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Partitioning of deformation within a subduction channel during exhumation of high-pressure rocks: a case study from the Western Alps

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Abstract

The metamorphic belt of the Western Alps was subjected to widespread extensional tectonism at the end of the Eocene (ca. 45–35 Ma). Extension was accommodated by hinterland-directed movements along gently inclined extensional shear zones, which facilitated rapid exhumation of high-pressure and ultra-high-pressure rocks. This deformation resulted in a normal metamorphic sequence. Extension in the inner parts of the Western Alps was coeval with shortening at the front of the belt (foreland-directed thrusts), which took place during decompression, and emplaced higher grade metamorphic units over lower grade metamorphic rocks, thus forming an inverse metamorphic sequence. Two mechanisms for this extensional episode are discussed: (1) collapse of an overthickened lithosphere, and (2) internal readjustments within the orogenic wedge due to subduction channel dynamics. We favour the latter mechanism because it can account for the development of the observed inverse and normal metamorphic sequences along foreland-directed thrusts and hinterland-directed detachments, respectively. This hypothesis is supported by published structural, metamorphic and geochronological data from four geological transects through the Western Alps. This study also emphasizes the importance of post-shearing deformation (e.g. horizontal buckling versus vertical flattening), which can modify the distribution of hinterland- and foreland-directed shear zones in orogenic belts.

Keywords: Exhumation; High-pressure metamorphism; Thrusts; Detachments; Strain partitioning

1. Introduction

In recent years, research into the evolution of the Western Alps has benefited significantly through the application of advanced geochronological techniques (Inger et al., 1996; Duchêne et al., 1997; Freeman et al., 1997; Gebauer et al., 1997; Rubatto et al., 1998, 1999; Amato et al., 1999; Reddy et al., 1999). Such geochronological data, when combined with structural and metamorphic data, have led many authors to conclude that major tectonic boundaries in the Western Alps resulted from crustal extension that was superimposed on the earlier thrust-related structures (e.g. Avigad et al., 1993; Wheeler and Butler, 1993; Jolivet et al., 1998; Reddy et al., 1999; Bousquet et al., 2002; Cartwright and Barnicoat, 2002).

Such is the case of the (east-) hinterland-directed kilometric-scale shear zones recognised over a significant portion of the internal part of the Western Alps. These tectonic contacts, e.g. the 'Gressoney Shear Zone' at the northwestern margin of the Sesia zone (Reddy et al., 1999; Fig. 1), were originally interpreted as shortening structures (e.g. Platt et al., 1989). However, the retrograde P–T paths and the exposure of low grade units above higher grade rocks along east-dipping detachments (normal metamorphic sequence) may indicate that many of these contacts operated as extensional faults. These extensional structures operated simultaneously with (west-) foreland-directed thrust in more external units of the Western Alps (e.g. Butler and Freeman, 1996; Ceriani et al., 2001; Wheeler et al., 2001; Bucher et al., 2003). The latter were responsible for the emplacement of higher grade metamorphic units over lower grade rocks (inverse metamorphic sequence)

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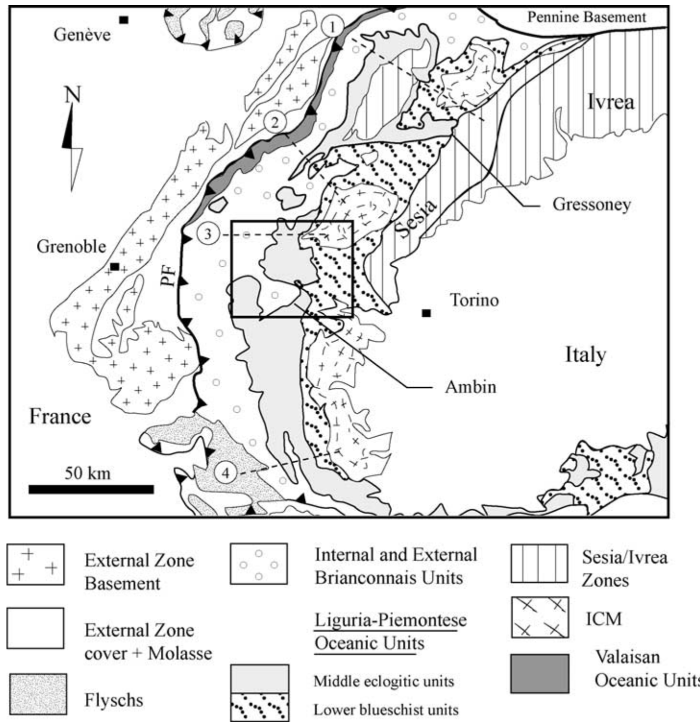

Fig. 1. Simplified geological map of the Western Alps showing locations of major extensional shear zones (Gressoney Shear Zone and Ambin Shear Zone). Inset indicates the location of the study area (Fig. 2). The locations of cross-sections–(Fig. 4) are also shown. PF, Penninic Front. ICM, Internal Crystalline Massifs.

during decompression. The late Eocene shearing event occurred after the culmination of ultra-high-pressure (UHP) metamorphism (Rubatto et al., 1998; Amato et al., 1999) in the Western Alps and played a major role in the exhumation of high-pressure (HP) and UHP rocks (Jolivet et al., 2003).

Kinematic links between extension, shortening and exhumation of Alpine HP rocks for the Eocene–Oligocene period is now well established (e.g. Platt, 1987). Proposed geodynamic models to explain this deformation include: a gravitational collapse model (Laubscher, 1983) and the expulsion of 'slivers' of buoyant material (Wheeler et al., 2001). In this contribution, we favour the latter model of Wheeler et al. (2001). We suggest, based on structural, metamorphic and geochronological data, that the distribution of foreland-directed thrusts and hinterland-directed detachments in the Western Alps can be explained by processes that occur within the subduction channel.

2. The high-pressure metamorphic belt of the Western Alps

The formation of the Alpine orogen is considered to result from the closure of a Neo-Tethyan ocean basin by SE-directed subduction, accompanied by the accretion of crustal fragments and the complex interleaving of various tectonostratigraphic units by regional overthrusting and extension (Coward and Dietrich, 1989; Schmid et al., 1996, 2004; Pfiffner et al., 1997; Stampfli et al., 1998; Schmid and Kissling, 2000; Reddy et al., 2003; Rosenbaum and Lister, 2005). The metamorphic internal zone of the Western Alps (commonly termed the Penninic Domain) consists of components of the European margin or the Briançonnais terrane (Internal Crystalline Massifs), allochthonous units of the Adriatic margin (Sesia–Austroalpine units) and at least two oceanic units (Liguria–Piemont and Valais; Fig. 1). The continental units and ophiolites are presently aligned parallel to the arcuate structure of the Western Alps, forming a complex orogenic structure that reflects the prolonged tectonic history of the orogen.

In this study, we focus on an E–W transect (the 'Vanoise transect') that runs through the Penninic Domain in the area around the South Vanoise and Ambin massifs (Figs. 1 and 2). Pre-Triassic crystalline basements are exposed along this transect in the South Vanoise and Ambin massifs (Briançonnais terrane) and in the Gran Paradiso and Dora–Maira massifs (Internal Crystalline Massifs). During Alpine orogeny, the basement massifs underwent blueschist- to eclogite-facies metamorphism (e.g. Ellenberger, 1960; Platt and Lister, 1985; Desmons et al., 1999). Peak metamorphic conditions have recently been estimated as 15 kbar and 500 °C in the Briançonnais (Ganne et al., 2003) and 18 kbar and 550 °C in the Internal Crystalline Massifs of the Gran Paradiso (Meffan-Main et al., 2004). These HP rocks are presently exposed in dome-shaped basement windows (Fig. 3), beneath ocean-derived allochthonous metamorphic units (Liguria–Piemont complex) (Escher et al., 1997).

