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The high-pressure metamorphic front of the south Western Alps (Ubaye-Maira transect, France, Italy)

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Abstract

The Ubaye-Acceglio transect (south Western Alps) is a unique rock section showing the roughly progressive evolution of metamorphism from low grade, high-pressure greenschist facies to carpholite-quartz (blueschist) facies to low-temperature eclogite facies. The section, corresponding to the Middle Penninic Briançonnais zone, is exposed along the upper Ubaye, Varaita and Maira valleys, and is ca. 10 km thick. Our P–T estimates indicate that from the petrological point of view the section exposed corresponds to a pressure interval of ca. 0.7 GPa. Depending on rock density this interval is expected to represent a 20 km thick portion of the Alpine orogen. Although an extensional discontinuity was not detected by us along the transect, its thickness is significantly reduced. This feature is particularly intriguing as post-peak metamorphism backthrusting dominates the structure of the section, so that excess thickening is expected.

We present a structural and petrologic study of the transect, including for the first time a metamorphic map of mineral occurrences. Along the transect, the Briançonnais zone consists of a number of rock slices (nappes) comprising duplexes that were piled through the operation of décollement in Lower Permian, Upper Scythian and Carnian levels. NW-directed D_1 deformation, associated to the duplex piling, occurred during the subduction of the Briançonnais lithosphere. This was followed by an oblique shortening (D_2) with longitudinal, reverse sinistral strike-slip faults and coeval NW-trending major folds. Further tightening (D_3) carried the external Briançonnais onto the frontal part of the Upper Penninic prism, i.e. the Helminthoid Flysch nappes (Briançonnais Front), and contemporaneously resulted in strong nappe backfolding and backthrusting. Both D_2 and D_3 occurred in greenschist-facies conditions. Contractional piling was followed by horizontal kink bands that may record late orogenic collapse.

At the top (west) of the pile, the external Briançonnais units equilibrated at ca. 0.6 GPa, 300 °C. Carpholite-quartz and/or lawsonite-glaucophane assemblages (ca. 1.1 GPa, 350 °C) occurs in the internal Briançonnais-external Ultrabriançonnais units as well as in the Schistes Lustrés of the Acceglio D_3 syncline. The most internal Ultrabriançonnais units of the Acceglio-Longet stripe and the underlying Schistes Lustrés at the base (east) of the section display jadeite-quartz and/or zoisite-jadeite-glaucophane assemblages (ca. 1.3 GPa, 430 °C).

The external Briançonnais units, including Middle–early Late Eocene flysch, reached their maximum depth not before ca. 35 Ma. Assuming 80 km for the restored width of the Briançonnais plateau and a subduction angle of ca. 45°, the most internal Briançonnais units potentially reached >60 km depth then, but petrological and geochronological data indicate that they accreted to the upper plate and equilibrated at ca. 40 km depth at 36–38 Ma. Exhumation began at the end of D_1 within the subduction channel-accretionary edifice, carrying deeply metamorphosed units onto, or closer to less metamorphic ones. Further decompression and cooling during the D_2 – D_3 shortening phases is interpreted as the result of extension in the upper levels, and erosion on top of the Schistes Lustrés wedge during the Oligocene Alpine collision.

Keywords: High-pressure metamorphism, blueschists, subduction, exhumation, Western Alps.

1. Introduction

Being one of the most extensively studied orogenic belts, the Western Alps has become a key region where the dynamics, kinematics and rates of subduction and exhumation of high-pressure, low-temperature (HP-LT) metamorphic rocks are topics of major interest. Whereas the character of high and ultra-high pressure metamorphism in the internal parts of the Western Alps has gained considerable attention in recent years, features of the more external parts of the orogen where metamorphism is waning from blueschist to low-greenschist facies, are still not well understood. At the scale of the entire Western Alps, this transitional domain, here referred to as the front

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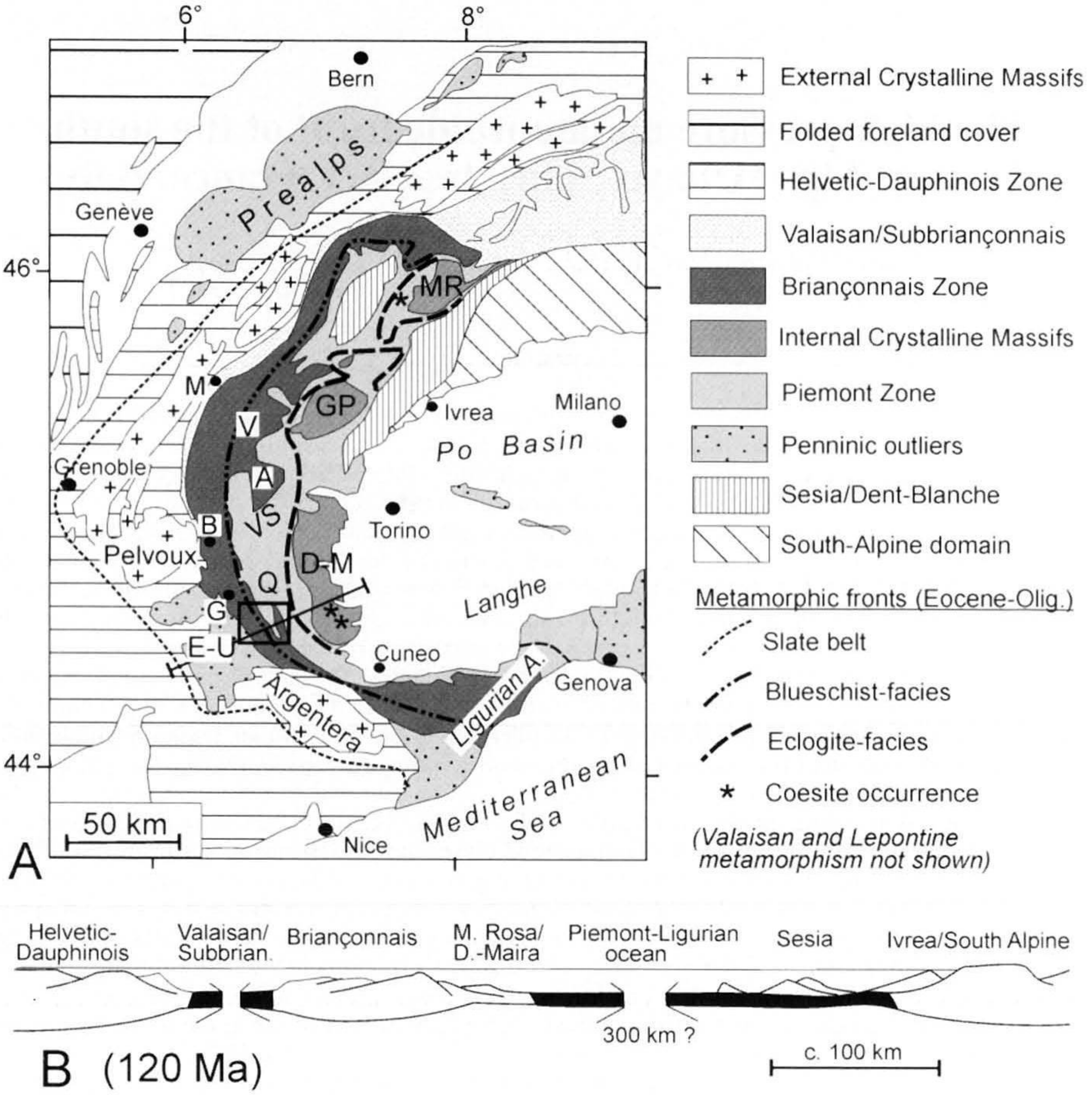

Fig. 1 (A) Sketch map of the Western Alps with location of the studied area (framed) and broad metamorphic zonation after Dal Piaz and Lombardo (1986) and Goffé and Chopin (1986). A: Ambin; B: Briançon; D-M Dora-Maira; E-U: Embrunais-Ubaye; G: Guillestre; GP: Gran Paradiso; M: Moûtiers; MR: Monte Rosa; Q: Queyras; V: Vanoise. (B) Current restoration of the Alpine domain in cross-section before plate convergence (e.g. Schmid et al., 1996; Michard et al., 1996; Dal Piaz et al., 2001).

of the HP-LT metamorphism, roughly corresponds to the Briançonnais or Middle Penninic Zone (Fig. 1A). The western boundary of blueschist-facies metamorphism runs within the Briançonnais Zone, or at its contact with the Piemont or Schistes Lustrés (Upper Penninic) Zone, whilst the eclogite-facies metamorphic equilibration occurs more easterly within the ophiolite-bearing Piemont-Ligurian units and underlying Internal Crystalline Massifs (ICM, Monte Rosa, Gran Paradiso and Dora-Maira massifs; e.g. Goffé and Chopin, 1986; Michard et al., 1996). Therefore, the Briançonnais appears as a key zone for understanding the evolution of the Alpine subduction as, in spite of being partly affected by high-pressure, low-temperature (HP-LT) metamorphism, it also includes Late Cretaceous to Late Eocene fossiliferous formations, coeval with the subduction-collision history of the belt.

The pre-orogenic Briançonnais domain is classically restored (Lemoine et al., 1986) as a submarine, Late Jurassic–Paleocene continental plateau between the Lower Penninic, Valaisan-Subbriançonnais trough and the Upper Penninic, Piemont-Ligurian ocean (Fig. 1B). This plateau was interpreted either as a major block (or horst) of the distal European margin (Lemoine et al., 1986, 2000), or as the northern tip of an Iberia/Briançonnais plate, separated from the European plate by the Pyrenean-Valaisan oceanic rift (Stampfli, 1993; Stampfli et al., 1998). On the other hand, the broad distribution of the HP-LT metamorphic zoning from the Briançonnais to the east is currently regarded as the manifestation of a SE-dip-

