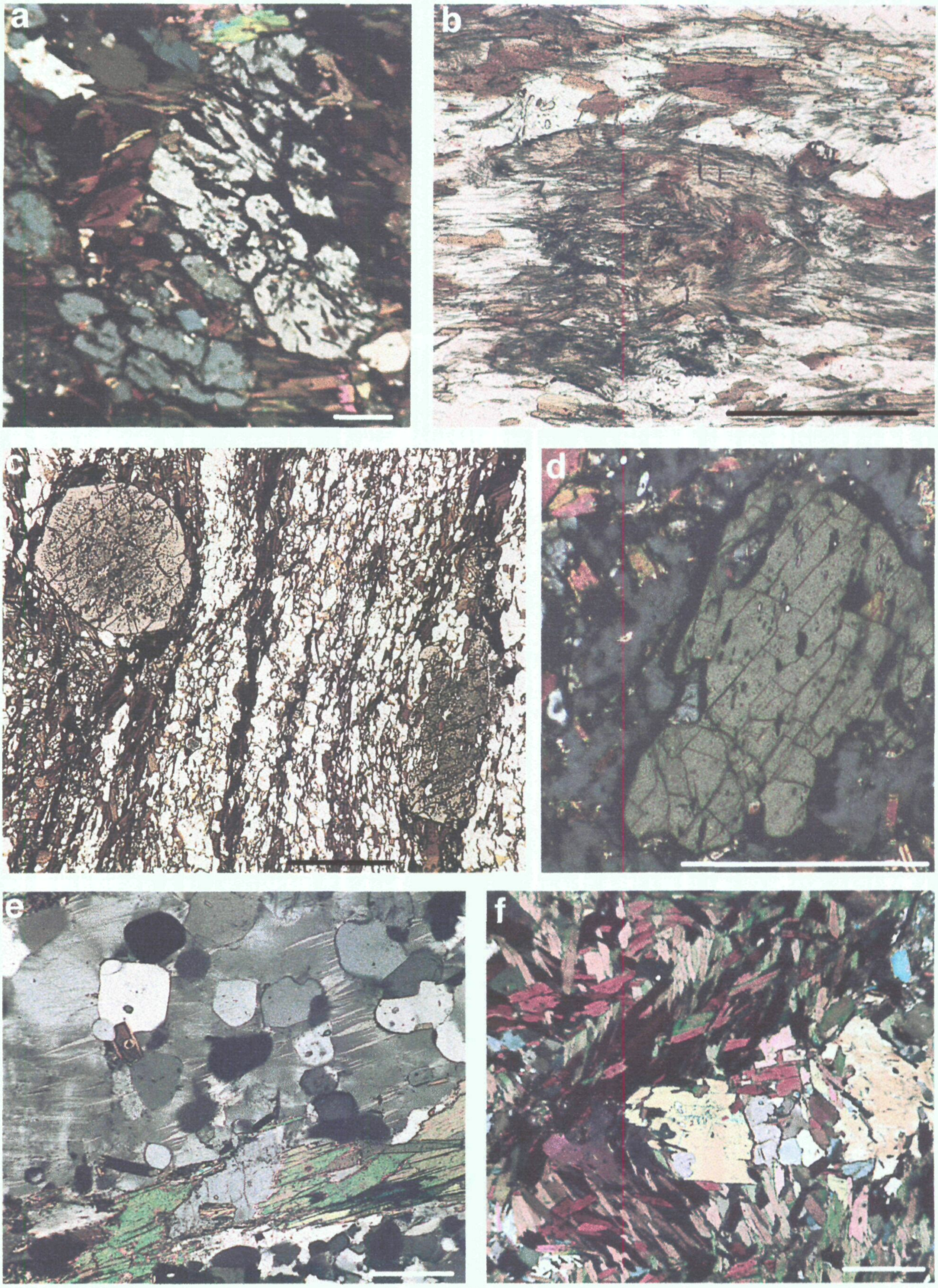


Fig. 15. Microphotographs from medium and high grade schists. Scale bar = 1 mm. (a): Cordierite-biotite schist. CPL. (b): Intergrowing sillimanite (fibrolite) and biotite. PPL. (c): Porphyroblasts of almandine garnet in a psammitic schist from the sillimanite-muscovite zone. Garnets show a weak zonation (prograde). PPL. (d): Staurolite relic in cordierite. CPL. (e): Microperthitic K-feldspar crystal in a schist from the sillimanite - K-feldspar zone. CPL. (f): Secondary muscovite in a micaschist from the sillimanite - K-feldspar zone. CPL.