Based on metamorphic and lithologic criteria, the Liguria–Piemont complex has been divided into two units (see Deville et al., 1992 and references therein): (1) a lower unit, the so-called Zermatt–Saas unit in the Swiss Alps (Escher et al., 1997), which constitutes an ophiolite-rich pelitic sequence and contains eclogite-facies assemblages indicating UHP metamorphism at conditions of up to 26–28 kbar (Reinecke, 1991); (2) an upper unit, the so-called Combin unit in the Swiss Alps

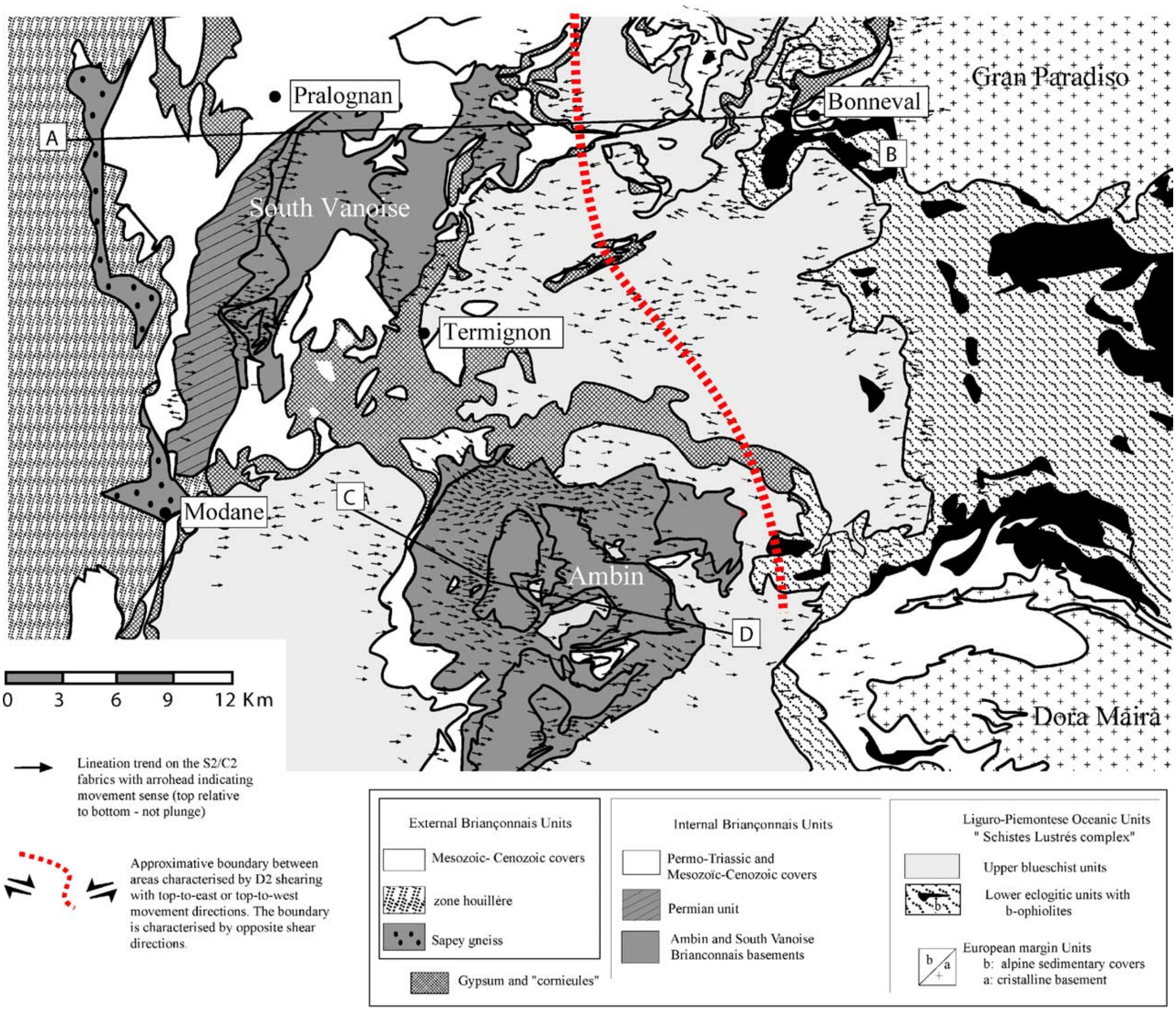

Fig. 2. Simplified geological map of the internal zone of the Western Alps in the area around the Ambin and South Vanoise massifs (modified after Ganne et al., 2005). Arrows represent the orientation of D2-related mineral lineations observed on S2/C2 blueschist and greenschist fabrics and interpreted to indicate shear sense (arrowhead indicates the sense of shear). D2 shear zones vary from east-verging shear zones in Ambin and South Vanoise to west-verging shear zones in Gran Paradiso and Dora Maira and to conjugate east- and west-verging shear zones in the Liguria–Piemont complex. Grey line (A–B) indicates the location of the cross-sections shown in Fig. 3. The dotted line marks the boundary between east- and west-directed movements on the S2/C2 fabric.

(Escher et al., 1997), which constitutes a pelite-rich sequence with scarce occurrences of ophiolitic rocks. The metamorphic grade of the upper unit is predominantly in greenschist-facies conditions (e.g. Wheeler et al., 2001). Agard et al. (2001) have shown that metamorphic conditions in the upper unit of the Liguria–Piemont complex increase from west to east, from ca. 12–13 kbar and 300–350 °C to 18–20 kbar and 400–450 °C (Fig. 4).

3. Structural evolution along the Vanoise transect

Structural observations from external parts of the Briançonnais terrane (Zone Houillère Briançonnais) show folded tectonic contacts within the monotonous pile of Carboniferous grits, conglomerates and black schists (Bertrand et al., 1996; Schmid and Kissling, 2000). The deformation associated with folding of these tectonic contacts may have also affected the crystalline basement further east, in the Ambin–South Vanoise massifs (Ganne et al., 2005). The folding event corresponds to the blueschist–greenschist transition (Ganne et al., 2003) and is manifested in the whole Penninic domain by ubiquitous growth of syn- to late-kinematic albite poikiloblasts.

At least two distinct deformational phases can be identified among major tectonic contacts in the region. The earliest deformational phase, D1, is associated with early thrust surfaces $\Phi1$ and related S1 schistosity. The schistosity is ubiquitous in the Ambin–South Vanoise region (in both basement and cover), but is rarely observed in the Zone Houillère Briançonnais. Within the cores of the Ambin and South Vanoise massifs, S1 corresponds to the main banding and its related garnet–blueschist-facies assemblages (Ganne et al., 2003). Structural observations indicate that transport directions are mainly towards the N or NW (Platt et al., 1989; Ganne et al., 2005). The early D1 chronology in the Ambin–South Vanoise massifs is not well constrained. Ganne (2003) has obtained preliminary results of 50–44 Ma ages using