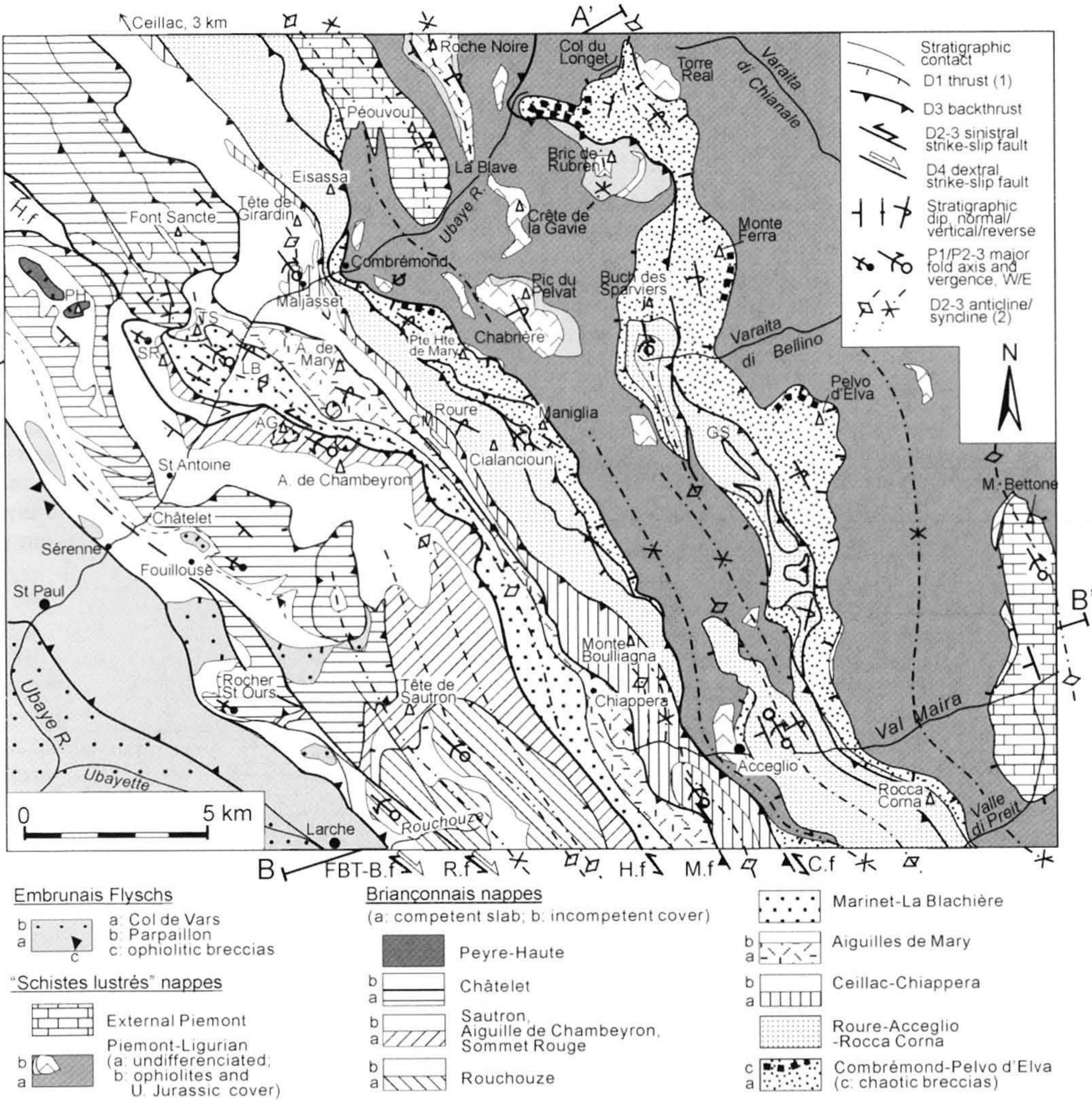

Fig. 2 Structural map of the Ubaye-Maira transect, after Gidon et al. (1994), Michard and Henry (1988), Lefèvre and Michard (1976) and unpublished field data. For location see Fig. 1. Abbreviations: AG: Aiguille Grande; BC: Brec de Chambeyron; B.f: Bersezio fault; C.f: Ceillac fault; CM: Col de Mary; FBT: Frontal Briançonnais Thrust; GS: Grangie Sagnères (polymetamorphic schists sliver); H.f: Houerts fault; LB: La Blachière; M.f: Col de Mary fault; Pa: Panestrel; PH: Pic des Houerts; PBM: Pointe Basse de Mary; R.f: Ruburent fault; SR: Sommet Rouge; TS: Tête du Sanglier (Tête du Seingle). AA'/BB': traces of combined cross-section of Fig. 4. Inverted Ω, ruled: Serpentinite cave near Combrémond, abandonned.

ping subduction of the European-Penninic plate beneath the margin of the Adriatic plate (Austroalpine domain), the ICM being located between the Briançonnais and Piemont-Ligurian domains (Goffé and Chopin, 1986; Avigad et al., 1991; Michard et al., 1996; Dal Piaz et al., 2001). However, alternative views have been also proposed concerning the direction of the subduction (Caby, 1996) or the initial location of the ICM (Gebauer, 1999; Froitzheim, 2001).

The aim of the present paper is to document the timing and mechanisms of subduction and exhumation in the Briançonnais units along the Ubaye-Maira transect (Fig. 2). This area offers good opportunities to disclosing the geometry and tectonic evolution of the frontal part of the Alpine HP-LT tectonic wedge through deep natural sections in the Briançonnais nappe stack and adjoining units. Previous studies (Gidon, 1962; Michard, 1967; Lefèvre and Michard, 1976; Michard and Henry, 1988; Platt et al., 1989; Gidon et al., 1994) yield detailed geologic mapping and structural descriptions along the transect. Additional structural and petrological investigations are available in the Acceglio-Longet stripe and adjoining Schistes Lustrés (Houfflain and Caby, 1987; Caby, 1996; Schwartz et al., 2000; Agard et al., 2000, 2001a) . In the following, we present for the first time a metamorphic study of the whole transect, combined with a description of the nappe structure. Finally, a tectonic interpretation of the subduction-exhumation history of the transect is proposed, based on the paleogeographic restoration of the Briançonnais domain and on the structural and mineralogical constraints concerning each of the Briançonnais units.

2. Geological setting

The studied Briançonnais transect (Fig. 2) is bounded to the west by the Embrunais-Ubaye nappes, which mainly consist in this area of Late Cretaceous–Paleocene Flysch units of Piemont-Ligurian (Upper Penninic) origin, i.e. the Cenomanian–Turonian flysch of the Col de Vars nappe, and the Senonian–Paleocene Helminthoid Flysch of the Parpaillon nappe (Gidon et al., 1994). The Subbriançonnais Zone is buried beneath the Briançonnais-Upper Penninic Flysch wedge, and only appears in the more uplifted transects of the Argentera and Pelvoux massifs, south and north of our transect respectively (Gidon, 1977; Barféty et al., 1995).

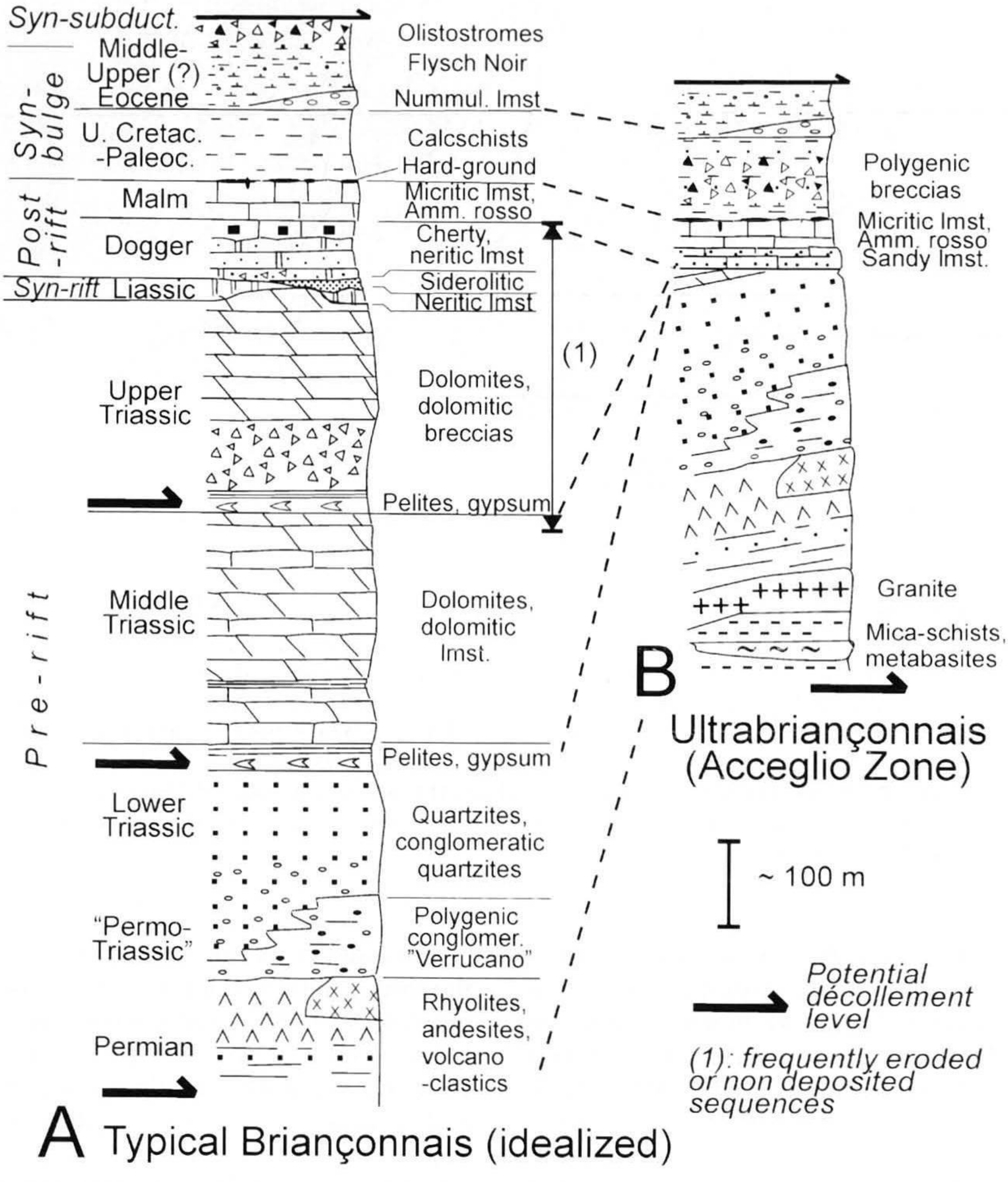

Fig. 3 Lithostratigraphy and décollements levels. (A) Ideally complete stratigraphic column of the external and median units. (B) Stratigraphy of the most internal units = Ultrabriançonnais or Acceglio zone. After Lefèvre and Michard (1976), Lemoine et al. (1986), Michard and Henry (1988) and Michard and Martinotti (2002).

The Briançonnais (sensu lato) units or nappes can be divided into two groups (Lefèvre, 1984; Gidon et al., 1994), i.e. the typical Briançonnais to the west, and the Ultrabriançonnais to the east. The typical Briançonnais display stratigraphic sequences which derive from an idealised, complete sequence (Fig. 3A) by detachment and/or erosion. The Peyre-Haute nappe (mostly developed in the Guillestre area to the north; Debelmas and Lemoine, 1962; Mégard-Galli, 1972a, b) is detached on the Carnian evaporites, and suffered few Liassic erosion before the sedimentation of the condensed Middle–Late Jurassic to Eocene sequence (Table 1). The underlying, Châtelet-Fontsancte, Aiguille de Chambeyron-Sommet Rouge and Sautron nappes are detached on the Upper Werfenian evaporites, and the Rouchouze nappe on a Lower Permian pelitic level. However, the latter four nappes of the external Briançonnais suffered a more important Liassic erosion than the Peyre-Haute nappe, so as part of the Triassic dolomitic carbonates are lacking in these nappes. The Marinet-La Blachière and Aiguilles de Mary nappes (median Briançonnais) widely expose Permian volcanics and Permo-Triassic silicoclastics at the bottom of a relatively thin Middle Triassic–Paleocene sequence. The Ceillac-Chiappera unit (internal Briançonnais) is detached in the south on a deeper level than in the north (Permian and Carnian levels, respectively). All typical Briançonnais units end upward with a "Flysch noir" formation, dated from the Middle to possibly early Late Eocene (Gidon et al., 1994; Barféty et al., 1995).

The most internal Briançonnais units (Acceglio Zone = Ultrabriançonnais: Debelmas and Lemoine, 1957; Lemoine, 1957; Lefèvre and Michard, 1976; Lefèvre, 1984) are characterised by an almost complete (Roure-Acceglio-Rocca Cor-

Table 1 Nomenclature and caracteristics of the Briançonnais nappes of the Ubaye-Maira transect, after Gidon et al. (1994), and personal observations. L.: Lower; M.: Middle; U.: Upper.

