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Structural analysis of the Subbriançonnais and Valais units in the area of Moûtiers (Savoy, Western Alps): paleogeographic and tectonic consequences

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Abstract Valais and Subbriançonnais units of the Western Alps of Savoie underwent a common structural evolution, postdating peak pressure conditions associated with high-pressure metamorphism of internal parts of the Valais units. The first two phases, due to roughly north/south-directed shortening, are interpreted to be related to a NNE/SSW-striking corridor of sinistral transpression between the internal Western Alps and the European foreland. Both phases led to nappe formation, isoclinal folding and north–south elongation. Only the third phase of deformation is related to WNW-directed orogen-perpendicular shortening, thus far regarded as the predominant thrusting direction in the Western Alps. Late (post 5 Ma) normal faulting, evidenced by fission-track dating, reactivated the Houiller Front in the north and the Penninic Front in the south. Kinematics of movement, observed along the present-day Houiller Front and Penninic Front, change from north to south. In the north the Houiller Front indicates post-D3 normal faulting while the Penninic Front preserved WNW-directed thrusting (D3). In the south the Houiller Front preserves syn-D2 north-directed thrusting, whereas the Penninic Front is partly reactivated by post-D3 normal faulting. Our observations clearly favor tectonic reasons for the disappearance of the Valais units south of Moûtiers in present-day map view.

Key words Western Alps · Valais domain · Subbriançonnais · Penninic Front · Houiller Front · Structural geology · Fission-track dating

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Introduction

According to the "multi-ocean" concept for the Alps, the North Penninic or Valaisan (Trümpy 1955) paleogeographic domain represents a second, more northerly located basin with respect to the South Penninic or Piemont-Liguria oceanic basin (e.g., Froitzheim et al. 1996; Schmid et al. 1996; Stampfli and Marchant 1997). This Valais basin is floored partly by oceanic crust and exhumed subcontinental mantle (e.g., Florineth and Froitzheim 1994), and partly by continental upper crust. The oceanic part opened in late Jurassic–Early Cretaceous time (Frisch 1979; Stampfli 1993). According to these authors this led to the separation of the Briançonnais microcontinent from Europe s.str.

Tectonic units attributed to the North Penninic paleogeographic domain can be found along most of the Alpine chain, from the eastern Alps in Austria (Rhenodanubian Flysch) through the classical areas in Switzerland (Graubünden and Valais) into the northernmost French Alps of Savoy near Moûtiers (Fig. 1). However, in map view the Valais units wedge out south of Moûtiers and they do not continue along the arc of the Western Alps further south. Instead, they are laterally replaced by Mesozoic cover nappes attributed to the Subbriançonnais paleogeographic domain (Fig. 1); the latter occupy the same tectonic position as the Valais units, i.e., they are also situated between a more external fault zone (the Penninic Front) and a more internal fault zone (the Houiller Front). This paper focuses on a comparative study of the structural evolution within the Valaisan and Subbriançonnais units (i.e., "unités penniques frontales" of Debelmas 1980), together with structural studies and fission track dating addressing the kinematics of movement along the Penninic Front and Houiller Front fault zones in the area of Moûtiers (Savoie). In light of these results we discuss (a) the arrangement of the different Valais tectonic units within the Valais paleogeographic domain, (b) the metamorphic zonation within the Valais and Subbriançonnais units (blueschist/eclogite