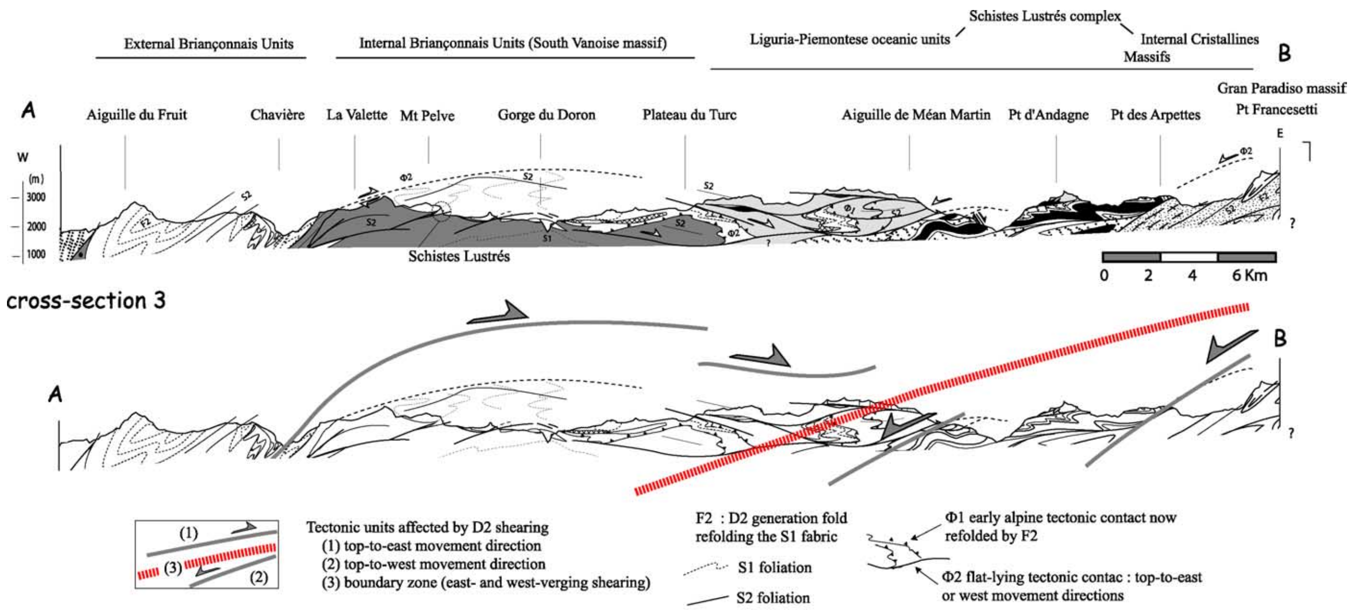

Fig. 3. Geological and schematic cross-sections through the Vanoise area (see Fig. 2 for location). A post-HP shearing deformation (D2) is responsible for the main structural pattern observed in both basement and cover units of Ambin and South Vanoise massifs. Syn-metamorphic deformation occurred under low blueschist conditions (glaucophane–chloritoid assemblage) down to greenschist facies conditions (chlorite–albite assemblage). Explanations for tectonic symbols (Φ1, Φ2, S1, S2, F1, F2) are given in the text.

^{40}Ar/^{39}Ar dating of S1-related white micas. These ages are younger than published ages for the D1 event in the surrounding Liguria–Piemont complex (62–55 Ma; ^{40}Ar/^{39}Ar dating of S1-related white micas) (e.g. Agard et al., 2002 and references therein).

The D2 deformational phase corresponds to late retrograde east-directed Φ2 shear zones, recognised at micro- to meso-scale, and related S2/C2 shear fabrics that cut the early D1 tectono-metamorphic boundaries (Fig. 3). The D2 fabrics are crosscut at high-angle by D3 brittle–ductile shear planes (Φ3). Along the western edge of the Ambin and South Vanoise domes, extensional Φ3 shear planes preferentially indicate a top-to-west sense of movement, whereas east of the domes the Φ3 shear planes are preferentially top-to-east (Ganne et al., 2005). This pattern indicates that the dome structure probably formed during the D3 stage. Ganne et al. (2005) suggested that the dominant simple shear regime (D2), which is characterised by top-to-east movement and is responsible for the development of large-scale gently-dipping shear zones, may have changed progressively with time to a pure shear regime (D3), which was characterised by high-angle conjugate extensional shear planes.

Preliminary argon dating, related to S2/C2 shear fabric in the Ambin and South Vanoise massifs, has yielded a minimum age of ca. 43–42 Ma (Ganne, 2003), which may correspond to the onset of greenschist-facies metamorphism (i.e. cooling ages; Reddy et al., 2003) during D2 deformation. This age is younger than the age of the D2 blueschist event in the surrounding Liguria–Piemont complex (ca. 45–52 Ma; Agard et al., 2002 and references therein) and older than the low-grade greenschist event discussed by Markley et al. (1998), Cartwright and Barnicoat (2002) and Reddy et al. (2003) in the Zermatt–Sass (^{40}Ar/^{39}Ar; ca. 40 Ma) and the Combin zone (Rb/Sr; ca. 36–34 Ma). Rb/Sr ages of ca. 34 Ma obtained in the Ambin massif (Ganne, 2003) are interpreted to represent the latest stage of operation along greenschist facies shear zones (D2 ± D3). Similar results have been obtained by Freeman et al. (1997) in the Entrelor shear zone farther north (Fig. 4b).

It should be mentioned that the quality of information extracted from ^{40}Ar/^{39}Ar and Rb/Sr dating predominantly depends on the microchemical and microstructural nature of the analysed micas. These methods require solid understanding of the petrogenesis of the white mica generations. The geochronological studies considered in this paper involved several levels of microstructural, microchemical and thermo-barometric control (e.g. see Agard et al., 2002; Challandes et al., 2003). A complete database of Ganne (2003) is also available at the following web address: http:/tel.ccsd.cnrs.fr/documents/archives 0/00/00/67/68 Index_fr.html.

4. Extensional structures in the Briançonnais basement massifs

Kinematic indicators in shear zones cannot easily be used to infer thrusting or extensional geometries at the time of shear formation, because such structures may have undergone tilting or rotation subsequent to shearing (Wheeler and Butler, 1994; Ring et al., 1999). In the study area, a late, gentle antiformal doming that reoriented earlier structures is recognized. Top-to-SE kinematic indicators on the western flank of the Ambin massif may therefore have a present-day apparent thrust geometry, while a similar shear sense in rocks at the eastern edge of the massif may have an apparent extensional orientation. In addition, the repetition or omission of tectonic units that can indicate thrust and extensional structures in

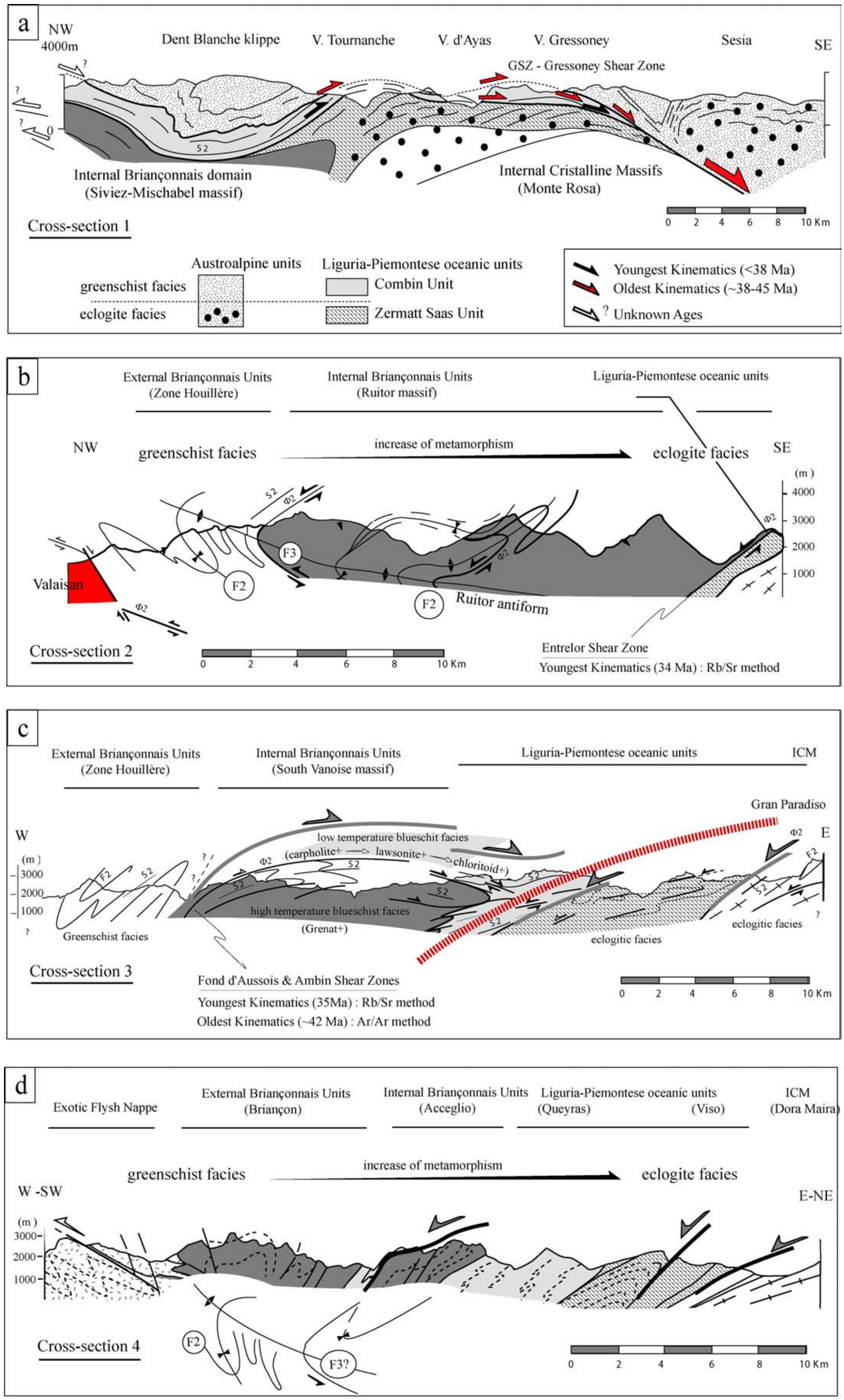

Fig. 4. Schematic cross-sections through the internal zone of the Western Alps (see Fig. 1 for locations). (a) D2 kilometric-scale shear zones show that hinterland-directed detachment faults are preferentially located in the inner part of the metamorphic belt (modified after Reddy et al., 2003). (b) Foreland-directed thrusts

