Pergamon

PII: S0191–8141(98)00017–0

Journal of Structural Geology, Vol. 20, No. 7, pp. 905 to 916, 1998

0191-8141/98 $19.00 + 0.00

Syntectonic anatexis and magmatism in a mid-crustal transpressional shear zone: an example from the Hercynian rocks of the eastern Pyrenees

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(*Received* 19 *December* 1996; *accepted in revised form* 20 *February* 1998)

Abstract—Hercynian metamorphism and magmatism in the Cap de Creus peninsula (eastern Pyrenees) occurred contemporaneously with non-coaxial deformation in a transpressive regime. An example of this has been taken from a migmatite complex in the northern coast. The studied area is formed by partially melted sillimanite schists together with two different suites of igneous rocks: a calc-alkaline magmatic sequence, consisting of small granitoid bodies, and later peraluminous leucogranites and pegmatites. All these rocks occur within a NE–SW- to E–W-trending sub-vertical high-strain zone, where a first schistosity is tightly folded and transposed. The relations between folds and crenulation cleavage in the metasediments and deformational fabrics in the granitoids and migmatites provide good evidence for the penecontemporaneity of deformational processes, magmatism and migmatization. Furthermore, the analysis of cross-cutting relationships between different granitoids as well as the observation of their different degrees of deformation verifies that magmatism in this area took place as synkinematic consecutive intrusions from the more basic magmas to the peraluminous acid dykes.

INTRODUCTION

There is now a well recognized spatial and temporal connection between granitic magmatism and deformation zones in the Earth's crust. At its simplest level this is generally taken to mean that such zones are preferred because of their ability to provide dilational space, during active deformation, for the ascent and emplacement of granitoids (Guineberteau *et al.*, 1987; Hutton, 1988; McCaffrey, 1992; Grocott *et al.*, 1994; Aranguren *et al.*, 1997). A second argument is that, in shear zones, dilatation is not entirely necessary for emplacement, and that syntectonic magmatism can occur in 'space denying' compressional/contractional settings (thrusts, transpressional shear zones, etc.). Associated with this, the utilization of crustal anisotropies by high magma overpressures has also been discussed (Blumenfeld and Bouchez, 1988; D'Lemos *et al.*, 1992; Hutton and Ingram, 1992; Ingram and Hutton, 1994; Hutton, 1997).

The more general scenario that goes beyond these geometric, temporal and kinematic relationships is that magmatism is often concentrated in these zones because of a close causal connection between tectonic stresses/deformation and granite petrogenesis itself. Aspects of this have been explored on the macro scale (e.g. Hutton and Reavy, 1992; D'Lemos *et al.*, 1992; Tommasi *et al.*, 1994) and on the outcrop and microscale (both analytically and experimentally) in a number of seminal contributions (Dell'Angelo and Tullis, 1987; Hand and Dirks, 1992; Davidson *et al.*, 1994; Sawyer, 1994, 1996; Brown *et al.*, 1995a,b; Rutter and Neuman, 1995; Rushmer, 1996; Lucas and St-Onge, 1995). A key aspect of many of these latter works is that it is the imposition of differential stresses rather than hydrostatic stresses alone, which aids either melt production (e.g. Hand and Dirks, 1992) and/or the more efficient segregation and movement of melt away from the source (e.g. Dell'Angelo and Tullis, 1988; Rutter and Neuman, 1995; Rushmer, 1996). However, elegant and convincing as such models are, there is a need to underpin them with field documentation on a variety of scales, in particular where enhanced differential stress is inferred in temporal and spatial coincidence with melting and melt products.

In what follows, we describe the relationships between deformational fabrics and structures, anatectic products and small-scale granitic bodies in a mid-crustal transpressional shear zone, beautifully exposed in the Hercynian basement of the Pyrenean Axial Zone.

REGIONAL GEOLOGY

The Cap de Creus peninsula in NE Spain (Fig. 1a), forms the most easterly outcrop of Hercynian basement exposed along the Axial Zone of the Pyrenees. Whilst the general characteristics of the metamorphism

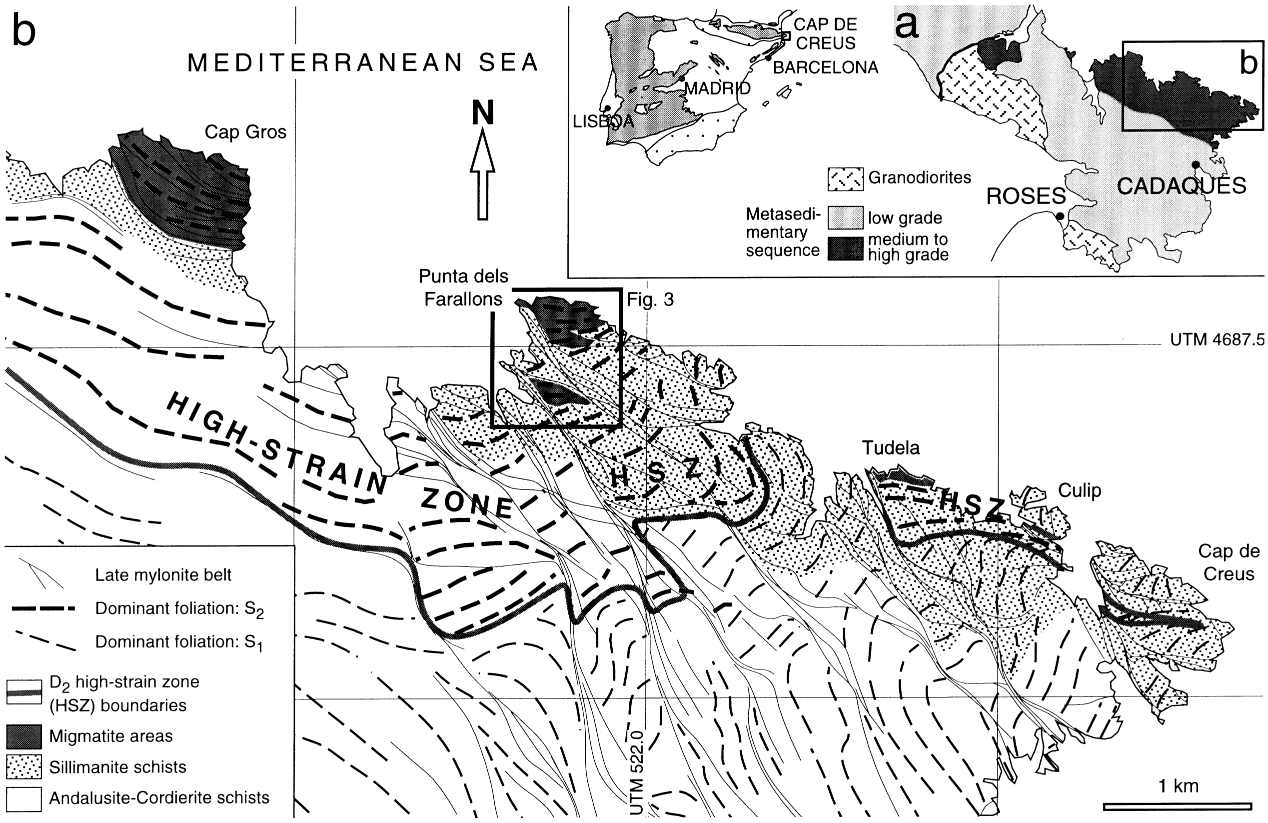

Fig. 1. Geological setting of the studied area. (a) Main lithological units in the Cap de Creus peninsula. (b) Structural sketch of the northern coast of the peninsula. Coordinate system refers to the Universal Transverse Mercator Grid (UTM), zone 31-European Datum.

and the structure in this area are rather similar to elsewhere in this general area (Autran *et al.*, 1970; Carreras and Capellà, 1994), some of the salient features of the local geology will be briefly described here.