Nappe	Detachment level	Litho-stratigraphy
Peyre-Haute	Carnian evaporites	Norian dolomites, M-U. Jurassic carbonates, U. Cretaceous-Paleocene calcschists, Eocene flysch
Châtelet -Fontsancte	Upper Werfenian evaporites	M.-U. Triassic & M.-U. Jurassic carbonates, U. Cretaceous-Paleocene calcschists, unconformable Eocene flysch with Triassic and Cretaceous flysch olistoliths
Aiguille de Chambeyron-Sommet Rouge	Upper Werfenian evaporites	M.-Triassic carbonates, U. Jurassic marbles, U. Cretaceous-Paleocene calcsch. & breccias (Triassic elements), Eocene wildflysch
Sautron	U. Werfenian evaporites	cf. Châtelet nappe, except the lack of U. Jurassic marbles (unconformable U. Cretac.)
Rouchouze	Lower Permian	U. Permian-L. Triassic quartzites, younger levels cf. Sautron nappe
Marinet-La Blachière	L.-U. Permian	North: rare Permian volcanites, U. Permian conglomeratic quartzites, L. Triassic quartzites. South: Permian-Triassic quartzites, M. Triassic-U. Jurassic carbonates, U. Cretaceous-Paleocene calcschists
Aiguilles de Mary	U. Carboniferous-L. Permian	Permian volcanites, U. Permian-L. Triassic quartzites, M.-U. Triassic carbonates, U. Jurassic radiolarites & carbonates, U. Cretaceous-Paleocene calcschists
Ceillac-Chiappera	North: Carnian evaporites; South: L. Permian	Permian-Triassic quartzites (S), M.-U. Triassic & M.-U. Jurassic carbonates, L. Cretaceous (?) clays & micrites, U. Cretaceous-Paleocene calcschists, Eocene flysch
Roure and Acceglio-Rocca Corna	Basement-Lower Permian unconformity	U. Permian volcanites & conglom. quartzites, L. Triassic quartzites, relics of M. Triassic carbonates & U. Jurassic marbles, U. Cretaceous-Paleocene calcschists, Eocene flysch
Combrémond and Pelvo d'Elva (Longet)	Basement schists	Basement schists and metabasites, Permian granites & volcanites, Permian-L. Triassic quartzites, U. Jurassic marbles, U. Cretaceous-Paleocene polygenic breccias & calcschists, Eocene flysch

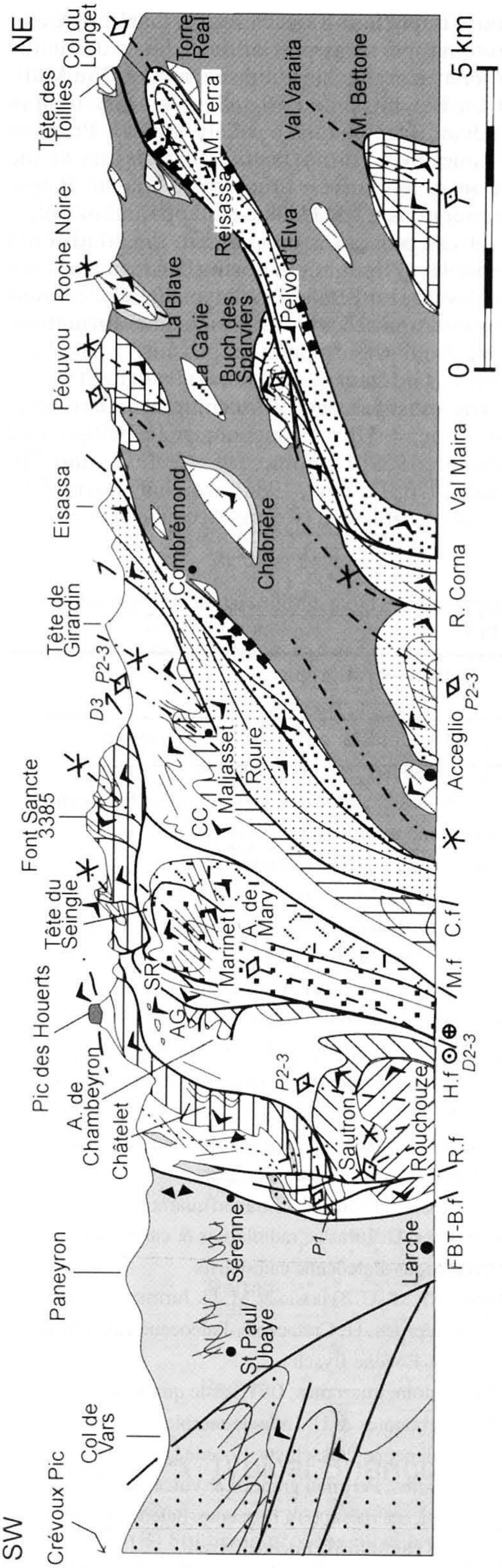

Fig. 4 Schematic cross-section of the Ubaye-Maira-Varaita transect. For location and legend see Fig. 2 (structures from the southern part of the map, trace BB', are projected onto the vertical plane following trace A, north of the Ubaye valley).

na nappe) or complete erosion (Combrémond-Pelvo d'Elva nappe) of the Triassic carbonates during the Liassic. They would have been sampled at the eastern shoulder of the Briançonnais rifted plateau (Fig. 1B; Lemoine, 1967). The Upper Cretaceous–Eocene (?) formations of the Ultrabriançonnais units include spectacular chaotic breccias reworking basement crystalline rocks and huge Triassic and Jurassic carbonate blocks ("Alpet breccias", Lemoine 1967; Gidon et al., 1994). The Pelvo d'Elva nappe includes slices of metamagmatic and polymetamorphic rocks originating from an Hercynian basement (Lefèvre and Michard, 1976; Monié, 1990).

The Ultrabriançonnais units are in contact with various units of the Schistes Lustrés, mainly with the ophiolite-bearing Piemont-Ligurian units, but also with the External Piemont units which include Upper Triassic dolomites and Liassic calcareous breccias and were formerly transitional between the Briançonnais and Piemont-Ligurian domains (Lemoine, 1967; Michard, 1967; Lemoine et al., 1986). Both the External Piemont and Piemont-Ligurian sequences include metasediments of Late Cretaceous age, partly equivalents to those of the Flysch nappe outliers (Lemoine, 2003). Critical for deciphering the Briançonnais-Schistes Lustrés tectonic relationships is the occurrence of the Acceglio-Longet antiformal window made up of most internal Briançonnais ("Ultrabriançonnais") units cropping out within the Schistes Lustrés domain (Debelmas and Lemoine, 1957; Lefèvre and Michard, 1976). A broadly similar setting is shown in the Vanoise-Ambin transect north of Briançon (Fig. 1).

3. Structure

3.1. Nature and geometry of the Briançonnais nappes

The studied Briançonnais transect is currently a nappe stack in which a number of rock slices having different lithologies and hence contrasting mechanical properties are associated. Rigid, thick and stiff carbonate or siliceous horizons are interleaved with weak evaporite, calcschist or finely-layered flysch horizons (Fig. 4). Each of the main tectonic units or nappes includes a competent slab, a few hundreds of metre thick, and an incompetent cover (Fig. 3; Table 1). The slabs mainly consist of Permian–Lower Triassic acidic-silicoclastic formations and/or Middle Triassic carbonates, except the Peyre-Haute slab which consists of Upper Triassic dolomites. The thin unconformable Middle–Upper Jurassic limestones are incorporated in the competent sequence. The incompetent cover, the thickness of which varies from a few metres to several hundred metres corresponds to the Upper Cretaceous–Paleocene calcschists and Eocene "Flysch noir". Therefore, one can assume after Gidon (1972) that these nappes formed as duplexes at the expense of the Briançonnais cover by shearing on the décollement levels quoted above (Fig. 3). The initial setting of the various nappes on top of the Briançonnais basement is conjectural. A pre-orogenic restoration at variance with that of Michard and Henry (1988) is proposed in this study (Fig. 12A), basically taking as a guide the metamorphic grade of the various nappes (section 4).

The map and cross-section (Figs. 2, 4) show that the original duplexes were dramatically deformed after their piling up, being affected by NW-trending major folds (Fig. 5), and by longitudinal, strike-slip and/or reverse faults. In the external part of the transect, a thick, folded pile of right-way up former duplexes is preserved (from bottom to top, Rouchouze, Sautron-Aiguille de Chambeyron, Châtelet, Peyre-Haute) in a pop-up structure between downward-converging faults (Frontal Briançonnais thrust, Ruburent and Houerts faults). The median part of the transect displays another folded pile of duplexes (La Blachière-Marinet and Aiguilles de Mary nappes) again bounded by steeply dipping faults (Houerts and Col de Mary faults), and overturned onto a more internal unit, the Ceillac-Chiappera nappe. The Châtelet nappe is backthrusted onto the Marinet, Aiguilles de Mary and Ceillac-Chiappera nappes in the Font Sancte massif. The most internal Briançonnais correspond to a major duplex (Combrémond-Pelvo d'Elva over Roure-Acceglio-Rocca Corna nappes) overturned in the Roure-Combrémond stripe, right-way up in the western limbs of the Acceglio, Buch des Sparviers and Col du Longet antiforms, and again overturned in the Monte Ferra-Pelvo d'Elva-Rocca Corna limb.

The overall aspect of the transect is a fan-like structure, well known in the Guillestre, Briançon and Moûtiers transects (Debelmas and Lemoine, 1965; Tricart, 1975, 1984; Caby, 1996; Bucher et al., 2003) as well as in the Ligurian Briançonnais (Vanossi et al., 1984). The westernmost, steeply E-dipping fault connects northwards to the Frontal Briançonnais Thrust (FBT) which carries the Briançonnais nappes onto the Helminthoid Flysch and underlying Dauphinois domain (Tricart, 1984; Barféty et al., 1995; Fügenschuh et al., 1999; Ceriani et al., 2001), whereas it merges southward with the Ruburent fault, and finally with both the Stura sinistral fault zone (Ricou and Siddans, 1986) and the younger, Bersezio dextral strike-slip fault (Gidon, 1977). The Houerts and Col de Mary faults at the axis of the fan structure likely operated as strike-slip faults, being associated with folded and laterally thrusted sub-units which evoke flower-structures (e.g. Aiguille Grande klippe and Aiguille de Chambeyron folds, Gidon, 1962, pl. 7; Sommet Rouge stripe). The latter faults connect southward with the Preit sinistral fault (Lefèvre, 1984) which parallels the more external Stura fault. In the following, we show that the progressive formation of this fan-like structure can be explained through the succession of three phases of ductile deformation (D_1–D_3) followed by a brittle phase (D4).

3.2. D_1 phase

Synmetamorphic foliation planes subparallel to the stratification plane S_0, and folded or sheared during subsequent events are observed in most of the Briançonnais rocks. This early foliation S1 is locally axial-planar to P_1 recumbent folds, either minor (Fig. 6) or major (Fouillouse fold on top of the Châtelet nappe, and Saint Ours overturned anticlinal hinge; Gidon 1962, Figures 28, 42; Platt et al., 1989). The corresponding deformation phase D_1 is associated with the early phase of décollement tectonics responsible for the piling up of the Briançonnais duplexes through mostly bedding-parallel thrusts, as shown by Tricart (1975) in the Briançon area, and by Carminati and Gosso (2000) in Liguria. It is possible to locally deduce a top-to-the WSW sense of shear (present coordinates) from P_1 fold asymmetry (e.g. Fig. 6), and from the S_1 obliquity with respect to S_0 (Tricart, 1975; Platt et al., 1989). This would correspond to roughly top-to-the NW thrust emplacement during the Eocene–Oligocene orogeny, as the Briançonnais units have been rotated anticlockwise by about 60° since the Oligocene in this area (Collombet et al., 2001).

In the Ultrabriançonnais units, an early synmetamorphic foliation S1 is locally preserved in microlithons or within synkinematic garnet crystals (see below, section 5.3), but otherwise transposed into S2. The linear D_1 structures (isoclinal fold axes and mineral lineation nearly parallel with these axes) are also strongly reoriented toward the L_{2-3}direction of shear (see next section; Lefèvre and Michard, 1976; Houfflain and Caby, 1987). D_1 is associated with prograde, high-pressure (HP) metamorphism in the entire Briançonnais-Ultrabriançonnais nappes (section 5).

