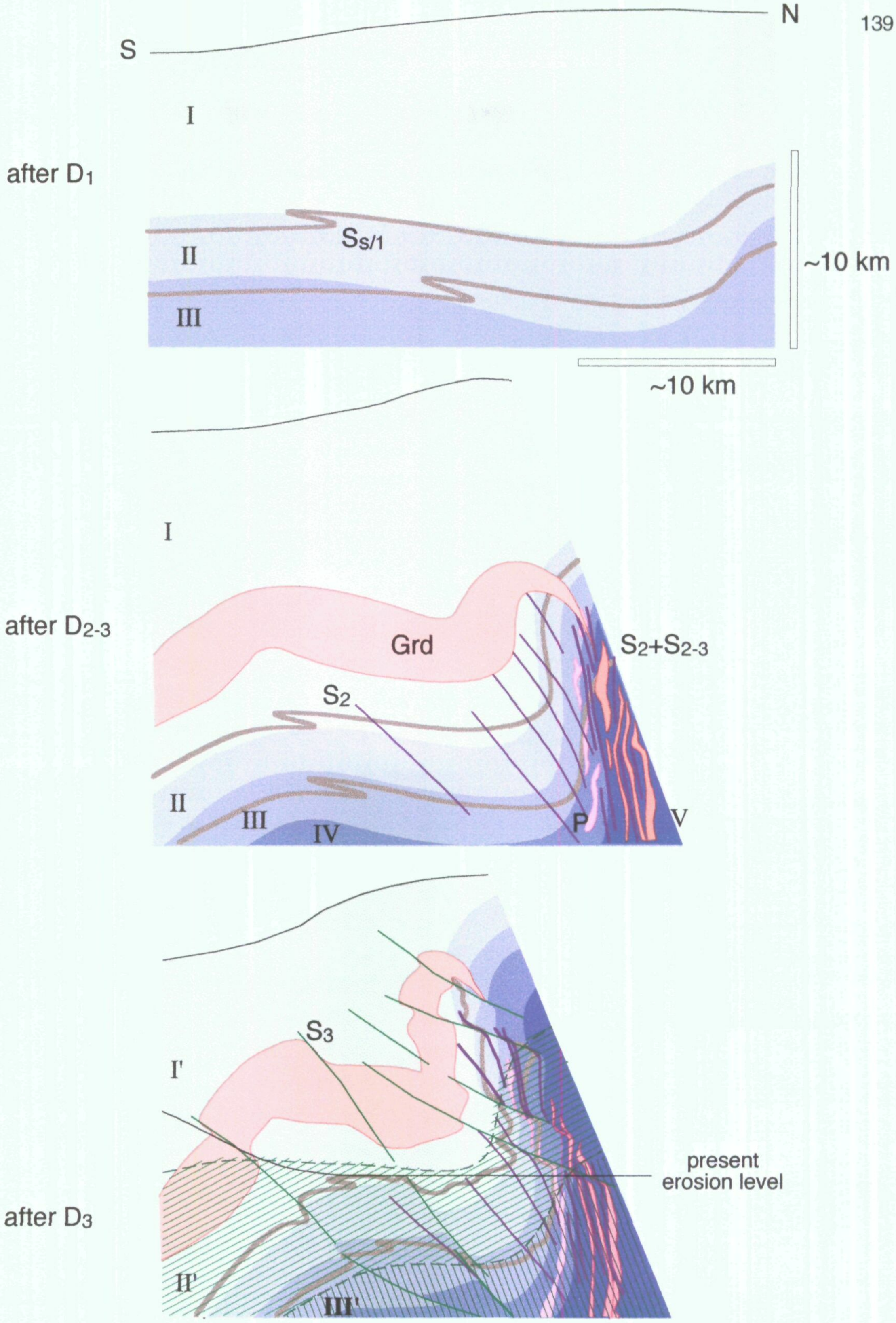

Fig. 104. Idealized cross-sections of a model for the development of "tectonoplutonometamorphism" in the Cap de Creus peninsula, with especial reference to the study area. Three ideal stages of the geological evolution have been considered: after D_1, after $D_{2\text{-}3}$ and after D_3 deformations.