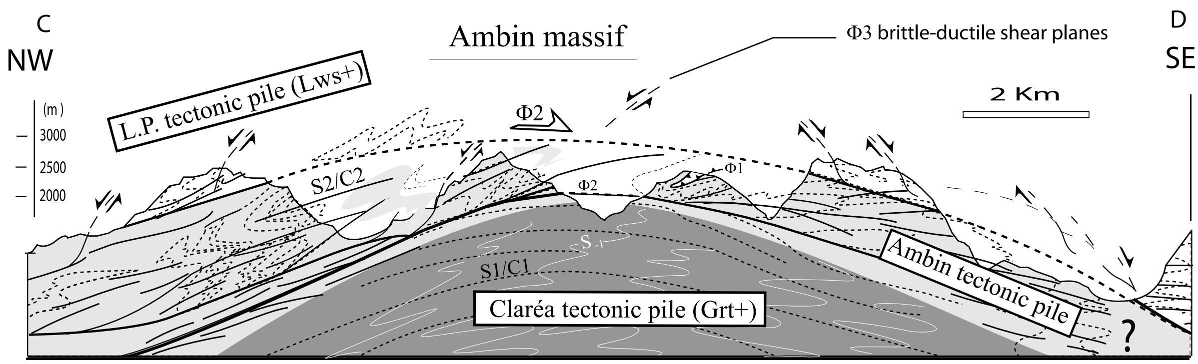

Fig. 5. E–W structural cross-section through the Ambin massif (see Fig. 2 for location), consisting of three superimposed tectonic piles: the Claréa, the Ambin and the Liguria–Piemont tectonic pile (modified after Ganne et al., 2005). The upper part of the Ambin massif was affected by a pervasive D2 shearing with top-to-east sense of movement. This D2 shearing is well expressed in the Ambin tectonic pile, where oceanic and Briançonnais cover units are strongly imbricated with slices of basement (Claréa + Ambin Groups). The deeper part of the massif (i.e. the Claréa tectonic pile), consisting exclusively of Claréa Group, preserves early HP structures (D1). The upper tectonic pile of the Liguria–Piemont complex is a lower-grade unit, overlying the higher-grade Claréa tectonic pile. The corresponding metamorphic gap suggests that the large-scale Φ2 shear zones, which are concentrated in the Ambin tectonic pile, have acted as a detachment fault. Explanations for tectonic symbols (Φ1, Φ2, Φ3, S1, S2, F1, F2) are given in the text. S_{-1} is a pre-Alpine fabric.

sedimentary sequences cannot be easily applied to areas where the original orientation of layering at the time of deformation is unknown (Wheeler and Butler, 1994; Reddy et al., 1999). Therefore, the best way to infer the regional significance of kinematic orientation is to examine the regional geometry of shear zones and to establish whether they cut up or cut down section in the metamorphic pile.

In this study, shear zones associated with D2 structures in the Briançonnais basement massifs are interpreted as extensional structures based on kinematic and geometric criteria (see also Wheeler and Butler, 1993; Ring, 1995; Ring et al., 1999). East-verging shear zones in the Ambin and South Vanoise massifs are interpreted as extensional structures because they record a retrograde P–T path at the time of operation and juxtapose lower-grade metamorphic units above higher-grade rocks along east-dipping planes.

Three main litho-tectonic units are distinguished within the different envelopes constituting the two massifs (Ganne et al., 2003) (Fig. 5). These are, from bottom to top: (1) the Claréa tectonic pile, consisting only of pre-Permian metapelites (Claréa Group); (2) the Ambin tectonic pile, consisting of slices of pre-Permian basement (Claréa + Ambin Groups) and Permo-Triassic and Triassic–Eocene metasediments; and (3) the Liguria–Piemont tectonic pile, consisting of metasediments originated from the Liguria–Piemont ocean. The three major units are separated by major metamorphic discontinuities, which are thought to have a tectonic significance (Ganne et al., 2005).

4.1. Metamorphic gap along the Ambin tectonic pile

P–T data and structural observations collected at the top of the domes indicate an overall thinning of the study area during D2 shearing. This crustal thinning began under blueschist facies conditions (10–12 kbar) (following D1 HP metamorphism) and culminated in greenschist-facies conditions at pressures of ca. 5 kbar. A retrograde P–T path, related to the D2 shearing, is shown in Ganne, (2003).

The Liguria–Piemont tectonic pile (lawsonite–blueschist facies; stability field of carpholite) lies above the Claréa tectonic pile (epidote–blueschist facies; stability field of garnet) with an interface (i.e. the Ambin tectonic pile) comprising the strongly deformed Ambin Group, a few slices of the Claréa Group and Mesozoic cover rocks. Thus, although the corresponding metamorphic gap is only a temperature gap, it suggests that the large-scale Φ2 shear zones may have acted together as a detachment fault that was subsequently deformed by open D3 antiforms (e.g. the Ambin and South Vanoise domes; Fig. 6).

4.2. The Ambin tectonic pile: a D2 east-dipping detachment

Rocks located at the hanging wall of the Ambin tectonic pile are remnants of the upper unit of the Liguria–Piemont complex and exhibit a strong metamorphic zonation (from west to east) of their D1-related HP assemblages (i.e. from 12–13 to 18–20 kbar; Agard et al., 2001). Therefore, at the time of D2 shearing, the Liguria–Piemont complex and Briançonnais basements were already exhumed to depths of less than 35 km (12 kbar; Fig. 6b). It is accordingly suggested that shear zones in the Ambin tectonic pile have operated as normal faults cross-cutting through the metamorphic isograds down to a pressure of 12 kbar. These normal faults affected an early D1 tectonic edifice consisting mainly of Liguria–Piemont complex. Since no intermediate deformation has been observed between the D1 and D2 events (Agard et al., 2001; Ganne, 2003), it must be assumed that the geometric pattern of the D1

outcrop in more external parts (modified after Bucher et al., 2003). (c) The structural boundary between thrusts and detachments (dotted line). (d) Foreland-directed thrusts in more external parts. Note that the metamorphic condition in the Liguria–Piemont complex increases from west to east, with carpholite-bearing assemblages in the west and chloritoid-bearing assemblages in the east. Explanations for tectonic symbols (Φ1, Φ2, S1, S2, F1, F2) are given in the text.