According to Reche, in BRGM-ITGE (1997), three main types of garnet chemical zonation can be differentiated in the Cap de Creus. Type I is a prograde zonation, type II is retrograde and type III is characterized by prograde cores followed by retrograde rims. Implications for geotermobarometry will be discussed later.

Few relics of staurolite in andalusite or cordierite have been found in the cordierite-andalusite and in the sillimanite-muscovite zones (Fig. 15d). This fact might indicate the existence of an earlier metamorphic stage which included stable staurolite. The present metamorphic pattern could be, in this sense, a later prograde overprinting in lower pressure conditions.

The already mentioned presence of secondary kyanite in both the cordierite-andalusite and sillimanite-muscovite zones is always restricted to certain lithologies (i.e. the rusty schists adjacent to the Sant Baldiri complex). It is also found within quartz segregation veins, in areas of intense late folding.

The **sillimanite-K-feldspar schists** are sometimes evidenced in the field by the appearance of conspicuous clusters made of fibrolite±quartz±K-feldspar. These rocks are very coarse grained and can be named either high grade schists or paragneisses. K-feldspar crystals (Or_{87}-Ab_{13} of average composition) are xenoblastic, usually microperthitic and frequently display many inclusions of quartz and biotite (Fig. 15e). Biotites from this metamorphic zone are similar in composition to those in the adjacent sillimanite-muscovite zone (Fig. 13).

Secondary muscovite and chloritization of biotite are rather common in the sillimanite-muscovite and sillimanite-K-feldspar zones, and are especially in relation to areas with extensive intrusion of pegmatite dykes. Muscovite forms coarse randomly oriented flakes, reemplacing previous minerals or growing in pressure shadows (Fig. 15f). Many times these crystals are kinked by late deformations. Tourmalinization of the micaschists also occurs close to the pegmatite dykes. These fenomena had also been described by Allaart (1959) and Wickham (1987) in the Trois Seigneurs massif, and are attributed to hydrothermal activity related to the emplacement of pegmatites.