The metasedimentary sequence is dominated by Precambrian or Cambro-Ordovician greywackes and subordinate pelites, and minor amounts of pre-Hercynian igneous rocks (sub-volcanic intrusions, lavas and volcanogenic sediments). These rocks are affected by a low-pressure regional Hercynian metamorphism which progressively changes grade across some 10 km of the area from the chlorite–muscovite zone in the south to the sillimanite zone in the north. Hercynian magmatism in the area consists of two major granodiorite stocks (the Roses and Rodes intrusions) which are emplaced in the low grade metasediments in the south, and numerous, much smaller, magmatic and migmatitic volumes which occur in the high grade metamorphic terrain in the north. These latter rocks are the subject of this paper. They include firstly, small bodies of quartz diorite, tonalite, leucogranite and pegmatite, which are all associated with migmatitic pods and which are restricted to the sillimanite zone metasediments, referred to hereafter as 'the migmatite complexes'. Secondly, these igneous rocks include a swarm of anatectic pegmatite dykes which occupy a wide area in the northern high grade zone, from the cordierite–andalusite zone northwards. These are referred to below as 'the pegmatite area'.

The Hercynian structure in the Cap de Creus peninsula is complex and is characterized by polyphase structures. The present study concentrates on the deformational features in the north, within the medium to high grade metamorphic domain (Fig. 1b). We find here a first schistosity (S_1), mainly parallel to bedding, which was formed prior to the metamorphic climax. In areas which are less affected by later overprinting deformation, the S_1 foliation is steeply dipping and has a predominantly N–S trend. In zones of high D_2 strain it is transposed into a NE–SW to E–W trend with steep dips. This second deformation, which is coeval with the peak metamorphism and, as will be shown, granitoid emplacement, is associated with folds and a crenulation cleavage and has a heterogeneous development (see below). Following this, just after the peak of metamorphism and in the subsequent retrograde metamorphic conditions, F_3 E–W- to NW–SE-trending folds developed. These also have a heterogeneous distribution. They become widespread in the medium metamorphic grade zone but, further northwards, these F_3 folds become progressively replaced by coplanar, predominantly dextral, NW–SE-trending shear zones (Carreras and Casas, 1987) which form the 'northern mylonite belt'. Individual shear zones within this belt have maximum offsets of about 1 km. In general terms, and as detailed by Druguet (1997), the overall D_2–D_3 Hercynian tectonic evolution of the area is characterized by (i) a progressive spatial (see below) and temporal clockwise rotation of structures from upright NE–SW to NW–SE with southern vergence; (ii) a broadly transpressive regional regime, evolving from contraction-dominated transpression (D_2) to wrench-dominated transpression (D_3); and (iii) progressive strain localization along narrow shear zones at low temperature conditions (D_2–D_3). Aspects of this history and development are discussed and enlarged upon below.

THE D_2 HIGH-STRAIN ZONES

Although D_2 consists of a general reworking of the S_1 schistosity in prograde metamorphic conditions, the intensity of D_2 varies systematically across the area and one can distinguish between D_2 structural domains characterized by high strains (high strain domains) and those characterized by low strains (low strain domains). On the largest scale, this 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 metamorphics (Fig. 1b).

In the northern area, although the D_2 strain is generally much higher than in the south, it is, nevertheless, heterogeneous, with high strain zones varying in width between 150 m (e.g. Culip area) and more than 1 km (in the studied area). The zones occur as NE–SW- to E–W-trending sub-vertical anastomosing bands within which the S_1 schistosity and bedding are entirely transposed by S_2. These high strain zones coincide with voluminous pegmatite dykes, and in the studied area they also coincide with the granitoids and migmatites of the migmatite complexes. It is remarkable that in spite of their marked compositional heterogeneity, all the intrusive rock types occur in the medium to high grade metamorphic area where there is intense D_2 deformation, and, as will be detailed below, where emplacement, metamorphism and deformation all occurred close in time.

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 (Fig. 2), especially visible in the Culip area (Fig. 1b) (Carreras and Druguet, 1994a; Druguet *et al.*, 1997), which Druguet (1997) interpreted as being associated with rotational components of the D_2 deformation. In this structure, the F_2 folds have moderately plunging to sub-vertical axes sub-parallel to the calculated fold axis for the cylindrical dextral flexure. 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 (Fig. 2 and see Druguet, 1997 for details) as a complex transpressive shear zone involving vertical extension, NNW–SSE

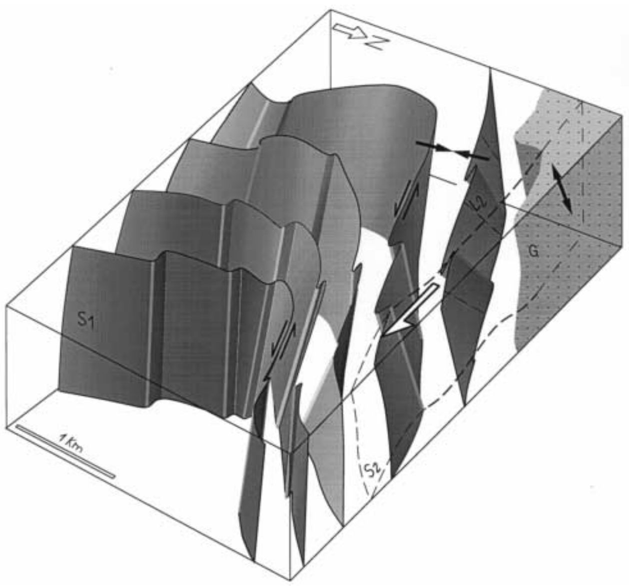

Fig. 2. Schematic block-diagram showing the arrangement of D_2 structures. Bulk dextral (white arrow) transpression involves sinistral S_1-parallel shearing (double arrows), sub-vertical extension and sub-horizontal shortening (black arrows). L_2: D_2 related stretching lineation, sub-parallel to F_2 fold axes. For simplicity, only one granitoid body (G) has been represented.

