Folds and Shear Zones at Cap de Creus

Pliegues y Zonas de Cizalla en el Cap de Creus

Field Trip Guide

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Cover photograph: aerial view of the Cala Culleró area in the Northern Cap de Creus shear belt. A swarm of syntectonic pegmatites is intruded into schists in a domain of high-temperature and high-strain deformation. Schists and pegmatites were subsequently heterogeneously affected by a late retrograding shear zone network. Photo by Jordi Carreras.

PREFACE

Located in the NE part of the Iberian peninsula, Cap de Creus represents the easternmost outcrops of the Pyrenees. It forms a peninsula in the Mediterranean and contains the most singular landscapes of the Costa Brava. Geological, biological and historical-cultural elements converge on the Cap de Creus peninsula to shape a site of great scientific and landscape value. The outstanding character of this site is mostly achieved from its geological configuration and from the Mediterranean clime, which is here extremely windy (tramuntana), giving rise to many spectacular landforms and beautiful rocky coasts. Moreover, the main heritage value of this area concerns the quality and quantity of geological outcrops that enable the understanding of geological processes far beyond their regional interest.

Nowadays, and after 45 years of geological research, this area has become a worldwide reference place for structural geology. The exceptional quality of the structures exposed, together with that in the 1990's the Cap de Creus stood as one of the unique landscapes preserved in the Catalan coast, lead the Catalan Government to protect large areas of the Cap de Creus Peninsula, designating the Cap de Creus Natural Park (1998). However, some outstanding geological outcrops are located in areas that suffered from uncontrolled touristic development. This is the case of the Roses lighthouse area, which, due to inadequate legislation, was not included in the Cap de Creus Natural Park because of its closeness to urban areas.

Most of the themes which have been object of research based on Cap de Creus are related to shear zones and associated structures and mylonites. One of the first papers relating folds in a shear zone with shear sense came out in 1973 using examples from the Cala Prona - el Llimac shear zone. Later, accurate studies on sheath fold evolution were presented using examples from here. The first published example of shear bands was taken from the Cala Sardina shear zone. A better understanding of shear zones in foliated rocks was also based on Cap the Creus examples. In 1977, the relationship between fabric asymmetries and shear sense was first reported at the Leiden Conference, using quartz-mylonite fabrics from this area, and the subsequent discussion on c-axis quartz fabric rotations versus constant orientation with regard to the shear plane was based on published data from Cap de Creus. The present-day ongoing discussion of shear zones nucleating as buckling instabilities or evolving from brittle precursors is based, among others, on Cap de Creus examples. Cap de Creus has recently also furnished many examples of the problems associated with relating local strain with regional strain regime, and the complex relations between rock symmetries and asymmetries and kinematics. It has also provided important insights into the relationships between deformation, metamorphism, anatexis and magmatism, and the significance of some magmatic structures, such as the apparent boudinage of dykes and anatectic leucosomes. Another aspect that transcends the regional interest of the area is the well documented LP/HT metamorphism and granitoid emplacement in a transpressive regime.

In 2019, the Tectonics Commission of the SGE revisits the Cap de Creus 30 years after the 1989 *I Reunión de la Comisión de Tectónica*. The former excursion, entitled *Zonas de Cizalla y Milonitas en el Cap de Creus*, covered the two classical itineraries on shear zones at Cap de Creus: the Rabassers-Cala Prona in the Northern shear belt and the Roses lighthouse itinerary in the Southern shear belt. The field guide of that excursion has been redited in a *Geo-Guías* volume (Carreras and Druguet, in Díaz-Azpiroz *et al*., 2019) which compiles field guides of most of the past 30 excursions of the Tectonics Commission of the SGE. The present 2019 excursion aims to combine observations on shear zones with the examination of structures related to the pre-shearing event (non relevant in the 1989 excursion), focussing on high-grade folds and syntectonic intrusions. The first day itinerary to Puig de Culip - Cala d'Agulles includes outcrops unknown in the former excursion, and on the second day we will visit the outcrops at Roses lighthouse. In both itineraries, aspects of relevance and debate in current tectonic studies will be emphasized, such as the problem of strain localization and partitioning. This field guide is based on Carreras and Druguet (2013) *Illustrated Field Guide to the Geology of Cap de Creus* and on later published research (Carreras *et al*., 2013; Ponce *et al*., 2013; Druguet *et al*., 2014; Carreras and Druguet, 2019).

The first day excursion runs on the Park reserve, and in this area restrictions are applied. Sampling is prohibited and permission for a supervised sampling for scientific purposes can only be solicited to the Park authorities. For this excursion, a Park permission has had to be solicited for walking outside pathways. Such strict restrictions based on biodiversity protection are not the most adequate, but prevent the area being spoilt as a result of dense tourism. Some outcrops are very fragile and we should be very careful, non stepping on them. Enjoy the rocks and their structures, photograph them, but please leave the outcrops undisturbed in the Park.

1. GEOLOGICAL SETTING

1.1. THE PYRENEES

Alpine tectonics in NE Iberia gave rise to the Pyrenees, the Catalonian Coastal Ranges and the Iberian chain, due to the collision between Iberia and the European plate. The Pyrenees is a chain of Paleogene age which extends from the NW Mediterranean to the Bay of Biscay. The chain is part of a larger structure, developed along the boundary between the Iberian and the European plates, which extends westwards from the Alps to the NW end of the Iberian Peninsula. The Pyrenees arose from crustal shortening between the European and Iberian continental crust, resulting into a fold and thrust belt which involves the Mesozoic-Tertiary sedimentary cover and the Variscan basement.

From north to south the following WNW-ESE trending structural units have been distinguished (Figs. 1 and 2):

1. The foreland Aquitania basin.
2. The North Pyrenean zone, which includes cover sequences and basement rocks. It is thrust to the north over the Aquitania foreland and bounded to the south by the vertical North Pyrenean Fault (NPF), a sinistral strike slip fault developed during the Cretaceous (≈100 Ma) associated with anticlockwise rotation of Iberia.
3. The Axial zone, made up of Variscan basement consisting of Precambrian and Paleozoic rocks.
4. The South Pyrenean zone, a south vergent fold and thrust belt involving cover and basement rocks.
5. The foreland Ebro basin.

The Pyrenean chain has a fan-like geometry, defined by northward thrusting in the north and southward thrusting in the south, the North Pyrenean Fault zone being the axis of the fan (Fig. 2). This fan is asymmetric, with the south-directed thrusts more developed than the north-directed ones. The Axial zone belongs to the south-facing domain of the fan.

There are different interpretations on the effects of Alpine deformation on the basement rocks. However, there is a broad acceptance that penetrative ductile structures, metamorphism and igneous activity pertain to the Variscan orogeny.

Neogene tectonics was responsible for the formation of predominantly NE-SW trending faults and tectonic grabens due to the western European extension (Fig. 1). These grabens are broadly parallel to the Catalonian Coastal Ranges and cut obliquely across the Pyrenees (e.g. the Cerdanya graben).

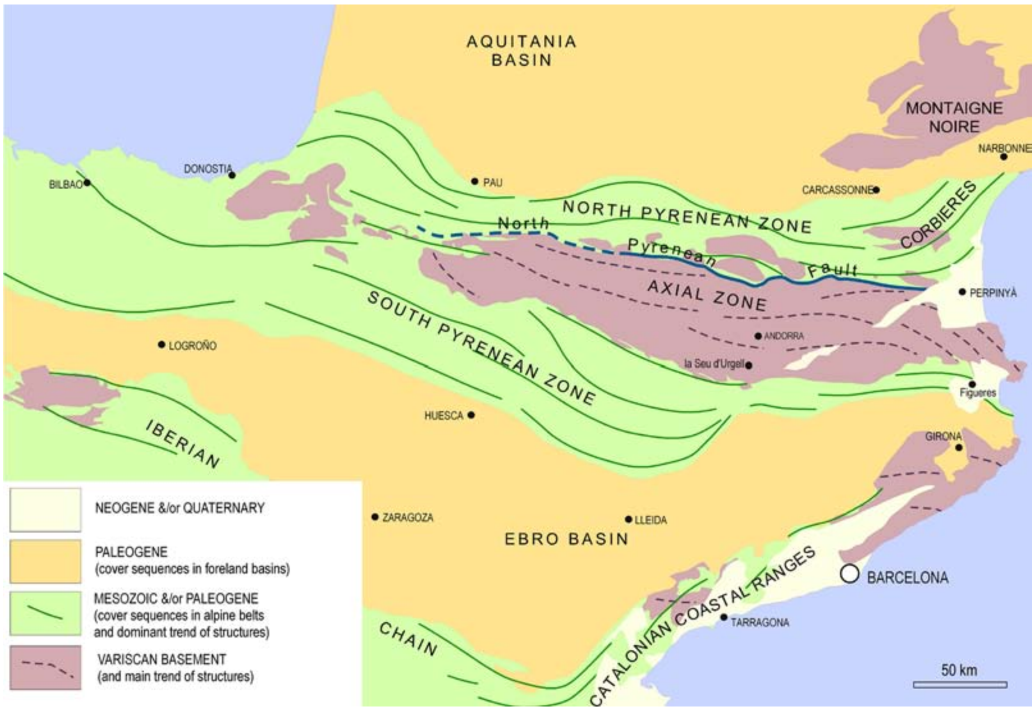


Fig. 1. Main geological units in NE Iberia and the Pyrenees (after Carreras and Druguet, 2013).

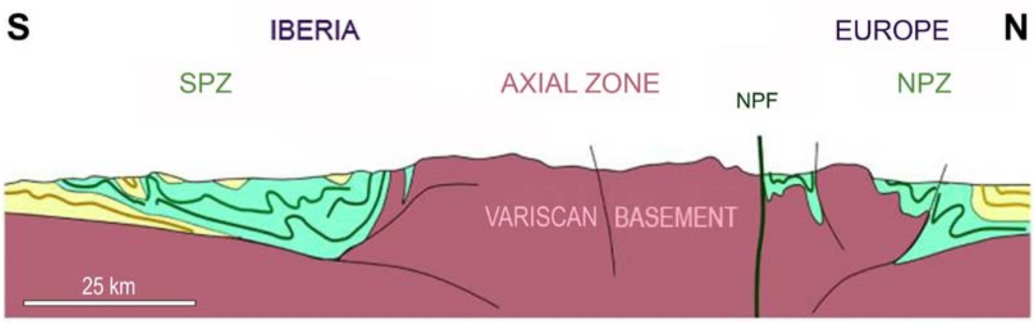


Fig. 2. Simplified cross-section of the Pyrenees (Colours according to legend in Fig. 1; NPZ: North Pyrenean Zone, SPZ: South Pyrenean Zone, NPF: North Pyrenean Fault). After Carreras and Druguet (2013).

1.2. VARISCAN OF NE IBERIA

The main lithological units of the Variscan basement of the NE Iberia are:

Sedimentary sequences

The upper part comprises sedimentary rocks from Upper Ordovician to Westfalian (Carboniferous), which are relatively well known from a chronostratigraphic point of view. The lower part consists of rather monotonous detritial sequences of unknown age. An unconformity between the Upper Ordovician and the underlying series has been reported (Santanach 1970; Casas and Fernández, 2007). The lower sequence is usually referred in the literature as Cambro-Ordovician, but geochronological data (Castiñeiras *et al*., 2008) indicate that this series also comprises Neoproterozic sequences.

Pre-variscan igneous rocks

Intrusive and volcanic rocks appear interlayered in the sedimentary pile. Magmatism is bimodal with predominance of acidic compositions. Concerning their ages, both Neoproterozoic-Early Cambrian (580-540 Ma) and Ordovician (475-460 Ma) events have been recognized (Castiñeiras *et al*., 2008). The Ordovician event is responsible for the emplacement of sheet-like bodies of granitoids transformed into orthogneisses during the Variscan orogeny. These gneiss bodies were formerly interpreted by Guitard (1970) and coworkers as derived from a Cadomian (Panafrican) granitic basement.

Variscan intrusions

These are mainly granitoids emplaced as sheet-like batholiths intruded into the metasedimentary pile. They range in composition from gabbros to leucogranites, with granodiorites being prevalent. Their ages range from 280 to 314 Ma (Guitard *et al*., 1996; Paquette *et al*., 1997; Roberts *et al*., 2000; Vilà *et al*., 2005; Olivier *et al*. 2008; Aguilar *et al*., 2014; Druguet *et al*., 2014). Batholiths are mainly located in low metamorphic grade shallow levels with development of a contact metamorphic aureole. At deeper levels migmatites appear in association with minor bodies of granitoids (from tonalites to granites) and often a widespread swarm of leucogranites and pegmatites.

Variscan events comprise: (i) polyphase tectonics with widespread foliations development, (ii) a LP/HT polyphase metamorphism and (iii) mainly syn-tectonic igneous activity.

Variscan tectonics

It is polyphasic, with tectonic style varying in space and time (Carreras and Capellà, 1994; Carreras and Druguet, 2014). Early tectonic events are responsible for thrusting in shallow levels and low dipping penetrative foliations in deeper levels. Main Variscan tectonic events took place under a transpressive regime, giving rise to folding and associated penetrative foliations with variable attitudes. In deeper tectonic levels foliations associated to the main phase appear as crenulation and/transposition foliations and occur close to the thermal peak of metamorphism. In shallow levels structures associated to the main phase appear as axial plane slaty cleavages or penetrative crenulation foliations. Late tectonic phases, developed under metamorphic retrograde conditions, gave rise to shear zones in highly crystalline rocks (granitoids, orthogneisses or high grade schists) and to crenulation foliations in lower grade rocks.

Variscan metamorphism

Variscan metamorphism in the Pyrenees is a classic example of low pressure, high temperature metamorphism. A widespread feature in different massifs is the distribution of prograde metamorphic zones, with large areas of low or very low grade metasediments bounding amphibolite to granulite facies metamorphic and migmatite cores. Most of these zones are concentric and dome-shaped, thus called thermal domes. These thermal domes often display an orthogneissic core, as in the Aston, St. Barthélemy, Canigó and Albera massifs. it was suggested that this gneissic cores acted as channels which caused and controlled the metamorphic gradient (Guitard, 1969; Fonteilles and Guitard, 1977). However, other massifs are cored by migmatitic schists, as in the Trois-Seigneurs and Cap de Creus. In these cases, the migmatite cores are usually associated with small funned-shaped intrusions of mantle-derived magmas. A high thermal gradient metamorphism is found in the interval between the beginning of the amphibolite facies and the beginning of anatexis. A metamorphic gradient of between 65 and 75°C/km is inferred from regional studies for the entire Variscan massifs (Zwart,1979; Guitard *et al*. 1996; Druguet, 2001).

Variscan igneous activity

This consist on a widespread emplacement of batholiths and stocks. Most plutons are sheet-shaped and emplaced in relative shallow levels. Granite to granodiorite compositions dominate although in many intrusions the composition range from minor gabbros to leucogranites. The gabbros, diorites and tonalites are the earlier and deeper seated intrusions, while leucogranites are the latest and occupy the apical parts of the intrusions. In deep seated levels, where medium to high metamorphic grade prevails, migmatites and associated granitoids appear as rather exiguous spots, commonly referred to as migmatite complexes. There, swarms of anatectic peraluminous leucogranites and pegmatites abound (referred to as perianatectic leucogranites by Autran *et al*., 1970). They are different from those occupying the apical parts of the large batholiths, which represent the ultimate differentiates. A comprehensive description of granitoids in the Eastern Pyrenean segment of the Variscides is given by Autran *et al*. (1970). Field and AMS studies of these granitoid plutons have lead to conclude that their emplacement was contemporaneous with the main Variscan transpressive event (Evans *et al*., 1997; Gleizes *et al*., 1997; Auréjac *et al*., 2004, Olivier *et al*., 2016). Migmatites and crustal anatexis leucogranites and pegmatites are also syntectonic with transpression (Druguet and Hutton, 1998; Druguet 2001).

2. CAP DE CREUS

The Cap de Creus peninsula forms the most easterly outcrop of Variscan basement exposed along the Axial Zone of the Pyrenees.