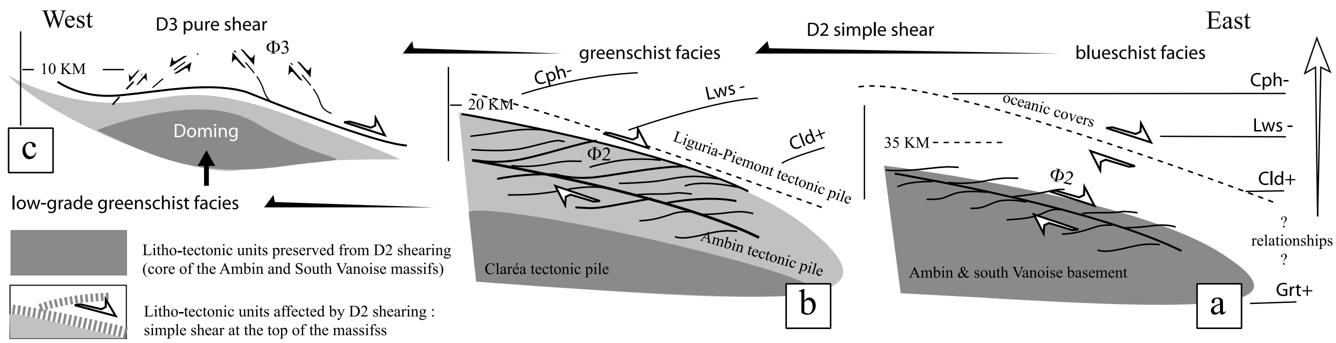

Fig. 6. Schematic illustration explaining the formation of the Ambin tectonic pile during the D2 deformation event. (a) In the proposed scenario, a post-HP simple-shear deformation (D2) is responsible for the development of major east-verging extensional shear zones enhancing the Ambin and south Vanoise basement units. The D2 extensional shear zones, occurring under low blueschist conditions (glaucophane–chloritoid assemblage) down to greenschist facies conditions (chlorite–albite assemblage), are responsible for exhuming the basement cores (i.e. the Claréa tectonic pile) within a stretching upper unit composed of oceanic cover (the Liguria–Piemont tectonic pile). The thin lines indicate D1-related metamorphic isograds subsequently crosscut by D2 extensional shear zones. Cph, carpholite; Lws, lawsonite; Cld, chloritoid; Grt, garnet. (b) During D2 shearing, slices of oceanic cover and basement constitute a tectono-metamorphic interface (i.e. the Ambin tectonic pile) equivalent to a detachment fault. (c) Finite strain analysis showing that the dome structure of the massif was probably formed during the D3 stage, when a dominant simple shear regime (D2) changed to a pure shear regime (D3) characterised by high-angle conjugate extensional shear bands.

tectonic edifice was characterized by a normal metamorphic sequence (Fig. 6a) at the time of the D2 shearing. Thus, shear zones in the Ambin tectonic pile were originally SE dipping, accounting for the top-to-east sense of movement during D2 shearing, and were subsequently folded into the present-day dome structure (Fig. 6c). Briançonnais basements and the upper unit of the Liguria–Piemont complex were juxtaposed by the D2 shearing at pressure conditions of 12–5 kbar, forming the Ambin tectonic pile (Fig. 5).

4.3. Horizontal and vertical magnitude of movement along the Ambin detachment fault

The temperature gap (≅100 °C) evidenced between the Claréa and the Liguria–Piemont tectonic piles does not enable us to determine the precise magnitude of relative movements along the Ambin detachments fault. However, the existence of originally deep rocks within the Ambin detachment fault suggests an important component of vertical movement. The oldest preserved rocks are D2 blueschist facies assemblages (42–43 Ma; Ganne, 2003) and the youngest are greenschist rocks (34 Ma; Ganne, 2003). If these rocks originated in a dipping detachment fault in which different strands had operated at different times, early sheared rocks may have been carried up passively between younger active strands of the shear zone, or redeformed to give younger deformation ages and lower pressures from recrystallized assemblages. Therefore 'old' tectonized rocks were brought from deeper levels than younger ones. Shear zones of the Ambin tectonic pile were active from blueschist (10–12 kbar) to greenschist facies (5 kbar), suggesting a vertical pressure gap of 5–7 kbar (Ganne, 2003). This is indicative of a minimum vertical shortening of 15–20 km, which was accommodated by the Ambin detachment fault. If the dipping portion of this detachment was nearly vertical, the horizontal component of extensional displacement would be close to zero. A more plausible dip of 45° leads to an estimate of 16 km of displacement on the dipping portion of the detachment to juxtapose the Briançonnais basement with the upper unit of the Liguria–Piemont complex and to decompress it from 12 to 5 kbar. Such movement remains modest compared with previous estimates given by Wheeler et al. (2001) for the Gressoney shear zone (60 km).

Based on these observations, we suggest that D1 slices that underwent HP metamorphism during the early–middle Eocene dominate the structure in the cores of Ambin and South Vanoise massifs. These slices were tectonically emplaced within Liguria–Piemont units during the Late Eocene extensional D2 ± D3 stages. The corresponding tectonic contacts (Φ2) that accommodated top-to-east sense of movement are responsible for the juxtaposition of lower-grade metamorphic units above the basements (Ganne et al., 2003) (Fig. 4c). Close to the Gran Paradiso dome, the Liguria–Piemont complex exhibits a similar geometry with a structural superposition of lower-grade rocks over higher-grade rocks along D2-type top-to-west shear-zones (Fig. 2). It is therefore possible that these top-to-west shear-zones formed simultaneously with the east-verging shear zones in the Ambin–South Vanoise massifs forming a regional-scale network of foreland- and hinterland-directed shear zones (Fig. 3). The large-scale implications of this possibility will be addressed in the following section.

5. Tectonic evolution along the Siviez–Mischabel transect

Structural analysis, performed by different authors, in the area approximately 100 km to the north of the Vanoise transect suggests regional extension associated with high-strain deformation (see Merle and Ballèvre, 1992; Wheeler and Butler, 1993; Barnicoat et al., 1995; Caby, 1996; Gebauer et al., 1997; Reddy et al., 1999, 2003). The overall orientation of foliations and lineations is relatively simple. However, the kinematic framework is complex, involving both top-to-NW and top-to-SE sense of shear that occurred at greenschist-facies