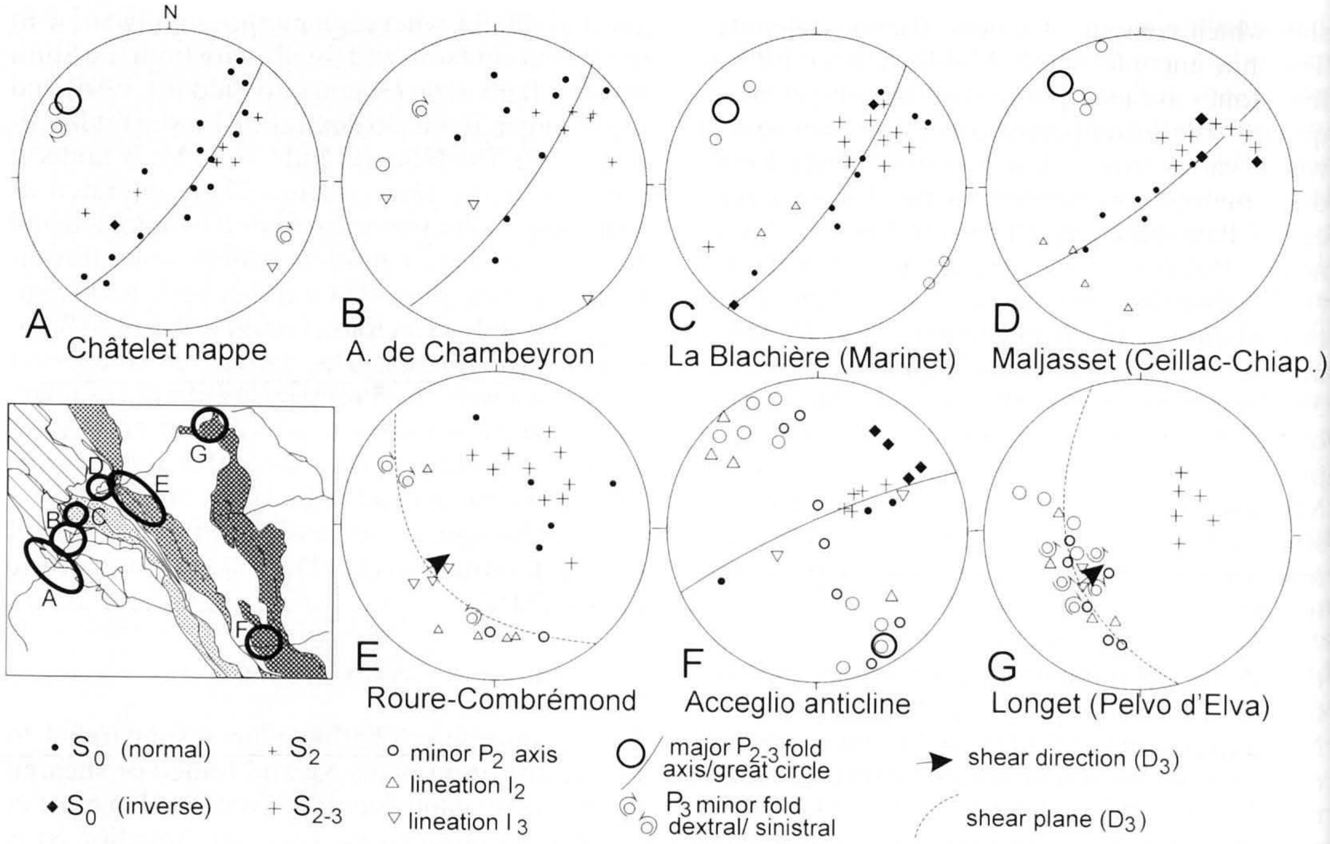

Fig. 5 Main structural elements of the Briançonnais nappes. (E) and (F) after Lefèvre and Michard (1976). Upper hemisphere equal area projection.

In the external, ophiolitic Schistes Lustrés, the prograde, HP metamorphic foliation S1 is a strain-slip cleavage, thus resulting from a more complex structural evolution than that of the Briançonnais-Ultrabriançonnais (Caron et al., 1973; Caron, 1977; Tricart, 1984). This evolution involves early, isoclinal shear folds reworked by NNE- to E-trending curvilinear folds (Schwartz, 2000; Agard et al., 2001a). The latter event of this complex D_1 "phase" would be associated with top-to-the N thrusting (Barféty et al., 1995).

3.3. D_2 and D_3 phases

The D_2 deformation phase is defined by the earliest structures associated to retromorphic recrystallisations, whereas we defer to the D_3 phase the ductile to brittle deformation of these early post-HP structures.

In the Briançonnais-Ultrabriançonnais nappes, the succession of two folding phases D_2 and D_3 is suggested by the complex geometry of the major folds (P_{2-3}) which affect the D_1 duplexes. These folds display rather constant, NW-trending axes, dipping about 10° to the NW (Fig. 5A, C, D), except the Acceglio anticline, the axis of which is virtually horizontal (Fig. 5F). By contrast, important variations characterise the dip of the axial planes. They are NE- or SW-dipping in the external part of the transect (Rouchouze-Sautron anticlinal pop-up; Aiguille de Chambeyron and Font-Sancte folds). The SW-dipping axial plane of the Marinet-Aiguilles de Mary folded duplex is curved in cross-section (steeper at depth than in upper levels). Such a curvature is also clear in the Roure-Combrémond reverse duplex and in the Pelvo d'Elva backfold, the reverse limb of which displays a SW dip of ca. 70° at the level of the Maira river, and of ca. 40° at the bottom of the Pelvo summit (Fig. 4). The SW dip of the major folds axial planes/reverse limbs also tends to steepen southeastward, i.e. close to the sinistral Preit fault. We assume that the reported curvatures, and at least part of the fan structure of the transect resulted from the deformation/reorientation (D_3 phase) of previously more open and upright folds (D_2 phase).

At the mesoscopic scale, east-verging, post- D_1 folds are observed in the Briançonnais units, that are in most cases referred to as " P_{2-3}" folds for lack of distinctive criteria, as already noted by Platt et al. (1989). They are more and more flattened and reclined as one goes from lower grade to higher grade units (Figs. 8A, B). An oblique, spaced crenulation cleavage is associated to these folds in the external and median units, whereas

Fig. 6 P_1 fold, Eocene flysch of the Aiguille de Chambeyron nappe between Sommet Rouge and Col des Houerts. Arrow: younging direction.

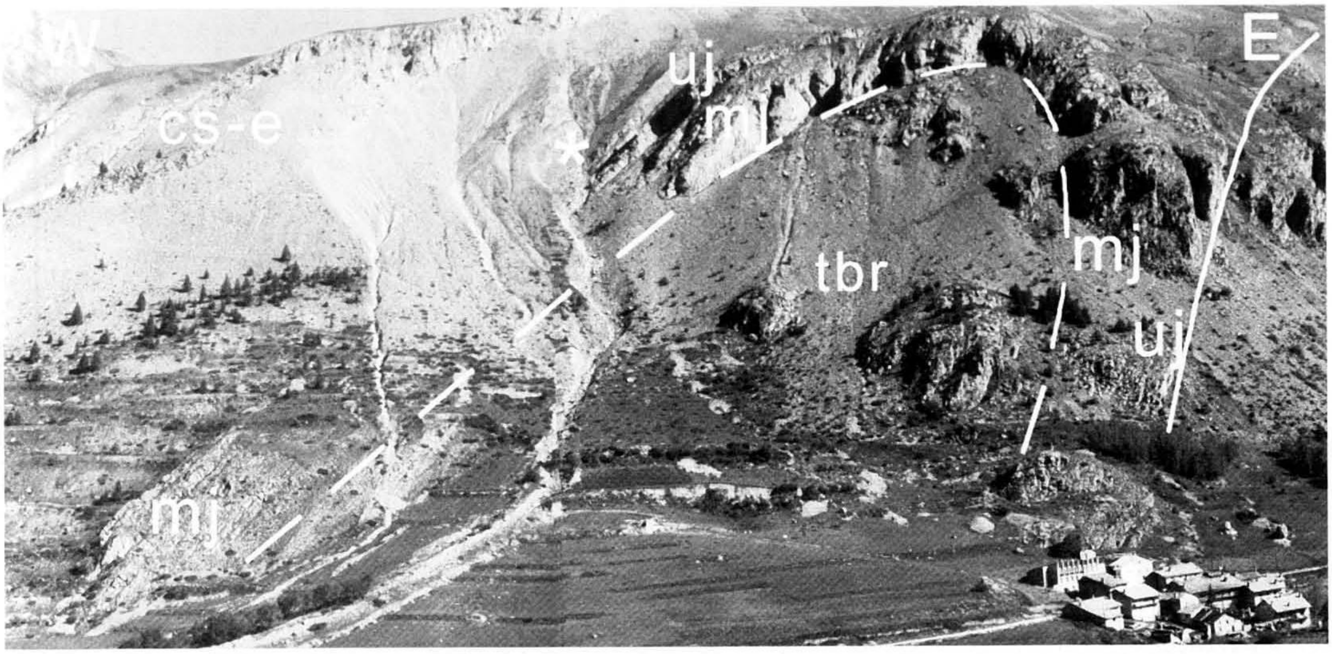

Fig. 7 Major P_{2-3} anticline above Maljasset, Ceillac-Chiappera nappe. Asterisk: location of the Early Cretaceous (?) metapelites with quartz-carpholite veins (see Fig. 10). tbr: Upper Triassic breccias; mj: Middle Jurassic carbonates; uj: Upper Jurassic marbles; cs-e: Upper Cretaceous-Eocene turbiditic calcschists.

the S_{2-3} cleavage becomes much more penetrative in the blueschist-facies units, and associated to a more conspicuous stretching lineation. S_{2-3} shows an open fan geometry in the major folds from the low-grade units (e.g. La Blachière-Marinet, Fig. 5C), whereas the fan aperture decreases in the major folds from the high-grade nappes (Maljasset, Fig. 5D). Only in the Acceglio-Longet units (Lefèvre and Michard, 1976) is it possible to recognize P_2 isoclinal shear folds (Fig. 8B) associated with L_2 stretching lineation almost parallel to the P_2 axes, folded by minor, cylindrical folds P_3. The P_3 folds are associated with a SW-dipping crenulation cleavage S3 within which S2 tends to be transposed. In the Col du Longet major fold, the dispersion of the minor P_3 axes in a great circle (Fig. 5G) and their asymmetry allow us to define a top-to-the NE sense of shear during D_3 (Hansen, 1971). A similar D_3 transport direction can be inferred from the minor folds in the Roure-Combrémond overturned duplex (Fig. 5E), which suggests that the N-trending lineations and fold axes must be deferred to D_2. NNW-trending P_2 axes are preserved in the Acceglio anticline, being dispersed in the axial plane of the major P_3 (Fig. 5F).

In the Val Maira Schistes Lustrés, D_2–D_3 superimposed deformation phases similar to that of the juxtaposed Ultrabriançonnais units are reported by Caron et al. (1973). The Monte Bettone E-verging anticline (Fig. 4) formed by the Upper

Table 2 Deformation phases in the Briançonnais/Ultrabriançonnais nappes and external Schistes Lustrés of south Western Alps: relative chronology, terminology and previous interpretations (see text for explanation). Note that Platt et al. (1989) describe in the Vanoise and Ubaye Briançonnais nappes a D1-D3 tectonic evolution substantially equivalent to that presented by Tricart (1975, 1984).