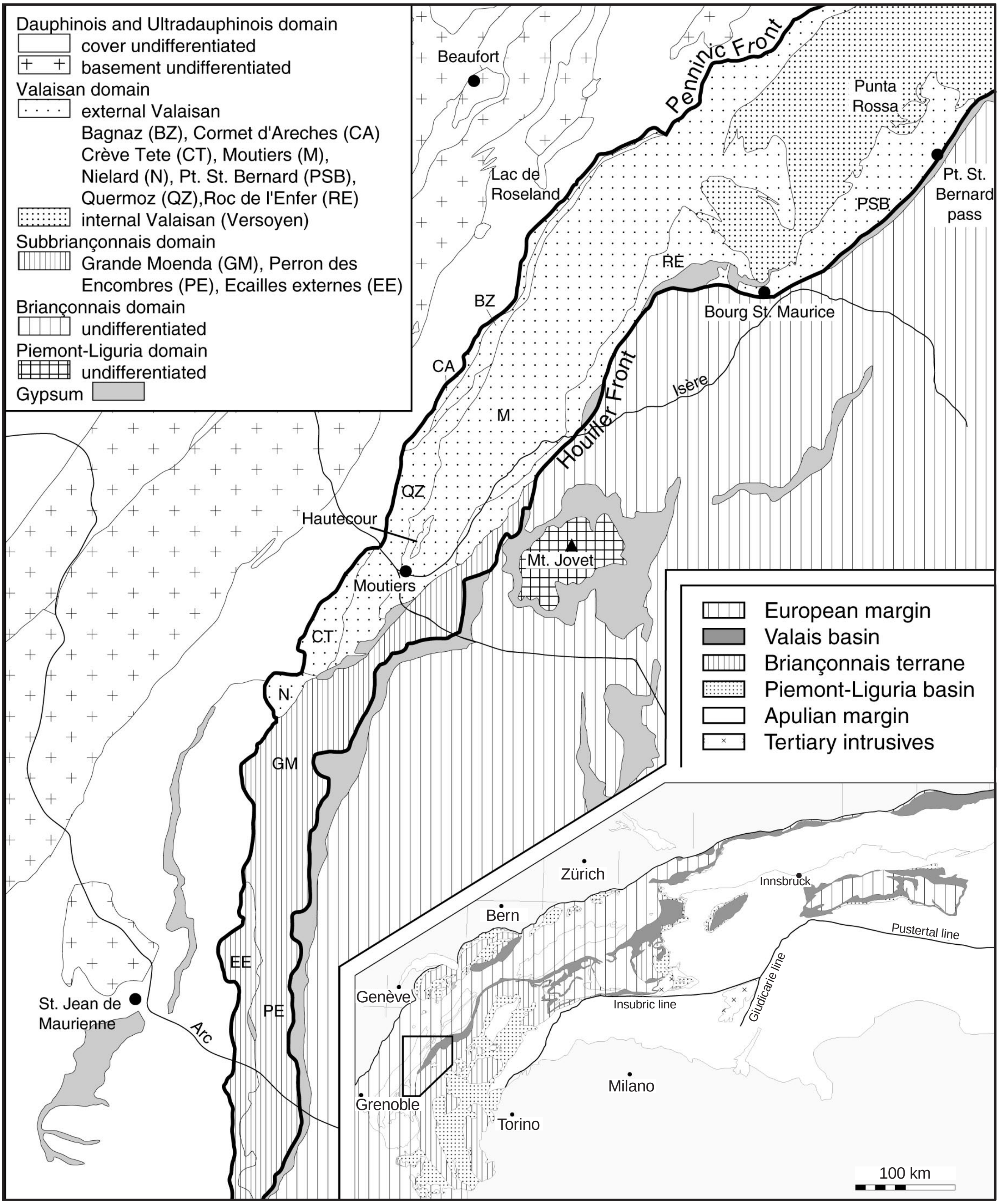


Fig. 1 Tectonic map of the area between Pt. St. Bernard pass (French–Italian border) and Arc Valley based mainly on the structural schemes of "Carte géologique de la France" feuille Bourg St. Maurice and feuille St. Jean de Maurienne. Inset shows distribution of the Valais paleogeographic domain redrawn after the "Structural model of Italy" (Bigi et al. 1983)

vs. greenschist or sub-greenschist facies), and (c) the kinematics along the Penninic Front, and Houiller Front fault zones, respectively.