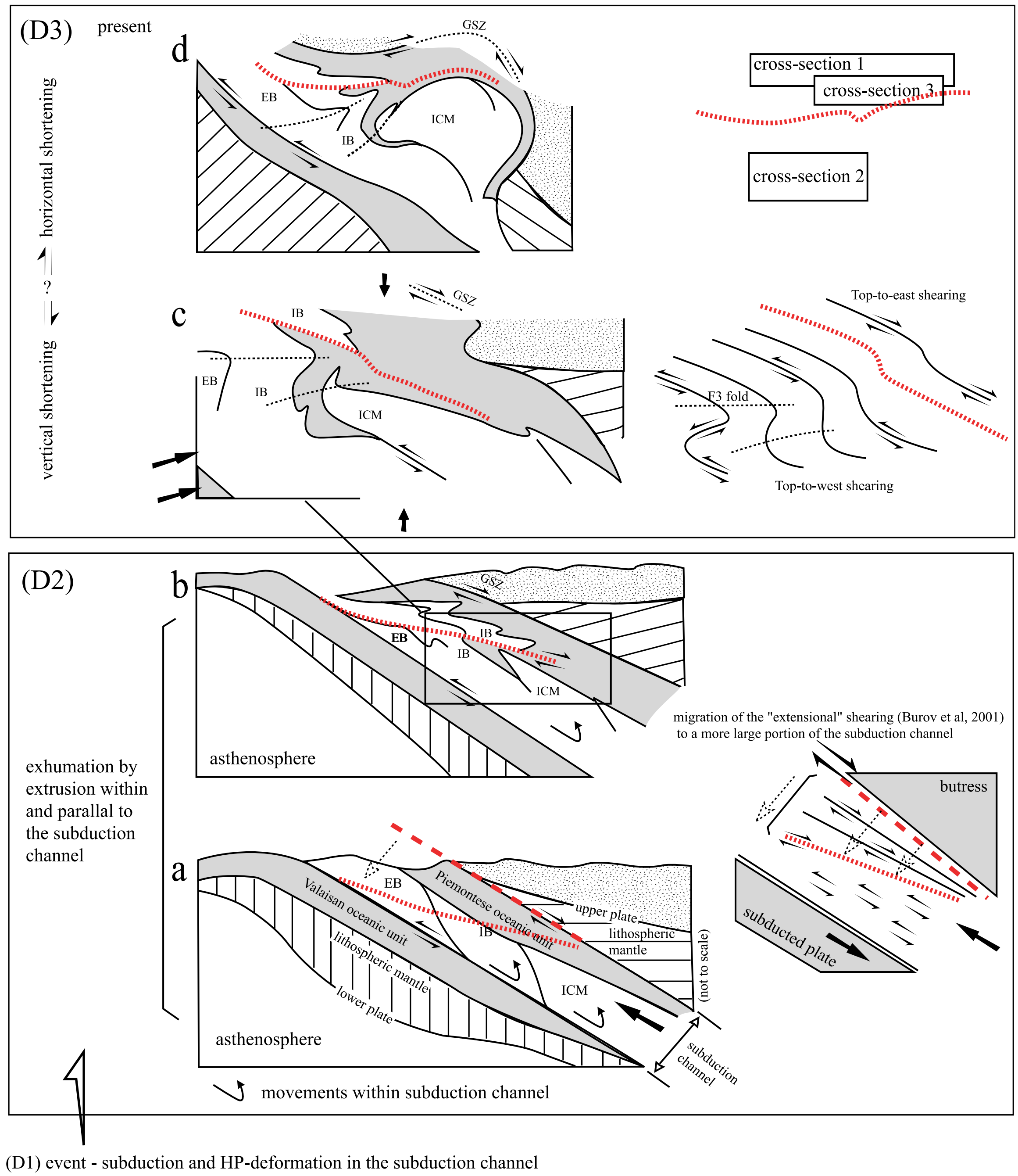

Fig. 7. Sketch of the tectonic evolution of the western Alps. (a) and (b) Exhumation by extrusion within and parallel to the subduction channel (D2) showing movements within the subduction channel (based on the model of Wheeler et al. (2001), modified by Bucher et al. (2003) for sketch (a)). Note the activity of kilometric-scale extensional detachments (east-hinterland-movements) at the top of the channel and thrusting (west-foreland-movements) at the bottom. The thick dotted line marks the boundary between the thrust domain and the detachment domain. We assume that D2 detachments were not strictly restricted to the contact with the buttress and progressively affected a larger portion of the subduction channel. (c) and (d) Evolution of the large-scale nappe refolding (D3). Note the refolding of nappe contacts at the bottom of the channel without inversion of the sense of shear. Similar structures are also shown in the model of Pfiffner et al. (2000). The position of the three cross-sections with respect to the thrust-detachment boundary (see above) is explained in the text. ICM, Internal Crystalline Massifs; IB, Internal Briançonnais units; EB, External Briançonnais units; GSZ, Gressoney shear zone.

metamorphic grade within the three major tectonic units (e.g. the Combin Zone, the Zermatt–Saas Zone and the Siviez–Mischabel Nappe). These units constitute the Siviez–Mischabel transect discussed below.

The foreland (northern part) of the middle Penninnic Siviez–Mischabel Nappe (Steck, 1984) is dipping gently to the south. Here most greenschist-grade deformation reflects imbrication beneath the overriding oceanic Piemont–Ligurian units (Zermatt–Saas Zone) and continental Austroalpine unit (Dent Blanche Nappe). Following imbrication, the Siviez–Mischabel was juxtaposed against units along strike (the lower Pennine Nappes and Lepontine Gneiss Dome to the east;

Mancktelow, 1992). At the hinterland (southern) part of the Siviez–Mischabel, the nappe dips steeply to the north and is juxtaposed against HP and UHP eclogites of the Zermatt–Saas Zone.

The Combin and Zermatt Saas Zones are lithologically complex and comprise a range of different rock types (Caby, 1981), which can be subdivided into a number of different units (Escher and Beaumont, 1997). Along the Siviez–Mischabel transect, the Combin and Zermatt–Saas Zones are dominated by carbonate-bearing rocks (calcschists) and metabasites that are intensively interbanded (Barnicoat et al., 1995). Microstructural observations performed by different authors on the oceanic-derived rocks from both zones (e.g. Merle and Ballèvre, 1992; Wheeler and Butler, 1993; Barnicoat et al., 1995; Caby, 1996; Gebauer et al., 1997; Reddy et al., 1999, 2003) allow us to distinguish three main tectonic and metamorphic stages.

D1 structures have been largely overprinted by subsequent deformation. In the Zermatt–Saas Zone, the D1 assemblage comprises garnet + jadeite + phengite + quartz + rutile ± glaucophane ± zoisite. This mineral association is typical for eclogite facies metamorphism and is consistent with the occurrence of jadeite + quartz + garnet + phengites associations described in meta-granites from the Monte Rosa massif. In the Combin Zone, the D1 blueschist assemblage is mainly represented by glaucophane + albite + lawsonite + phengite. This association is consistent with temperatures lower than 400 °C and pressures lower than 9 kbar, which is roughly the boundary between blueschist and greenschist facies conditions (Reddy et al., 1999).

D2 represents the dominant deformation event, characterized by isoclinal folds and E- to SE-verging S2/C2 shear fabrics at all scales (Reddy et al., 2003). This fabric lies subparallel to the early D1 fabric. The orientation of F2 fold axes varies from ESE to WNW along the Siviez–Mischabel transect (Reddy et al., 1999). D2 assemblage in the Zermatt–Saas Zone comprises glaucophane + phengite + quartz + zoisite + albite + ilmenite ± garnet. This mineral association is characteristic of the blueschist facies conditions (Evans, 1990). D2 assemblage in the Combin Zone comprises albite + lawsonite + actinolite + chlorite + phengite ± titanite ± calcite ± stilpnomelane. The association lawsonite + albite + phengite is consistent with temperatures lower than 400 °C and pressures between 4 and 7 kbar, therefore at the boundary between blueschist- and greenschist-facies conditions. The latter is achieved by the final crystallization of chlorite + albite + actinolite + calcite + titanite.

The D2 (blueschist to-) greenschist-facies deformation (i.e. the main tectonic event along the Siviez–Mischabel transect) is mainly localized in the Combin zone and is present over a distance of almost 50 km from Val Sesia in the SE to the northwestern part of Dent Blanche massif (Reddy et al., 2003). It should be noted that syn-kinematic greenschist facies metamorphism also occurred in the immediately adjacent rocks of the overlying Austroalpine basement and the structurally lower eclogite facies rocks of the Zermatt–Saas Zone in the east, and the greenschist facies rocks of the Siviez–Mischabel Nappe in the west (Barnicoat et al., 1995). The D2 (blueschist to-) greenschist contacts are deformed at the kilometric-scale by D3 undulations. Localized, pervasively developed, dynamically recrystallized sub-greenschist facies fabrics are developed within the hinge of these dome and basin structures (Reddy et al., 1999).

The (blueschist to-) greenschist facies contacts and their relationship to the formation and exhumation of HP metamorphic units have been debated in the geological literature (see Merle and Ballèvre, 1992; Wheeler and Butler, 1993; Barnicoat et al., 1995; Caby, 1996; Gebauer et al., 1997; Reddy et al., 1999, 2003). In particular, Wheeler et al. (2001) have specifically addressed the kinematic link between top-to-NW shear-zones and the (possibly simultaneous) SE-verging shear zones. The Siviez–Mischabel transect constitutes a key in this debate because it bridges two radically different tectono-metamorphic setting: (1) the low-grade rocks of the foreland domains (the Helvetic Nappes), characterized by top-to-NW displacements, and (2) the high-grade rocks of the hinterland domains (Piemont–Ligurian and Austroalpine units), which are mainly characterized by top-to-SE displacements.

5.1. Hinterland domains of the Siviez–Mischabel transect

Geological mapping and observations of meso-scale structures performed by Reddy et al. (1999, 2003) in the northern ends of the Val Sesia to Val Tournanche valleys (Piemont and surrounding rocks) indicate a generally similar chronology of deformation as evidenced along the Vanoise Transect. The kinematic pattern in this area is dominated by top-to-SE deformation. To the NW of Dent Blanche massif, within the Cime Bianchi in the eastern side of Val Tournanche and in Val Sesia, only top-to-SE kinematic indicators are recorded. In contrast, top-to-NW kinematic indicators are recognized in a single packet of rocks in Val Gressoney and on the western side of Val Tournanche, within or directly beneath the Dent Blanche, but structurally above the Cime Bianchi unit.

In order to infer the regional significance of the top-to-SE shear zones, Reddy et al. (1999, 2003) have proposed examining the regional geometry of structures and establishing whether they cut up or cut down section in the kinematic direction. Based on this geometric analysis and the observation of metamorphic gaps between units (greenschist Dent Blanche klippe vs. eclogite Penninic basement), the authors suggested that the most significant top-to-SE deformation was associated with extensional deformation. This interpretation contrasts earlier work that has linked top-to-SE deformation with ‘backthrusting’ (Trümpy, 1980; Baird and Dewey, 1986).

The cross-section proposed by Reddy et al. (2003) shows that the Combin Zone is a zone of extension that has undergone subsequent D3 undulation to produce a regional antiformal dome in the Piemont Zone and a synformal structure located in the Dent Blanche (Fig. 4a). The common limb of this antiformal–synformal pair passes through the western side of Val Tournanche. A similar geometry has been described at similar structural levels farther south (Philippot, 1990).