Zones of prograde metamorphism (in blue): I non-metamorphic to chlorite-muscovite zone, II biotite zone, III cordierite-andalusite zone, IV sillimanite zone, V migmatite areas. Intrusive rocks: Grd granitoids (from the migmatite complexes and Roses and Rodes stocks), P pegmatites. Zones of retrograde metamorphism (in green): I' non- or very low metamorphic, II' lower greenschist facies, III' upper greenschist facies.

After D_1: early deformations produce a gently east dipping schistosity and onset of regional metamorphism.

After $D_{2\text{-}3}$: folding and developing of high strain zones in the north, with granitoid sheets and pegmatites intruded vertically and equally sub-vertical high thermal gradient. The Roses and Rodes granodiorites are speculated to intrude stratoidally in shallower levels, having their root in the migmatite areas. 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).

The Roc de Frausa massif displays a rather similar tectonic evolution than the Cap de Creus, but contains a orthogneissic dome-shaped gneissic core (Autran et al. 1970, Liesa & Carreras 1989, Liesa 1994). The structure of this massif is attributed to three main deformational phases: an originally flat-lying regional schistosity and two steep fold systems, trending NE-SW and NW-SE. If these fold systems correlate with F_2 and F_3 folds in the Cap de Creus massif, then it follows that there is a manifest difference between the structural domes with metamorphic cores (where flat-lying foliations are most prevalent) and the Cap de Creus massif, where the metamorphic belt coincides with a domain with prevalent steep foliations. However, one should not ignore some style similarities between flat S_2 in the Canigó and steep S_2 in the Cap de Creus area and the penecontemporaneity of both fabrics with the metamorphic climax.

Concerning the bulk tectonic setting of the Hercynian of the Pyrenees, evidence presented from the Cap de Creus structure is in agreement with the transpressive regime proposed for the main Hercynian event in the Pyrenees by Carreras & Capellà (1994) and also by Gleizes et al. (1997). According to the first authors, the structural history is interpreted as a gradual evolution from an earlier compressive to a late transpressive and finally transcurrent regime, implying a continuous horizontal crustal shortening. Furthermore these authors have noticed a on-going predominance in time of transpressive shear zones, especially located towards the eastern Axial Zone, which might represent a more internal part of the Pyrenean segment of the Hercynian belt. A shear zone belt would connect the Hercynian segment in the eastern Pyrenees with the SW French Massif Central and the South Armorican zone. (Carreras & Capellà 1994, Fig. 8b). Analysis of structures in the Cap de Creus area supports this hypothesis. No evidence has been found suggesting a regional extensional event nor the existence of lithological omission.

7.2. CONCEPTUAL ASPECTS

On the correlation of structures in complex deformation zones

The structural analysis of the Cap de Creus exemplifies the limitations of the classic methods of identification and correlation of deformation events in complex mid crustal domains undergoing progressive non-coaxial inhomogeneous deformation interactive with metamorphic and magmatic processes. Orientation and style criteria have to be carefully handled, as both can change significantly in space for a given deformation event. Overprinting criteria in all scales, together with relationships with metamorphism and magmatism, are very useful tools.

However, as evidenced in this work, as a consequence of the strong gradients of metamorphism related to plutonometamorphic processes, the relationships between foliations and blastesis should be treated with great caution. In such settings, a deformational event can develop at different metamorphic conditions over relatively short distances (i.e. a few kilometers). Syntectonic intrusive rocks, rather localized in space and time, especially the pegmatite dykes, can be used as indicators of relative timing of deformations. In this way, dykes enable to distinguish between events pre-dating, contemporaneous and post-dating their emplacement. These criteria enables to show that a specific event can give rise to significant different structures over short distances (e.g. refolded folds and crenulated crenulations in one locality, in contrast with an unique fold system and an associated crenulation cleavage in another place).

On the relations between geometry, kinematics and tectonic transport in high strain zones

It is commonly accepted that deep crustal high-strain zones differ in many aspects from well known greenschist facies upper crustal shear zones (e.g. Ramsay 1980; Carreras 1997; Passchier & Trouw 1996). In deeper high strain zones, deformation is not by plane strain flow, and volume changes associated with igneous intrusion and partial melting are more common. In some cases, even the familiar relationship of parallelism between stretching lineations and displacement direction

(Ramsay 1980), common in low and medium grade mylonites deformed by simple shear, is absent, and lineations may appear to lie oblique to the displacement direction of the shear zone (Robin and Cruden 1994, Tikoff & Greene 1997).