This discussion will help to answer the following important question: Do the Valais units wedge out south of Moûtiers primarily for paleogeographic reasons (the "classical" view, which considers the Briançonnais of the French Alps as the distal European margin), or is this wedging out due to Alpine tectonic displacements? According to the latter option the Valais paleogeographic domain would originally have extended further to the south, the opening of this partially oceanic domain being kinematically linked with the opening history of the Atlantic Ocean (Frisch 1979; Stampfli 1993; Stampfli and Marchant 1997).

Geological setting

The investigated area extends from the Pt. St. Bernard pass in the NE (Italian–French border) to the Arc Valley in the SW (Fig. 1). Enclosed between two major fault zones, namely the Penninic Front (Bertrand et al. 1996) and the Houiller Front (referred to as Briançonnais Front by Bertrand et al. 1996), a series of tectonic units belonging to the Valaisan and Subbriançonnais paleogeographic domain have been mapped out in Fig. 1. The Penninic Front separates the Penninic units from units attributed to the Dauphinois and Ultradauphinois paleogeographic domain (i.e., Europe s.str.), whereas the Houiller Front delimits the western edge of the Zone Houillère (more internal Penninic units of predominantly Permo-Carboniferous age, attributed to, from south to north, the Briançonnais, Subbriançonnais, and Valais paleogeographic domains; Escher 1988; Escher et al. 1997; Stampfli and Marchant 1997).

In Fig. 1 the Valaisan domain or "Nappe des Brèches de Tarentaise" (Barbier 1948) is divided into external and internal Valaisan respectively. The external Valaisan includes, from west to east, the following units: Cormet d'Arêches, Bagnaz, Crève Tête, Niélard, Quermoz, Moûtiers, and Roc de l'Enfer (Antoine 1971; Antoine et al. 1992; Antoine et al. 1993). The Pt. St. Bernard unit, according to our findings also part of the Valais domain, was previously attributed to the Subbriançonnais paleogeographic domain (e.g., Elter and Elter 1957, 1965). The internal Valaisan is composed of the Versoyen unit. The Valais tectonic units represent detached cover nappes, with only small relics of continental basement preserved, namely the Hautecour crystalline near Moûtiers and the Punta Rossa crystalline near the Pt. St. Bernard pass, both of questionable tectonic and stratigraphic position. Sediments which are generally taken to be typical for the Valais domain are of Cretaceous (Barrêmian) to Tertiary age and often referred to as "flysch" (Antoine et al. 1993). We consider the basal part of this "flysch" ("Couches d'Aroley") to represent post-rift sediments unconformably overlying and reworking older formations. Mid-Cretaceous quartzites and black shales ("Couches des Marmontains") are followed by a thick sequence of Upper–Cretaceous to Tertiary sediments ("Couches de St. Christophe"). Only the latter are flysch-type sediments. The base of the Couches d'Aroley either overlies a partly eroded sedimentary substratum deposited on continental crust (Permo-Triassic continental deposits or Triassic to Liassic carbonate platform, typical for the Moûtiers unit) or a "complexe antéflysch" (Antoine 1971) which often contains abundant mafic sills, pillow lavas and rare occurrences of serpentinite (representing the ophiolitic or at least partly oceanic part of the Valaisan referred to as the Versoyen). Intermediate stratigraphic levels of unknown age (post-Liassic, pre-Barrêmian) may be interpreted as Upper Jurassic syn-rift breccias (Brèche du Grand Fond formation of the Moûtiers unit and Brèche du Collet des Rousses of the Versoyen).

The proposed subdivision into internal and external Valaisan, respectively, is based partly on structural evidence outlined herein and partly on significant differences in the stratigraphic record. These differences are depicted in Fig. 2 where the external Valaisan represents a basin floored by continental crust, the post- and syn-rift sediments having been deposited onto an eroded carbonate platform. In the internal Valaisan (i.e., Versoyen) these sediments are interpreted to have directly overlain a substratum of oceanic crust situated near the continent–ocean transition. Remnants of this substratum are only very rudimentarily preserved because most of the oceanic crust of the internal Valaisan was subducted.