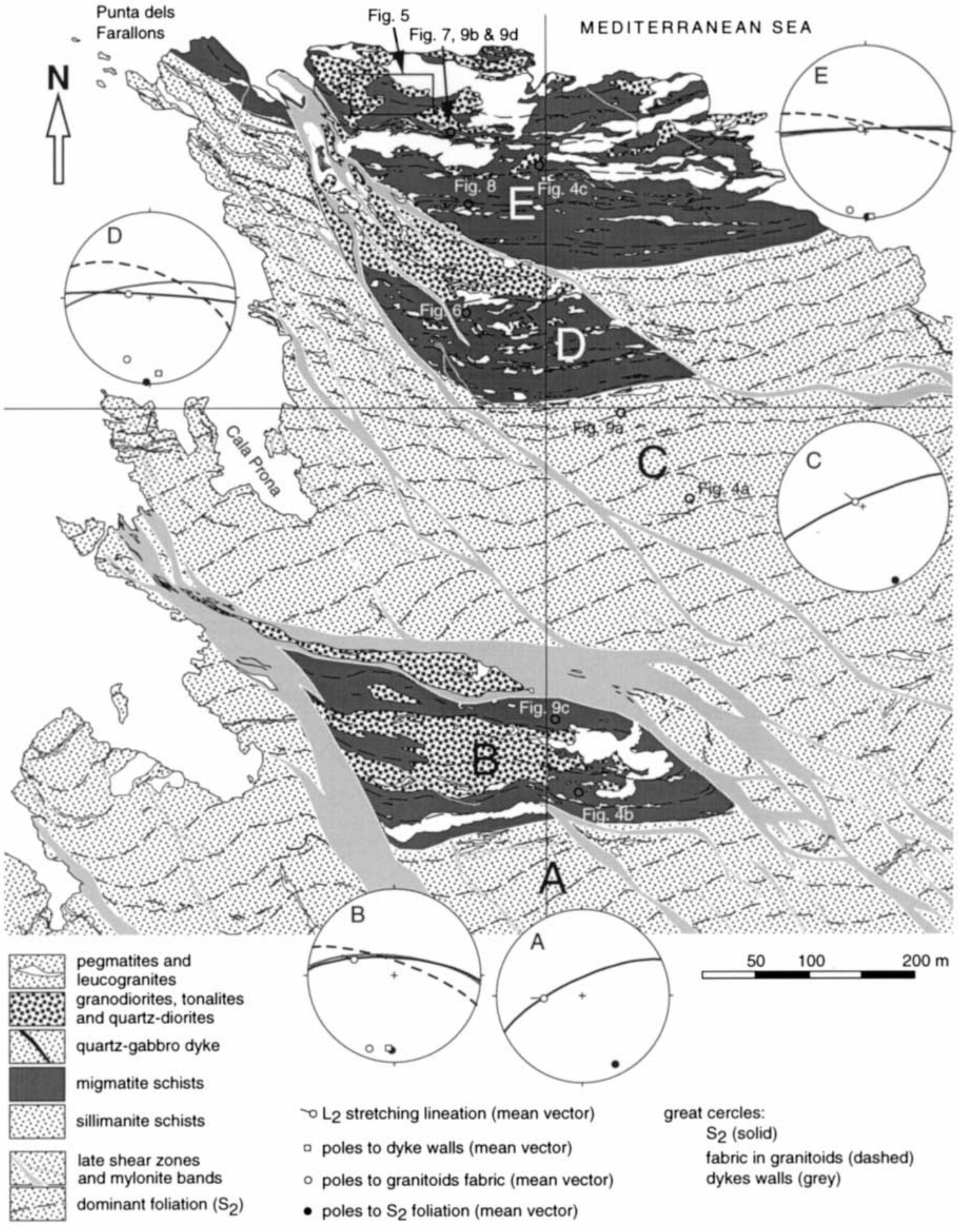

Fig. 3. Structural map of the Punta dels Farallons migmatite complex and equal-area lower-hemisphere projections for

subhorizontal bulk shortening with a dextral component, and layer (S_1) parallel sinistral flexural flow.

MIGMATITES AND DEFORMATION

There are three migmatite complexes in the Cap de Creus peninsula (Fig. 1b): at Cap Gros (Ramírez, 1983); Punta dels Farallons (Druguet, 1992; Druguet *et al*., 1995); and Tudela, and these are all located in zones of very high D_2 strain. The Punta dels Farallons complex, which is presented here in detail (Fig. 3), is separated by late (D_3) shear zones into two outcrop areas. Moving northwards towards either of the migmatite outcrops, the S_2 foliation increases in intensity and the foliation patterns show a smooth dextral drag

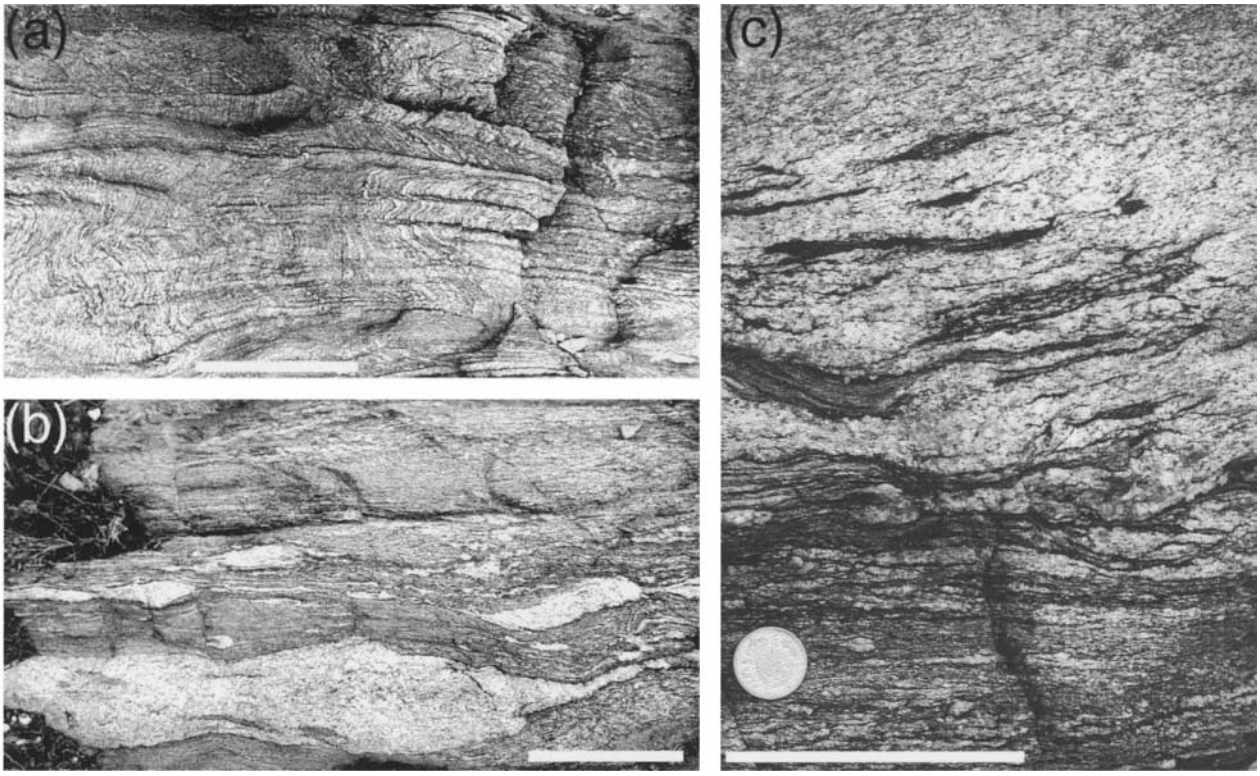

Fig. 4. Field photographs. Scale bar is 10 cm. Locations are shown in Fig. 3. (a) F_2 folds and related crenulation cleavage affecting the metagreywackes. (b) Metapelites are transposed by S_2 foliation and beaded leucosome veins develop in close parallelism with S_2. (c) Transition from a stromatic migmatite (lower part of the photograph) to a schlieren migmatite and more homogeneous granite (upper part). The layered structure corresponds to the S_2 foliation and is well defined by biotite-rich melanosomes.

(Fig. 3) in which the axial traces of the F_2 folds and the S_2 foliation change from NE–SW trending (stereoplots from domains A and C) to E–W trending in the higher strain zones to the north (stereoplots from domains B, D and E). Within the zones of highest strain, a steeply plunging stretching lineation is associated with the steeply N-dipping fabric. The steeply plunging F_2 folds become tighter northwards towards these zones (Fig. 4a), until they are transposed by S_2 (Fig. 4b). As reported above, the smaller scale shear-sense indicators show dominant horizontal sinistral kinematics associated with the bulk dextral flexural flow. This pattern is complicated by F_3 folds and shear zones (Figs 1b & 3). These folds have E–W-trending axial surfaces and sub-vertical axes, and are synchronous with the intrusion of the very latest pegmatites. The later D_3 retrograde shear zones are of dextral-reverse sense (Carreras and Druguet, 1994b) and these overprint all earlier structures and produce local mylonitization in the migmatite complexes.