5.2. Foreland domains of the Siviez–Mischabel transect

Farther to the west, in the less metamorphosed Penninic units, Barnicoat et al. (1995) have interpreted the frontal part of the Siviez–Mischabel transect as a sequence of upright D2 thrusts imbricates. This is based on preserved sedimentary structures, which indicate that cover units are upright. This observation is consistent with other studies of Penninic nappe geometry in the eastern Alps (Baudin and Marquer, 1993; Marquer et al., 1996; Schmid et al., 1996; Markley et al., 1998; Froitzheim, 2001). These studies describe Penninic units similar to the frontal part of the Siviez–Mischabel transect as a succession of imbricate post-HP D2 thrusts in a normal stratigraphic position rather than as folds. The interpretation in term of HP-nappes, gently-folded above and within D3 domes and basins, led Barnicoat et al. (1995); Markley et al. (1998) to interpret these steepened nappe contacts as representing original top-to-NW thrusts formed during the final stages of nappe stacking (D2). These thrusts emplaced higher grade greenschist metamorphic units over lower grade metamorphic units (Wheeler et al., 2001). This reverse metamorphic field gradient, together with decompression documented during D2 (Cartwright and Barnicoat, 2002), suggest that the higher metamorphic units came to lie over the lower metamorphic units during exhumation and during the final stages of top-to-W nappe stacking.

Hence, along the Siviez–Mischabel transect, exhumation of HP units is both related to inverse faulting along D2 kilometric shear zones in its foreland part and normal faulting in its hinterland part (Wheeler et al., 2001; Reddy et al., 2003). This interpretation has been tested with geochronological data, showing that the age of top-to-NW shear deformation at the front of the Siviez–Mischabel Nappe was roughly the same as the age of gently folded top-to-SE extensional deformation in the hinterland zone of the Combin Zone (Barnicoat et al., 1995; Markley et al., 1998; Reddy et al., 1999, 2003). This partitioning of D2 shearing is not observed 30 km to the south, along the Ruitor transect.

6. Tectonic evolution along the Ruitor transect

Recent structural and metamorphic studies (e.g. Fügenschuh et al., 1999; Ceriani et al., 2001; Bucher et al., 2003) have highlighted the significance of post-HP kilometric-scale shearing and very late mega-folding events in the tectonic pattern of the Ruitor massif (Briançonnais units) and adjacent domains (Valaisan and Zone Houillère units in the west and Piemont–Ligurian units in the east). The area studied by these authors is located in a region where the dip of the main schistosity changes from SE to NW (Caby, 1996; Bucher et al., 2003). In earlier works, this change in the dip direction was referred to as 'fan structure of the Briançonnais' (Caby, 1968) and was explained by hinterland-directed shear zones that followed foreland-directed displacements. This view was supported by different authors invoking the SE-verging shear structures (i.e. backthrusting), as described by Butler and Freeman (1996, 1999) on the Entrelor shear zone and SE-verging folds structures (i.e. backfolding), as described by Baudin (1987) in the Ruitor massif.

Bucher et al. (2003) reinterpreted the so-called hinterland-verging 'backthrusting' and 'backfolding' structures (Fig. 4b). Their cross-section displays an inverse metamorphic stack of foreland-directed thrusts that have been overturned and steepened by late kilometric-scale folds. The new set of structural and metamorphic data brought by Bucher et al. (2003) strongly support our observations along the Vanoise transect. These authors have grouped this set of data into three deformation stages (D1–D3).

As discussed by Caby (1996) and Bucher et al. (2003), D1 structures have been largely overprinted by subsequent deformation. The D1 mineral assemblage (garnet, phengite, chloritoid) is associated with peak pressure condition, that range from 10–14 kbar (at temperatures around 450 °C) in the Piemont units to 5 kbar (at around 400 °C) in the Zone Houillère unit (Bucher et al., 2003). D2 represents the dominant deformation event, characterized by isoclinal F2 folds at all scales (see Fig. 4b). In most cases, the main S2/C2 shear fabric developed in the F2 axial planes is a composite of D1 and D2. The L2 stretching lineation, carryied by the S2/C2 fabrics, indicates a direction of shear toward the WNW. S2/C2 in the Piemont units is usually defined by the mineral assemblage garnet + phengite + epidote + chlorite that replaces the peak pressure assemblage (Caby, 1996). P–T estimates performed by Bucher et al. (2003) indicate that D2 is contemporaneous with decompression from 14 to 5 kbar for temperatures around 500 °C. D3 is characterized by open meso-scale parasitic folds, refolding the composite S1/S2 main shear fabric and causing a wide range of orientation for the F2 folds axes (Bucher et al., 2003). D3 fold axes are moderately NE or SW plunging with gently, SE-dipping axial planes (see Fig. 4b).

Based on mapping of the vergence of mesoscopic D3 folds, Bucher et al. (2003) identified the axial trace of two D3 mega-folds in map and cross-section views. These folds affect all previous structures, including D2 thrusting nappe-contacts. The gradual change of D1/D2 main shistosity (apparent 'fan-structure'), from a SE-dip over a subvertical orientation into a NW-dip, is a consequence of such mega-folding (Bucher et al., 2003). The first mega-fold (Ruitor mega-fold) is W-closing. The second and tectonically higher D3 mega-fold closes toward the east. This fold can be correlated with the Valsavaranche 'backfold' of Argand (1911). It returns the whole nappe stack back into an upright position, as observed by Bucher et al. (2003) in the area of the Grande Sassière, a Piemont klippe is tectonically overlying internal Briançonnais units.

Since no significant metamorphic gap has been observed by Bucher et al. (2003) across the nappe contacts, post-metamorphic normal faulting must be excluded. The interpretation in term of nappe refolding around D3 mega-fold have let Bucher et al. (2003) interpret these folded nappe contacts as representing original top-to-NW thrusts formed during the final stages of nappe stacking (D2). These thrusts emplaced a higher grade metamorphic unit over lower grade rocks. This inverse metamorphic field gradient, together with decompression documented during D2 (Bucher et al., 2003), suggests that the HP units came to lie over the lower-pressure units during exhumation and during the final stages of top-to-W nappe stacking. Hence, along this cross-section, exhumation of HP units is unrelated to normal faulting. The same conclusion is given by a geological transect located 100 km to the south.

7. Tectonic evolution along the Queyras–Monviso transect

Overturned nappe stacks linked directly to large-scale nappe refolding are recognised also along other transects of the Alpine arc (Milnes et al., 1981). In the external southern Western Alps, similar large-scale nappe refolding and refolded tectonic contacts with a top-to-W sense of shear are described by Tricart (1984), Philippot (1990), Henry et al. (1993), Schwartz (2002) and Tricart et al. (2004).

Here we briefly discuss the Queyras–Monviso transect (Fig. 4d), which shows evidence for ductile top-to-W shearing. The western part of this transect mainly consists of metasediments (e.g. the Queyras Schistes Lustrés units; Lemoine and Tricart, 1986), enclosing boudinaged ophiolitic bodies (Tricart and Lemoine, 1986). Thick ophiolitic nappes are dominant in the Monviso massif (Lombardo et al., 1978; Philippot, 1988). The Queyras and Monviso nappe piles presently dip westwards in conformity with the western flank of the Dora–Maira dome. In a regional scale, the HP–LT metamorphism grades up eastwards with jumps along tectonic contacts at various structural levels. These contacts records different metamorphic conditions (and probable different ages; Agard et al., 2002) associated with HP greenschist-facies in the Briançonnais, low grade (west) to high grade (east) blueschist-facies in Queyras (Caby, 1996; Schwartz, 2002), eclogite-facies in Monviso (Lardeaux et al., 1987; Schwartz et al., 2000) and UHP eclogite-facies in Dora–Maira (Chopin et al., 1991).