A high-pressure/low-temperature metamorphic overprint ($\leq$18 kbar, 350–400 °C) has been demonstrated for the ophiolitic Versoyen unit (Schürch 1987; Cannic et al. 1996) as well as for the Pt. St. Bernard nappe (Goffé and Bousquet 1997), in response to a subduction scenario. For the more external parts of the Valaisan (i.e., Moûtiers unit) estimated pressure conditions did not exceed 10 kbar (Goffé and Bousquet 1997).

The Subbriançonnais cover nappes south of Moûtiers are composed of, from external to internal: "Ecailles externes," Grande Moëndaz unit, and Perron des Encombres unit (Barbier 1948). These cover nappes are detached along Carnian evaporites (Grande Moëndaz and Perron des Encombres units) and Oxfordian schists (Ecailles externes), respectively. The Grande Moëndaz and Perron des Encombres units comprise a continuous stratigraphic record from the Carnian evaporites up to the Oxfordian schists. Paleogeographically these two units individualized during the middle to upper Liassic, characterized by the formation of horst-graben structures related to the opening of the Piemont-Liguria ocean (Loreau et al. 1995). Whereas the Grande Moëndaz unit represents a basin characterized by the deposition of Lower Liassic marly limestones, a relatively shallower position is indicated by the coeval deposition of massive limestones within the Perron des Encombres unit. A further strati-