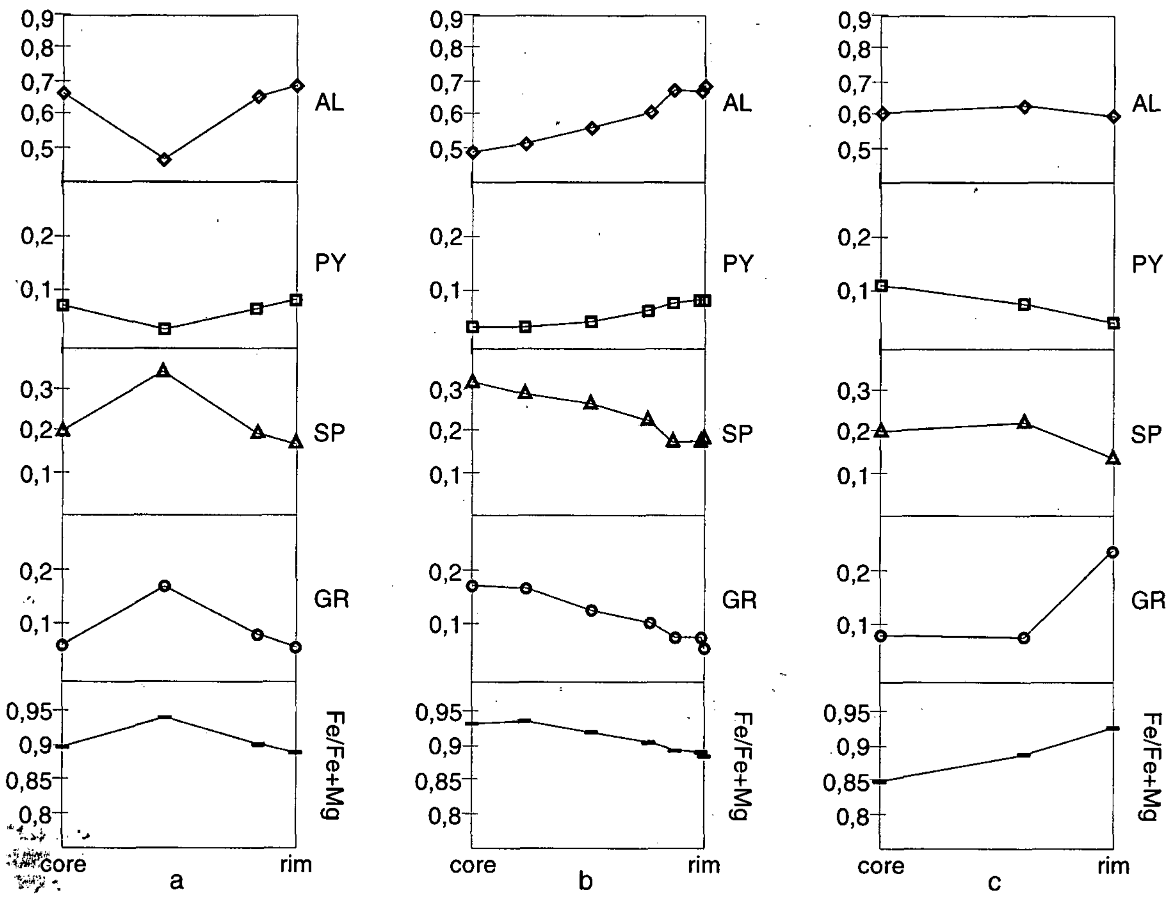


Fig. 16. Compositional diagrams of garnets of the sillimanite-muscovite zone. (a) and (b) show prograde zonations and (c) shows a retrograde zonation. AL, PY, SP and GR are the almandine, pyrope, spessartine and grossular components.

3.2.2. INTERLAYERED ROCKS: PLAGIOCLASE-AMPHIBOLE ROCKS, QUARTZITES AND THE SANT BALDIRI COMPLEX

Thin interlayers of plagioclase-amphibole rocks in the metasedimentary sequence are very common and have been found from low to high grade metamorphic zones. They bear perceptible amphibole crystals displaying "bow tie" textures (Fig. 17), getting an appearance of the so called amphibolites "en gerbes" (Capdevila 1969). The amphibole type varies with the metamorphic grade, although cummingtonite types dominate. Plagioclase varies between bytownite and anortite. Biotite usually accompanies this mineral assemblage, and Ca-rich almandine garnet is relatively frequent. Epidote and sphene are common accessory minerals, and secondary calcite can also be present. Although the chemical analyses indicate a basic composition (Table 2), the protoliths of these rocks could be either volcanoclastic sediments (ashes), carbonate bearing psammites or marls. Capdevila (op. cit.) attributed similar rocks from the Precambrian of NW Spain to original sandstone protoliths with carbonate and clay-rich matrix. However, in Cap de Creus, the presence of these rocks in very low metamorphic domains suggests that the protholith was already an anortite rich rock, and supports their volcanic origen. Furthermore, their rather uniform composition and the fact that they have never been found associated with marbles prevents to assign these rocks to a pure sedimentary origin.

Other metamorphic rocks present in the metasedimentary sequence are quartzites (Rabassers and Culip types) and various lithotypes of the Sant Baldiri complex. The black streaks interbedded in the Rabassers quartzite type usually contain garnet as an accessory mineral. A great variety of new minerals developed in the rocks of the Sant Baldiri complex, as a consequence of metamorphism, especially those of calc-silicate composition (see Ramírez 1983a). It is important to point out that the marbles from Mas de la Birba have wollastonite and that the black schists nearby contain chloritoid.

Chemical analysis of a leucogneiss from the Sant Baldiri complex is shown in Table 2. It is more acid in composition than the Port de la Selva gneiss, and might correspond to a leucogranitic or riolitic protolith.

Fig. 17. Microphotograph of a plagioclase-amphibole rock, with amphiboles displaying "bow-tie" microstructure. Width of view 7.5 mm. Plane polarised light.

3.2.3. THE PORT DE LA SELVA GNEISS AND THE METABASITES

The Port de la Selva gneiss is located within the cordierite-andalusite zone. It displays a blastoporphyric gneissic microstructure (Fig. 18), with relic phenocrystals of microcline and plagioclase (Ramírez 1983a). According to several analyses (Table 2) it has a granitic to quartz monzonitic composition.

Fig. 18. View of the Port de la Selva gneiss, displaying a blastoporphyric texture.