The amount of quartzofeldspathic veins and migmatitic leucosomes mirrors the D_2 strain gradient by increasing towards the cores of the migmatite complex outcrops (Fig. 4). In general terms, it can be seen that as the D_2 strain and transposition increase towards the north so the more pelitic layers progressively become migmatites. First, the metapelites develop a penetrative S_2 foliation, marked by the preferred orientation of biotite, which wraps around early cordierite porphyroblasts. New prismatic sillimanite grows parallel to this fabric. Secondly, S_2-parallel quartz veins which are abundant on the margins of the pelitic layers are, on moving northwards into the migmatite zone, progressively replaced by quartzofeldspathic veins or migmatitic leucosomes (composed of quartz, plagioclase $\pm$ K-feldspar $\pm$ sillimanite $\pm$ almandine), forming stromatic migmatites. In the northernmost outcrops the migmatites themselves consist of lensoid or beaded leucosomes surrounded by mafic selvages (Fig. 4b & c), mainly composed of biotite ($\pm$ sillimanite, cordierite, almandine). These are typical metatexite migmatites (Brown, 1973; Sawyer, 1996) where the bulk of the rock has not melted and, given the amphibolite facies setting, they were probably formed in the presence of a water-rich volatile phase (Sawyer, 1996). In some parts of the high strain zone, small volumes of diatexitic schlieren and nebulitic migmatites are seen, often in close association with tonalitic and dioritic bodies. These indicate that more complete melting has taken place (Sawyer, 1996), although on a local scale.

The close correspondence between the spatial gradients of the D_2 strain and the development and abundance of the migmatitic features strongly suggests that the two are related dynamically. Brown *et al.* (1995a, b), have produced a model for the formation of stromatic migmatites in which they show that gravity (buoyancy) driven compaction across the stromatic layers is too slow and inefficient a process to segregate

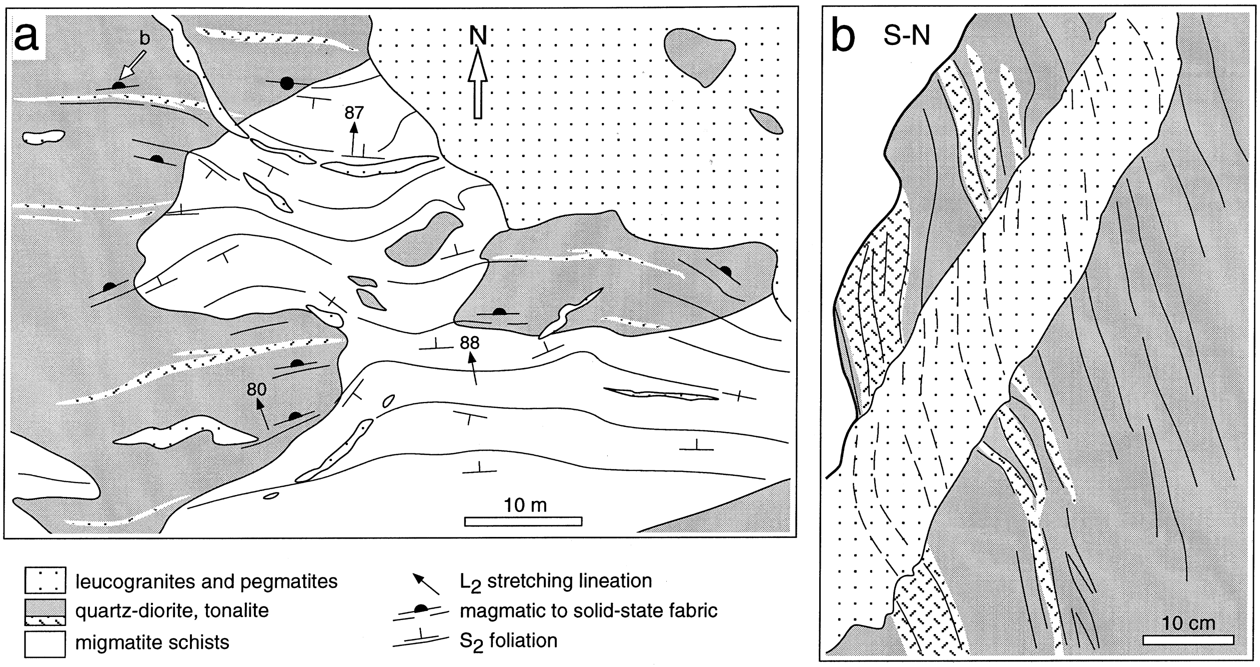

Fig. 5. Detailed map (a) and field sketch (b) of one of the basic–intermediate bodies. Locations are shown in Fig. 3. (a) The external S_2 foliation is deflected around the intrusion but at the same time is axial planar to the fold and can be correlated with the internal magmatic to solid state fabric and the compositional banding in the granitoid. (b) The magmatic to solid state fabric in the quartz-dioritic body is parallel to different compositional heterogeneities. A leucogranite dyke cross-cuts it with sharp contacts but includes some of the deformational fabric in the more basic intrusion.

melts in geologically reasonable time scales and volumes. The addition of tectonically induced differential stresses however makes the melt segregation process much quicker and more efficient (see also Sawyer, 1994) and it is believed that the tectonic structures themselves are favourable for melt escape (Brown *et al.*, 1995a, b). The general relationships in the Cap de Creus between the shear zones, which are structures produced by the operation of tectonic differential stresses, and the enhanced spatial abundance of the migmatites is in accord with this model.

MAGMATISM AND DEFORMATION

The magmatic rocks in the Punta dels Farallons migmatite complex (Fig. 3) vary from small bodies of hornblende-rich quartz-gabbros and quartz-diorites to garnet-bearing peraluminous leucogranites and pegmatites. The rocks have been divided into two associations (Druguet *et al.*, 1995). In order of emplacement these are: (a) a calc-alkaline sequence which includes quartz-gabbros, quartz-diorites, tonalites, granodiorites and granites; and (b) a later peraluminous association, which is anatectic in origin and comprises leucogranites and pegmatites.

The calc-alkaline sequence

These are heterogeneous in composition and the more acid types contain xenoliths or enclaves of the more basic types. As well as this, mafic schlieren structures (Fig. 4c) and compositional banding (Fig. 5) are common. These rocks occur as small bodies (up to 60 m × 50 m), which may be lensoid or have fold-like shapes (Figs 5a & 6a). However, the more acid units (granodiorites and granites) also occur as veins (Fig. 6b).

The basic–intermediate types contain an internal deformational fabric consisting of an alignment of plagioclase, hornblende and biotite. This is a magmatic or sub-magmatic-state fabric (i.e. formed during or at the latest stages of crystallization) and it is often preserved in the centre of the bodies (Fig. 7a). It is progressively modified into a high temperature solid-state fabric (i.e. early post-crystallization deformation) as it intensifies towards the contacts with the enclosing schists (Fig. 7b), where it can be correlated with the pervasive S_2 fabric (Fig. 5a). This magmatic to solid-state fabric (which trends between E–W and NW–SE, and dips steeply north) is parallel to the different compositional heterogeneities (Fig. 5b) and to the margins of the lensoid bodies. In the granitoids with the fold-like shapes the fabric is axial planar to these folds (Fig. 5a). A predominantly steeply plunging stretching lineation is sometimes seen associated with the granitoid fabric and whereso it is sub-parallel to the lineation in the enclosing metasediments. These relationships imply that magmatic deformation was coeval with the regional D_2 deformation.