The present work confirms the need to be careful when interpreting the relations between geometry (large and minor structures) and kinematics, especially in the use of kinematic indicators in complex strain zones. The detail analysis of the high strain zones developed during the D_2 to D_{2-3} event reveal an apparent contradiction between shear sense deduced with reference to the bedding-S_1 foliation (sinistral) and the shear component involved in bulk deformation of the S_1 surfaces, considering an external reference frame (dextral). Furthermore, a three-dimensional examination of the structure reveals that, although larger rotations occurred, the main stretching direction was sub-vertical, showing that, in fact, main extension and main shearing directions would make a low angle with the rotation or bend axis of different elements. In addition, and in analogy to what observed in horizontal view, a normal component of shearing in the vertical section (north downward) is in apparent contradiction with the deduced components of the displacement direction (north southward and predominantly upwards), related to the real tectonic transport (i.e. vergence of structures). It is considered that geometry and kinematics are highly influenced by the presence of a previous anisotropy, which controls the way in that a given deformation is accommodated. It follows that in this setting, and presumably in analogous settings, the concept of dextral and sinistral, normal or reverse is poorly useful when understanding the bulk kinematics and when deducing the tectonic regime operating at crustal scale.

On the metamorphic setting

The Cap de Creus massif is an illustrative example of the interaction between deformation, metamorphism and magmatism in an orogenic belt governed by a regional transpressive regime. The high thermal gradients, evidenced here and in many other similar settings, can be explained on the basis of strong horizontal strain gradients superimposed on already narrow metamorphic zones. The continuity of some stratigraphic markers (e.g. quartzites), from low to high grade metamorphic zones, and the absence of significant truncation of the structures and zones of isometamorphism, exclude the existence in this area of any detachment or major fault responsible for the strong metamorphic gradients. Late shear zones, although locally affect the isogrades and the boundaries of the migmatite complexes, involve offsets of a few hundred meters of displacement. Progression of deformation during cooling is responsible for narrowing of the metamorphic zones. The presumably quick evolution from high to low temperature deformation requires rapid uplift and denudation of the crustal domains overlying the high strain metamorphic belt.

On syntectonic intrusions

An evident feature of this mid-crustal setting is the close connection between deformation and emplacement of igneous bodies (granitoids and pegmatites). Small bodies and dykes emplace syntectonically in high strain zones developed in a transpressive setting, during and just after the metamorphic peak. It appears that deformation in the crust favours opening of spaces irrespectively of the tectonic regime, in a similar way as hydrothermal or segregation veins do in deforming low grade rocks. However, in a horizontal view, there is a striking parallelism between the predominant orientation of lensoid magmatic bodies and dykes and the trend of the axial planes of folds developed synmagmatically. This orientation appears to contradict a priori the expected orientation of dykes, at a high angle to the extension direction. Such parallelism between traces of dykes and traces of axial planes (i.e. planes of flattening) can be in part explained in several ways:

1) Synmagmatic folds have their axes parallel to the stretching lineation. Thus, taking into account the attitude of the expected opening fractures (i.e pure extensional or shear-extensional) in regard to the strain ellipsoid, these are oriented in such a way that, in a plane close to the YZ section, the fractures tend to form a small angle with Y (i.e. the trace of the foliation in the horizontal surface).

2) It appears that most dykes emplace along hinge zones or short limbs of incipient folds.

Overpressures of pegmatite melts might exceed the tensile strengths across the anisotropy at these intrusive sites (i.e. the axial planes) before intrusion took place. The required small differential stress is consistent with the ductile character of strain at this deep levels, where deformation takes the form of plastic flow.

3) Rotation of the short limbs might cause that the mechanical anisotropy enters in the shortening field. In consequence, a layer perpendicular stretching would propitiate decollement between the foliation planes, and emplacement of dikes along dilatation holes. This could explain dykes formed by arrays of boudins, which most likely intruded in that shape since there is a continuity of layers across the interboudins.

4) Syntectonic dykes record an amount of deformation (change in length and orientation) depending on their initial orientation. Except for dykes parallel to the eigenvectors, a rotation towards parallelism with the flow plane would imply a obliquity decrease. However, there is no evidence of parallelism being achieved by large rotations.

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