Meta-morphism	*Lefèvre & Michard 1976 (Acceglio-Longet)*	*Tricart 1975, 1984 (Briançonnais, external Schistes Lustrés, Helminthoid flyschs)*	*Houfflain & Caby 1987; Caby 1996 (Acceglio-Longet, Briançonnais)*	*Carminati & Gosso 2000 (Ligurian Briançonnais*	*Schwartz 2000 (Queyras Schistes Lustrés, Longet Ulrabrianç.)*	*Agard et al. 2001 (Susa and Ubaye Schistes Lustrés)*	*This study*
		Pre-metamorphic obduction (cf. Caron 1977)			Obduction of the low-grade Chenaillet massif	Obduction of the low-grade Chenaillet massif	
Prograde (Sch. Lustrés)		HP foliation, W-verging folds (cf. Caron et al. 1973)			HP foliation, isoclinal folds, L//P axes, coaxial deformation	HP, complex "D1" (cf. Caron 1977)	D1
Prograde (Sch. Lustrés and Briançonnais)	Relic HP foliation S1 (Penninic thrusting)	- Lower Helminthoid flysch gravity emplacement - Isoclinal folds, WNW lineation (P2, L2 in Tricart 1975) = thrusting event	Syn-HP, fan-shaped P1-S1. Acceglio-Longet: syn-HP, east-verging shear folds, W-dipping S1, transverse L1	D1: S-verging overturned P1 folds, N-dipping S1 foliation, duplications	HP strain-slip cleavage, NE-trending curvilinear folds, N-S shear strain, thrusting of Schistes Lustrés over Briançonnais	"D1" (continuation): ~ E-W trending folds, S1 foliation, N-verging thrusting (cf. Barféty et al. 1995)	D1
Retrograde	Syn- to late-HP P2 folds, crenulation cleavage S2 L2 ~NNW (nappe forming event)	- Upper Helminthoid flysch gliding	Syn-greenschist reorientation of P1, S1, L1 Local S2 crenulation	D2: Open/tight back-folds P2, crenulation cleavage S2	Top-to-the-E folds and extensional shear bands (vertical shortening) on previously E-dipping foliation	D2: W-dipping foliation, ~ N-S trending folds, top-to-the-E shear bands (synorogenic extension)	D2
Retrograde	Syn-greenschist shear folds P3, W-dipping crenulation cleavage S3 (backthrust event)	Fan-shaped folding, crenulation cleavage, frontal thrusting, backthrusting (P3, S3 in Tricart 1975)	Sinistral torsion of L1-L2 along the Preit fault (Lefèvre 1984)	D3/D4: kink bands dipping S or E; SW translation (Frontal Thrust)	Additional extension followed by 30°-40° westward tilting of foliation and shear bands	D3: ductile-fragile top-to-the-W shear bands (late synorogenic extension)	D3
Post-metamorphic	Cataclastic faults, late backthrusting	Increasing E-W shortening		Normal and strike-slip faults	Dextral torsion of L1-L2 along the Stura-Durance fault zone	Fragile faulting	D4

Triassic dolomites of the External Piemont zone (Michard, 1967) appears as a major P_{2-3} fold, equivalent of the Acceglio anticline in the Briançonnais zone. The major P_{2-3} folds point to a NE–SW shortening direction (present coordinates) in the Briançonnais, as well as in the juxtaposed Schistes Lustrés, which form an asymmetric synform ("Acceglio syncline") between the Roure and Acceglio-Longet stripes. However, the mesoscale structures associated to early retromorphic assemblages in the external Schistes Lustrés have been interpreted by Schwartz (2000) and Agard et al. (2000) as formed by extension. We discuss this interpretation in section 6.

The horizontal backthrust at the bottom of the Châtelet-Font Sancte dolomites occurred during the D_3 phase, as it crosscuts the P_{2-3} folds of its foot-wall, while large P_2 folds are seen on top of the backthrusted unit. This backthrust could result from differential shortening between the Châtelet-Font Sancte dolomitic slab and the juxtaposed, more ductile Sautron-Chambeyron, Marinet-Aiguilles de Mary and Ceillac-Chiappera units. We assume that the main longitudinal faults quoted above (chapter 4.1) operated both during the D_2 and D_3 phases. Of particular interest is the fact that, in the Acceglio-Longet band and juxtaposed Schistes Lustrés, and going from the Varaita valley to the Maira and Preit valleys, the trajectories of the L_2 and L_3 stretching lineations progressively curve from a transverse, NE direction to a N–S, and finally NW direction (Houfflain and Caby, 1987). This strongly suggests a sinistral strike-slip movement along the Preit fault (and then, along the Houerts fault that branches on it) during D_2 and D_3.

3.4. D_4 events

Several semi-brittle or brittle structures postdate D_3 and are referred to D_4 events. Horizontal kink bands are observed in the calcschist lithologies, particularly in the Tête de Girardin area (Ceillac-Chiappera nappe). Top-to-the-west small-scale shear bands are described by Caby (1996) in the retrogressive rocks from the Pelvo-d'Elva nappe. All these structures record a late extensional deformation or collapse of the previously thickened tectonic prism. Brittle, steeply-dipping normal faults are also observed by place (Saint Antoine area). However, the FBT and Ruburent-Bersezio faults were reactivated as dextral strike-slip faults during the late orogenic evolution, with horizontal throw of a few kilometres and downthrow of the northeastern block (Gidon, 1977). This implies a rotation of the regional shortening direction towards the N.

4. Metamorphism

4.1. Metamorphic map

The metamorphic map of the Ubaye-Maira transect (Fig. 9) is based both on key mineral assemblages and on the typical, penetrative fabric of each unit. Metamorphism of the Upper Penninic nappes thrust over the Briançonnais complex is also considered to constrain the age of the tectonic contact between both nappe complexes during the Alpine subduction history.

The metamorphic grade progressively increases from west to east, and towards tectonically low-

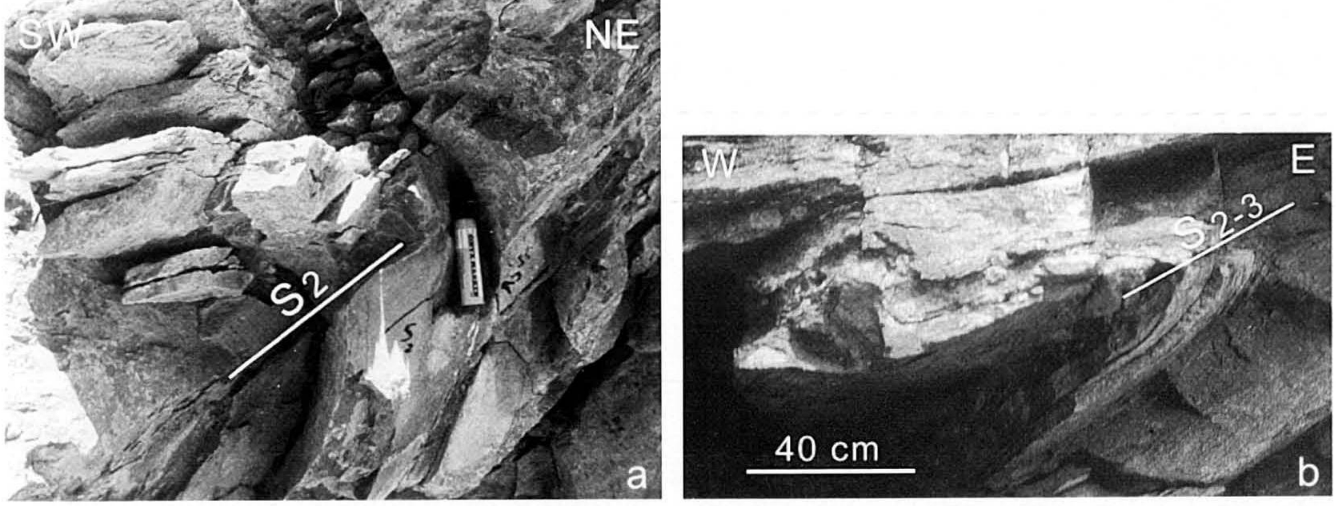

Fig. 8 Minor P_2 folds. (a) Upper Cretaceous–Eocene calcschists, southwest slope of Tête de Girardin, Ceillac-Chiappera nappe. (b) Permian metagreywackes, south slope of Monte Ferra, Pelvo d'Elva nappe.

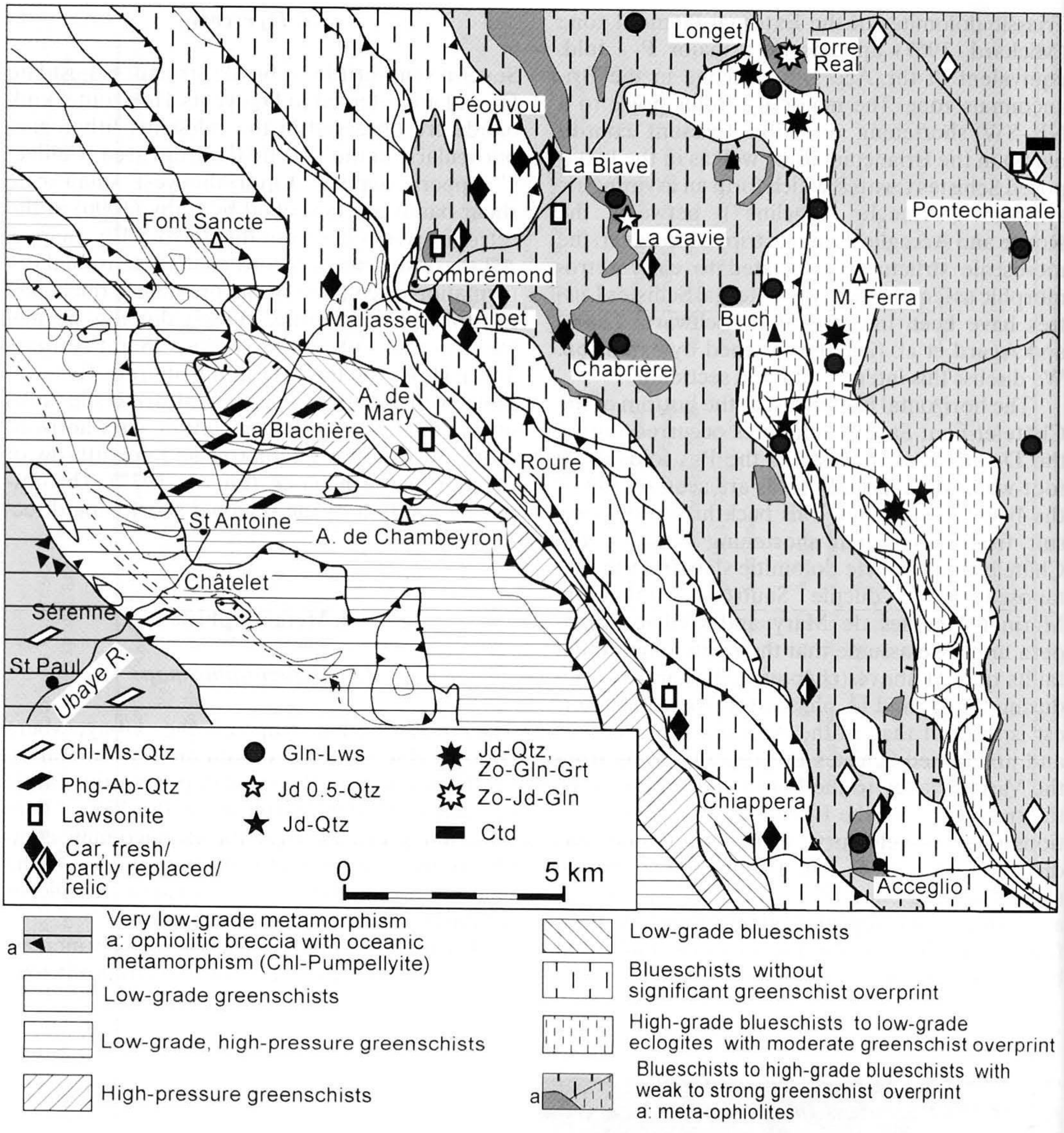

Fig. 9 Metamorphic map of the Ubaye-Maira transect. Torre Real mineral assemblage after Schwartz (2000).

er units of the exposed section. A succession of seven metamorphic zones has been recognized, each involving one or two major nappes. In the Col de Vars Flysch unit, white mica and chlorite lamellae mainly correspond to detrital, altered crystals. Recrystallisation is incipient, which points to a very low-grade greenschist-facies metamorphism. In the underlying Châtelet nappe, the assemblage white mica + chlorite developed in pressure shadows of the coarse detrital grains of the Eocene flysch. Likewise, low-grade greenschist-facies recrystallisation characterise the anastomosing foliation planes in the Jurassic nodular marbles from the same unit. Recrystallisation appears more important in the underlying Aiguille de Chambeyron nappe, as fibrous quartz–calcite veins become abundant in the calcschist lithologies (Saint Antoine). Si content in the newly formed white micas (phengites) corresponds to the HP-greenschist facies conditions in the presence of albite (see below). A further increase in intensity of metamorphism is recognized entering the Marinet nappe where blasto-mylonitic foliations containing phengite and chlorite developed in quartzitic-conglomeratic lithologies. Nevertheless, the Permian meta-andesite situated at the base of the Permo-Triassic conglomerates west of La Blachière do not contain any sign of HP mineralogy. This contrasts with the overlying duplex, i.e. the Aiguilles de Mary nappe, where the equivalent meta-andesites contain abundant lawsonite in a chlorite-albite-epidote matrix.