The more acid granodiorite and granite types, although commonly cross-cutting the more basic types,

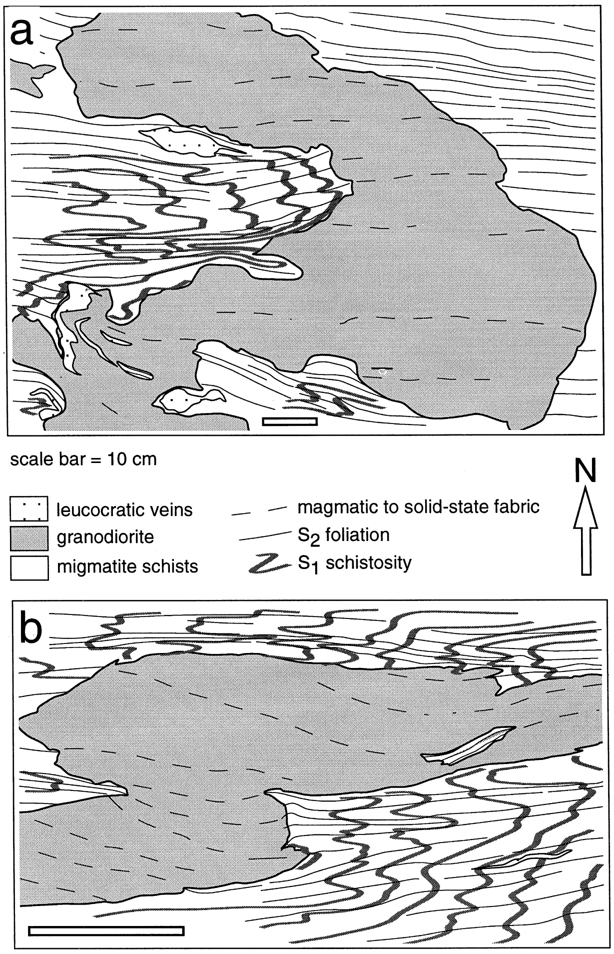

Fig. 6. Field sketches. Locations are shown in Fig. 3. (a) A small granodiorite body with a fold-like shape. The internal fabric is parallel to the external S_2 foliation and axial planar to the folds. (b) A granite vein slightly oblique to S_2. In this case, the internal magmatic fabric is also slightly oblique to the vein walls and to S_2 foliation. The vein contains a small metasedimentary xenolith which carries S_2 deformation features and is aligned parallel to the internal (D_2) magmatic fabrics.

often do so with diffuse and indistinct contacts, suggesting that they are penecontemporaneous. These more acid magmas are intruded as small bodies (up to 50 m × 30 m) as well as subparallel veins between 1 cm and 4 m wide (Fig. 6). The bodies and veins all contain magmatic to solid-state deformation fabrics which may be correlated with the S_2 foliation in the enclosing metasediments. A variety of features, related to the timing of emplacement and D_2 deformation, are displayed by these minor intrusions. The veins and sheets may be sub-parallel or slightly oblique to the external S_2 (in these cases they are straight or boudinaged), or else highly oblique to S_2 (in which case they are folded). In the folding case, the internal magmatic fabric is continuous with the external S_2, as in the situation of the basic–intermediate igneous bodies (Fig. 6a). In the slightly oblique veins and sheets, the internal magmatic fabric is, in most cases, also slightly oblique to the vein walls (Fig. 6b). The bodies usually contain small xenoliths of the enclosing metasediments which carry D_2 deformation features (crenulation cleavage and folds, Fig. 8a & b). These xenoliths may additionally be sometimes aligned parallel to, or elongated by, the internal (D_2) magmatic fabrics (Fig. 6b). These relationships suggest that the igneous bodies were intruded while the D_2 deformation event was occurring and were subsequently deformed by it.

The leucogranites and pegmatites

The latest igneous intrusions are the peraluminous leucogranites and pegmatites. These are related to anatectic processes (Carreras *et al.*, 1975; Damm *et al.*, 1992; Pau, 1995). They are not restricted to the migmatite area and the leucogranites are found also in the sillimanite zone. The pegmatites are much more widespread and are found in the above areas and in the adjacent cordierite–andalusite zone. Both types increase in quantity and size towards the higher grade and higher strain areas. The leucogranites are relatively small (between 5 cm and 5 m in width) and the pegmatites are often larger (between 5 cm and 50 m in width). All these bodies are vein or dyke-like, but many, in addition, have beaded or boudinaged shapes,

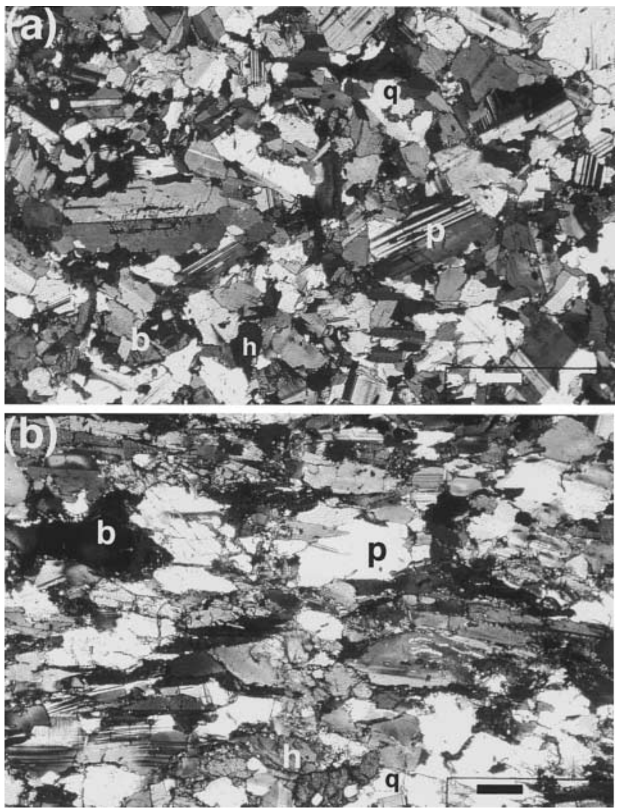

Fig. 7. Photomicrographs of granitoid fabrics. Location of samples are shown in Fig. 3. (a) Quartz diorite showing a pre-full crystallization fabric defined by the preferred orientation of plagioclase, hornblende and biotite crystals. (b) Within the same intrusion but closer to the contacts with the enclosing schists, quartz diorite shows more intense deformation as a high temperature solid-state fabric. Sections are perpendicular to the main fabric and parallel to lineation. Photographs taken in cross polars. p: plagioclase; q: quartz; b: biotite; h: hornblende. Scale bar is 1 mm.

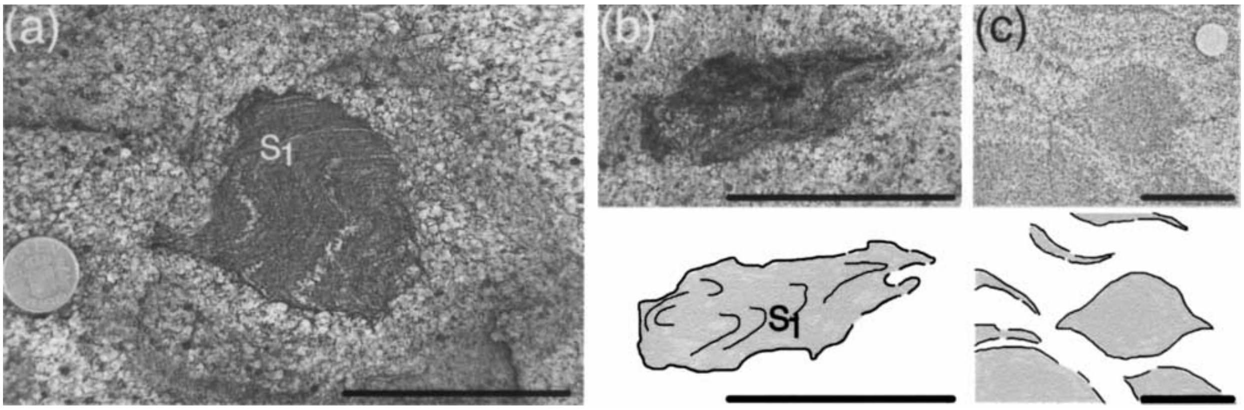

Fig. 8. Field photographs and sketches of metasedimentary xenoliths. Scale bar is 5 cm. Locations are shown in Fig. 3. (a) Garnet-rich granodiorite–tonalite containing a small metasedimentary xenolith which displays F_2 folds. (b) Within the same intrusion, a more diffuse country rock xenolith displays relics of F_2 folds. (c) Xenoliths where the internal D_2 structures have completely disappeared.

or occur in networks of foliation-parallel/foliation-oblique veins and dykes which suggest bridge structures related to dyke emplacement. These mainly subvertical E–W intrusions are in close parallelism with the dominant S_2 foliation in the country rocks (Figs 3 & 9a). In the leucogranites, an internal magmatic-state fabric, marked by the preferred orientation of biotite and feldspar, occurs as a continuation of the country rock foliation and, where leucogranites rarely cross-cut the external fabric, they may be folded with the magmatic state fabric parallel to the axial planes of the folds (Fig. 9b), indicating syntectonic emplacement.