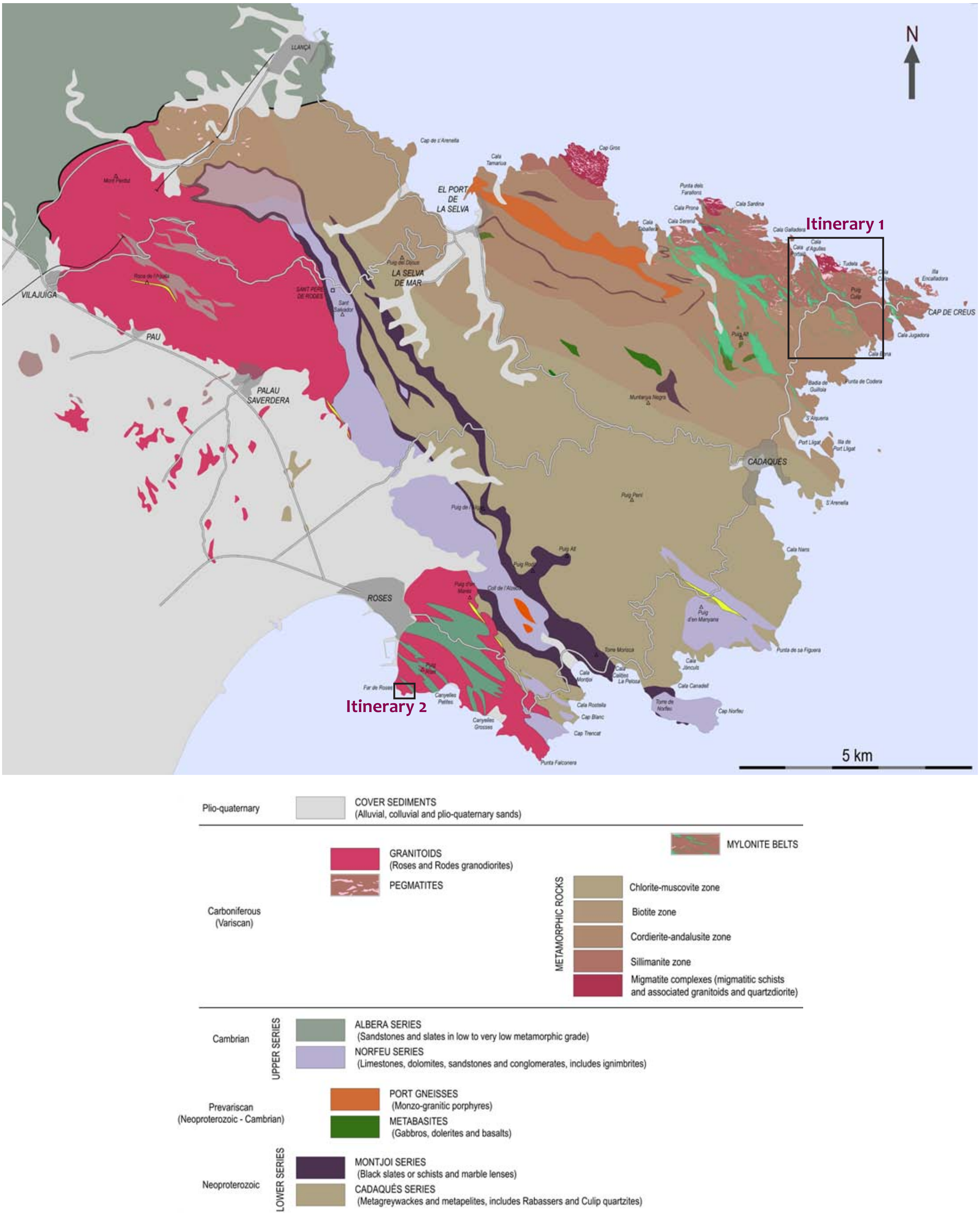


Fig. 3. Geological map and of the Cap de Creus peninsula (after Carreras and Druguet, 2013).

2.1. LITHOLOGICAL UNITS

In the Cap de Creus peninsula two main lithological units can be delimited (Fig. 3):

I) An essentially psammitic-pelitic monotonous metasedimentary sequence of uncertain age, but probably Neoproterozoic to lower Cambrian, that includes pre-Variscan igneous intercalations, and

II) Two small stocks of granitoids (mainly granodiorites) which form the Rodes and the Roses Massifs.

2.1.1. The metasedimentary sequence and the pre-Variscan igneous intercalations

The lowermost metasedimentary sequence (Cadaqués Series, Fig. 4) consists of a thick and rather monotonous succession of alternating metapsammites (metagreywackes) and minor metapelites, scarce intercalations of quartzites and abundant thin layers of plagioclase-amphibole rocks. The quartzites form distinct layers whose thickness range between a few centimetres and a few meters. They are well banded and may appear either predominantly dark (named Rabassers quartzite) or light coloured (named Culip quartzite). A Neoproterozoic age (Ediacaran) is assumed for the Cadaqués Series on the basis of the age of intrusive porphyries dated as 580-540 Ma age (Castiñeiras *et al*., 2008).

Towards upper stratigraphic levels, the metasediments become gradually darker and more pelitic, until they form a well distinguishable unit of black slates, phyllites and schists with minor marble lenses, which are known as the Montjoi Series, after a locality in the SE of the Cap de Creus peninsula. The ensemble including Cadaqués and Montjoi Series forms the Lower Series in the Cap de Creus area (Fig. 4).

The Lower Series are unconformably covered by a siliciclastic-carbonate series, known as the Upper or Norfeu Series, outcropping essentially in the south-eastern corner of the Peninsula and along a NW-SE trending band bounding the Roses and Rodes granodiorite massifs. These series lie unconformably over the lower series and include calcareous and dolomitic marbles and conglomerates (Fig. 4). On the basis of correlation with similar rocks from near areas, a lower Cambrian age is assumed for these series.

Sporadically, an intercalation known as the Sant Baldiri Complex appears in the Cap de Creus Series. This is heterogeneous in both thickness and composition and it is mainly made of black schists (carbon-rich metasediments), calc-silicate rocks, marbles, white quartzites, leucogneisses, and amphibolites. This complex is interpreted as tectonic slices of the Upper Series.

Two main types of pre-Variscan igneous rocks are present in the area: acidic and basic, both affected by later Variscan events and transformed respectively into gneisses and metabasites. The gneisses predominantly correspond to the Port de la Selva gneiss (Fig. 3) and to some thin lenses of leucocratic gneisses located in the upper part of the Montjoi Series. The protolith of this orthogneiss has a granitic to quartz-monzonitic composition. It has an age of ≈553 Ma (Castiñeiras *et al*., 2008) and has an intrusive character in the Cadaqués Series. The metabasites form discontinuous bodies located at different positions within the Cadaqués Series (Fig. 3).The largest ones are those of the northern slopes of Muntanya Negra and at the Puig Alt Petit summit. They were originally gabbro-dolerite intrusions, now forming lens-shaped bodies which have been transformed into greenschists and amphibolites due to Variscan metamorphism.

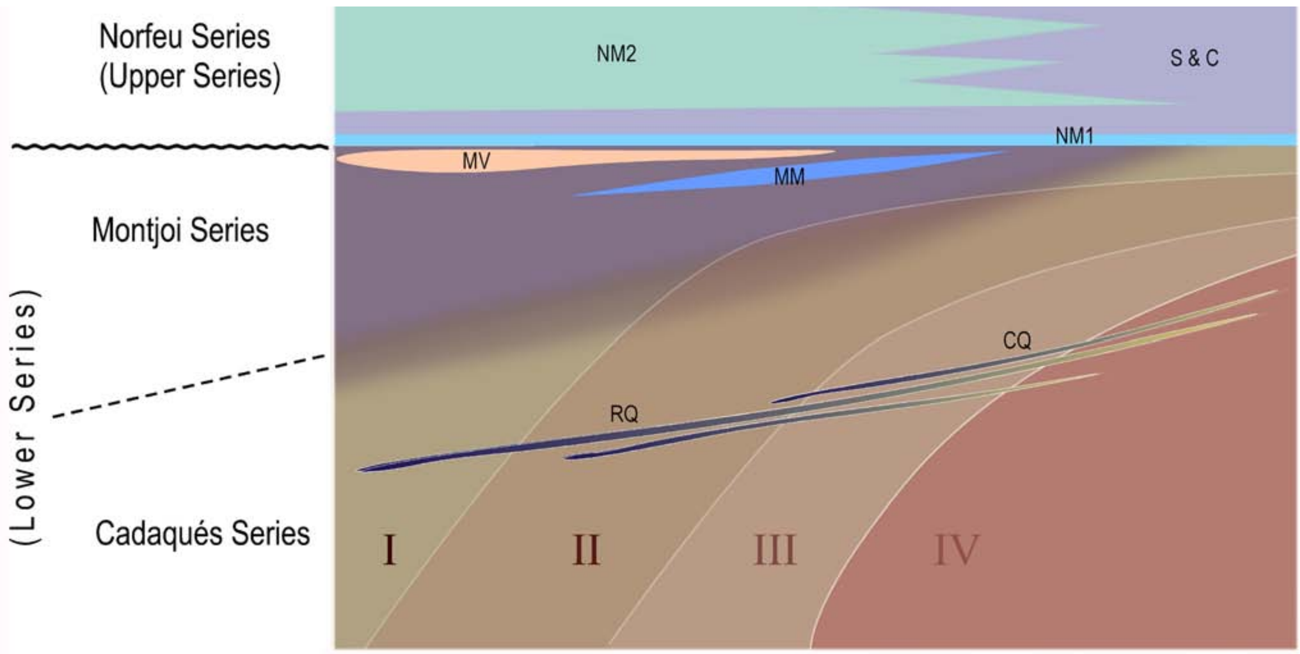


Fig. 4. The Cap de Creus series and metamorphic zones (after Carreras and Druguet, 2013).
RQ: Rabassers quartzite, CQ: Culip quartzite, MM: Montjoi marbles, MV: Montjoi volcanics, NM1: Norfeu marbles (basal layer) NM2: Norfeu marbles (upper marbles and dolomitic marbles), S & C: Norfeu sandstones and conglomerates.
I: Chlorite-muscovite zone, II: biotite zone, III: Andalusite-cordierite zone, IV: Sillimanite zone.

2.1.2. The Variscan intrusives

In the Cap de Creus, the Variscan igneous activity is represented by three groups, on the basis of the level of emplacement and the volume of the intrusions (Figs. 3 and 5): (i) the Roses and Rodes granodiorite stocks; (ii) a pegmatite dyke swarm and (iii) small intrusions of quartzdiorites and granitoids.

(i) Two major syntectonic stocks of Roses and Rodes, located south and southwest of the peninsula. 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. The granodiorites are rich in microquartzdioritic enclaves and display a magmatic foliation, being similar to other calcalkaline granitoid stocks and batholiths of the Variscan basement of the Pyrenees. The Roses granodiorite was dated as Early Permian (≈291 Ma) by Druguet *et al*. (2014).

(ii) From the cordierite-andalusite zone to the north, a swarm of anatectic pegmatite dykes forms the so called pegmatite area or perianatectic domain. Pegmatites have been dated as ≈299 Ma by Van Lichtervelde *et al*. (2017).

(iii) In the sillimanite zones of the northern highest grade metamorphic zone, there are small intrusions of quartzdiorites, tonalites and leucogranites, usually surrounded by small migmatite pods, the whole sets being called migmatite complexes. A quartzdiorite from the Tudela migmatite complex was also dated as ≈299 Ma by Druguet *et al*. (2014).

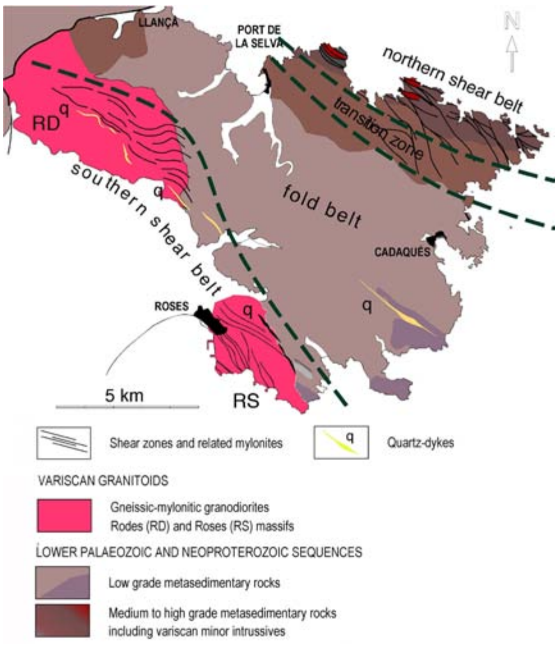


Fig. 5. Schematic map of the shear belts in the Cap de Creus peninsula (modified from Carreras, 2001).

2.2. THE STRUCTURE

The structure in the northern Cap de Creus area is mainly attributed to Variscan tectonics. The effect of Alpine tectonics is likely restricted to the overturning of the Variscan structures along the southern border of the peninsula (i.e. Roses, Fig. 6). Variscan structures arise from a complex polyphase tectonics, with at least two main deformation events during the prograde metamorphic episode and late folding and shearing events in retrograde conditions. The overall structure is essentially the result of late folds affecting ubiquitous penetrative foliations, but also results from the effect of shear zone-related mylonite belts, which cut across the previous folded foliations and also across the Variscan intrusives (Fig. 5). This structural arrangement is considered to be responsible for the distribution of different tectonic and lithostratigraphic levels in the Cap de Creus peninsula. Shallow levels consisting of Upper Series of very low metamorphic grade outcrop mainly on the SE corner of the peninsula and along a NW-SE trending zone bounding the Roses and Rodes granodiorites. On the other hand, deeper seated structural levels, consisting of the medium to high grade Cadaqués Series, are located along the northern side of the Peninsula (Fig. 3).

The Variscan structures can be ascribed to three main deformation episodes:

D_1 is responsible for a penetrative foliation where associated major structures are difficult to depict. The presence of recumbent folding and associated thrusting is presumabily associated to this deformation episode.

D_2 is characterized by a highly heterogeneous deformation with domains of very tight folds with an associated transposition schistosity and domains where D_2 is manifested by a light crenulation or even absent. D_2 developed closely in time with the metamorphic climax.

D_3 developed under retrogressive metamorphic conditions and is also characterized by a highly heterogeneous deformation with both folds and shear zones depending on the nature of the deforming rocks. The shear belts in Cap de Creus are interpreted as the result of D_3 deformation in domains where rocks achieved high crystallinity (granitoids and medium to high grade schists and associated intrusives).

Later Neogene tectonics is related to the development of a low angle extensional fault (Valleta fault) which separates medium grade metasediments and Rodes granitoids from the low grade Albera metasedimentary sequences (see Fig. 3).

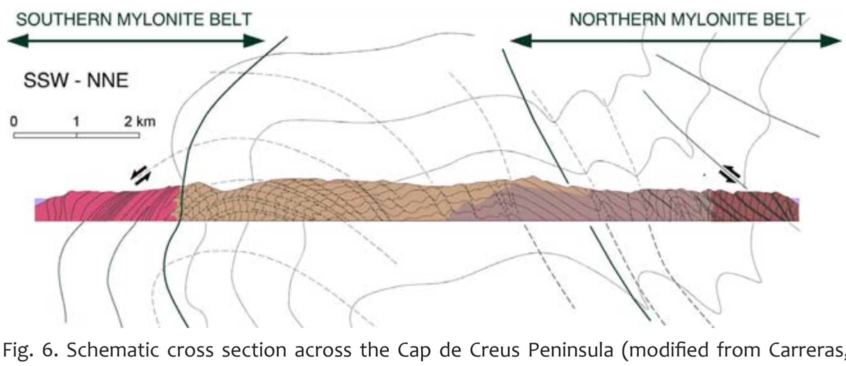


Fig. 6. Schematic cross section across the Cap de Creus Peninsula (modified from Carreras, 2001).