Blueschist-facies metamorphic conditions are reached in the Ceillac-Chiappera nappe which contains Fe-Mg carpholite-quartz veins in Lower (?) Cretaceous meta-argilites (Fig. 10A) interbedded with siliceous limestones on top of the Maljasset anticline, and in Permian–Triassic metapelites close to Chiappera. Similar carpholite-quartz veins occur in the Ultrabriançonnais Roure and Combrémond nappes and in the External Piemont Péouvou unit, although the carpholite alteration into white mica + chlorite takes some importance there. In the Schistes Lustrés of the Acceglio syncline, metapelitic lithologies contain both fresh and partly replaced carpholite. Lawsonite is widespread in the metapelites and calcareous metapelites, while the glaucophane–lawsonite–albite assemblage is observed in most of the metabasites (e.g. Acceglio pillow basalts, Bearth, 1962; Chabrière metagabbros, Steen, 1972; Schwartz, 2000). Lawsonite–pumpellyite assemblage occurs in the Combrémond pillow basalt lense, while the assemblage jadeitic pyroxene (Jd 0.5)–quartz is observed in the meta-albitites at La Gavie. Ferrocarpholite from the Chabrière Lower Cretaceous meta-argilites was the first occurrence of this mineral described in association with blueschist-facies metabasites (Steen, 1972, p. 183; Steen and Bertrand, 1977).

West of the Acceglio syncline, the Ultrabriançonnais domain crops out again in the Acceglio-Longet stripe (Acceglio and Pelvo d'Elva nappes), which displays higher-grade HP-LT assemblages. The Monte Ferra outcrops are well known for their large jadeite–quartz pseudomorphs after perthitic felspars (Lefèvre and Michard, 1976). The jadeite–quartz assemblage is widespread in the sub-alkalic metagranites and metagraywackes, and also occur in the Mesozoic–Eocene cover rocks at Col du Longet (Michard, 1977). Lensoid mafic sills are converted into zoisite/lawsonite–glaucophane–garnet assemblages (Lefèvre and Michard, 1976; Houfflain and Caby, 1987; Caby, 1996; Schwartz et al., 2000). East of, and below the Acceglio-Longet overturned antiform, the Torre Real metabasites yield zoisite–jadeite–glaucophane associations (Schwartz, 2000), whereas the metapelite lithologies are characterised by relic, quartz-hosted carpholite needles and fibrous micaceous pseudomorphs. Chloritoid that formed at the expense of carpholite appears at Pontechianale, thus registering a similar positive thermal gradient as observed in the Schistes Lustrés of the Briançon-Ambin transect (Agard et al., 2000, 2001a).

4.2. P–T conditions

Peak pressure-temperature (P–T) metamorphic conditions have been estimated for most of the studied units (Fig. 11). It should be noted that within the lower grade rocks (mainly in the west of the section) P–T estimates are hampered by the lack of index minerals. In the Saint Antoine calcschists (Aiguille de Chambeyron nappe) and the adjacent La Blachière conglomerates (Marinet nappe) P–T conditions are close to a maximum of 0.6 GPa, 310 °C and 0.7 GPa, 330 °C, respectively, according to Si substitution in phen-

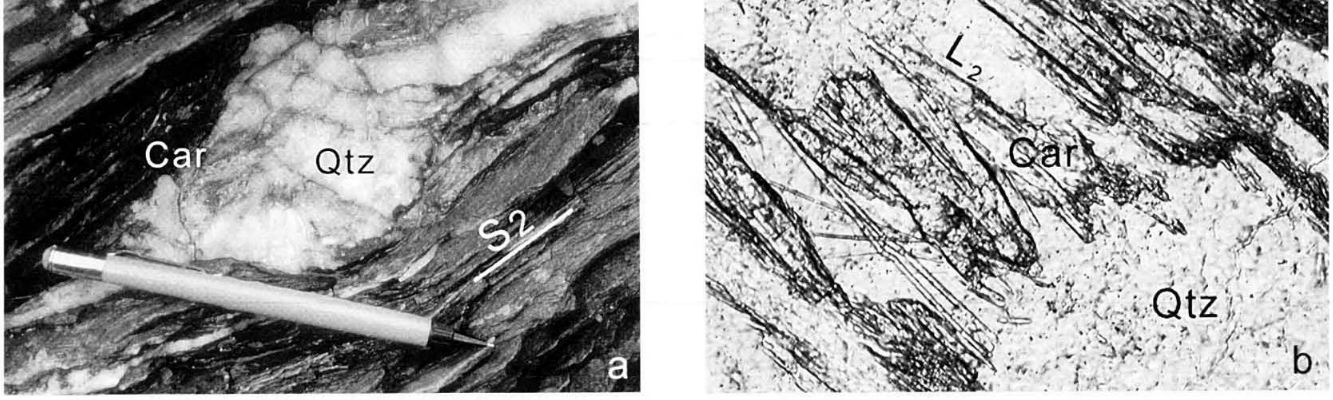

Fig. 10 Fresh carpholite-quartz assemblages from Maljasset (Ceillac-Chiappera nappe). (a) Early carpholite–quartz veins transposed in the S_2 tight crenulation cleavage (location: Fig. 7). (b) Micrography of another vein from the same outcrop. Car: Fe-Mg carpholite; Qtz: quartz.

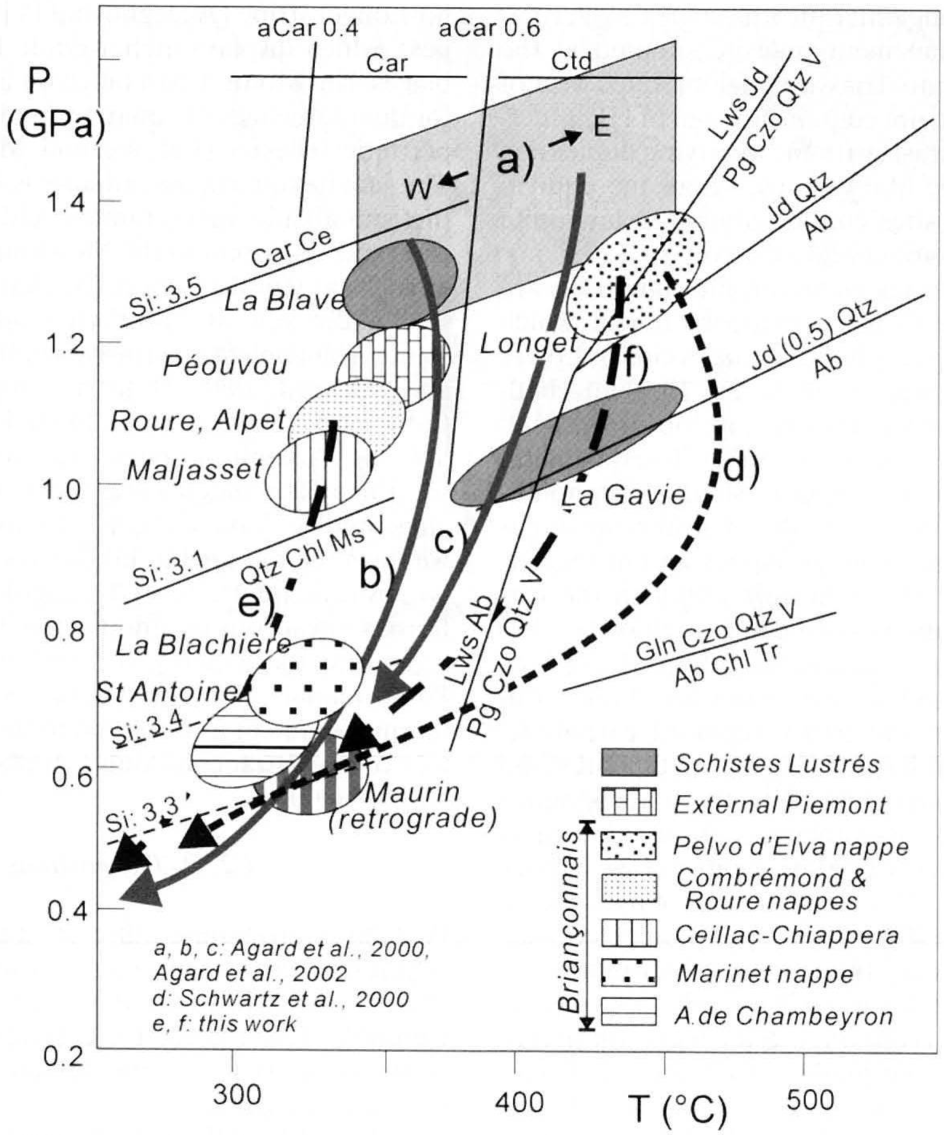

Fig. 11 P–T conditions of typical metamorphic assemblages from the studied transect. Si 3.3/3.4: Si in phengite isopleths in presence of Ab and Chl; Si 3.1/3.5: Si in phengite isopleths in the reaction Qtz+Chl+Ms+V=Car+Ce, as a function of Mg-carpholite activity (aCar); after Bousquet et al. (1998). Jd + Qtz = Ab calculated using Thermocalc (Powell and Holland, 1988). a–c) P–T estimates for different localities of the external Schistes Lustrés and corresponding retrograde paths (Agard et al., 2000, 2002); d) retrograde path for the Col du Longet unit after Schwartz et al. (2000); e, f) retrograde paths suggested for the Briançonnais blueschist-facies units and the Longet-Pelvo d'Elva nappe (this work).

gite, in the presence of albite (see e.g. Goffé and Bousquet, 1997; Bousquet et al., 1998). In the carpholite-bearing units, Si in phengite isopleths do indicate higher pressure conditions of approximately 1.0 GPa in the Maljasset anticline (Ceillac-Chiappera nappe) and 1.1 GPa in the Roure and Alpet outcrops (Roure and Combrémond nappes respectively), for temperatures slightly below 350 °C. Si substitution in phengite indicates higher P–T conditions in the External Piemont (Péouvou) and ophiolitic Schistes Lustrés (La Blave) east of Combrémond, i.e. about 1.2 GPa, 350 °C and 1.3 GPa, 370 °C respectively. This is consistent with the P–T conditions (Fig. 11, left part of field a) calculated by Agard et al. (2000) for these most external Schistes Lustrés units where chloritoid is lacking.