There are rarely any magmatic-state fabrics in the pegmatites because of the massive nature of their crystals and the common occurrence of crystal growth-related preferred orientations at right angles to vein walls. However, in some of the largest pegmatites, an S_2-parallel magmatic banding can be observed. In the small number of cases where dykes have intruded at high angles to the dominant S_2 foliation they have been folded (Fig. 9c), but often locally cross-cut the external fabric. 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. A minor number of dykes are straight, undeformed and cross-cut S_2 foliation, probably indicating post-D_2 emplacement.

Consequently, as with the calc-alkaline suite, the leucogranites and most of the later pegmatites were also emplaced during the D_2 deformation history. However, whereas the earlier calc-alkaline units often cross-cut each other with rather vague and diffuse contacts, suggesting a small temperature interval between the respective times of emplacement, the sharp and often fine grained contacts of the later peraluminous bodies where they cut the earlier calc-alkaline units (Figs 5b & 9d), suggests that the peraluminous units intruded after the thermal maximum and crystallization of the earlier calc-alkaline units. This would be consistent with temperatures of crystallization of the pegmatites in the migmatite area, estimated by Alfonso (1995) to be around 580°C.

Xenoliths of the country rocks

On a smaller scale, a common feature of the granitoid bodies is their association with melting and assimilation of the deformed country rocks. Often one can see within a single outcrop: (a) country rock schists and metagreywackes containing D_2 deformational features (folds and cleavages); (b) within the igneous bodies and close to their margins there may be angular xenoliths of the country rocks containing obvious D_2 structures (Fig. 8a); (c) further within the intrusions the country rock xenoliths have lost their angular nature and have become more rounded; (d) within these more rounded xenoliths, the D_2 structures are very vague and vestigial (Fig. 8b) and the original deformational texture has almost disappeared; (e) there are other such xenoliths with diffuse contacts where the internal D_2 structures have completely disappeared (Fig. 8c). This suggests that the xenoliths have come close to complete melting before crystallizing again, i.e. they have been 'mobilized' (Pitcher and Berger, 1972). In addition, the igneous bodies often carry a D_2 fabric which may deform the mobilized xenoliths. These relationships again indicate that there was a strong overlap in time between deformation and intrusion.

CONCLUSIONS AND DISCUSSION

The general context of our findings in the Cap de Creus area is a few-km-wide, mid-crustal, steeply inclined transpressional shear zone associated with transcurrent shear and a steep stretching lineation. This structure began to develop during the main D_2 syn-metamorphic deformation and produced a strain gradient across the whole area from south to north. In the high strain northern part heterogeneous deformation led to second-order hectometric E–W zones of very high strain. The general deformation continued in late to post-metamorphic conditions (D_3) with the progressive localization of deformation along narrow mylonite bands. Associated with this geometry and history we have come to the following conclusions.

1. There is an increase in metamorphic grade across the area from muscovite–chlorite pelitic and metagreywacke assemblages in the south to cordierite–andalusite and then sillimanite grade assemblages in the high strain area in the north.

2. In the highest D_2 strain zones migmatites are found developed in close association with D_2 transpositional processes. These range from relatively widespread metatexitic/stromatic types, produced by incomplete melting of the metasediments to more restricted diatexitic/nebulitic types, indicative of more complete melting.

3. Two associations of magmatic rocks occur in the area. First, a suite of small (up to 50×60 m) quartz gabbro, quartz diorite, tonalite and granodiorite–granite bodies which are restricted to the highest metamorphic grade/highest strain zones (the migmatite complex). Secondly, a later suite of anatectic garnet-bearing leucogranites and pegmatites which occupy a wider, less restricted, area in the medium to high grade zones (the pegmatite area).

4. A wide variety of detailed field observations show that the early magmatic bodies were emplaced during the general syn-metamorphic D_2 deformational episode and that the bulk of the later leucogranites and anatectic pegmatites were emplaced during, and a small volume after, the main metamorphic peak and before the retrograde D_3 event.