Rock	Gn-P			LGn	PA	Mg
SiO_2	65,3	65,4	65,23	69,52	52,6	51,49
TiO_2	0,52	0,32	0,37	0,89	1,66	1,36
Al_2O_3	17,33	16,9	17,06	14,76	14,67	13,89
Fe_2O_3	2,77	2,43	2,6	2,0	12,03	12,64
MnO	0,04	0,04	0,05	0,02	0,22	0,17
MgO	1,41	1,34	1,4	1,54	5,39	6,39
CaO	3,05	2,75	3,28	2,17	6,41	8,08
Na_2O	5,23	5,37	4,92	2,87	4,19	4,19
K_2O	3,13	3,16	3,38	3,12	0,46	0,32
P_2O_5		0,1	1,12	0,25	0,26	0,22
H_2O	0,72	0,68	0,8	2,47	1,91	1,07
Total	99,5	98,53	99,21	99,61	99,8	99,82
	1	2	3	3	3	3

Table 2. Chemical composition of pre-Hercynian acid and basic rocks from the Cap de Creus. Gn-P: Port de la Selva gneiss; LGn: leucogneiss (Sant Baldiri Complex); PA: plagioclase-amphibole rock; Mg: metagabbro; 1: after Ramírez (1983a); 2: after Druguet (1992); 3: after Navidad and Carreras (1995).

The metabasites (metagabbros or metadolerites) from Muntanya Negra correspond to greenschists, while those located around Puig Alt Petit, in the cordierite-andalusite zone, are amphibolites (located in Fig. 5). Metagabbros display diablastic or lepido-nematoblastic textures (Fig. 19b), overprinting original diabasic textures. Coarse-grained pegmatoid differentiates are occasionally preserved (Fig. 19a). Their mineral assemblages vary depending on the achieved metamorphic degree. Those in the cordierite-andalusite zone are composed by plagioclase (partially albitized) and zoned amphibole with a ferrotschermakitic hornblende core and an actinolitic margin, transformed in some instances into clinozoisite (Navidad and Carreras 1995). Accessories are ilmenite, allanite and sphene. One analysis of metagabbro from Puig Alt Petit is shown in Table 2.

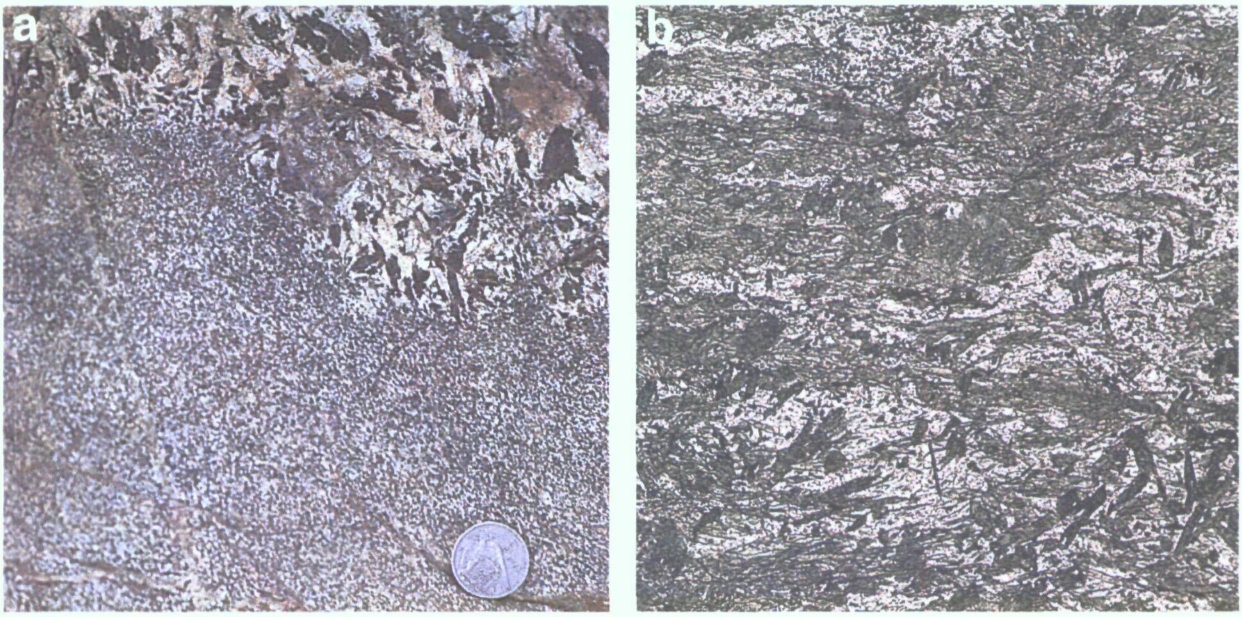


Fig. 19. Photographs of amphibolites from Puig Alt Petit. (a): Field photograph of an amphibolite with a pegmatoid differentiate in the upper side. (b): Microphotograph of an amphibolite displaying lepido-nematoblastic texture. Width of view 8 mm. Plane polarised light.