In the Acceglio-Longet antiform, a minimum pressure of 1.2 GPa is indicated by the jadeite (Jd 95 to 99)–quartz association, while the co-stability of zoisite and glaucophane and the garnet-phengite thermometer suggest temperatures higher than 400–430 °C (Schwartz et al., 2000). Lefèvre and Michard (1976), Houfflain and Caby (1987) and Caby (1996) assumed that lawsonite was stable from the prograde phase up to the beginning of the retrograde path, which suggests temperatures not in excess of 430 ± 20 °C. In contrast, Schwartz et al. (2000) showed that lawsonite can be a late retrograde phase in the Longet rocks, and suggests heating to 450 ± 25 °C at 1.3 ± 0.1 GPa. The preservation of pre-Alpine ages in relic muscovite grains from meta-aplites and meta-granites (360–340 and 270 Ma, respectively) is consistent with the interpretation of low temperature conditions during the Alpine metamorphism, combined to low fluid mobility in the basement (Monié, 1990).

In view of the fairly perfect preservation of carpholite in the Maljasset and Alpet outcrops, we infer a cooling during decompression for the internal Briançonnais nappes (Fig. 11, path e). In the case of the Acceglio-Longet unit, Schwartz et al. (2000) suggest a reheating episode up to 465 ± 25 °C at the beginning of the decompression path (Fig. 11, path d), mainly based on the notion that lawsonite did not crystallise during prograde metamorphism. However, based on the alternative observations quoted above, we rather suggest a decompression path at decreasing T (Fig. 11, path f). This is *a priori* consistent with the presence of epidote cores in lawsonite as also observed by this study, and with the overall lack of any other sign of heating. Note that similar retrograde P–T paths were determined by Agard et al. (2000) and Agard et al. (2002) for the external Schistes Lustrés units farther to the north (Fig. 11, b, c). The P–T estimates that we obtained for the La Gavie meta-albitite and even more clearly for the Maurin (Combrémond) metabasalts (Fig. 11) likely indicate re-equilibration stages of the Schistes Lustrés units during decompression.

4.3. Deformation–crystallisation relationships

In the lower-grade Briançonnais units, as well as in the overlying Heminthoid Flysch, micas and chlorite seem to crystallise both in S_1 and S_2. In the blueschist-facies Briançonnais units, the early (syn-D_1) veins are strongly deformed and transposed within the dominant S_2 foliation (Fig. 10A) which curves around the hinge of the major P_{2-3} fold (Fig. 7). Moreover, the carpholite fibres are folded by minor P_{2-3} folds, and boudinaged along the L_2 stretching lineation (Fig. 10B). This indicates that carpholite crystallised during D_1, and was just poorly altered during D_2 and D_3.

In the higher-grade Acceglio-Longet units, the Jd–Qtz pseudomorphs clearly formed before D_2 as they are rotated towards the S_2 planes, and most often boudinaged parallel to the L_2 stretching lineation. Glaucophane is associated to S_1 planes in S_2 microlithons, but also occurs in S_2 shear planes. Garnet occasionally displays intergrowths with jadeite and glaucophane, and helicitic structures which testify to its broadly syn-D_1 crystallisation. This does not support the hypothesis of a pre-kinematic HP-LT stage of crystallisation there (Schwartz et al., 2000). The fact that the relict magmatic muscovite grains are poorly oriented in metagranitic lithologies suggests that deformation was strongly heterogeneous during D_1. Well-shaped lawsonite grains superimposed to S_2 crystallised after incipient alteration of glaucophane into chlorite, but lawsonite growth would have begun earlier according to the presence of deformed, lawsonite-bearing S_1 foliations (Lefèvre and Michard, 1976) and of garnet-lawsonite intergrowths (Caby, 1996).

In the external Schistes Lustrés, HP recrystallisations occurred during a " D_1 " phase (Agard et al., 2000, 2001a; Schwartz, 2000) more complex than the Briançonnais D_1 phase (see above, section 3.2, and Table 2). Low-grade blueschist- to greenschist-facies assemblages are developed in the dominant, W-dipping foliation; in pervasive, mostly transverse stretching lineation structures, associated with N-trending, curvilinear folds, and finally in top-to-the E extensional shear bands. All of these structures are ascribed to a D_2 phase by Agard et al. (2000, 2001a). Late greenschist recrystallisations are associated to top-to-the W extensional shear bands D_3 (Agard et al., 2001a).

4.4. Age constraints

The argillaceous matrix of microbreccias at the bottom of the Flysch noir from the classical Briançonnais yielded early Bartonian planktonic foraminifers (Barféty et al., 1995), about 40 Ma (Gradstein and Ogg, 1996). Assuming a rapid accumulation of the overlying, ca. 100 m thick turbidites, we infer an age of 38 ± 1 Ma (mid-Bartonien to earliest Priabonian) for the olistostrome which terminates the flysch sequence (Kerckhove, 1969; Barféty et al., 1995; Michard and Martinotti, 2002). This bounds metamorphism to be younger than 38 ± 1 Ma, consistent with the $^{40}Ar/^{39}Ar$ plateau age released by phengite separate from the Pelvo d'Elva Triassic quartzites, 37 ± 1 Ma (Monié, 1990). Likewise, this stratigraphic dating fairly fits the recently published, HP white mica ages from more northern parts of the internal Briançonnais, ~38 Ma (Rb–Sr; Freeman et al., 1997) and 41–36 Ma ($^{40}Ar/^{39}Ar$; Markley et al., 1998).

5. Tectonic interpretation

Unfolding of the present-day cross section allows us to restore, at least approximately, the original architecture of the Briançonnais units (Fig. 12A). The Brianconnais was originally a submerged plateau separated into basins and highs by normal

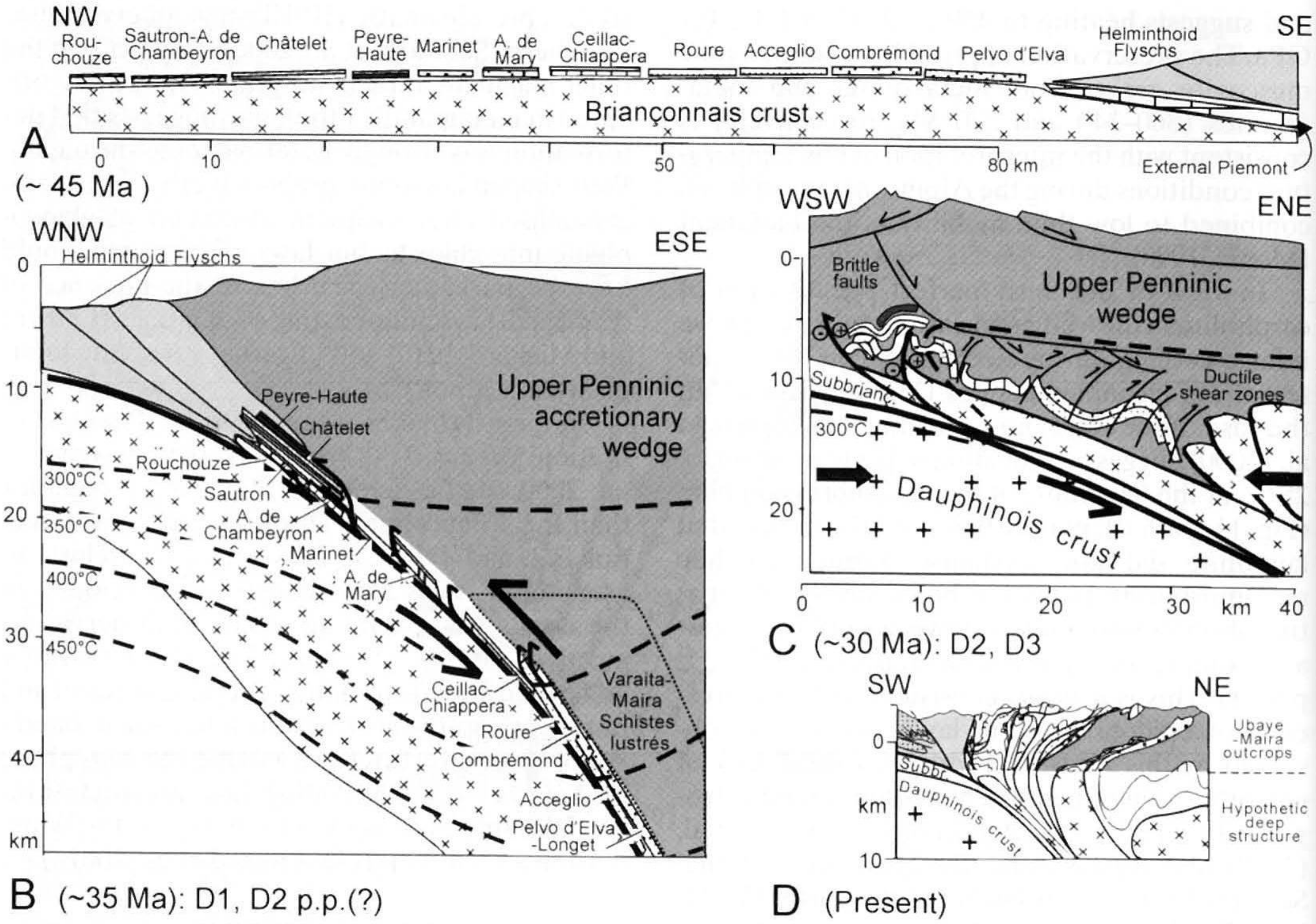

Fig. 12 Interpretation of the tectonic-metamorphic evolution of the Ubaye-Maira transect. (A) Pre-orogenic setting of the Briançonnais cover units; the actual width of the transect was possibly less than shown here, because transcurrent faults make the pre-orogenic restoration uncertain. (B) Subduction stage; the Briançonnais units have been detached from the plunging slab and form duplexes (D_1 deformation, and incipient exhumation D_2?). (C) Intermediate stage of exhumation; shortening and backfolding are in progress at depth (D_2 and D_3 deformations). The upper part of the uplifted wedge collapses, and is submitted to erosion. (D) Present-day cross-section (see enlarged cross-sections Figs. 4, 13).

faults inherited from the Jurassic rifting (Lemoine et al., 1986; Jaillard, 1988), and reactivated in the Late Cretaceous–Eocene while overstepping the frontal bulge of the Alpine subduction zone (Stampfli et al., 1998; Michard and Martinotti, 2002). The corresponding sketch cross-section (Fig. 12A) is oriented NW–SE, consistent with the assumed kinematics of the Adria-Europe convergence in the Western Alps during this time span (e.g. Schmid and Kissling, 2000).

Once the Upper Penninic ocean was totally subducted (including its External Piemont margin), ongoing plate convergence pushed the Briançonnais lithosphere below the internal accretionary wedge. During the subduction process, the Briançonnais cover units detached and formed duplexes at varied depth. Detachment was straight forward in the external units, where evaporite levels were abundant. By contrast, detachment of the Ultrabriançonnais units lagged until temperatures rose high enough to allow Permian and basement rocks to become ductile. Local structural constraints such as the presence of inherited steep ramps likely controlled the depth at which detachment occurred. Figure 12B shows the state of the detached and duplicated Briançonnais units at about the end of the top-to-the NW thrusting phase D_1 defined above (section 3.2). The detachment depth of the varied units is shown in such a way as each unit meets the maximum P conditions indicated by its mineral assemblages. The external units are piled up beneath the front of the wedge at about 10 km depth. Assuming that the dip of the subduction plane was about 45°, we may observe that the Acceglio-Longet units detached before reaching the maximum depth (60 km) permitted by their initial location at ca. 80 km east of the most external units. In Figure 12B, the depth at which the isotherms cross the subduction channel results from the P–T estimates for the corresponding units. The observed T/depth relationships are similar to those computed by Goffé et al. (2003) in their model 4, i.e. subduction of oceanic crust beneath a thick accretionary prism at low rate (0.5–1 cm/y), which is consistent with the broad geodynamic setting of the Ubaye-Maira transect, except the occurrence of a thinned continental crust in the lower plate. Assuming a subduction rate of 1 cm/y (Stampfli et al., 1998), the age of stage B (Fig. 12) is constrained at about 36 ± 1 Ma by that of the olistostrome on top of the external Briançonnais Flysch noir, admittedly 38 ± 1 Ma (see section 4.4).