2.2.1. D_1-related structures

The oldest deformation recorded in the area (D_1) led to folding of bedding and to the development of a widespread first penetrative foliation (S_1 , Fig. 7), formed at prograde metamorphic conditions, prior to the metamorphic climax. S_1 is sub-parallel to bedding (S_s) almost throughout the entire area. No clear D_1 macrostructures have been recognized, although some repetitions of quartzite beds and the presence within the Lower Series of slices of rocks belonging to the Upper Series (Sant Baldiri complex) could be due to D_1 tight folds and thrusts. 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 east dip.

Fig. 7. S_1 Foliation and associated D_1 folds and intersection lineations in the Cadaqués metasedimentary sequence (Punta des Fosso, Cadaqués).

Fig. 8. D_2 folds and crenulations in high-grade schists. Left: low strain domain, south Puig de Culip. Centre: mid strain domain, east Puig de Culip. Right: High strain domain of tight folds and transposition foliation, Volt Andrau. All photographs correspond to sub-horizontal surfaces.

2.2.2. D_2-related structures

Later, heterogeneous D_2 deformation led to folding of S_s and S_1 , and also to folding of early segregated quartz veins (Fig. 8). In the northern part of the Peninsula D_2 folds are characterized by vertical or steeply inclined axial surfaces and an associated crenulation cleavage, trending approximately NE-SW in less deformed domains and about ENE-WSW in domains of intense D_2 deformation (Fig. 9). In highly deformed domains, the crenulation cleavages leads to the transposition of the previous S_1 foliation. D_2 folds often have steeply plunging axes parallel or subparallel to the stretching lineations. D_2 deformation took place about peak metamorphism, and lasted until the emplacement of the pegmatite dyke swarm.

The D_2 structural zonation applies to the entire area, with a generally low strain domain dominated by the S_1 fabric and weak D_2, occupying the south where it is associated with the lowest grade of metamorphism, and a generally high strain domain characterizing the north where it is associated with the high grade metamorphism (Fig. 9). These high strain zones coincide with voluminous pegmatite dykes, and locally with the granitoids and migmatites of the migmatite complexes. Traced from the less deformed domains towards the D_2 high strain zones, the N-S trending S_1 foliation together with the S_2 crenulation cleavage describe a km-scale dextral flexure, especially visible in the Culip area (see Itinerary 1), interpreted as being associated with rotational components of the D_2 deformation. Moreover, the stretching lineations, which are best developed in the high strain zones, also plunge moderately to steeply to the east or to the west, with the west direction being prevalent. Although the major structure is dextral in character, small scale asymmetric structures (e.g. deformed quartz veins and inclusion patterns in porphyroblasts), which are better developed on horizontal planes than on L_2-parallel steeper faces, show a prevalence of sinistral shear throughout the zone. The whole structure is interpreted as a complex transpressive shear zone involving vertical extension, NNW-SSE sub-horizontal bulk shortening with a dextral component, and bedding/S_1 parallel sinistral flexural flow.

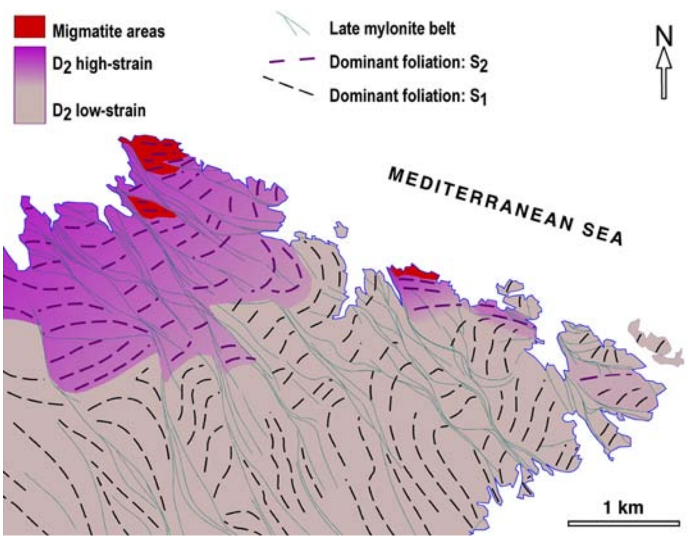


Fig. 9. Map of distribution of D_2 strain gradients in NE Cap de Creus (modified from Druguet and Hutton, 1998).

2.2.3. D_3-related structures

D_3 developed under retrograde metamorphic conditions and gave rise to E-W to NW-SE trending folds and shear zones both with a high degree of heterogeneity.

Folds develop preferentially in metapsammites and crenulations in metapelites (Fig. 10) and may display an associated crenulation cleavage that affects either S_{s-1} or S_2 or both.

Shear zones form predominantly where crystalline rocks (granitoids or medium high grade schists) were involved in deformation (Figs. 11 and 12). Shear zones affecting the Roses and Rodes stocks display predominant south dips and form the southern mylonite belt, while shear zones affecting medium and high metamorphic grade domains display north dips and form the northern mylonite belt (Figs. 5 and 6).

Fig. 10. D_3 folds affecting metapsammites and crenulations in metapelites with a previous layer-parallel foliation (S_{s-1}). Mas Rabassers de Baix.

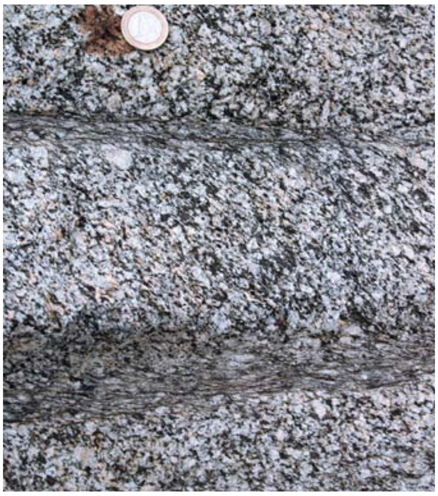


Fig. 11. Sinistral shear zones across the Roses granodiorite (Roses lighthouse).

Fig. 12. The Cala Prona shear zone consisting on a lower shear zone involving a dextral reverse shear with an associated displacement of ≈25 m, and an upper splaying shear with a rapidly decreasing displacement.

The Cap de Creus shear belts

Analogously to other mylonite belts in the Pyrenean domain (Carreras *et al*., 1980), the Cap de Creus mylonites form in bands related to prevailing ductile shear zones of millimetric to hectometric thickness. These are preferentially located across coarse crystalline rocks (i.e. orthogneisses, granitoids and medium-high grade metasediments).

In the Cap de Creus peninsula, two WNW-ESE trending mylonite belts exist (Fig. 5). The **Northern shear belt** cuts across medium to high grade schists and the **Southern shear belt** cuts across Variscan granitoids emplaced in low grade metasediments (the Rodes and Roses plutons). Both belts are separated by D_3-folded low grade metasediments (Carreras, 1975a).

A stretching lineation is usually well marked and its kinematic signification was well established by Carreras and Santanach (1973), later corroborated by quartz fabric studies (Carreras *et al*. 1977). The main attitude of this stretching lineation is quite constant not only in each belt but also across different belts of the Pyrenean realm. NW-SE directions prevail, with predominance of NNW-low plunges.

Sense of displacement across both belts can be deduced and corroborated by the presence of shear sense indicators inside the mylonite bands (Figs. 11, 15 and 16), and also by the inflexion of previously existing foliations (e.g. in schists, Fig. 12) or the newly formed shearing- related mylonitic foliation (e.g. in granitoids, Fig. 11). In some instances cut off and displaced dykes enable to determine the amount of bulk displacement on a single shear zone.

The northern and southern belts in the Cap de Creus peninsula are considered to be contemporaneous on the basis of similar relative timing with regard to the structural and metamorphic evolution, as well as of their similar geometry and kinematics. However the later effect of Alpine tectonics caused an overturning of structures in the south of the area (see Fig. 6) and, in consequence, the predominantly dextral shear zones appear as sinistral ones in the Roses granodiorite.

Shearing produces a drastic grain size reduction accompanied by a partial regressive mineral readjustment to greenschist facies assemblages. Mylonites are well foliated and commonly banded, with bands forming as a result of high strain of initial inhomogeneities.

The Northern shear belt

The medium to high metamorphic grade schists occupying the northern part of the Cap de Creus peninsula are cut by an anastomosing network of shear zones forming a WNW-ESE trending belt with predominantly dextral-reverse movement (Figs. 13 and 14). Individual shear zones vary in size, but most have a width ranging from a few centimetres to a few tens of meters. Broader shear zones of decametric width exhibit a composite structure being formed by anastomosing minor-sized shear zones. Individual shear zones are often discontinuous joining or dying with common splay geometry. The shear zones cut across schists which underwent a previous polyphase tectonics (D_1 and D_2) with development of inhomogeneously distributed penetrative foliations synchronous with a prograde LP-HT metamorphism. The metamorphic climax occurred during D_2 and, thus, prior to shear zone development. Metamorphic grade reached upper amphibolite facies conditions and local anatexis, followed by the widespread intrusion of pegmatite bodies (Carreras and Druguet 1994a). Shearing occurred in lower grade retrograde conditions (greenschist facies) and partial re-equilibritation of minerals is often observed in mylonites with growth of new biotite and muscovite, albite, epidote, chlorite and extensive recrystallization of quartz.

Mylonitic equivalents of schists and related igneous rocks are penetratively foliated and often banded, as the result of strong deformation of rather heterogeneous schists containing segregation quartz nodules and veins and locally intruded by pegmatites and small bodies of granitoids.

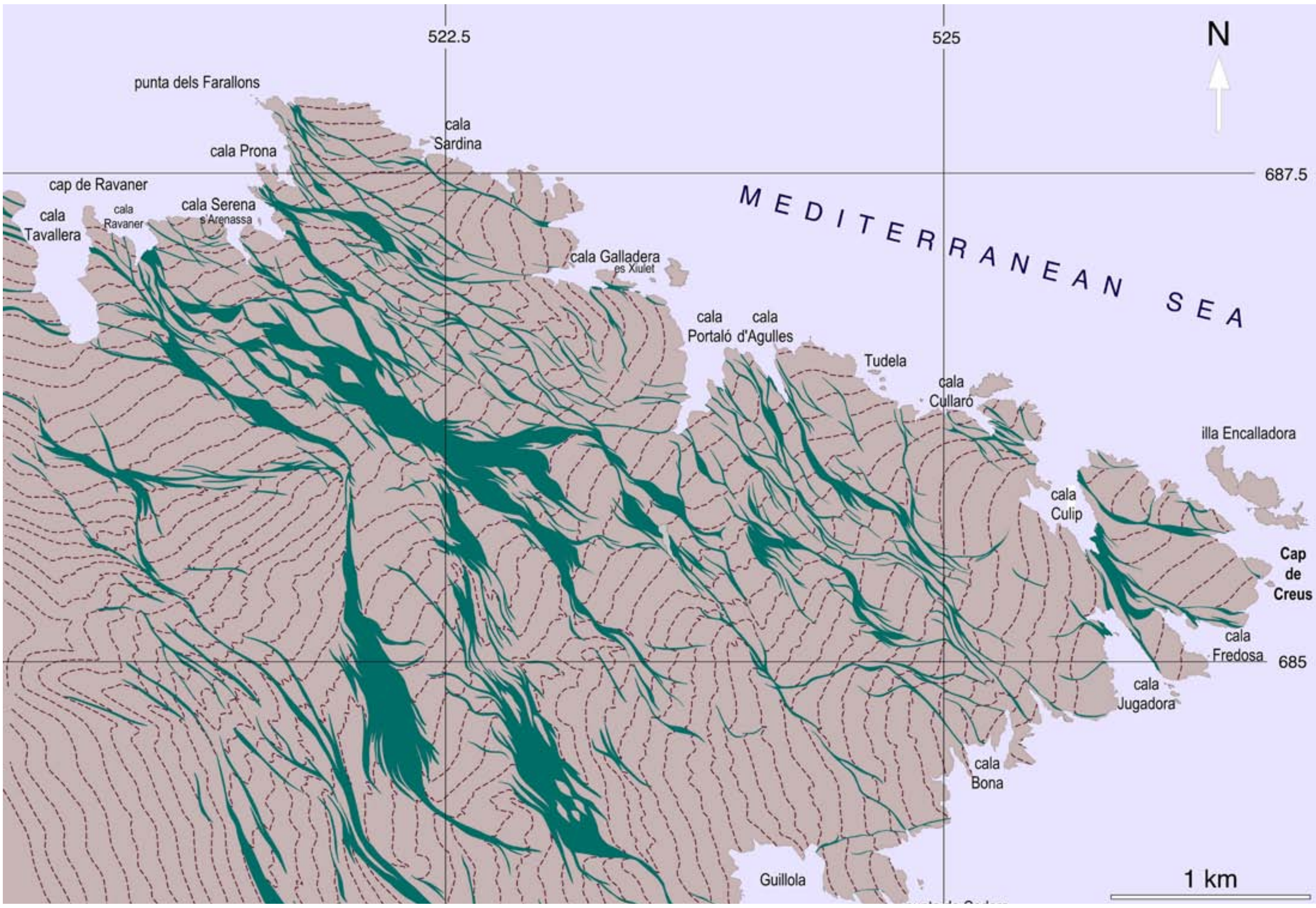


Fig. 13. Shear zones and transposition bands (in green) cutting across the schists and the associated dominant foliation: S_1, S_2 or $S_{1/2}$. Eastern part of the Northern shear belt (modified from Carreras, 2001; Grid refers to UTM coordinates).

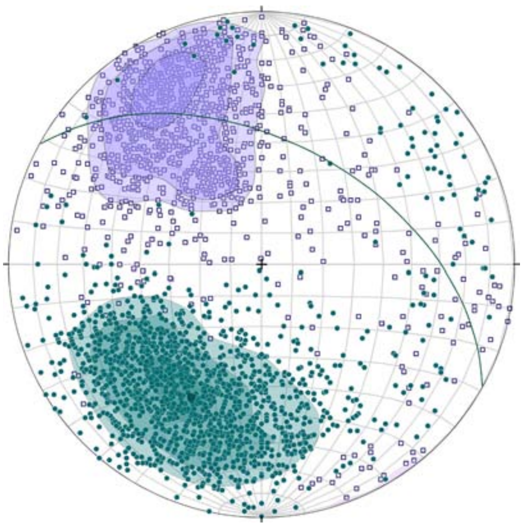


Fig.14. Stereogram of 1850 mylonitic foliation planes (cerulean dots) and 1339 associated stretching lineations (lavender squares) corresponding to the Northern Cap de Creus shear belt. Great circle corresponds to the mean foliation plane. After Carreras and Druguet (2013).

Fig. 15. Sheared pegmatite showing asymmetric pinch-and-swell. Note variation in mylonitization from the cores of the swells towards the margins an tails (Cala Serena ravine).

Fig. 16. Shear bands indicating dextral shearing along the Cala Sardina shear zone.

2.2.4. D_2- D_3 relationships

D_2 and D_3 structures can in most cases be mutually differentiated because they generate fold interference patterns due to superposition of deformation (Fig. 17). Another criteria for discriminating D_2 and D_3 structures is the marked difference in metamorphic conditions. However, in some areas, structures formed during an intermediate time-span can be found. These are labelled as $D_{2\text{-}3}$ structures (Druguet 1997). It is thus assumed that deformation was progressive from D_2 to D_3, and from prograde to retrograde metamorphic conditions (Fig. 18).

In a similar sense, not all D_3 structures seem to be contemporaneous, as, for instance, older E-W trending D_3 folds and shear zones are overprinted by later NW-SE trending dextral shear zones.

Fig. 17. D_2/D_3 fold interference pattern (Cala Prona).