3.2.4. MYLONITES

High strain deformation in late shear zones produced a retrograde dynamic metamorphism under greenschist facies conditions, with consequent microstructural and mineralogical transformations. Retrograde mylonitic bands are more easily identifiable in medium and high grade metamorphic zones, because of the drastic grain size reduction in regard to the non-mylonitic equivalents and also because of the preferred development of these structures in more crystalline rocks. Mylonitization can affect the metasedimentary sequence, the interlayered metabasites and orthogneisses, and all the Hercynian intrusives.

There is abundant literature on this mylonitization, especially devoted to microstructural and microfabric aspects, and mostly concentrated on quartz and quartzfeldspathic mylonites (Carreras 1974, 1979; Carreras et al. 1975, 1977; Carreras & García 1982; García 1982, 1983; Guofei et al.1994; Norton 1982; Carreras et al. 1997a, b, c).

Mylonites usually display a banded structure which is mainly a consequence of huge stretching of existing lithological heterogeneities within the metasedimentary sequence. The mechanical behaviour of each mineral phase is the same irrespective of the rock type. In all mylonites, quartz forms granoblastic aggregates of new grains produced by a combination of intracrystalline slip, polygonization and primary recrystallization. New quartz grains tend to conform ribbons, preserving their granoblastic polygonal microstructure and developing a strong cristallographic preferred orientation (see Carreras & García 1982). Feldpars tend to remain as porphyroclasts which are deformed by twinning, bending and fracture. Micas are usually bent, have experienced a grain size reduction, and show a preferred orientation parallel to the mylonitic fabric. This is achieved by a combination of intracrystalline deformation, fracture and new crystal growth. Relic micas develop into mica-fishes. Accessory minerals as tourmaline and garnet remain as rigid porphyroclasts preserving an euhedral shape, even after large strains. Cordierite is totally altered into pinite, forming stretched aggregates.

Taking into account the high ductility, deformation of these rocks is accomplished essentially by intracrystalline deformation in quartz and phyllosilicates. The rocks which contain a major proportion of these two minerals become easily sheared and record larger deformations. By contrary, feldspar-rich rocks show little intracrystalline plastic deformation and, as a whole, they undergo less mylonitization, with frequent development of relatively undeformed lozenges or pods wrapped by the mylonitic foliation. New minerals in mylonites are those typical of greenschist facies: albite, chlorite, muscovite, epidote, clinozoisite and occasionally biotite and garnet.

In addition to shear induced stretching, complex strain histories in phyllosilicate-rich mylonites often lead to oyster-shell structures.

From a petrological point of view Carreras et al. (1975) distinguished three main mylonite types:

Mylonitic schists and phyllonites: They derive from metapelites and mica-bearing metagreywackes, with chlorite, muscovite ± biotite ± garnet as neoformed minerals (Fig. 20a). Feldspar-rich schists give rise to fine grained mylonites with augen-shaped microporphyroclasts. Usually, these mylonites are banded, some of the bands being quartz ribbons derived from quartz segregation.

Quartz mylonites: These are mylonites derived from quartz segregation veins in the schists, mainly made up of microcrystalline quartz aggregates (Fig. 20b). They usually form very thin bands, interlayered in mylonitic schists. Due to the high strain and the size of original quartz veins, quartz mylonites are usually of millimetric-scale thickness, reaching a maximum thickness of a few centimeters. Some quartz mylonites are derived from the Rabassers or Culip quartzites, and these may be thicker than those derived from the segregation veins. When the mylonite derives from pure quartz, it consists of a granoblastic polygonal aggregate, with steady-state grain size of 100 μm. The presence of an additional phase (e.g. phyllosilicates) results in grain size reduction by difficulting grain growth.

Quartzofeldspathic mylonites: These are mylonites derived from granitoids, pegmatites and from the Port de la Selva gneisses. They display a gneissic to porphyroclastic mylonitic microstructure, with augen-shaped feldspar porphyroclasts and quartz ribbons wrapping around them (Fig. 20c).