The retrograde, shortening events D_2 and D_3 could hardly be explained without tilting the duplexes and their foot-wall towards a shallower dip, this tilting being combined with the exhumation of the whole system. Collisional underthrusting of the thick Dauphinois crust beneath the Penninic wedge (included the studied Briançonnais units and underlying Subbriançonnais suture zone) is a good candidate to account for the tilting of the subduction channel towards shallow dips (Fig. 12C). The resulting uplift of the orogenic wedge would have triggered synorogenic extension tectonics in the upper levels of the wedge, as illustrated by part of the D_2 and D_3 structures described by Agard et al. (2000, 2001a) and Schwartz (2000) in the Ubaye and Val de Susa Schistes Lustrés.

Thrusting of the Penninic nappes onto the Dauphinois-Helvetic domain was directed to the WSW or SW in south Western Alps (Tricart, 1984; Lickorish and Ford, 1998). An ENE–WSW orientation of convergence (Fig. 12C) is consistent with the occurrence of sinistral strike-slip movement on the Preit-Stura fault zone (Lefèvre, 1984; Ricou and Siddans, 1986) and on its northwest projections (Ruburent and Houerts faults in the Ubaye transect). These collisional events occurred after the end of the accumulation of the internal Dauphinois flysch, dated at about 31 Ma (Ruffini et al., 1994; Boyet et al., 2001). Consistently, Early Oligocene molasse deposits include ophiolite and blueschist clasts in the External Dauphinois (De Graciansky et al., 1971; Evans and Mange-Rajetsky, 1991) and the southern Po basin (Gelati and Gnaccolini, 1996).

The present-day cross-section (Figs. 12D, 13) essentially results from, (i) accentuation of shortening in the deeper levels, with D_3 backthrust and backfold development. We assume that this could be related to the westward thrusting of Briançonnais/Dora-Maira crustal slices; (ii) Oligocene-Miocene transport of the deforming Penninic wedge towards the external Dauphinois with formation of the Frontal Briançonnais Thrust, and (iii) Neogene, late orogenic collapse (Sue and Tricart, 2003) and excision/erosion of ca. 10 km thick rock units from the top of the wedge.

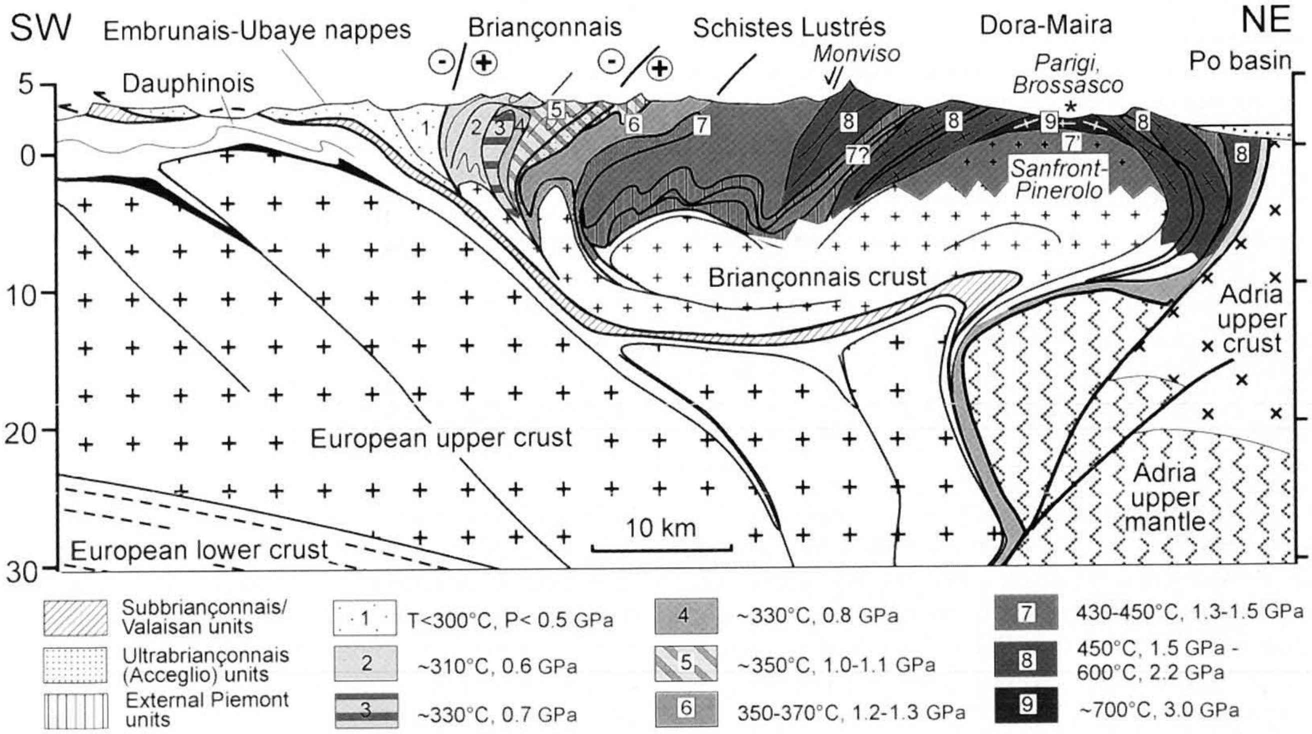

Fig. 13 Crustal scale, tectonic-metamorphic cross-section of the Embrunais-Ubaye – Dora-Maira transect (for location see Fig. 1). Sources of geological and petrologic (P–T) data: Briançonnais and juxtaposed Upper Penninic units after this work; Monviso after Schwartz et al. (2001) and references therein; Dora-Maira after Michard et al. (1993) and Avigad et al. (2003). Structural interpretation at depth inspired from Schmid and Kissling (2000).

6. Discussion and conclusion

During the progressive burial of the Briançonnais plate, the cover and uppermost basement rock units detached at various depth and accreted to the accretionary Upper Penninic (Schistes Lustrés) wedge in front of the Adria plate. Thus syn- to post-peak metamorphism duplexes formed through flat-and-ramp reverse, top-to-NW faulting, with only rare and moderate reverse metamorphic gaps between the juxtaposed units (e.g. Aiguilles de Mary over Marinet, and Pelvo d'Elva over Acceglio units). This early, prograde evolution is labelled D_1 in this study, although it clearly includes, particularly in the Schistes Lustrés, a complex tectonic-metamorphic evolution (Table 2). D_1 began in the external Briançonnais at 38 ± 1 Ma (end of Flysch noir sedimentation; section 4.4), and ended at 36 ± 1 Ma (subduction at ca. 20 km, assuming a subduction rate of 1cm/year; Fig. 12B). Note that HP to UHP metamorphism of the Dora-Maira crystalline massif (Fig. 13) occurred within the same time segment (Gebauer, 1999, with references therein), which supports its location at the leading edge of the Briançonnais plate (Michard and Martinotti, 2002). In contrast, prograde metamorphism of the more internal, Piemont-Ligurian units began earlier, and evolved from ~50 Ma to ~38 Ma (Monié and Philippot, 1989; Cliff et al., 1998; Lapen et al., 2003, with references therein).

Exhumation tectonics developed within the further collisional setting. According to our structural analysis of the D_2 and D_3 phases, bringing the higher-grade units closer and closer to the lower-grade ones at the bottom of the deforming wedge (within the former subduction channel) resulted from inward nappe refolding, with a component of ductile backthrusting associated with longitudinal strike-slip faults. For example, the carpholite-bearing Ceillac and Roure units, equilibrated at 1.0–1.1 GPa, 350 °C, were brought at a distance of about 4 km beneath the Aiguille de Chambeyron nappe, equilibrated at 0.6 GPa, 300 °C, which would represent a vertical omission of ca. 10 km thick rock units. We suggest (section 5) that this juxtaposition of formerly distant units resulted from shortening of the former subduction channel, tilted westward towards shallow dip above the buoyant, and progressively sliced and thickened Dauphinois crust (Figs. 12, 13). This implies vertical escape of the rock material, particularly of the incompetent material from the synform structures, resulting in the thickening of the base of the wedge. Thinning of the upper part of the wedge (Schistes Lustrés) occurred contemporaneously through both erosion and extension.

It is not *a priori* out of question that the NE-directed P_{2-3} structures formed when the package was still dipping SE in the subduction zone, and that they were extensional, as postulated for other "backthrusts" (Wheeler and Butler, 1993, 1994). Later tilting would give them the appearance of backthrusts. Post-extensional westward tilting in the range of 30°–40° have been indeed suggested by Schwartz (2000) for the Queyras and Longet outcrops, and ascribed to the domal deformation of Dora-Maira (cf. D4 phase of Bucher et al., 2003). However, it must be noted that the Ubaye-Maira P_{2-3} major backfolds display rather steeply dipping axial planes in upper levels (ca. 60° to ca. 40°, Fig. 4), and nearly vertical axial planes at depth (Figs. 4, 12D, 13). Therefore, a late westward rotation in the range proposed by Schwartz (2000) would not allow us to restore a SE dip for these structures. Assuming that they represent extensional, "pseudo-backthrust" structures lately rotated by 60–80° to the west would imply that they would have formed in a steep subduction zone, which is unlikely in a collisional scenario (Bucher et al., 2003).

In the Moûtiers-Gran Paradiso transect of the Briançonnais (Fig. 1), Bucher et al. (2003) also reach the conclusion that the fan-structure which carries the most internal Briançonnais units on top of the Schistes Lustrés results from nappe refolding, combined with subsidiary backthrusting, during a D_3 phase, contemporaneous with activation of the Frontal Briançonnais Thrust (Fügenschuh et al., 1999; Ceriani et al., 2001). This phase would be linked to the lithosphere scale thrusting of the Briançonnais microcontinent over the European margin (their Fig. 7b; cf. our Fig. 12C). Bucher et al. (2003) argue that the D_3 phase overprints an early exhumation phase D_2 which in turn transposes the D_1 subduction-related structures – an evolution which compares with that of the Ubaye-Maira transect. However, in the northern transect, the authors assume that D_2 occurred by extrusion within the SE-dipping subduction channel, whereas we rather relate the Ubaye-Maira D_2 phase to incipient backfolding in a westward tilted subduction channel. At a larger scale in the Ubaye-Maira transect, we do admit (Avigad, 1992; Michard et al., 1993; Avigad et al., 2003) that extrusion tectonics occurred within the SE-dipping subduction channel, carrying eclogite-facies nappes over blueschist-facies Briançonnais basement in the Dora-Maira massif (Fig. 13). Indeed, the fact that the most internal Briançonnais units of the Ubaye-Maira transect display metamorphic grades (1.1 GPa, 350 °C at Combrémond; 1.3–1.4 GPa, 430 °C at Col du Longet) only slightly weaker than, or equal to the juxtaposed Schistes

Lustrés (1.2 GPa, 370 °C and 1.3–1.4, 430 °C, respectively; see section 4.3 and Fig. 11) could record an extrusion event which would have affected even the upper part of the subduction channel. In that case, the early transverse lineations L_2 and associated shear folds of the Acceglio-Longet stripe could have formed during this extrusion event before being overprinted by the P_{2-3} back-folds.

Therefore, exhumation history at the HP-LT metamorphic front (i.e. in the Briançonnais nappes) appears to have been different from that of the more internal, eclogitic nappes, which indeed corresponds to a quite distinct geometry in the present-day cross-section (Fig. 13). Whereas exhumation of the Dora-Maira UHP-HP eclogitic crystalline rocks and overlying Monviso meta-ophiolites mostly occurred through extrusion in the subduction channel, then extensional tectonics, the Briançonnais nappes were exhumed partly by early extrusion, then mostly through transpressional deformation at the bottom of a collapsing and eroded orogenic wedge.

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