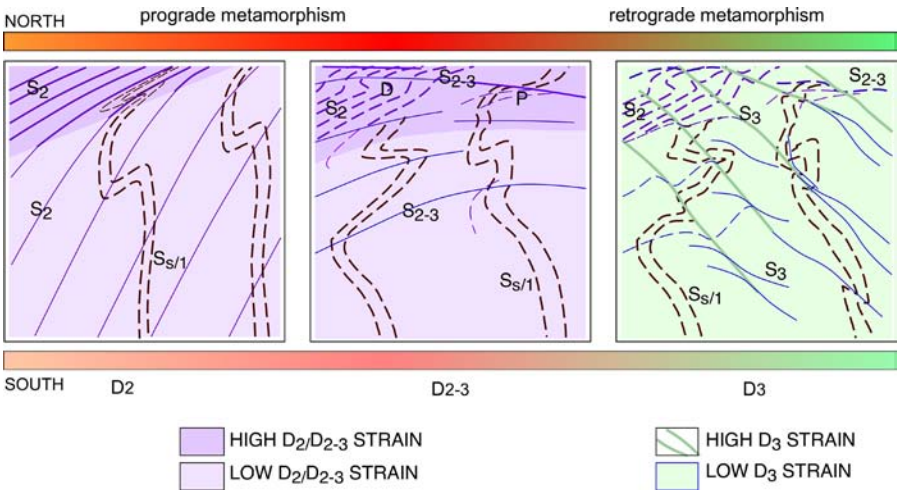


Fig. 18. Sketches of map views to show three stages in the progressive structural evolution from D_2 to D_3 in the NE Cap de Creus (after Druguet, 2001). Thick lines represent dominant foliations at each stage. Solid lines represent developing foliations and dashed lines are foliations being deformed. D_2 deformation took place during prograde metamorphism and D_3 occurred at retrograde conditions. An intermediate stage between D_2 and D_3 has different spatial correlations to metamorphism: while in the north it developed just around peak metamorphism, in the central and southern domains it developed at more clear retrograde conditions.

2.3. METAMORPHISM

The area is characterized by the presence of a metamorphic gradient increasing from south to north, which reflects a prograde low-P regional metamorphism, initiated during the early Variscan deformation events and affecting all the pre-Variscan lithologies. Over a horizontal distance of ≈5 km, the metapelites show a gradient from the chlorite-muscovite zone (greenschists facies) in the south to the sillimanite-K-feldspar (upper-amphibolite facies) and migmatite zones in the north (Fig. 3). Although prograde zoning reflects low-P mineral assemblages developed about the metamorphic peak (e.g. Fig. 19), medium pressures must have been reached according to: (i) the sporadic presence of relics of staurolite inside andalusite or cordierite, which would indicate an early stage of medium-P conditions, and (ii) the localized existence of kyanite pseudomorphosing andalusite and partially reemplaced by muscovite produced during a late metamorphic stage.

A superimposed retrograde metamorphism (greenschists facies) is strongly heterogeneous in distribution, preferentially developed along the mylonitic zones.

Few quantitative data on P-T conditions of metamorphism have been obtained by geothermometry (Druguet, 2001). The peak metamorphism P-T conditions vary from 670°C and 470 MPa (sillimanite zone) and 700°C and 740 MPa (Punta dels Farallons migmatite complex).

Fig. 19. Vertical section of schist with andalusite porphyroblasts (partially replaced by sillimanite and muscovite) showing straight inclusion trails of S_1. All porphyroblasts show a relative rotation (clockwise in this section) with regard to $S_{S/1}$, attributed to D_2 deformation.

2.4. MAGMATISM

The Variscan igneous activity in the Cap de Creus massif is represented by the three groups mentioned before (section 2.1.2, Fig. 3): The granodiorite stocks, the pegmatite dyke swarm, and the small intrusions from the migmatite complexes.

The **migmatite complexes** consist of partially migmatized sillimanite schists, small heterogeneous bodies of quartz-gabbros, quartzdiorites and granitoids, and a voluminous pegmatites dyke swarm (Fig. 20). The igneous rocks can be divided in two associations (Druguet *et al*. 1995): a mantle derived calc-alkaline association which includes quartz gabbros, quartz diorites, tonalites, granodiorites and granites, and a peraluminous association, anatectic in origin, which comprises leucogranites and pegmatites. The increase in temperature associated to the emplacement of the basic-intermediate magmas likely induced local anatexis, which may be the reason for migmatites being restricted to a few small areas around the intrusions. Three migmatite complexes has 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 complexes (Druguet 1997, 2001).

The **pegmatite dyke swarm** occupies a wider area, from the cordierite-andalusite zone to the north. The pegmatites are vein or dyke-like, but many have beaded (apparent boudinage) shapes, or occur in networks of foliation-parallel/foliation-oblique veins and dykes, suggesting bridge structures related to dyke emplacement (Druguet and Hutton, 1998; Bons *et al*., 2004; Fig. 20). Most pegmatites show a general E-W trend which is in close parallelism with the dominant S2 foliation in the country rocks. In the small number of cases where dykes have intruded at high angles to the dominant S_2 foliation they have been folded (Carreras and Druguet, 1994a). The folds of the dykes are coaxial with those in the country rocks but more open, indicating that some D_2 deformation had occurred before pegmatite emplacement and that this deformation continued after, i.e. the pegmatites are syntectonic.

Because its close relationship to metamorphism, pegmatites have been interpreted as derived from anatexis of a metapelitic source (Carreras *et al*. 1975), later corroborated geochemically by Damm *et al*. (1992). On the basis of mineralogical and textural criteria, four types of peraluminous, LCT (lithium-cesium-tantalum)-family pegmatites were distinguished (Corbella, 1990; Corbella and Melgarejo, 1993; Alfonso *et al*., 2003). Different types are distributed roughly parallel to the metamorphic zones, from the high-grade metamorphic zones in the north (types I and II) towards the cordierite-andalusite zone (Types 3 and 4). Type II pegmatites belong to the beryl-columbite subtype, type III to the beryl-columbite-phosphate subtype, and type IV belong to the albite subtype. Type I pegmatites are nearly sterile in LCT. In addition to quartz, albite and K-feldspar, muscovite and/or biotite, pegmatites from the high-grade metamorphic zones may have other peraluminous minerals such as Al-silicates (andalusite and/or sillimanite), cordierite and garnet. Tourmaline can be abundant, specially at the boundaries of dykes, forming tourmaline-rich rims developed from boron-rich fluids which invaded the wall rocks of the dykes.

In spite of this marked compositional heterogeneity, both the anatectic dykes and the igneous rocks from the migmatite complexes show close space and time links with metamorphism and tectonics. They all appear in medium to high grade metamorphic perianatectic domains with intense D_2 deformation, and were all emplaced during the D_2 transpressional event (Druguet, 2001). U-Pb geochronological data from quartzdiorite (Druguet *et al*., 2014) and pegmatites (Van Lichtervelde *et al*., 2017) thus constrain the age of D_2 tectonic phase to about 300 Ma.

The **Roses and Rodes granodiorite stocks** (Fig. 3) correspond to the emplacement of calc-alkaline magmas into low grade metasediments. They recorded a magmatic- to solid-state fabric which has been interpreted as a result of progressive deformation during cooling (Fig. 11; Carreras *et al*., 2004).

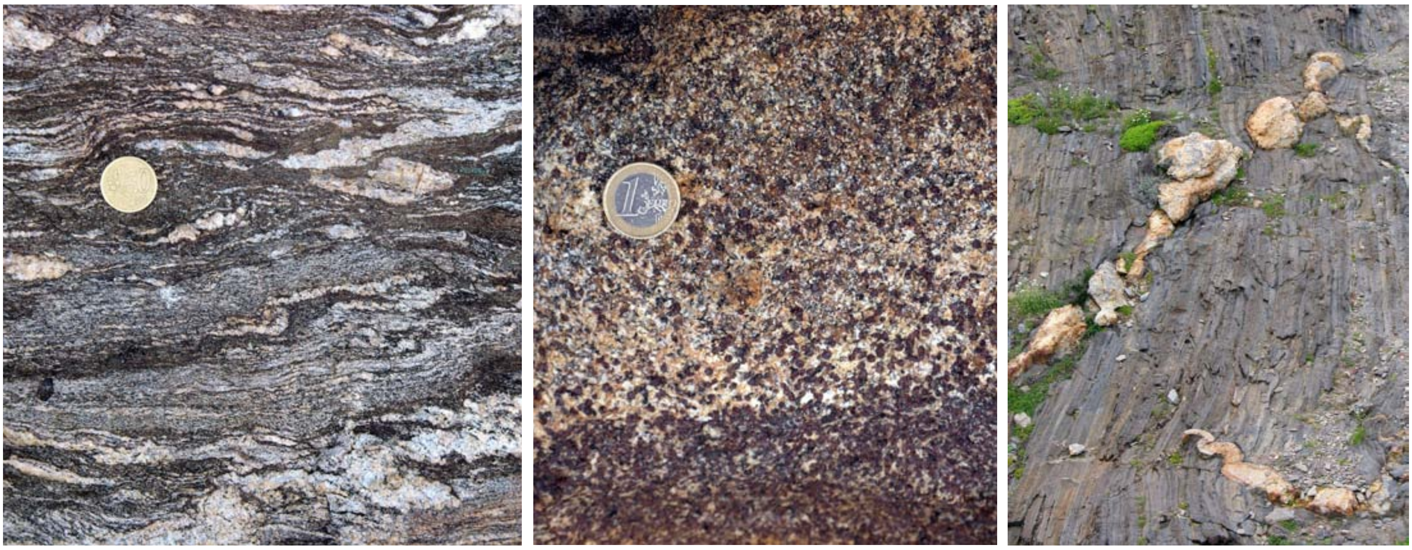


Fig. 20. Left: stromatic migmatites (with cordierite, sillimanite and garnet assemblages). These are affected by tight D_2 folds coeval with metamorphic peak and anatexis. These migmatites consist on alternating layers of leucosome, melanosome and mesosome. Punta dels Farallons migmatite complex. Center: Garnet (almandine)-rich tonalite from the Punta dels Farallons migmatite complex. Right: Apparently boudinaged (above) and folded (below) pegmatite dykes. Notice that layering is continuous across the uppermost pegmatite beads. Cap de Creus lighthouse area.

2.5. SUMMARY OF THE CAP DE CREUS VARISCAN EVOLUTION

The Cap de Creus massif is an illustrative example of the interaction between deformation, metamorphism and magmatism in an orogenic belt under a regional transpressive regime.

Metamorphism started to develop high gradients at the time that deformation also displayed high strain gradients, and intensification of deformation and metamorphism coincides in space and time with magmatism (Fig. 21). The interaction of these three processes would explain the high thermal gradients observed in the northern area (Druguet, 2001). The rocks now located in the north would have experienced a temperature rise due to the vicinity of hot magmas that would have risen towards upper levels and along high strain domains, possibly favoured by vertical extension. Zones of isometamorphism became rather steep and thin towards de core of the sub-vertical “high strain intrusion zone”. As deformation proceeded, just after the peak of metamorphism, horizontal shortening (>50% in domains of intense deformation) might account also for the narrowing of the isometamorphic zones. Progression of deformation during cooling is responsible for narrowing of the metamorphic zones. Taking into account the bulk structure of the Cap de Creus Peninsula and the geometry drawn for other eastern Pyrenean plutons, it is conceivable that the Roses and Rodes granodiorites were linked to the migmatite domains but lying at a higher, now eroded, structural level.

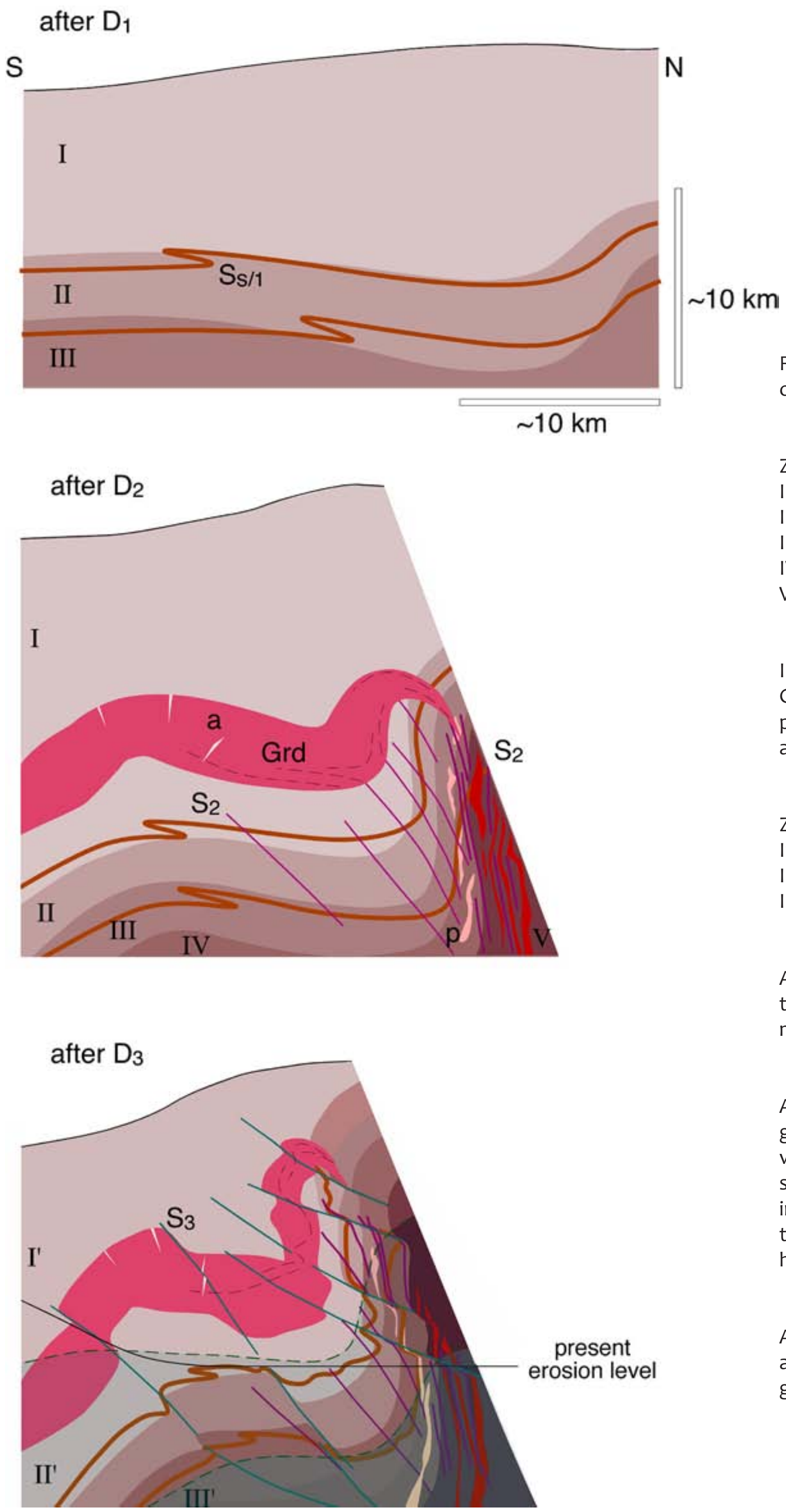


Fig. 21. Cross-sections of three ideal stages in the geological evolution of the Cap de Creus peninsula (modified from Druguet, 2001).

Zones of prograde metamorphism (in brown):
I non-metamorphic to chlorite-muscovite zone
II biotite zone
III cordierite-andalusite zone
IV sillimanite zone
V migmatite areas

Intrusive rocks:
Grd granitoids (migmatite complexes and Roses and Rodes stocks)
p pegmatites (anatectic and peri-anatectic)
a aplites (late differentiates in the granodiorite)

Zones of retrograde metamorphism (superposed greenish color):
I‘ non- or very low metamorphic
II’ lower greenschist facies
III‘ upper greenschist facies

After D_1: early deformations produce a gently diping N-S trending schistosity (apparently flat in this N-S cross-section) and onset of regional metamorphism.

After D_2: folding and developing of high strain zones in the north, with granitoid sheets and pegmatites intruded vertically and equally subvertical high thermal gradient. The Roses and Rodes granodiorites are speculated to intrude stratoidally in shallower levels, having their root in the migmatite areas; magmatic foliation (dashed lines). Regional transpression with a south vergence produces vertical extension and horizontal shortening (crustal thickening).

After D_3 : Rapid erosion and uplift causes exhumation of metamorphic and igneous rocks, while resetting the normal thermal gradient (in green).