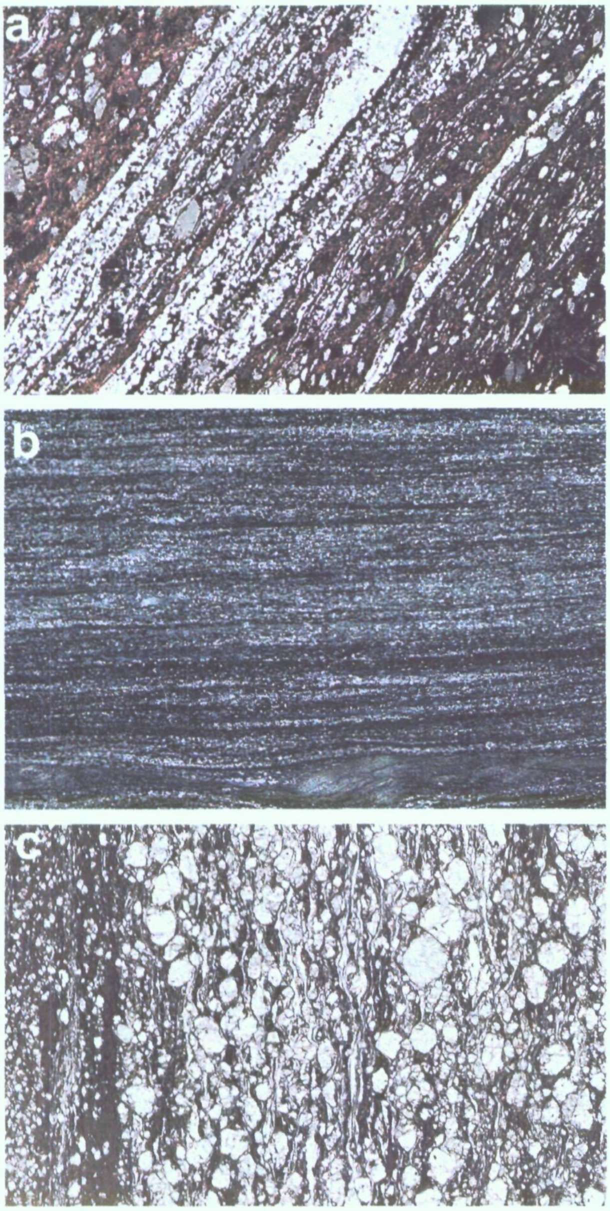


Fig. 20. Microphotographs of different types of mylonites. (a): Mylonitic schist with quartz ribbons. Width of view 20 mm. CPL. (b): Quartz mylonite. Width of view 40 mm. CPL. (c): Mylonitic tonalite. Width of view 40 mm. PPL.

An additional type of mylonites can be distinguished. It corresponds to the greenschist mylonites derived from the occasional mylonitization of metabasites.

A preliminary petrological study of the metamorphism associated to mylonitization has been done for mylonites from Cala Serena shear zone (Bossière et al. 1995, 1996). A new mylonitic paragenesis is distinguished by the generation of green biotite, garnet, plagioclase, muscovite and Mn-ilmenite. The chemical evolution of biotite is characterized by a correlation between Fe and Mg, indicating an increase of Fe/Fe+Mg ratio (Fig. 21). Relic almandine-rich garnets tend to reequilibrate towards the composition of new idiomorphic garnets, which are poorer in almandine component. New muscovites are slightly richer in paragonite than the relics.

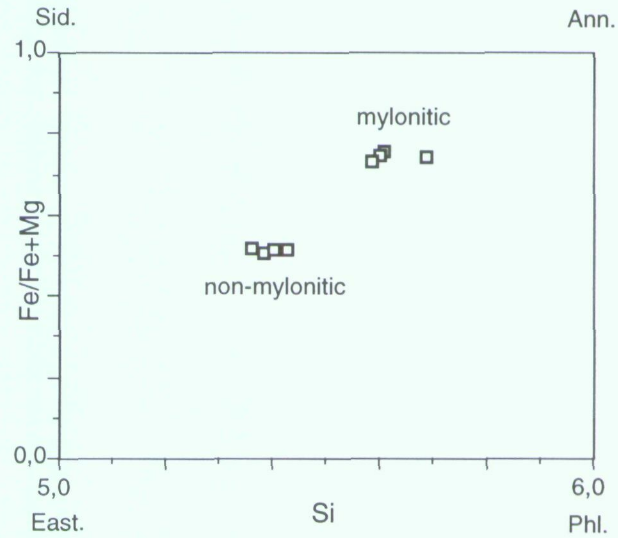


Fig. 21. Fe/Fe+Mg versus Si compositions of biotites from the Cala Serena shear zone.

3.3. MIGMATITES AND MAGMATIC ROCKS

As introduced above, there are three migmatite complexes, with associated granitoid bodies, in the northern Cap de Creus peninsula. Two of these complexes lie in the study area: the Punta dels Farallons and the Tudela complexes. The pegmatite dyke swarm (also called pegmatite area) occupies a more extensive area, from the cordierite-andalusite zone to the north (Fig. 11).