In the western part, the Queyras–Monviso transect, this tectonic and metamorphic pattern resulted from three deformation stages (D1–D3). Detailed mapping performed by Tricart et al. (2004) in the Queyras Schistes Lustrés units showed tilted piles of isoclinal folds, which affected early imbricated thin tectonic sheets. Regional-scale isoclinal folds (D1 generation) are associated with the primitive blueschist foliation. They have been refolded by D2 while metamorphism evolved towards greenschist conditions. Hence, in most cases, D2 ductile deformation is accompanied by the development of a strong omnipresent stretching lineation (Schwartz, 2002). Underlined by the last greenschist minerals, it lies within the plane of the regional foliation (S2). The regional trend of D2 structures is almost E–W, plunging westwards because of the general dip of the structure (Caby, 1996). Folds of second generation (D2) are north-verging plurikilometric recumbent folds, contrasting with the south-verging third generation of folds (D3). These D3 folds were traditionally attributed to 'back movements' (e.g. Tricart, 1975, 1984; Caron, 1977). However, in eastern Queyras, Tricart et al. (2004) have recently suggested that they are most likely representing drag folds resulting from bed-to-bed gliding within the foliation, in a top-to-west general shear.

Unlike the Ruitor transect, we have not been able to identify the axial trace of D3 mega-folds in a map view. Nevertheless, the Queyras–Monviso cross-section strongly suggests the existence of a late, post-nappe stacking, D3 mega-refolding. The hinge of this mega-fold should occur in the western part of the Queyras–Monviso transect, within the Briançonnais units (Fig. 4d). It is likely to affect all previous structures, including D2 west-verging thrusting nappe-contacts (Tricart, 1984). The gradual change of D2 main schistosity (apparent 'fan-structure') from E-dipping over a subvertical orientation into W-dipping more to the east (Dora–Maira massif) could be the consequence of such mega-folding.

This D3 event, considered as an episode of extensional shearing by Bucher et al. (2003) and Tricart et al. (2004), has also been recognised further north (Agard et al., 2001; Ganne et al., 2005). It corresponds with thinning of the structure through tectonic denudation (see Tricart et al. (2004) for discussion). The onset of denudation tectonics was confirmed by the subsequent development of west-dipping extensional shear zones between eastern Queyras and Dora Maira (west-Viso and west-Dora–Maira ductile normal faults). They guide the extensional denudation of the Dora–Maira basement nappes below the Monviso ophiolites, which are denuded below the eastern Queyras Schistes Lustrés (Ballèvre et al., 1990; Black and Jayko, 1990).

8. Discussion

8.1. The tectonic significance of D2 deformation

Structural observations from the internal zone of the Western Alps indicate that the development of kinematic indicators associated with regionally high-strain shearing deformation and syn-kinematic recrystallization operated from blueschist- to greenschist-facies conditions (Reddy et al., 1999; Agard et al., 2002; Bucher et al., 2003; Tricart et al., 2004; Ganne et al., 2005). This deformational event (D2) extends laterally for almost 100 km, from the eastern side of the Penninic Front (Fügenschuh et al., 1999; Ceriani et al., 2001) to the western side of the Dora–Maira, Gran Paradiso and Monte Rosa massifs, across the Liguria–Piemont complex (this study) (Fig. 4). Despite local complexities that are attributed to the role of D3 refolding (Bucher et al., 2003) or D3 doming (Reddy et al., 2003), the distribution of east-verging extensional shear zones (normal metamorphic sequence) and west-verging thrusts (inverse metamorphic sequence) is relatively systematic. Foreland-directed thrusts are restricted to the western part of the internal zone (see fig. 4 of Bucher et al., 2003), whereas hinterland-directed detachments outcrop in the eastern parts (see fig. 8 of Reddy et al., 2003). These thrusts and extensional shear zones were possibly synchronous (Wheeler et al., 2001). They represent a large-scale shearing event in the western Penninic Domain that took place during the late Eocene and terminated at ca. 34–36 Ma (Barnicoat et al., 1995; Freeman et al., 1997; Markley et al., 1998; Agard et al., 2002; Cartwright and Barnicoat, 2002; Challandes et al., 2003). This heterogeneous shearing event and crustal deformation partitioning has played an important role in the exhumation of HP metamorphic rocks.

Kinematic links between extension, shortening and exhumation of Alpine HP rocks have been explained in different geodynamic models, such as post-orogenic collapse (Laubscher, 1983), or expulsion of 'slivers' of buoyant material (Wheeler et al., 2001). At the front of the belt of the Western Alps, HP units are partly exhumed along low-angle thrust (Fig. 4b), which operated during decompression and emplaced higher-grade metamorphic units over lower grade rocks (Bucher et al., 2003). This tectonic configuration does not seem to support exhumation during extensional collapse (e.g. Jolivet et al., 2003) because, in this model, the exhumation of high-grade rocks is not likely to occur along thrusts, but solely along extensional detachments. Therefore, subduction dynamics and tectonically assisted erosion are the most likely mechanisms accounting for the development of inverse and normal metamorphic sequences along west-directed thrusts and east-directed detachments, respectively.

A model involving buoyancy-driven extrusion within a subduction channel (e.g. Burov et al., 2001) is presented in Fig. 7. This model accounts for the activation of extensional detachments at the top of the channel and thrusts at the bottom during an overall compressional regime. Wheeler et al. (2001), Reddy et al. (2003) and then Jolivet et al. (2003) have suggested a similar model in the tectonic framework of the Western Alps. According to this model, during the late Eocene, a body of rock (corresponding to the internal zone) was subjected to extrusion in a subduction channel (Fig. 7a) and was bounded by extensional detachments above and thrust faults below. Given the top-to-west sense of thrusting during extrusion, the study areas of Bucher et al. (2003) (Fig. 4b) and Schwartz (2002) (Fig. 4d) must have been located near the lower interface of the extruding HP units (Fig.7d). Conversely, the top-to-east sense of movement that was mapped by Reddy et al. (2003) along an extensional detachment (Fig. 4a) indicates a position near the upper interface (Fig. 7d).

We favour the subduction channel originally proposed by Wheeler et al. (2001) as the most likely syn-orogenic Alpine model for the exhumation of HP rocks along D2 kilometric-scale shear zones. As predicted by the model, east-directed detachments were preferentially located at the inner part of the Alpine metamorphic belt, whereas foreland-directed thrusts outcrop in the more outer parts. Along the Vanoise transect (Figs. 2 and 3), it seems that this tectonic pattern was modified by subsequent (D3) deformation that inverted the distribution of east-directed detachments and west-directed thrusts.

8.2. Horizontal vs. vertical shortening during the D3 event

Zircon fission track ages (e.g. Hurford et al., 1989) indicate that the whole Penninic domain was exhumed to upper crustal levels at the Eocene–Oligocene boundary. Thereafter, there is no evidence for the exhumation of HP rocks along shear zones. It therefore seems that subduction dynamics ceased during that time, either as a result of slab breakoff (e.g. Von Blanckenburg and Davies, 1995), or due to an increase in buoyancy as progressively more low-density continental lithosphere entered into the subduction zone.

Bucher et al. (2003) proposed that mega-folding occurred at this time in front of and above the relatively rigid Gran Paradiso massif. This stage of folding involved the combination of inhomogeneous horizontal simple shearing (e.g. buckling of Burg et al. (2002)) and vertical shortening (Fig. 7c). This partitioned brittle–ductile deformation, called D3 by different authors (Bucher et al., 2003; Reddy et al., 2003; Tricart et al., 2004; Ganne et al., 2005), took place during the differential WNW-directed movement of the Internal Crystalline Massifs with respect to both to the Valaisan suture and the overlying Austroalpine units (Bucher et al., 2003). As shown in Fig. 4a and c, D3 mega-folding is of only local significance at the scale of the Western Alps. However, it is strongly expressed at the bottom of the remnant D2 subduction channel (Fig. 7). Elsewhere, D3 mega-folds have apparently evolved toward large-scale open antiforms and synforms (domes and basins geometry) at the top of the channel (Fig. 7c).

9. Conclusion

Exhumation of HP and UHP rocks in the Western Alps was facilitated by the widespread D2 shearing event during the late Eocene (ca. 45∼35 Ma). This event took place in a subduction channel (corresponding to the present internal zone). It involved the decompression and the activation of east-directed kilometric-scale extensional detachments at the top of the channel and west-directed thrusts at the bottom.

As shown in the four geological transects across the Western Alps, the distribution of thrusts and detachments was controlled by the position of the extruding high-pressure unit within the subduction channel. The Vanoise transect crosscut the boundary between the domain of thrusts and the domain of detachments (the boundary does not outcrop along the other transects). The original geometry of this boundary has been gently tilted by D3 doming, which locally modified the distribution of the hinterland and foreland-directed shear zones in the orogen.

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