3. FIELD TRIP ITINERARIES

FIRST DAY - Itinerary 1

PUIG DE CULIP - CULLERÓ - TUDELA - CLOT DE LA VELLOSA - CALA D'AGULLES:

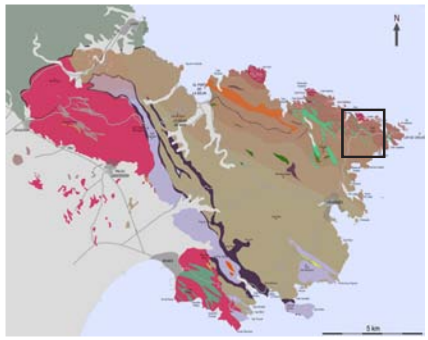
D_2 structures developed under medium-high grade metamorphism
Syntectonic granitoids and pegmatites
D_3 retrograde shear zones and strain partitioning

SECOND DAY - Itinerary 2

ROSES LIGHTHOUSE:

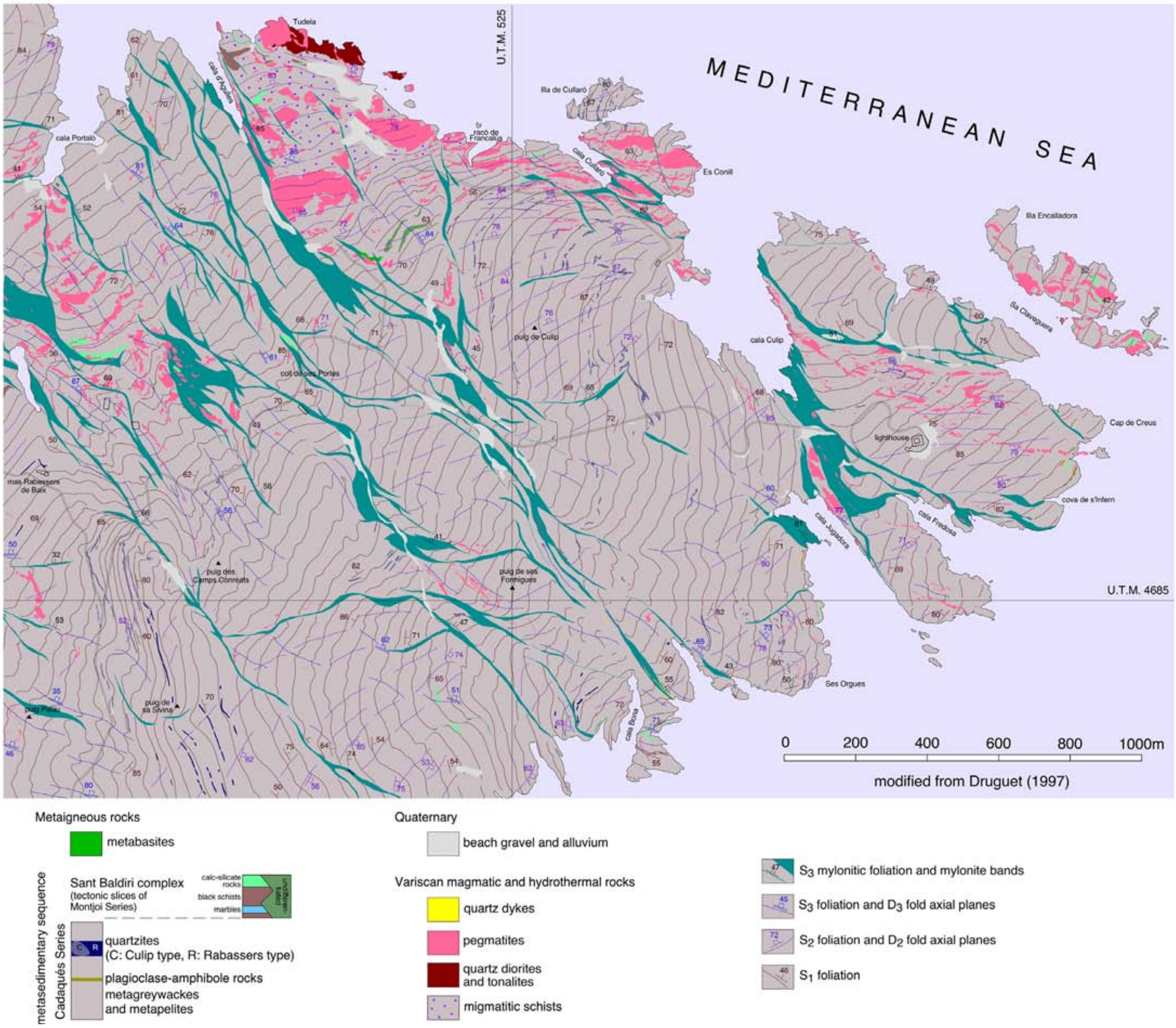
Progressive deformation history during and after pluton cooling
From D_2 magmatic structures to D_3 retrograde shear zones in the Roses granodiorite

Itinerary 1



PUIG DE CULIP - CULLERÓ - TUDELA - CLOT DE LA VELLOSA - CALA D'AGULLES:

D_2 structures developed under medium-high grade metamorphism
Syntectonic granitoids and pegmatites
D_3 retrograde shear zones and strain partitioning

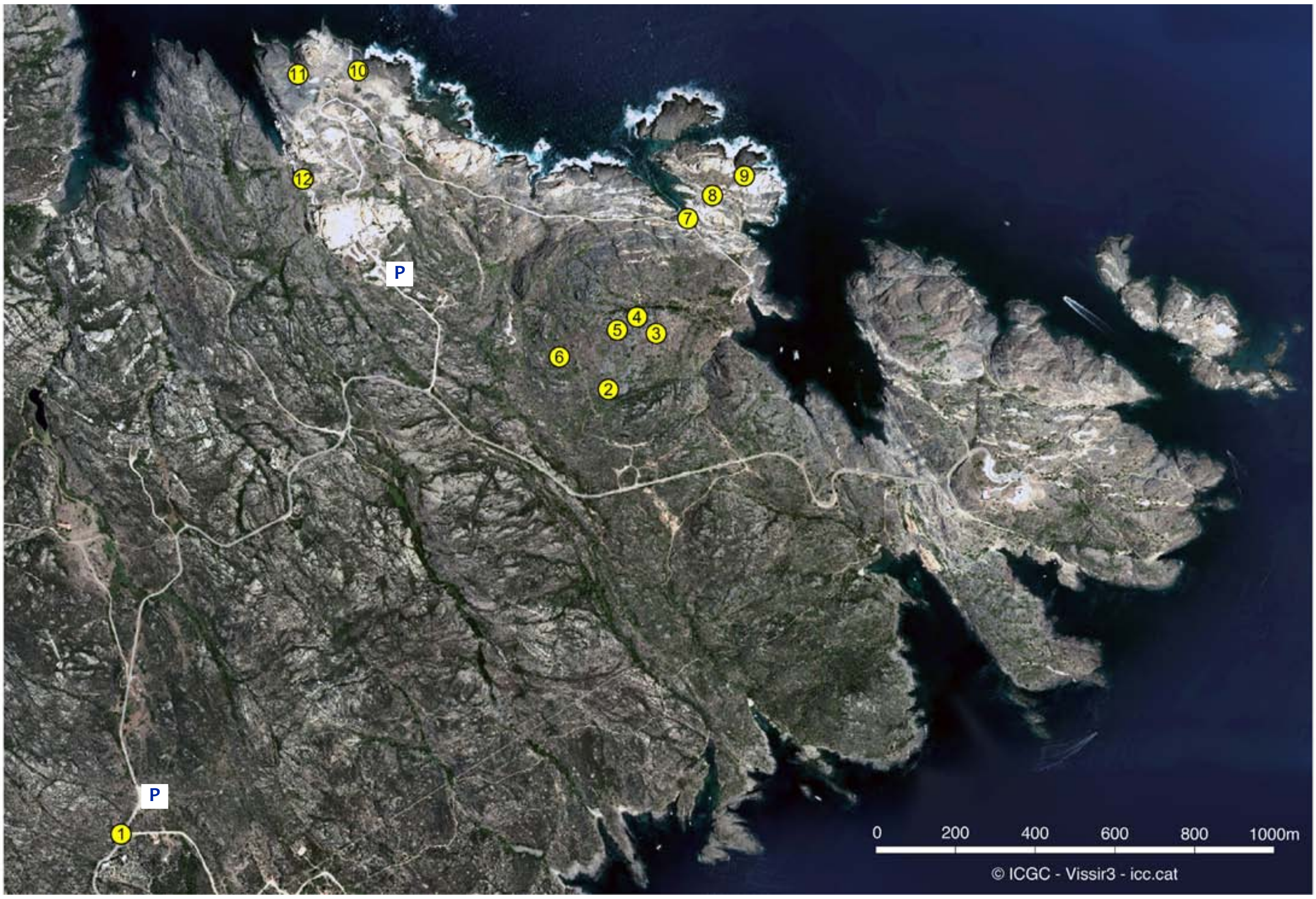


Geological-structural map of the northeasternmost segment of the Cap de Creus peninsula (modified from Druguet, 1997).

We will first drive from Cadaqués to **Locality 1 at Guillola crossroad (42°18'43.06"N , 3°17'18.98"E)**, which will serve for giving a brief geological introduction to the area and to observe some D_3 folding structures.

Then, we will drive to the **parking lot at Paratge de Tudela (42°19'26.20"N , 3°17'47.00"E)** from where we will start our day walk towards Localities 2 to 12. Localities 2 to 9 pass around Puig de Culip and Cala Culleró. The area is a good example of heterogeneous D_2 deformation. The outcropping rocks correspond to the metasedimentary sequence consisting of an alternance of metapsammites and minor metapelites, and some thin layers of light quartzites an plagioclase-amphibole rocks. The area is located in the medium- to high-grade metamorphic zone, mainly formed by sillimanite-bearing micaschists, although andalusite is still present. A strain gradient across the area defines, in horizontal view, a shear zone-like geometry, with two main structural zones of relatively high and low strain. High strain deformation is associated in space and time to the syntectonic emplacement of a pegmatite dyke swarm. The D_2 folds have sub-vertical or steeply plunging axes, which are closely parallel to lineations. Lineations are usually stretching lineations defined by the alignment of quartz grains or mineral lineations defined by sillimanite, tourmaline or biotite. These lineations indicate that the X axis of the finite D_2 strain is sub-vertical. The geometric relationships (steep $S_{S/1}$ and S_2 foliations, and sub-vertical fold axes and stretching lineations) imply that most features such folds, boudinage and asymmetric structures are better marked in flat-lying outcrop surfaces than in steep ones, i.e. the vorticity axis is sub-vertical. The whole D_2 structure is interpreted as a complex transpressive shear zone involving vertical extension, NNW-SSE sub-horizontal bulk shortening with a dextral component, and bedding-S_1 parallel sinistral flexural flow.

During subsequent D3 deformation, the metasedimentary rocks and pegmatites were affected by a late retrograding network of shear zones. Shear zones are rather heterogeneous in distribution, orientation and kinematics, giving rise to a complex anastomosing system. This will be observed in the magnificent outcrops of Cala Culleró (localities 7, 8 and 9). In Tudela-Cala d'Agulles area (localities 10, 11 and 12), the shear zone network also affects a complex multilayered sequence of schists, marbles and calc-silicate rocks, and the late Variscan quartzdiorites and migmatitic schists of the Tudela migmatitic complex.



Aerial orthophoto showing the location of the main localites of Itinerary 1 (Institut Cartogràfic i Geològic de Catalunya, ICGC-Vissir).

Locality 1 Guillola crossroad

park the car at 42°18'43.06''N , 3°17'18.98''E

The road to the Cap de Creus lighthouse cuts across the Cadaqués Series. On the junction to Guillola housing, the metasedimentary sequences belonging to the cordierite-andalusite zone show D_3 folds affecting bedding/S_1 foliation. Two distinct lineations (a folded intersection lineation and a D_3 fold-related crenulation lineation) are observable on the folded S_1 surface. The main structure observable in this outcrop is a quasi-cylindrical reclined D3 fold.

D_3 fold with associated crenulation lineation. The S_1 foliation contains a folded intersection lineation. The schists derive from the Cadaqués Series in the andalusite-cordierite zone.

Locality 2 Puig de Culip S - la Llosa *(the Slab)*

park the car at Paratge de Tudela:

42°19'26.20"N , 3°17'47.00"E

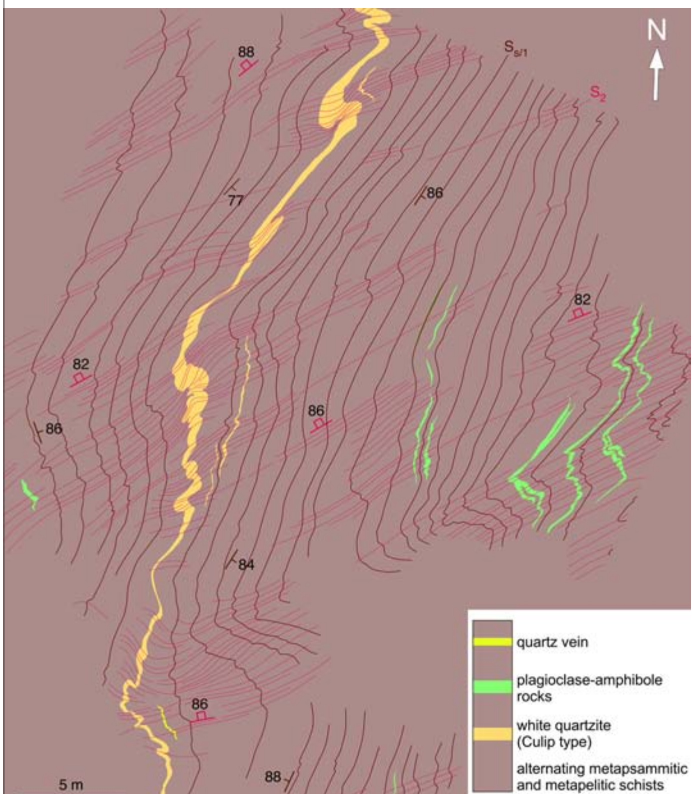
In these outcrops located in the area of relative low D_2-strain, the original stratigraphic features are still recognizable despite the effects of metamorphism. S_1 is a bedding-parallel foliation, while S_2 is a crenulation cleavage, best visible in the pelitic layers. Most segregation quartz veins are wrapped by S_1 foliation and folded by D_2. In the eastern side of the outcrop, andalusite porphyroblasts are visible. They grow over the S_1 foliation while are wrapped and sinistrally rotated during D_2. Also during D_2, andalusite is partially replaced by sillimanite (see e.g. Fig. 19).

High-grade psammitic-pelitic schists affected by a slight D_2 crenulation. A relict graded-bedding is recognizable.

Locality 3 Puig de Culip E

In the eastern slope of Puig de Culip we can see the effects of D_2 deformation on different lithologies, in a domain of relative low strain. Asymmetrical folds and associated crenulation cleavage are widespread in the metapelite-metapsammite sequence. These folding structures contrast with those observed in the plagio-amphibole rocks and in beds of light quartzites (Culip quartzite). Highly disharmonic polyclinal folds are developed in the quartzite beds. A set of lensoidal quartz veins, deformed during D_1, is also affected by D_2, giving rise to rod structures. During D_2, as a result of general shortening, they were folded and rotated, developing an asymmetrical shape with the widest part of the lenses or swells located in the short limbs. They display an anticlockwise sense of rotation with respect to the S_1 foliation. Multiple examples can be observed of the geometrical transition from folded quartz veins of uniform thickness to trains of isolated quartz rods.

Disharmonic folds in the quartzite from the eastern slope of Puig Culip. The complex folding parttern is localized in the thickened hinge.



Detail map of the D_2 structures in this locality (modified from Carreras and Druguet, 2013).

Asymmetric folds affecting a plagioclase-amphibole layer.

Sinistrally rotated quartz rods enclosed in asymmetrically (S-shaped) folded schists and quartz veins.