3.3.1. THE PEGMATITE DYKE SWARM

A swarm of predominantly sub-vertical pegmatite dykes extends over the northern part of the Cap de Creus peninsula (Fig. 11). As explained before, pegmatites are intruded into the metasediments from the cordierite-andalusite zone to the migmatite complexes, defining an approximately E-W trending band nearly parallel to the trend of the metamorphic zones. Pegmatite bodies have variable sizes, from centimetric veins to hectometric dykes up to 100 m wide and 200 m long. Dykes are larger and more abundant in higher grade metamorphic zones, especially within the migmatite complexes, where they occupy 15-20% of the total outcrop surface.

The pegmatite area has been the object of several studies, especially devoted to petrological and mineralogical aspects. Because of their close relationship to the metamorphism, they have been considered as perianatectic pegmatites (Carreras et al. 1975). In fact, the presence of pegmatites and leucogranites in perianatectic domains is a common feature in other zones in the Hercynian basement of the Pyrenees (Autran et al. 1970). The anatectic origin of the Cap de Creus pegmatites was later corroborated by Damm et al. (1992) and Pau (1995), who interpreted them as derived from the anatexis of a metapelitic source. Moreover, Corbella (1990) and Alfonso et al. (1995) have revealed the presence of four mineralogical types of pegmatites which comprise a zoned pegmatite field. There is a clear relationship between zoning of pegmatites and metamorphism, with Types I and II outcropping in the migmatite and sillimanite zones, and types III and IV occurring in the cordierite-andalusite zone. However, according to these authors, zoning of pegmatites responds to a fractionation trend from a source magmatic body that might be presently situated offshore, north of the study area. A similar zoned pegmatite field has been described in the Albera Massif (Malló et al. 1995). In this case the authors propose an origin by differentiation of anatectic muscovite-biotite leucogranites.

In the study area, the pegmatites in the sillimanite and higher grade zones are peraluminous and leucogranitic in composition, and are located close to the composition of the minimum ternary Qtz-Ab-Or. On the other hand, the pegmatites in less metamorphic areas move away from these compositions, being more albitic. Their K-content decreases progressively in a south direction, probably due to a more important hydrotermal participation. A proposed reconciling interpretation would situate the leucogranitic magmas (i.e. type I pegmatites and leucogranites close to the migmatite complexes) as resulting from partial melting of the pelitic schists in zones of high grade metamorphism and migmatization, subsequently differentiated into type II to IV pegmatites.

Major element oxide compositions of several pegmatites are shown in Table 3.

Rock	P H-K		P M-K		P L-K	
SiO_2	73,1	74,24	73,46	74,45	73,12	73,18
TiO_2	0,1	0,08	0,08	0,08	0,1	0,1
Al_2O_3	14,19	14,13	14,78	14,62	16,3	15,62
Fe_2O_3	0,21	0,67	0,7	0,4	0,15	0,23
MnO		0,02	0,14	0,06		
MgO	0,18	0,09	0,05	0,04	0,17	0,17
CaO	0,46	0,19	0,13	0,23	0,34	0,46
Na_2O	4,38	3,29	4,82	4,73	6,45	6,94
K_2O	5,08	5,5	3,72	3,73	1,46	1,42
H_2O	0,55	0,6	0,46	0,36	0,87	0,84
Total	98,27	98,81	98,34	98,7	98,96	98,96

Table 3 Chemical composition of pegmatites from the Cap de Creus. P H-K: high-K pegmatites; P M-K: medium-K pegmatites; P L-K: low-K pegmatites. After Druguet et al. (1995).

In addition to quartz, albite and K-feldspar, muscovite and biotite are present in most pegmatites. Other peraluminous minerals, such as aluminium silicates (andalusite and/or sillimanite), *cordierite and garnet are also locally present* (Fig. 22a). Internally, most pegmatites have complex textural relationships among different sized minerals. An internal compositional banding parallel to the border of dykes is frequent in the largest pegmatites (Fig. 22b, c). Tourmaline is abundant at dykes boundaries, forming tourmaline-rich rims, developing from boron-rich fluids which invade the wall rocks.