5. Our principle general conclusion is that in a considerable crustal section across this transpressional

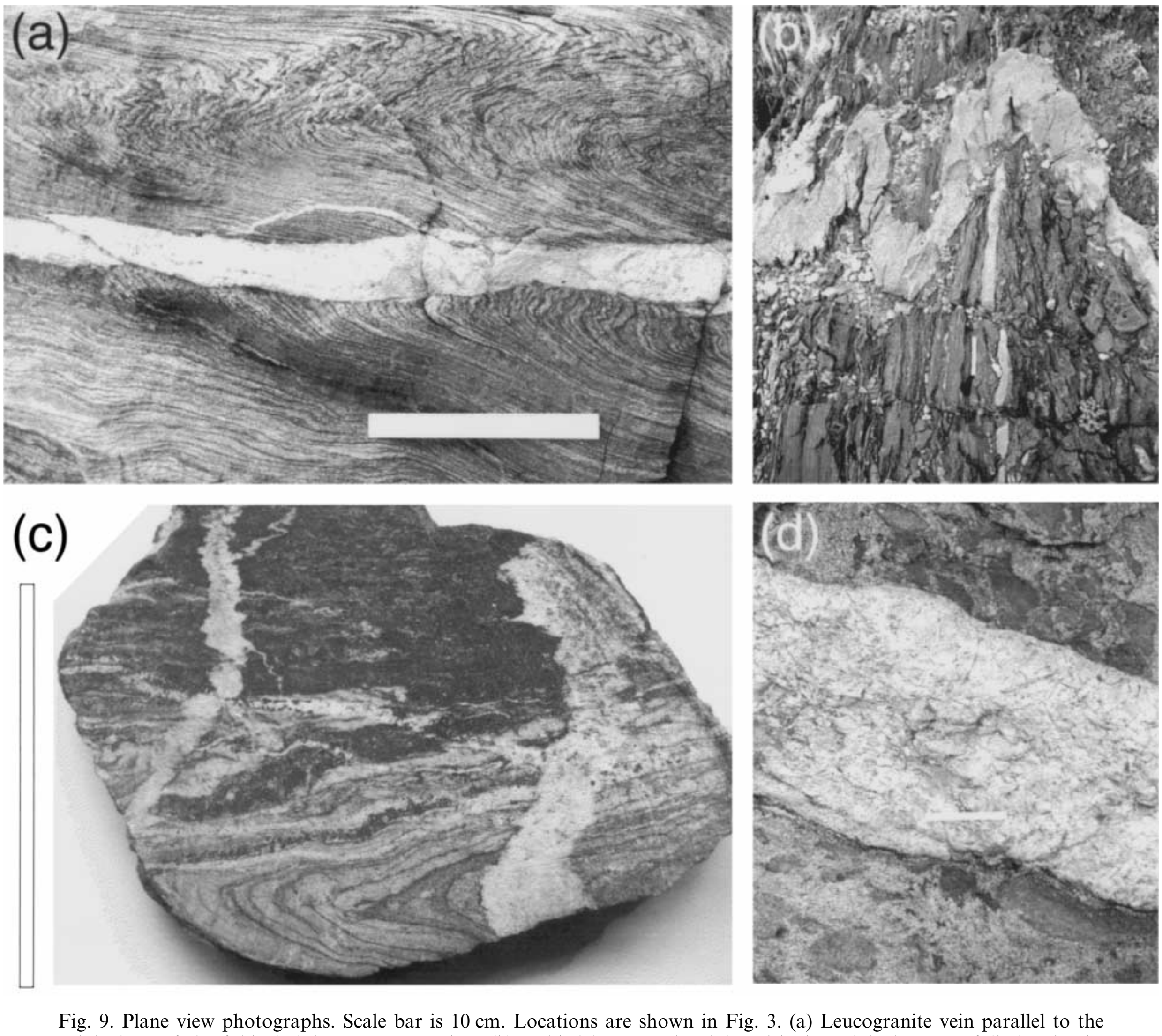

Fig. 9. Plane view photographs. Scale bar is 10 cm. Locations are shown in Fig. 3. (a) Leucogranite vein parallel to the axial plane of the folds (F_2) in metagreywackes. (b) Folded leucogranite dyke with the axial planar S_2 foliation in the enclosing migmatite schists. The internal magmatic state fabric may be correlated with the S_2 foliation. In the centre of the photograph, a boudined leucocratic vein occurs parallel to S_2 foliation. (c) Pegmatite veins coaxially folded with the enclosing metasediments but in a more open attitude. (d) Pegmatite dyke intruded with sharp contacts into an enclave-rich granitoid.

shear zone we can observe a strong spatial and temporal correlation between increasing strain, increasing metamorphic grade, increasing melting of the country rocks and the occurrence of magmatic bodies themselves.

This paper joins a growing number of contributions which supports the 'syntectonic granite' paradigm (Karlstrom, 1991; Brown, 1994) which states that not only do tectonic structures and deformation affect the ascent and emplacement of granites but they influence the generation of the magmas themselves. There are a number of ways in which it is thought this could happen. The first is that melting is enhanced in high strain zones because additional strain energy is provided there to allow early overstepping of the activation energy (Hand and Dirks, 1992). Once melt is present in a deformation zone, it lowers the matrix free energy of the system and aids progressive deformation by being a strain recovery process. This may be the fundamental reason why magma zones may become weak zones in the deforming lithosphere (Hollister and Crawford, 1986; Hollister, 1993; Tommasi *et al.*, 1994; Holdsworth, 1994; Fitzsimons, 1996). Secondly, the presence of tectonically induced differential stresses in melt regions is thought to be crucial in determining melt extraction rates and segregation processes since: (i) fabric anisotropy and differential stress in the source will create structural heterogeneities and pressure gradients to drive melts from 'sources to sinks' (Sawyer, 1994; Lucas and St-Onge, 1995; Brown *et al.*, 1995b; Rutter and Neumann, 1995; Watt *et al.*, 1996); (ii) confining pressure, buoyancy, volume change during dry melting and differential stress will combine to create an overpressure in the source, thus allowing rapid fracture-controlled escape of melts (Dell'Angelo and Tullis, 1987; Clemens and Mawer, 1992; Davidson *et al.*, 1994).

In the Cap de Creus example it is the progressive nature of the metamorphic, melting and magmatic effects as the strain increases across the zone that is striking. Whilst this may be because of a combination of the above processes, it may more likely reflect the progressive increase in differential/shear stress across the area in association with the development of the shear zone. In terms of ascent and emplacement, we note that the rocks of the calc-alkaline sequence probably intruded as sub-vertical lensoid bodies, both parallel and oblique to S_2 and along D_2-related tension fractures, Riedel shears and/or following the S_1 fabric. These latter features at this stage may have acted, in the scenarios described above, as conduits or feeders for migrating melts. These intrusions were folded by the progressive D_2 deformation and the sub-parallelism of many of them to S_2 may be purely because of post-emplacement deformation. During the later stages of the D_2 deformation, leucogranite and pegmatite dykes were injected parallel to the axial surfaces of F_2 folds (cf. Hand and Dirks, 1992), along S_2-oblique tension fractures, or into bridges between P-shears (cf. Tikoff and Teyssier, 1992), but also by using the already existing S_2 surfaces as planes of weakness (cf. Lucas and St-Onge, 1995). High strain zone anisotropy in this strongly contractional setting was probably an important factor controlling dyke orientations during the latest stages of emplacement (see Hutton, 1992, 1997) and implies high magma pressures combined with strong anisotropies. The steeply plunging stretching direction (i.e. transport direction) may have facilitated melt ascent (Wolf and Wyllie, 1995; Tikoff and Saint Blanquat, 1997).

Turning to more regional considerations, most workers have agreed that the general WNW–ESE Hercynian structural trend is the result of a crustal shortening event (Matte, 1976; Zwart, 1979). Recently, Carreras and Capellà (1994), Leblanc *et al.* (1996), Gleizes *et al.* (1997) and Evans *et al.* (1997) have determined that this main event is the result of dextral transpression and is synchronous with relatively high level crustal emplacement of the main Pyrenean Hercynian granite plutons. Our findings here, in the deeper levels of the Hercynian of the Pyrenees, are complementary to and in accord with these conclusions and it seems likely that we are looking at the deeper level processes of stress and strain controlled melting, melt migration and early ascent in transpressional shear zones that led to the higher crustal level accumulation and shear zone related emplacement of granite plutons.

Acknowledgements—The field work was financed by the D.G.I C.Y.T. (Spain) PB 91-0477 project. The elaboration of the paper has been financed by a D.G.I C.Y.T. grant for a stay in the University of Durham (E. Druguet) and partially financed by the D.G.I C.Y.T. (Spain) PB 94-0685 project. We thank M. Handy and A. Vauchez for helpful comments on a previous version of this manuscript. Profitable suggestions from J. Carreras and A. Teixell are also much appreciated. R. Vaquer (University of Barcelona) is thanked for the photomicroscopy facilities. The stereographic analysis was done with Stereonet, a program by R. W. Allmendinger.

REFERENCES

Alfonso, M. P. (1995) Aproximación a la petrogénesis de las pegmatitas del Cap de Creus. Unpublished Thesis, University of Barcelona.

Aranguren, A., Larrea, F., Carracedo, M., Cuevas, J. and Tubia, J. M. (1997) The Los Pedroches batholith (Southern Spain): polyphase interplay between shear zones in transtension and setting of granites. In *Granite: From Segregation of Melt to Emplacement Fabrics*, eds J. L. Bouchez, D. H. W. Hutton and W. E. Stephens, pp. 199–214. Kluwer Academic Publishers, Dordrecht.

Autran, A., Fonteilles, M. and Guitard, G. (1970) Relations entre les intrusions de granitoïdes, l'anatexie et le métamorphisme régional considerées principalment du point de vue du rôle de l'eau; cas de la chaîne hercynienne des Pyrénées Orientales. *Bulletin de la Societé géologique de la France* **12,** 673–731.

Blumenfeld, P. and Bouchez, J. L. (1988) Shear criteria in granite and migmatite deformed in the magmatic and solid states. *Journal of Structural Geology* **10,** 361–372.

Brown, M. (1973) Definition of metatexis, diatexis and migmatite. *Proceedings of the Geologists Association*, pp. 371–382.

Brown, M. (1994) The generation, segregation, ascent and emplacement of granite magma: the migmatite to crustally derived granite connection in thickened orogens. *Earth Sciences Reviews* **36,** 83–130.

Brown, M., Rushmer, T. and Sawyer, E. W. (1995a) Introduction to special section: mechanisms and consequences of melt segregation from crustal protoliths. *Journal of Geophysical Research* **100,** 15,551–15,563.

Brown, M., Averkin, Y. A., McLellan, E. L. and Sawyer, E. W. (1995b) Melt segregation in migmatites. *Journal of Geophysical Research* **100,** 15,655–15,679.