Locality 4 Puig de Culip N

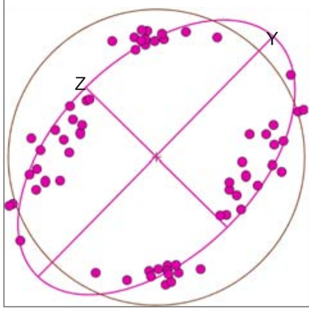
In the northern slope of Puig Culip one of the few pegmatite dykes is outcropping showing the syntectonic character with regard to the D_2 event.

Syn-D_2 emplaced pegmatite. Notice the relative fold tightness in the schists and pegmatite and the fact that some folds in the schists are cut by the intrusive contact while both surfaces share the same axial plane.

Locality 5 Puig de Culip NW

Quartz veins are abundant in these area. They form different sets which record different amounts of deformation depending on their orientation. Most veins show ptygmatic folds, indicating a high competence contrast with regard to the enclosing schists. All these characteristics make this veins suitable for strain determination.

Example of a D_2-ptygmatically folded vein in sub-horitzontal outcrop. S_2 crenulation cleavage has a SW-NE trend, almost perpendicular to the vein. Stretch of the vein is S≈0.49, thus the vein has been shortened by a 50%.



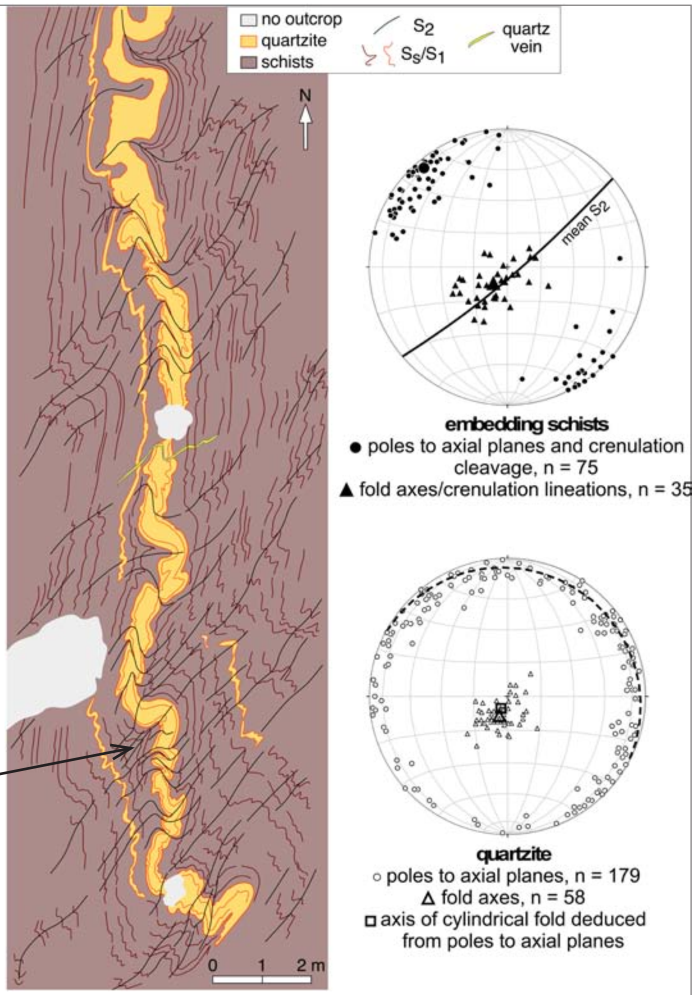
Synthetic best-fit finite strain ellipse determined from different stretch analysis performed in this D_2 low strain domain. Area reduction in the sub-horitzontal section ranges between 25 and 35 %. This is coherent with the fact that the analysis is performed in a section close to the YZ finite strain plane and the fact that D_2 stretching lineation is sub-vertical. Data from Druguet and Griera (1998).

Locality 6 Puig de Culip W

Another quartzite bed displaying complex D_2 polyclinal folding is outcropping in the western slope of Puig Culip. This particular complex fold pattern is characterized by strongly curved axial surfaces but straight and parallel hinges. While the fold hinges remain in close parallelism with extension direction, the axial planes become dextrally sheared as result of progressive deformation (Carreras and Druguet, 2019).

This folding type falls into type 3 hook-shaped interference pattern of Ramsay (1967) classification, and has been usually interpreted as resulting from the superposition of two deformation phases. Folds in the embedding schists are harmonic and asymmetric (S-shaped), revealing sinistral layer-parallel shearing.

Complex disharmonic folding in the Culip Quartzite. Strain localization in folded layer induces anti-clockwise rotation of hinges and associated dextral shearing of limbs and D_2 axial planes. Fold hinges are sub-parallel to the sub-vertical extension direction.



Detail map and stereoplots of the D_2 folds affecting the Culip quartzite and embedding schists in West Puig Culip (modified from Carreras and Druguet, 2019).

Locality 7 Cala Culleró S

An anastomosing NW-SE trending shear belt affects the E-W trending $S_{1/2}$ transposition foliation of the D_2 high strain domain. Shear zones are mainly dextral, with low NNW plunging stretching lineations.

These shears tend to bound more competent pegmatitic bodies which may develop a strong mylonitic foliation at their margins and tails. The mylonitic equivalents of pegmatites consist on leucocratic gneisses. In this area vertical and horizontal sections of small shear zones with marked longitudinal gradients are well exposed. Lozenges of non-mylonitic schists wrapped by the anastomosing shears are very well exposed (Ponce *et al.*, 2013).

Dextral shears cutting obliquely across the $S_{1/2}$ foliation corresponding to a high D_2 strain domain leaving lozenges of non-mylonitic schists.

Locality 8 Cala Culleró N - s'Eixugador *(the Dryer)*

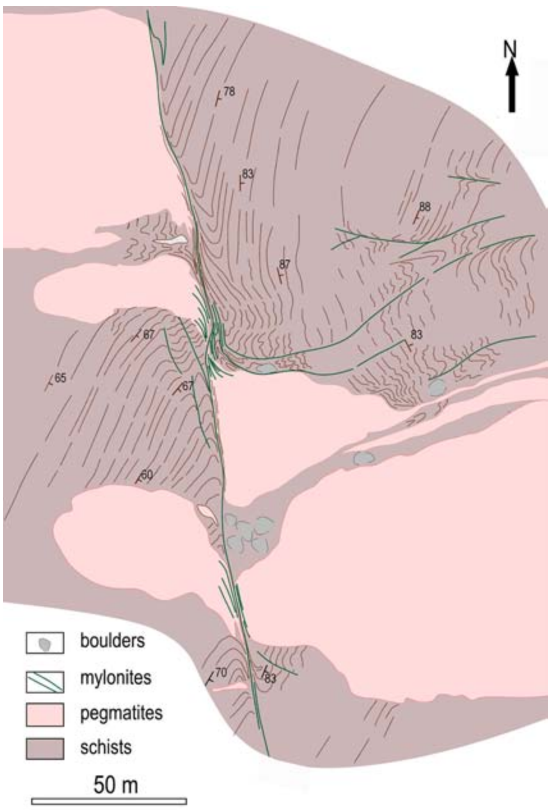
The same shear belt is visible on the northern part of the bay, here including spectacular examples of lozenges. At this locality some of the millipede-shaped lozenges result from the coalescence of sinistral and dextral shears.

On the eastern prolongation of this shear belt pseudotachylites cutting across the mylonitic foliation are found.

View of the anastomosing shear zones at Cala Culleró. This is one of the most spectacular localities to observe anastomosing shears and associated lozenges. Notice the intensely foliated mylonitic pegmatite on the right.

Locality 9 Sa Ferradura *(the Horseshoe)*

This spectacular dextral strike-slip shear zone has a nearly N-S trend and cuts across large pegmatite bodies producing a bulk horizontal displacement of about 75 m. The S_2 foliation lies in the shortening field forming an acute angle to the shear zone, being rotated nearly 160° towards parallelism with the shear plane.



Map of the "Sa Ferradura" shear zone (from Carreras and Druguet, 2013).

Detail of the shear zone (looking southwards).

Intramylonitic folds of foliated pegmatite.

Locality 10 Tudela

A rather homogeneous quartzdiorite-tonalite body (≈300x50m of surface size) appears on the northern coast of Tudela. The intrusion is elongated sub-parallel to the WNE–ESE S_2 foliation and intruded into strongly deformed migmatitic schists. The whole area is known as the Tudela migmatite complex. A D_2 magmatic to solid-state foliation is weakly defined by the preferred orientation of amphibole, biotite and plagioclase crystals. A sample of weakly deformed quartzdiorite was dated as ≈299 Ma (Druguet *et al*., 2014; see section 2.1.2).

These intrusive rocks are cut by the pegmatite dykes, and the whole set is cut by a network of heterogeneously distributed D_3 shear zones.

Massive quartzdiorite-tonalite forms the northern coast of Tudela.

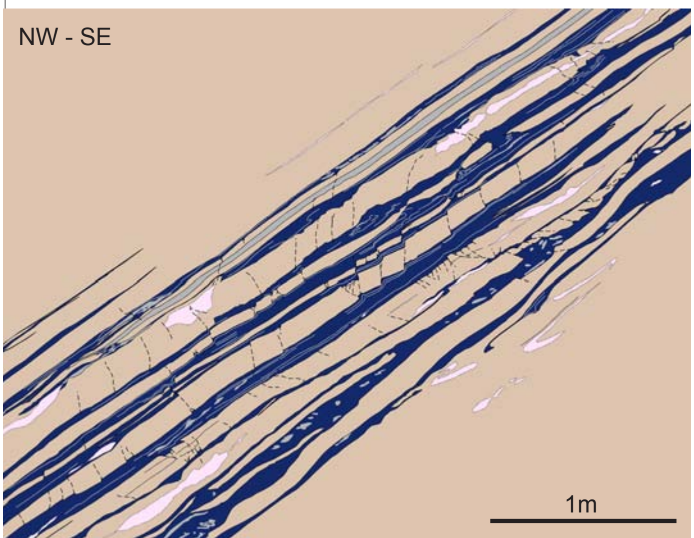
Locality 11 Clot de la Vellosa

West of locality 10 and still within the Tudela migmatite complex, there is an outstanding outcrop which shows clear evidence of a strong lithological control on strain partitioning during D_3. A lithological heterogeneous sequence consisting on metapsammites, marbles, calc-silicate rocks and black schists is superbly exposed here. This intercalation, called the marble-metapsammite unit (MMU) by Druguet *et al*. (2009) is equivalent to the Sant Baldiri tectonic slice (see section 2.1). D_3 deformation of this unit of contrasting rheological properties causes the presence of coeval brittle and ductile structures.

The dominant effect of D_3 on this unit is the localization of a layer-parallel sinistral shearing inducing both contractional and extensional structures, while in the underlying metapsammites and migmatitic schists display steep dextral D_3- shears.

The MMU can be followed westwards on the other side of the bay. There, the effects of D_3 deformation are far more complex and folded layers become subsequently broken, giving rise to breccias with a matrix of marble.

On the western slopes, gently dipping shear zones oblique to the bedding and $S_{1/2}$ foliation are visible. The particularity of this shears zones is that there is dilation across the shear plane with quartz veins being emplaced parallel to it.



Drawing of the structures in the marble-metapsammite unit (MMU) on the eastern slopes of Clot de la Vellosa (modified from Druguet *et al*., 2009). Blue: marbles; light blue: calc-silicate rocks; ochre: dominant metapsammitic layers; pink: quartz and quartz.feldspathic veins.

Boudinaged metapsammite and calc-silicate layers. Asymmetric boudins are delimited by a W-down set of fractures. The interlayered incompetent marbles accommodate the slip along fractures by ductile flow.

Asymmetric D_3 folds in layered marbles exhibiting west (left) vergence compatible with layer-parallel sinistral shearing

South of the MMU, D_3 shear zones abound across migmatitic schists and pegmatites. One is located along the eastern slopes of the Clot de la Vellosa valley and can be recognized by the metric dextral displacement visible by the offset of a cross-cut pegmatite dyke. The unsheared domains belong to a D_2 high strain zone, where the migmatitic schists display a banded high temperature S_2 fabric with abundant leucosome pods.

Locality 12 Cala d'Agulles

This narrow valley and sea entrance is eroded along a straight anastomosing belt on NNW-SSE trending shear zones. Although the bulk predominant kinematics is dextral, some conjugated anthithetic sinistral shears can be recognized. Along the eastern slopes, reacheable from the Tudela area, good examples of shear zones and related mylonites can be observed.

Following the valley towards the south, there are many other superb examples of mylonitic pegmatites and anastomosing shear zones.

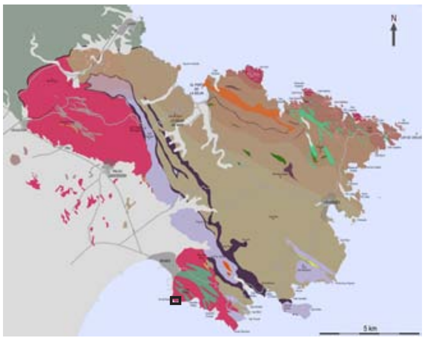
Tourmaline-rich rigid pod wrapped by mylonites derived from schists and pegmatites.

Small dextral shear zone across $S_{1/2}$ foliation in the migmatitic schists. Note the longitudinal gradient from ductile (right) to brittle (left) shear.

Detail of individual dextral-reverse shear zones cutting across the sub-vertical $S_{1/2}$ foliation of the schists.

After that, we will get back to the parking lot in upper Tudela, where the Camel sits on the largest pegmatite intrusion in Cap de Creus.

Itinerary 2



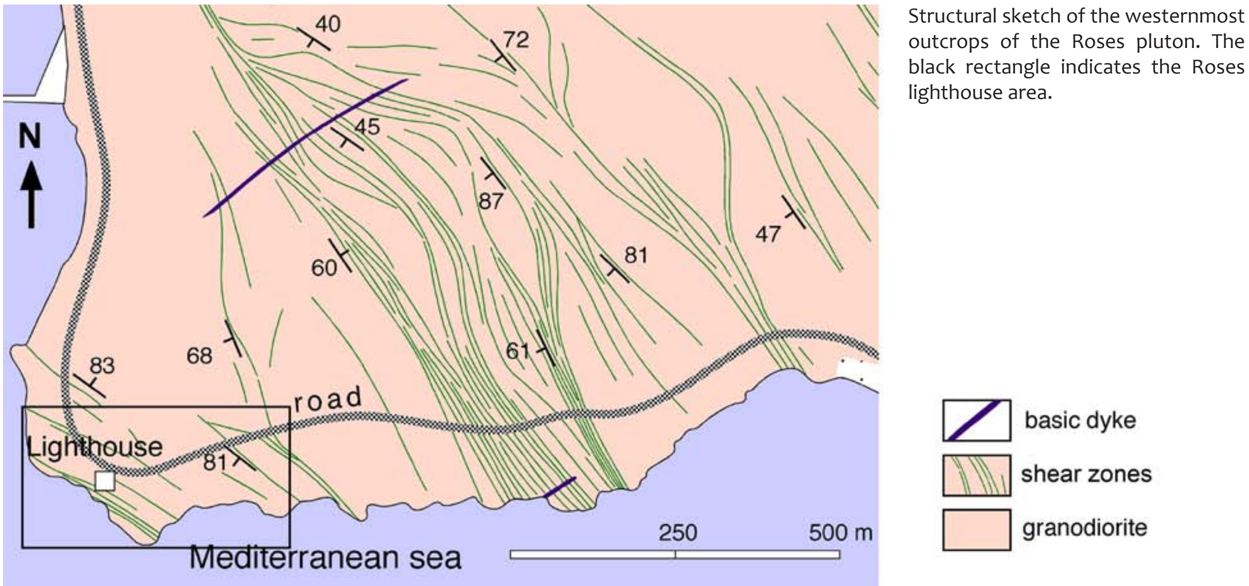
ROSES LIGHTHOUSE:
park the car at 42°14'45.35"N , 3°10'57.88"E

Progressive deformation history during and after pluton cooling
From D_2 magmatic structures to D_3 retrograde shear zones in the Roses granodiorite

The Roses granodiorite displays syn- and post-magmatic structures formed under a continuous deformation history during and after magma cooling (Carreras *et al*., 2004). The deformation history is divided on the basis of mechanical behaviour into two stages: an early one with development of pre-full crystallization fabrics and a late stage with development of mylonites along shear zones. These shear zones were also studied by Carreras and Losantos (1982), Simpson *et al*. (1982) and Simpson (1983). The Roses granodiorite has been dated as Early Permian (≈291 Ma) by Druguet *et al*. (2014).