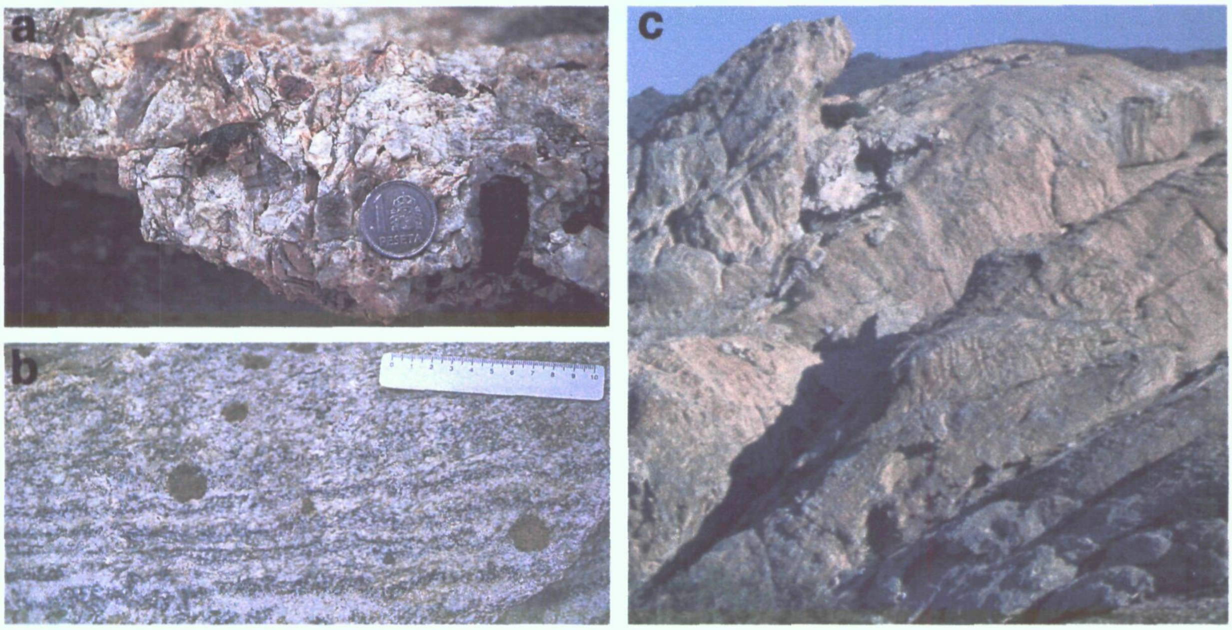


Fig. 22. Field photographs of pegmatites. (a): Garnet, cordierite and tourmaline-bearing pegmatite vein. Punta del Molí. (b): Garnet-bearing pegmatite displaying a compositional banding. Canal Guillosa. (c): Large pegmatite body with a marked compositional banding. Tudela.

3.3.2. THE MIGMATITE COMPLEXES

The larger variety of magmatic types is present in the Punta dels Farallons migmatite complex. In the Tudela complex there is only one granitoid body (quartz dioritic to tonalitic in composition), surrounded by locally migmatized sillimanite-K-feldspar schists (Fig. 23a) and some large bodies of pegmatite (Fig. 11 and Structural Map).

In what follows, the Punta dels Farallons migmatite complex will be described. It is separated by late shear zones into two outcrop areas: Punta dels Farallons-Volt Andrau and east Cala Serena. These complexes incorporate partially melted sillimanite schists and a calc-alkaline magmatic sequence, the latter consisting of small and heterogeneous granitoid bodies. The youngest magmatic types are leucogranites and pegmatites. The association of schists, granitoid intrusions and pegmatites, caused heterogeneous mixing at different scales. From both the geometrical and genetic points of view one may distinguish several types of migmatites.

Anatectic migmatites: There are two types of these: stromatic and nebulitic. The first are present in the pelitic layers, displaying a gradual northwardsl transition from schists with quartzofeldspathic veins (with associated plagioclase ± sillimanite) to banded or stromatic and phlebitic migmatites (Fig. 23b). Leucosomes are 2-5 mm in grain size and leucogranitic to trondhjemitic in composition (composed of quartz, plagioclase +/- K-feldspar +/- sillimanite). Melanosomes or mafic selvages are mainly formed by biotite, sillimanite and porphyroblasts of cordierite. Biotites from these migmatites are similar in composition to those from the sillimanite zone, with a Fe/Fe+Mg ratio around 0.5 (Fig. 24). Both parts, leucosome and melanosome, may contain considerable concentrations of almandine garnet (Fig. 23c). Individual crystals show retrograde rims (Fig. 25). Locally, garnet-rich restites are developed. They could be the result of partial melting of biotite and sillimanite-rich schists through the reaction:

$$Bt + Sil + Qtz \;\; H_2O = (Crd) + Gt + L$$

Nebulitic migmatites are seen in the most psammitic layers, which gradually lose their metamorphic structure developing into granodiorites (Fig. 23d). Two whole rock chemical compositions of migmatites are shown in table 4.