Carreras, J. and 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. and 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. and 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. and 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 de la Sociedad geológica de España* **7,** 21–29.

Carreras, J., Orta, J. M. and San Miguel, A. (1975) El area pegmatítica del litoral Norte de la península del Cabo de Creus y su contexto metamórfico y estructural. *Revista del Instituto de Investigaciones Geológicas* **30,** 11–34.

Clemens, J. D. and Mawer, C. K. (1992) Granitic magma transport by fracture propagation. *Tectonophysics* **204,** 339–360.

D'Lemos, R. S., Brown, M. and Strachan, R. A. (1992) Granite magma generation, ascent and emplacement within a transpressional orogen. *Journal of the Geological Society, London* **149,** 487–490.

Damm, K., Harmon, R. S., Heppner, P. M. and Dornseipen, U. (1992) Stable isotope constraints of the Cabo de Creus garnet–tourmaline pegmatites, massif des Alberes, Eastern Pyreenes, Spain. *Geological Journal* **27,** 76–86.

Davidson, C., Schimdt, S. M. and Hollister, L. S. (1994) Role of melt during deformation in the deep crust. *Terra Nova* **6,** 133–142.

Dell'Angelo, L. N. and Tullis, J. (1987) Experimental deformation of partially melted granitic aggregates. *Journal of Metamorphic Geology* **6,** 495–516.

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, Universitat Autònoma de Barcelona.

Druguet, E., Enrique, P. and Galán, G. (1995) Tipología de los granitoides y las rocas asociadas del complejo migmatítico de la Punta dels Farallons (Cap de Creus, Pirineo Oriental). *Geogaceta* **18,** 199–202.

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.

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

Fitzsimons, I. C. W. (1996) Metapelitic migmatites from Brattsstrand Bluffs, East Antarctic—Metamorphism, melting and exhumation of the mid crust. *Journal of Petrology* **37,** 395–414.

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

Grocott, J., Brown, M., Dallmeyer, R. D., Taylor, G. J. and Treloar, P. J. (1994) Mechanisms of continental growth in extensional arcs: an example from the Andean plate-boundary zone. *Geology* **22,** 391–394.

Guineberteau, B., Bouchez, J. L. and Vigneresse, J. L. (1987) The Mortagne granite pluton (France) emplaced by pull-apart along a shear zone; structural and gravimetric arguments and regional implication. *Geological Society of America Bulletin* **99,** 763–770.

Hand, M. and Dirks, P. H. G. M. (1992) The influence of deformation on the formation of axial-planar leucosomes and the segregation of small melt bodies within the migmatitic Napperby Gneiss, central Australia. *Journal of Structural Geology* **14,** 591–604.

Holdsworth, R. E. (1994) Structural evolution of the Gander–Avalon terrane boundary: a reactivated transpression zone in the NE Newfoundland Appalachians. *Journal of the Geological Society, London* **151,** 629–642.

Hollister, L. S. (1993) The role of melt in the uplift and exhumation of mountain belts. *Chemical Geology* **108,** 31–48.

Hollister, L. S. and Crawford, M. L. (1986) Melt-enhanced deformation: a major deformation process. *Geology* **14,** 558–561.

Hutton, D. H. W. (1988) Granite emplacement mechanisms and tectonic controls: inferences from deformation studies. *Transactions of the Royal Society of Edinburgh: Earth Sciences* **79,** 245–255.

Hutton, D. H. W. (1992) Granite sheeted complexes: evidence for the dyking ascent mechanism. *Transactions of the Royal Society of Edinburgh: Earth Sciences* **83,** 377–382.

Hutton, D. H. W. (1997) Syntectonic granites and the principle of effective stress: a general solution to the space problem. In *Granite: From Segregation of Melt to Emplacement Fabrics*, eds J. L. Bouchez, D. H. W. Hutton and W. E. Stephens, pp. 189–197. Kluwer Academic Publishers, Dordrecht.

Hutton, D. H. W. and Ingram, G. M. (1992) The Great tonalite Sill of southeast Alaska and British Columbia: emplacement into an active contractional high angle reverse shear zone. *Transactions of the Royal Society of Edinburgh: Earth Sciences* **83,** 383–386.

Hutton, D. H. W. and Reavy, R. J. (1992) Strike-slip tectonics and granite petrogenesis. *Tectonics* **11,** 960–967.

Ingram, G. M. and Hutton, D. H. W. (1994) The great Tonalite Sill: Emplacement into a contractional shear zone and implications for Late Cretaceous to early Eocene tectonics in southeastern Alaska and British Columbia. *Geological Society of America Bulletin* **106,** 715–728.

Karlstrom, K. E. (1991) Towards a syntectonic paradigm for granitoid. *EOS, Transactions of the American Geophysical Union* **70,** 762.

Leblanc, D., Gleizes, G., Roux, L. and Bouchez, J. L. (1996) Variscan dextral transpression in the French Pyrenees: new data from the Pic des Trois-Seigneurs granodiorite and its country rocks. *Tectonophysics* **261,** 331–345.

Lucas, S. B. and St-Onge, M. R. (1995) Syn-tectonic magmatism and the development of compositional layering, Ungava Orogen (northern Quebec, Canada). *Journal of Structural Geology* **17,** 475–491.

Matte, Ph. (1976) Raccord des segments hercyniens d'Europe sud-occidentale. *Nova Acta Leopoldina* **224,** 239–262.

McCaffrey, K. (1992) Igneous emplacement in a transpressive shear zone. Ox Mountains igneous complex. *Journal of the Geological Society, London* **149,** 221–235.

Pau, B. (1995) Anàlisi de lantànids mitjançant ICP-AES. Aplicació a l'estudi de les terres rares de les roques ígnies i metamòrfiques del Cap de Creus. Unpublished Tesi de Llicenciatura, Universitat de Barcelona.

Pitcher, W. S. and Berger, A. R. (1972) *The Geology of Donegal: A Study of Granite Emplacement and Unroofing*. Wiley Interscience, New York.

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

Rushmer, T. (1996) Melt segregation in the lower crust: how have experiments helped us? *Transactions of the Royal Society of Edinburgh: Earth Sciences* **87,** 73–83.

Rutter, E. and Neumann, D. (1995) Experimental deformation of partially molten Westerly granite under fluid-absent conditions with implications for the extraction of granitic magmas. *Journal of Geophysical Research* **100,** 15,697–15,716.

Sawyer, E. W. (1994) Melt segregation in the continental crust. *Geology* **22,** 1019–1022.

Sawyer, E. W. (1996) Melt segregation and magma flow in migmatites: implications for the generation of granite magmas. *Transactions of the Royal Society of Edinburgh: Earth Sciences* **87,** 85–94.

Tikoff, B. and Teyssier, C. (1992) Crustal-scale, en échelon 'P-shear' tensional bridges: A possible solution to the batholithic room problem. *Geology* **10,** 927–930.

Tikoff, B. and Saint Blanquat, M. (1997) Transpressional deformation and changing kinematics in the Late Cretaceous Sierra Nevada magmatic arc, California. *Tectonics* **16,** 442–459.

Tommasi, A., Vauchez, A., Fernandes, L. A. D. and Porcher, C. C. (1994) Magma-assisted strain localisation in an orogen-parallel transcurrent shear zone of southern Brazil. *Tectonics* **13,** 421–437.

Watt, G. R., Burns, I. M. and Graham, G. A. (1996) Chemical characteristics of migmatites: accessory phase distribution and evidence for fast melt segregation rates. *Contributions to Mineralogy and Petrology* **125,** 100–111.

Wolf, M. B. and Wyllie, P. J. (1995) Liquid segregation parameters from amphibolite dehydration melting experiments. *Journal of Geophysical Research* **100,** 15,611–15,621.

Zwart, H. J. (1979) The geology of the central Pyrenees. *Leidse Geologische Mededelingen* **50,** 1–74.