The abundance of quartz dioritic enclaves allows the use of shape analysis to characterize the magmatic fabric, and the presence of leucocratic dykes (aplites and pegmatites) allows to determine the displacement associated to the late shear zones.

Several excellent outcrops of shear zones affecting the granodiorite and the leucocratic dykes can be observed along the coastal cliffs. The granodiorite has a magmatic pre-full crystallization fabric which is particularly evident in domains with concentration of enclaves. Dykes were emplaced along



Structural sketch of the westernmost outcrops of the Roses pluton. The black rectangle indicates the Roses lighthouse area.

Detail aerial photography showing the main localities (1, 2 and 3) and some included individual outcrops (small yellow dots) in the Roses lighthouse area.

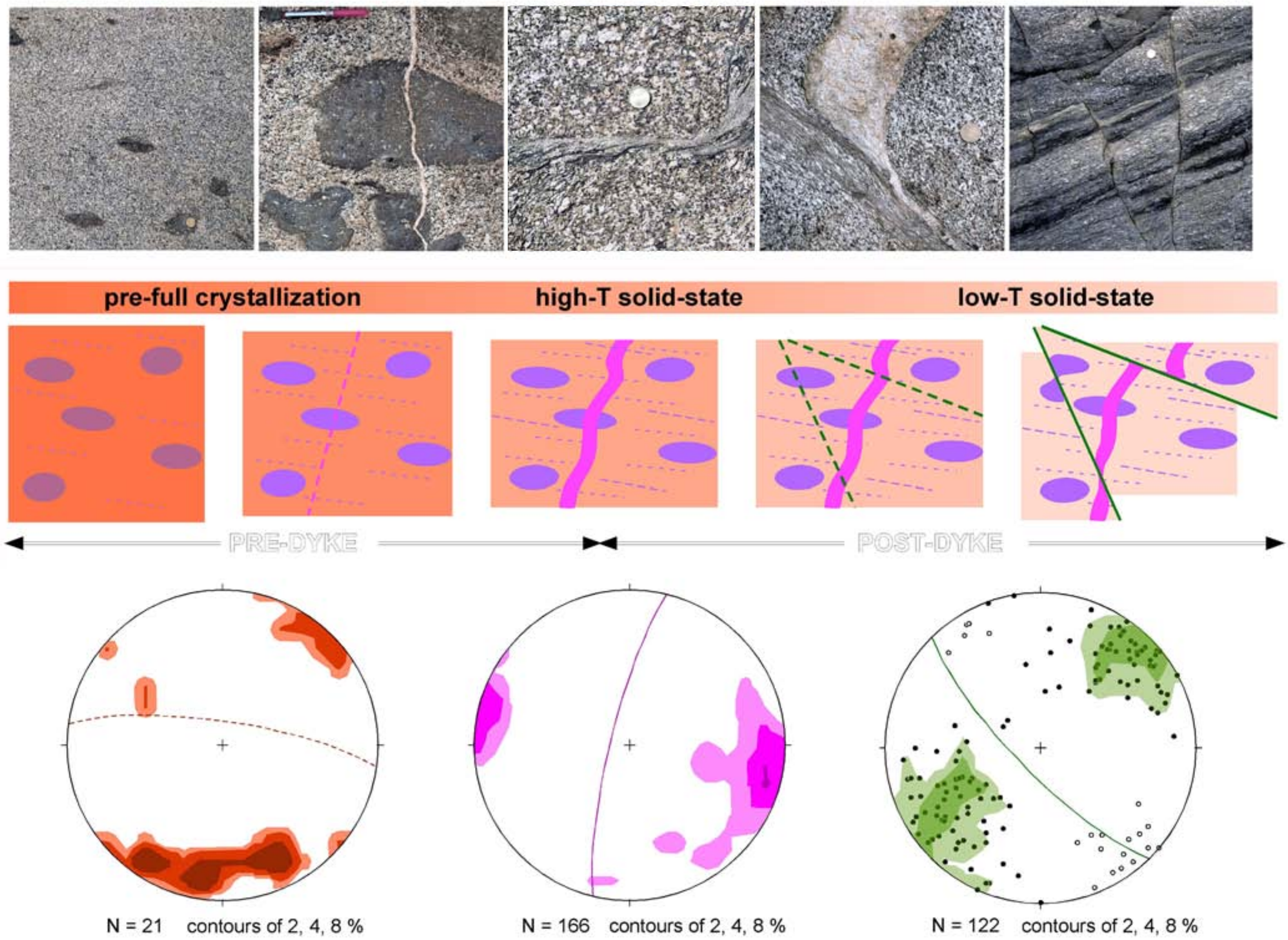
extension fractures cutting at a high angle to the preexisting magmatic fabric.

Shear zones have a predominantly sinistral kinematics due to the km-scale overturning of the structure along the southern border of the peninsula (see section 2.2 and Fig. 6).

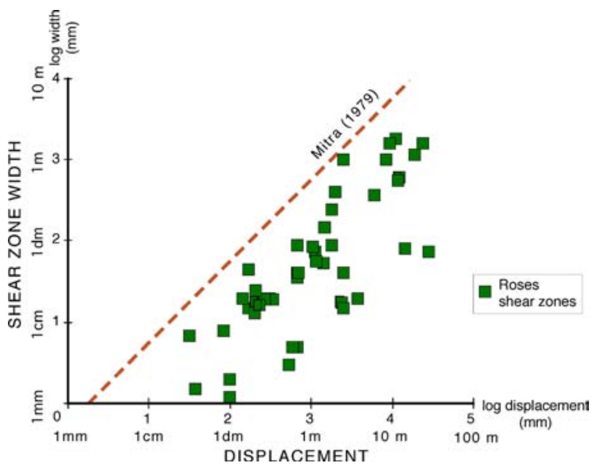
The amount of displacement is in average an order of magnitude greater than the width of each shear zone. Shear zones were developed under greenschist facies metamorphic conditions. Locally, mylonitization is associated with quartz depletion and formation of albite-chlorite mylonites.

A later event of brittle shear fractures is also recognizable.

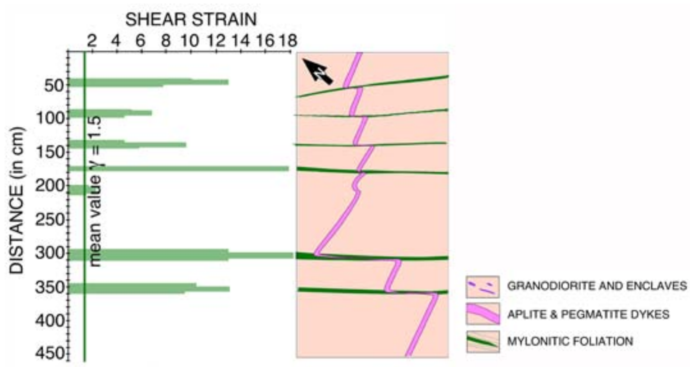
All these structures are interpreted as result of progressive deformation of the granodiorite during syntectonic emplacement and gradual cooling. Strain determinations for the magmatic state and the shearing event indicate similar amounts of bulk shortening, with a rather homogeneous deformation at high temperature and an accentuated strain localization during the greenschist facies conditions shearing.



Interpretation of the structural evolution of the syntectonic Roses granodiorite. (Modified from Carreras *et al*., 2004).



Displacement/width relation in the Roses shear zones (modified from Carreras *et al*., 2004).



Shear strain profile assuming heterogeneous simple shear (modified from Carreras *et al*., 2004).

Locality 1 Roses lighthouse (W)

Several outcrops enable to observe excellent examples of shear zones affecting the granodiorite, their enclaves and leucocratic dykes.

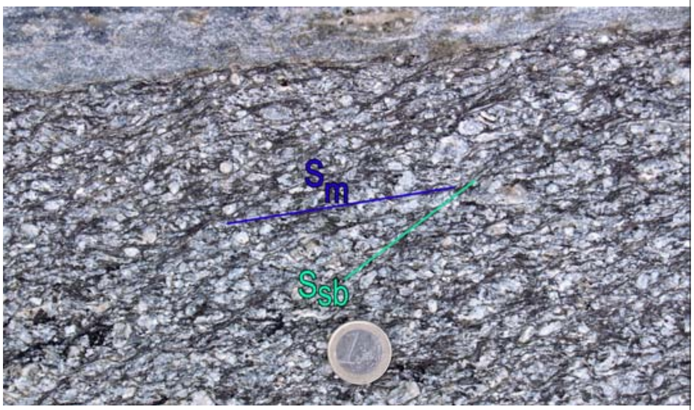
The pre-shearing features of the granodiorite can be widely observed due to the discrete and localized character of the shear zones. Among these, the presence of a magmatic fabric that is cut at a high angle by undeformed or slightly deformed aplite-pegmatite dykes.

Offset dykes and the sigmoidal mylonitic fabric enable to determine the shear sense and amount of displacement. Moreover, some shear zones contain shear bands coherent with the sinistral shearing deduced from vein displacement and the sub-horizontal attitude of the stretching lineation. Shear zones commonly involve manifest longitudinal strain gradients. In some examples brittle splays at the tips of shear zones are displayed.

Later fractures with associated cataclasites cut across and displace the mylonite bands.

Two parallel sinistral shear zones, the upper one showing a marked longitudinal shear gradient.

Sinistrally sheared leucocratic vein.



Shear bands (S_{sb}) in a sinistral shear zone (S_m: mylonitic foliation).

Locality 2 Roses lighthouse (S and E)

A broader shear zone with a northern sharp boundary can be observed south of the Lighthouse. This shear zone shows a complex kinematics with nearly parallel dextral and sinistral shearing. Inside the shear zone, the mylonitic granodiorite shows extremely stretched enclaves.

Towards the east, minor sized shear zones are observable with also associated dextral and sinistral displacements. Both sets form marked obtuse angles facing the shortening direction. Sets of sinistral shear zones also enable displacement determinations. Some shear zone terminations grade into brittle fractures (see Simpson, 1983; Segall and Simpson, 1986).

Stretched enclaves inside the high strain domain of the shear zone.

Shear bands across the mylonitic foliation in a dextral shear zone.

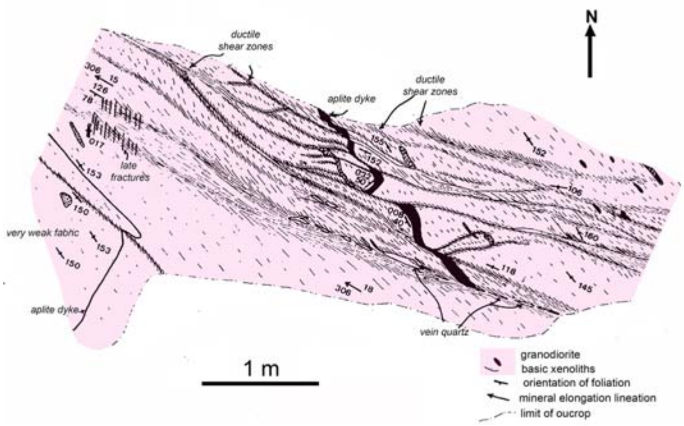
Set of discrete sinistral shear zones sharply cutting across a leucocratic vein and an enclave.

Locality 3 Coastal cliffs E from Punta Bateria

Several shear zones are well exposed in this locality. Some of them are cut by late small brittle fractures.

Peculiar curved shears filled with fine-grain undeformed granite can also be observed.

End of the field trip



Detail map of sinistral shear zones (modified from Simpson *et al.*, 1982).

4. REFERENCES

This list includes works on the Cap de Creus geology and other works that are related to the area by subject or by regional setting

Aguilar, C., Liesa, M., Castiñeiras, P., Navidad, M. (2014). Late Variscan metamorphic and magmatic evolution in the eastern Pyrenees revealed by U–Pb age zircon dating. Journal of the Geological Society 171, 181-192.

Alfonso, M.P. 1995. Aproximación a la petrogénesis de las pegmatitas del Cap de Creus. Unpublished Thesis, Univ. de Barcelona.

Alfonso, M.P., Corbella, M., Melgarejo, J.C. 1995. Nb-Ta-minerals from Cap de Creus pegmatite field, eastern Pyrenees: distribution and geochemical trends. Mineralogy and Petrology 55, 53-69.

Alfonso, M.P., Corbella, M., Melgarejo, J.C. 2003 Fluid evolution in the beryl-columbite-phosphate pegmatites of Cap de Creus (Catalonia, Spain). Journal of Geochemical Exploration 78-79, 17-21.

Alsop, G.I., Carreras, J. 2007. The structural evolution of sheath folds: A case study from Cap de Creus. Journal of Structural Geology 29, 1915-1930.

Auréjac, J-P., Gleizes, G., Diot, H., Bouchez, J-L. 2004. Le complexe granitique de Quérigut (Pyrénées, France) ré-examiné par la technique de l'ASM : un pluton syntectonique de la transpression dextre hercynienne. Bulletin Société Géologique de France 175, 157-174.

Autran, A., Fonteilles, M., Guitard, G. 1970. Relations entre les intrusions de granitoides, l'anatexie, et le métamorphisme régional considerées principalment du point de vue de l'eau: cas de la Chaine hercynienne des Pyrénées Orientales. Bulletin Societé Géologique de France 7, 673-731.

Bons, P.D., Druguet, E., Castaño, L.-M., Elburg, M.A. 2008. Finding what is now not there anymore: recognizing missing fluid and magma volumes. Geology 36, 851-854.

Bons, P.D., Druguet, E., Hamann, I., Carreras, J., Passchier, C.W. 2004. Apparent boudinage in dykes Journal of Structural Geology 26, 625-636.

Carreras, J. 1974. Progressive mylonitization in quartzitic and quartz-feldspathic rocks in the shear belts of the Cabo de Creus (E-Pyrenees, Spain). M.Sc. thesis, Imperial College, London.

Carreras, J. 1975a. Las deformaciones tardi-hercínicas en el litoral septentrional de la península del Cap de Creus (prov. Gerona, España): la génesis de las bandas miloníticas. Acta Geológica Hispánica 10, 109-115.

Carreras, J. 1975b. Determinación de las relaciones ángulares y de la deformación por cizalla, para cizallamientos en materiales con una heterogeneidad planar. Acta Geológica Hispánica 10, 141-145

Carreras, J. 1997. Shear zones in foliated rocks: geometry and kinematics. In: Sengupta, S. (Ed.), Evolution of geologic Structures in Micro- to Macro-scales. Chapman and Hall. London, pp. 185-201.

Carreras, J. 2001. Zooming on Northern Cap de Creus shear zones Journal of Structural Geology 23, 1457-1486.

Carreras, J., Capellà, I. 1994. Tectonic levels in the Palaeozoic basement of the Pyrenees: a review and a new interpretation. Journal of Structural Geology 16, 1509-1524.

Carreras, J., Casas, J.M. 1987. On folding and shear zone development: a mesoscale structural study on the transition between two different tectonic styles. Tectonophysics 135, 87-98.

Carreras, J., Casas, J.M., Losantos, M. 1998a. Some mesoscopic aspects of mylonitic shear zones. In: Snoke, A.W., Tullis, J., Todd, V.R. (eds) 1998. Fault-Related Rocks. A Photographic Atlas. xv + 617 pp. Princeton: Princeton University Press, pp. 234-235.

Carreras, J., Casas, J. M., Losantos, M. 1998b. Banding and foliation in mylonites. In: Snoke, A.W., Tullis, J., Todd, V.R. (eds) 1998. Fault-Related Rocks. A Photographic Atlas. xv + 617 pp. Princeton: Princeton University Press, pp. 296-297.

Carreras, J., Cosgrove, J.V., Druguet, E. 2013. Strain partitioning in banded and/or anisotropic rocks: implications for inferring tectonic regimes. Journal of Structural Geology 50, 7-21.

Carreras, J., Druguet, E. 1994a. Structural zonation as a result of inhomogeneous non-coaxial deformation and its control on syntectonic intrusions: an example from the Cap de Creus area (eastern-Pyrenees). Journal of Structural Geology 16, 1525-1534.

Carreras, J., Druguet, E. 1994b. El papel de las zonas de cizalla en la configuración estructural del complejo migmatítico del sector septentrional de la península del Cap de Creus (Girona). Revista Sociedad Geológica de España 7, 21-29.

Carreras, J., Druguet, E. 2011. Photograph of the Month: Ductile-brittle shear fractures in quartzites from Cap de Creus, Spain , Journal of Structural Geology 33, p.1.

Carreras, J., Druguet, E. 2013. Illustrated Field Guide to the Geology of Cap de Creus. Servei de Publicacions de la Universitat Autònoma de Barcelona, Bellaterra, 123 pp.

Carreras, J., Druguet, E. 2014. Framing the tectonic regime of the NE Iberian Variscan segment. In: Schulmann, K., Martínez Catalán, J.R., Lardeaux, J.M., Janousek, V., Oggiano, G. (eds): The Variscan Orogeny: Extent, Timescale and the Formation of the European Crust. Geological Society, London, Special Publications, 405, 249–264.

Carreras, J., Druguet, E. 2019. Complex fold patterns developed by progressive deformation. Journal of Structural Geology 125, 195-201.

Carreras, J., Druguet, E., Griera, A. 2005. Shear zone-related folds. Journal of Structural Geology 27, 1229-1251.

Carreras, J., Druguet, E., Griera, A., Soldevila, J. 2004. Strain and deformation history in a syntectonic pluton. The case of the Roses granodiorite (Cap de Creus, Eastern Pyrenees). In: Alsop, G.I., Holdsworth, R.E., McCaffrey, K.J.W. Hand, W. (Eds.), Flow Processes in Faults and Shear Zones. Special Publications 224. Geological Society, London, pp. 307-319.

Carreras, J., Estrada, A., White, S.H. 1977. The effects of folding on the c-axis fabrics of a quartz-mylonite. Tectonophysics 39, 3-24.

Carreras, J., García-Celma, A. 1982. Quartz c-axis fabric variation at the margins of a shear zone developed in schists from Cap de Creus (Spain). Acta Geológica Hispánica 17, 137-149.

Carreras, J., Julivert, M., Santanach, P. 1980. Hercynian Mylonite Belts in the Eastern Pyrenees: an example of shear zones associated with late folding. Journal of Structural Geology 2, 5-9.

Carreras, J., Julivert, M., Santanach, P. 1997. The Pyrenean Basement: Hercynian Structure and Effects of Alpine Reworking. In: Sinha, A.K., Sassi, F.P., Papanikolaou, D. (eds): Geodynamic Domains in the Alpine-Himalayan Tethys. A Publication of IGCP Project 276. xv + 441 pp. Rotterdam, Brookfield: A.A. Balkema.

Carreras, J., Losantos, M. 1982. Geological setting of the Roses granodiorite (E-Pyrenees,Spain). Acta geológica Hispánica 17(4), 219-225.

Carreras, J., Orta, J.M., San Miguel, A. 1975. El área pegmatítica del litoral N. de la península del cabo de Creus y su contexto metamórfico y estructural. Revista Instituto Investigaciones Geológicas 30, 11-34.

Carreras, J., Ramirez, J. 1984. The geological significance of the Port de la Selva Gneisses (Eastern Pyrenees, Spain). Newsletter I.G.C.P.-5, 6, 27-31.

Carreras, J., Santanach, P. 1973. Micropliegues y movimiento en los cizallamientos profundos del cabo de Creus (prov.Gerona). Estudios Geológicos 29, 439-450.

Casas, J.M., Fernández, O. 2007. On the Upper Ordovician unconformity in the Pyrenees: new evidence from the La Cerdanya area, Geologica Acta 5, 193-198.

Castiñeiras, P., Navidad, M., Liesa, M., Carreras, J., Casas, J.M. 2008. U-Pb zircon ages (SHRIMP) for Cadomian and Early Ordovician magmatism in the Eastern Pyrenees: new insights into the pre-Variscan evolution of the northern Gondwana margin. Tectonophysics 461, 228-239.

Corbella, M. 1990. Estudi metal.logenètic del camp pegmatític del Cap de Creus. Unpublished Tesi de Llicenciatura, Universitat de Barcelona.

Corbella, M., Melgarejo, J.C. 1993. Rare-element pegmatites of Cap de Creus Peninsula, northeast Spain: a new field of the beryl–phosphate subtype. Proc. 8th IAGOD Symp. E. Schweizerbart'sche Verlagsbuchhandlung (Nägele u. Obermiller), Stuttgart, Germany, 295–302.

Damm, K.W., Harmon, R.S., Heppner, P.M., Dornsiepen, U. 1992. Stable isotope constraints of the Cabo de Creus garnet-tourmaline pegmatites, Massif des Alberes, Eastern Pyrenees, Spain. Geological Journal 27, 76-86.

Druguet, E. 1992. Petrologia del complex migmatític de l'àrea de la Punta dels Furallons (Cap de Creus). Unpublished Tesi de Llicenciatura, Universitat de Barcelona.

Druguet, E. 1997. The structure of the NE Cap de Creus peninsula. Relationships with metamorphism and magmatism. Unpublished Thesis, Univ. Autònoma de Barcelona.

Druguet, E. 2001. Development of high thermal gradients by coeval transpression and magmatism during the Variscan orogeny: insights from the Cap de Creus (Eastern Pyrenees). Tectonophysics 332, 275-293.

Druguet, E., Alsop, I.G., Carreras, J. 2009. Coeval brittle and ductile structures associated with extreme deformation partitioning in a multilayer sequence. Journal of Structural Geology 31, 498-511.

Druguet, E., Carreras, J. 2006. Analogue modelling of syntectonic leucosomes in migmatitic schists. Journal of Structural Geology 28, 1734-1747.

Druguet, E., Castro, A., Chichorro, M., Pereira, M.F., Fernández, C. 2014. Zircon geochronology of intrusive rocks from Cap de Creus, Eastern Pyrenees. Geological Magazine 151, 1-20.

Druguet, E., Enrique, P., Galán, G. 1995. Tipología de granitoides y rocas asociadas del complejo migmatítico de la Punta dels Farallons (Cap de Creus, Pirineo oriental). Geogaceta 18, 199-202.

Druguet, E., Griera, A. 1998. Strain analysis using deformed quartz veins. Geogaceta 24, 119-122.

Druguet, E., Hutton, D.H.W. 1998, Syntectonic anatexis and magmatism in a mid-crustal transpressional shear zone: an example from the Hercynian rocks of the eastern Pyrenees. Journal of Structural Geology 20, 905-916.

Druguet, E., Passchier, C.W., Carreras, J., Victor, P. and den Brok, S. 1997. Analysis of a complex high-strain zone at Cap de Creus, Spain. Tectonophysics 280, 31-45.

Eeckhout, B. v. d., Zwart, H. J. 1988. Hercynian crustalscale extensional shear zone in the Pyrenees. Geology 16, 135-138.

Enrique, P. 1995. Una nueva interpretación sobre las relaciones entre el emplazamiento de granitoides y el metamorfismo regional hercinianos en el Pirineo oriental. Geogaceta 18, 203-206.

Evans, N. G., Gleizes, G., Leblanc, D., Bouchez, J. L. 1997. Hercynian tectonics in the Pyrenees: a new view based on structural observations around the Bassiès granite pluton. Journal of Structural Geology 19, 195-208.

Fonteilles, M., Guitard, G. 1977. Influence des noyaux de socle pré- cambrien sur le métamorphisme et la structure profunde de l'orogène hercynien des Pyrénnées orientales. Comparaisons avec les réggions voisines. Coll. Intern. C.N.R.S. La Chaîne Varisque d'Europe Moyenne et Occidentale, vol. 243. Rennes, pp. 81-87.

Fusseis, F., Handy, M.R. 2008. Micromechanisms of shear zone propagation at the brittle-viscous transition. Journal of Structural Geology 30, 1242-1253.

Fusseis, F., Handy, M.R. and Schrank, C. 2006. Networking of shear zones at the brittle-to-viscous transition (Cap de Creus, NE Spain. Journal of Structural Geology 28, 1228-1243.

García-Celma, A. 1982. Domainal and fabric heterogeneities in the Cap de Creus Quartz mylonites. Journal of Structural Geology 4, 443-456.

García-Celma, A. 1983. c-axis and shape fabrics in quartz mylonites of Cap de Creus (Spain); their properties and development. Ph.D. Thesis, Institute of Earth Sciences, University of Utrecht.

Gleizes, G., Leblanc, D., Bouchez, J.L. 1997. Variscan granites of the Pyrenees revisited: their role as syntectonic markers of the orogen. Terra Nova 9, 38-41.

Gomez-Rivas, E., Bons, P.D., Griera, A., Carreras, J., Druguet, E., Evans, L. 2007. Strain and vorticity analysis using small-scale faults and associated drag folds. Journal of Structural Geology 29, 1882-1899.

Griera, A. 2005. Estructures tectòniques i la seva relació amb la vorticitat cinemàtica: casos reals i models. Unpublished Thesis, Univ. Autònoma de Barcelona.

Guitard, G. 1970. Le métamorphisme hercynien mésozonal et les gneiss oeillés du massif du Canigou (Pyrénées Orientales). Mém. Bur. Rech. Géol. Min., 63, 353 pp.

Guitard, G., Vielzeuf, D., Martínez, F. (coord.) *et al*. 1996. Métamorphisme hercynien. In: Barnolas, A., Chiron, J.C. (Eds.), Synthèse Géologique et Géophysique des Pyrénées, vol. 1, BRGM-IGTE, pp. 501-584.

Hudleston, P.J., Lan, L. 1993. Information from fold shapes. Journal of Structural Geology 15, 253-264.

Hudleston, P.J., Treagus, S.H. 2010. Information from folds: A review. Journal of Structural Geology 32, 2042-2071.

Institut Cartogràfic de Catalunya, Generalitat de Catalunya. 1997. Mapa Geològic de Catalunya 1:25.000, fulls: 259-1-1 Roses; 259-2-1 Cap de Creus; 259-1-2 Far de Roses; 221-1-2 Llançà.

Julivert, M., Martinez J.F. 1987. The structure and evolution of the Hercynian fold belt in the Iberian Peninsula. In: J.P. Schaer, J. Rodgers (Eds.), The anatomy of mountain belts. Princeton Univ. Press, 65-103.

Mariani, I. 2009. Rotazione dei porfiroblasti sotto differenti regimi deformativi. Modellizzazione analogica e studio di un caso reale, Puig de Culip (Catena Costiera Catalana, Spagna). Unpublished Tesi di Laurea. Università di Pisa.

Montomoli, Ch., Carosi, R., Frassi, Ch., Ardito N., Carreras, J., Druguet, E. 2008. Shear partitioning across two parallel shear zones in Roses Granodiorite (Eastern Pyrenees, Spain). Rendiconti della Società Geologica Italiana 3, 569-570.

Navidad, M., Carreras, J. 1994. Pre-Hercynian magmatism in the Cap de Creus and Albera Massifs (Eastern Pyrenees) and their geodynamical setting. Geologie en Mijnbouw, 74, 65-77.

Norton, M. 1982. The kinematic and microstructural development of some shear zones. Ph.D. Thesis. Imperial College, London.

Olivier, Ph., Gleizes, G., Paquette, J.L., Muñoz Sáez, C. 2008. Structure and U-Pb dating of the Saint-Arnac pluton and the Ansignan charnockite (Agly massif): a cross section from the upper to the middle crust of the Variscan Eastern Pyrenees. Journal of the Geological Society 165, 141-152.

Olivier, Ph., Druguet, E., Castaño, L.M., Gleizes, G. 2016. Granitoid emplacement by multiple sheeting during Variscan dextral transpression: The Saint-Laurent–La Jonquera pluton (Eastern Pyrenees). Journal of Structural Geology 82, 80–92.

Paquette, J.-L., Gleizes, G., Leblanc, D., Bouchez, J.-L. 1997. Le granite de Bassiès (Pyrénées): un pluton syntectonique d'âge westphalien. Géochronologie U-Pb sur zircons. Comptes Rendus de l'Académie des Sciences, Paris 324, 387-392.

Piazolo, S., Passchier, C.W. 2002. Controls on lineation development in low to medium grade shear zones: a study from the Cap de Creus peninsula, NE Spain. Journal of Structural Geology 24, 25-44.

Ponce, C., Carreras, J., Druguet, E. 2010. Development of "lozenges" in anastomosing shear zone networks in foliated rocks. Geogaceta 48, 207-210.

Ponce, C., Druguet, E., Carreras, J. 2013. Development of shear zone-related lozenges in foliated rocks. Journal of Structural Geology 50, 176-186.

Ramirez, J. 1983. Zonació metamòrfica de les roques metapelítiques i metapsamítiques del litoral nord del Cap de Creus (Port de la Selva-Cala Taballera). Revista Instituto Investigaciones Geológicas 36, 7-24.

Roberts, M.P., Pin, C., Clemens, J.D., Paquette, J.L. 2000. Petrogenesis of mafic to felsic plutonic rock associations: the calc-alkaline Quérigut complex, French Pyrenees. Journal of Petrology 41, 809-844.

Santanach, P. 1972. Estudio tectónico del Paleozoico inferior del Pirineo entre la Cerdaña y el río Ter. Acta Geológica Hispánica 7, 44-49.

Segall, P., Simpson, C. 1986. Nucleation of ductile shear zones on dilatant fractures. Geology 14, 56-59.

Seguret, M., Proust, F. 1968. Tectonique hercynienne des Pyrénées centrales: signification des schistosités redressées, chronologie des déformations. C. R. Acad. Sci. Paris (sér. D) 266, 984 -987.

Simpson, C. 1983. Displacement and strain patterns from naturally ocurring shear zone terminations. Journal of Structural Geology 5, 497-506.

Simpson, C., Carreras, J., Losantos, M. 1982. Inhomogeneous deformation in Roses granodiorite. Acta Geológica Hispánica 17, 219-226.

Van Lichtervelde, M., Grand'Homme, A., de Saint-Blanquat, M., Olivier, Ph.; Gerdes, A., Paquette, J.-L., Melgarejo Draper, J.C., Druguet, E., Alfonso Abella, M.P. 2017. U-Pb geochronology on zircon and columbite-group minerals of the Cap de Creus pegmatites, NE Spain. Mineralogy and petrology, 111, 1-21.

Vilà, M., Pin, C., Enrique, P., Liesa, M. 2005. Telescoping of Three Distinct Magmatic Suites in an Orogenic Setting: Generation of Hercynian Igneous Rocks of the Albera Massif (Eastern Pyrenees). Lithos 83, 97-127.

Vissers, R.L.M. 1992. Variscan extension in the Pyrenees. Tectonics 11, 1369-1384.

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

White, S.H. 1979. Large strain deformation: report on a tectonic studies group discussion meeting held at Imperial College, London on 14 november 1979. Journal of Structural Geology 1, 333-339.

White, S.H., Burrows, S.E., Carreras, J. 1978. Textural and Microstructural development in a naturally deformed quartzite: a metallurgical approach. In: Gottstein, G., Lucke, K. (Eds.), Proceeding of the Fifth International Conference on Textures of Materials. March 28-31, 1978, Aachen, Germany. Springer-Verlag, Volume II, 211-220.

White, S.H., Burrows, S.E., Carreras, J., Shaw, N., Humphreys, J. 1980. On mylonites in ductile shear zones. Journal of Structural Geology 2, 175-187.

Zwart, H.J. 1963. The structural evolution of the Paleozoic of the Pyrenees. Geologishe Rundschau 53, 170-205.

Zwart, H.J. 1968. The Paleozoic crystalline Rocks of the Pyrenees in their structural setting. Kristalinikum 6, 125-140.

Zwart, H.J. 1979. The geology of the central Pyrenees. Leidse Geologische Mededelingen 50, 1-74.