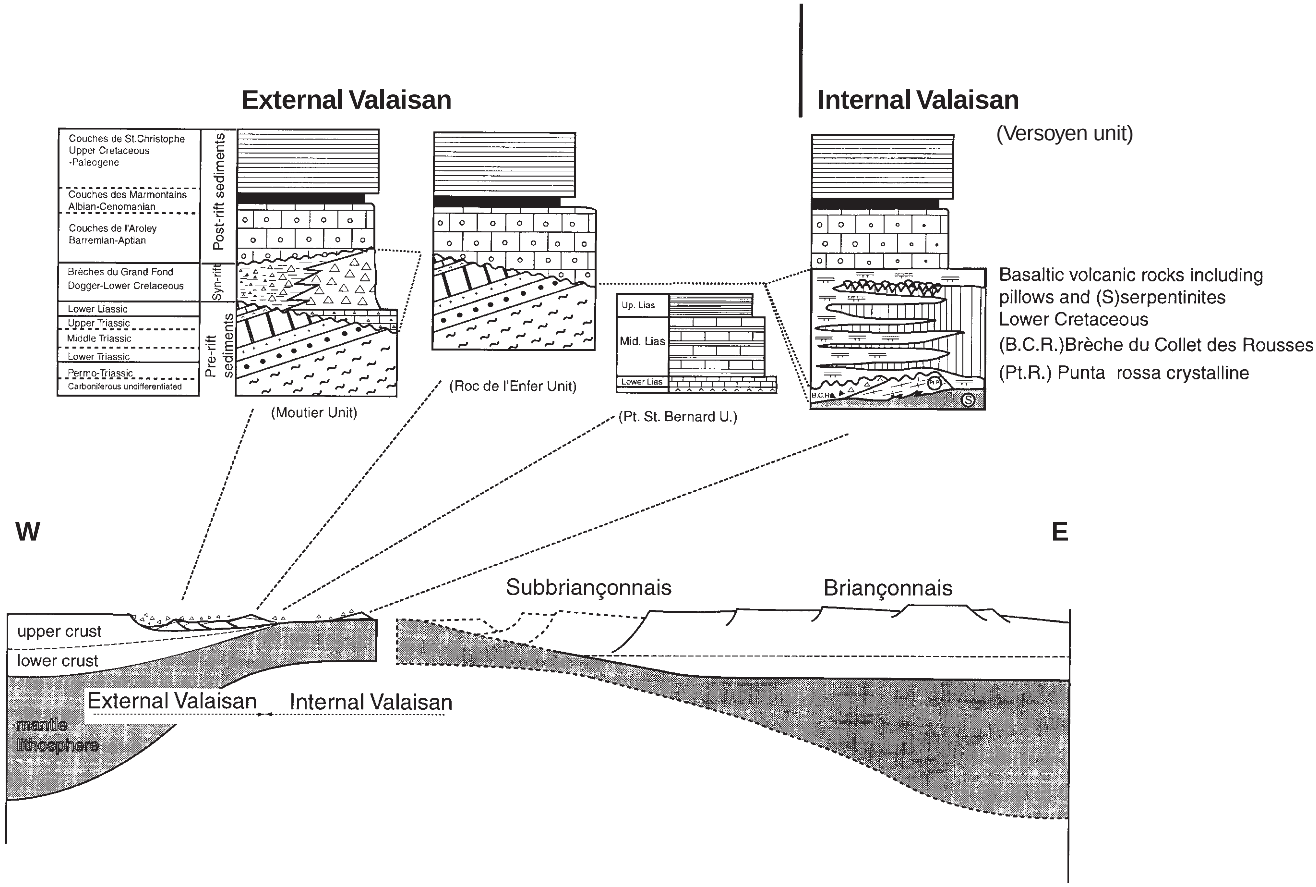


Fig. 2 Proposed palinspastic reconstruction of the different units based on our structural analysis, together with their stratigraphic record. (After Antoine et al. 1971)

graphic difference concerns the Oxfordian “Brèches du Télégraphe” (Barbier 1948; Perez-Postigo 1988), only present within the Perron des Encombres unit. With respect to metamorphic grade, preliminary illite crystallinity data indicate temperatures $<300\,°C$, with no indication for a significant pressure overprint.

Different models have been proposed for the formation of the fault zones bounding the Valais and Subbriançonnais units (Penninic Front and the Houiller Front). While some authors regard the Penninic Front as a top-to-the-WNW-directed thrust (e.g., Butler et al. 1986), others argue for normal fault displacements (e.g., Seward and Mancktelow 1994). Normal faulting along the Houiller Front has been proposed by Bertrand et al. (1996), in contrast to thrusting as deduced, for example, by Freeman et al. (1998). Major sinistral strike-slip faulting along both these tectonic contacts has been proposed as well (e.g., Ricou and Siddans 1986).

Structural evolution of the Valais zone

Detailed structural re-mapping (Fig. 3) forms the base for the construction of a series of cross sections (Fig. 4). These data are presented in order to unravel the hitherto unrecognized complex structural evolution of the Valaisan in the area between the Roselend pass (Penninic Front fault zone) and Bourg St. Maurice (Houiller Front fault zone). The reinvestigated area roughly follows the ECORS-CROP seismic line (Nicolas et al. 1990). The structural analysis led to the reinterpretation and reorganization of the classical units (mentioned previously, e.g., Antoine 1971) mapped out in Fig. 3. Three phases of deformation, associated with folding, affect the entire Valaisan and have already been described previously (e.g., Lancelot 1979; Spencer 1989); however, the overprinting relationships and their tectonic impact have remained unclear so far.

D1 is characterized by a penetrative bedding-parallel schistosity (S1) related to F1 isoclinal folds. Due to the extreme attenuation of the F1 folds they are hardly ever observable on the outcrop (e.g. Fig. 5). The regional impact of large-scale F1 folds and associated thrusts can only be deduced from opposing younging directions observed in sediments subsequently affected by major F2 folds, overprinting normal and overturned F1 fold limbs, respectively. Thereby it is possible to map the axial trace of a major F1 fold within the internal Valaisan (Fig. 4). D1 stretching lineations (L1) can be inferred from the alignment of minerals (e.g., chloritoid), whereas the directions of principle exten-

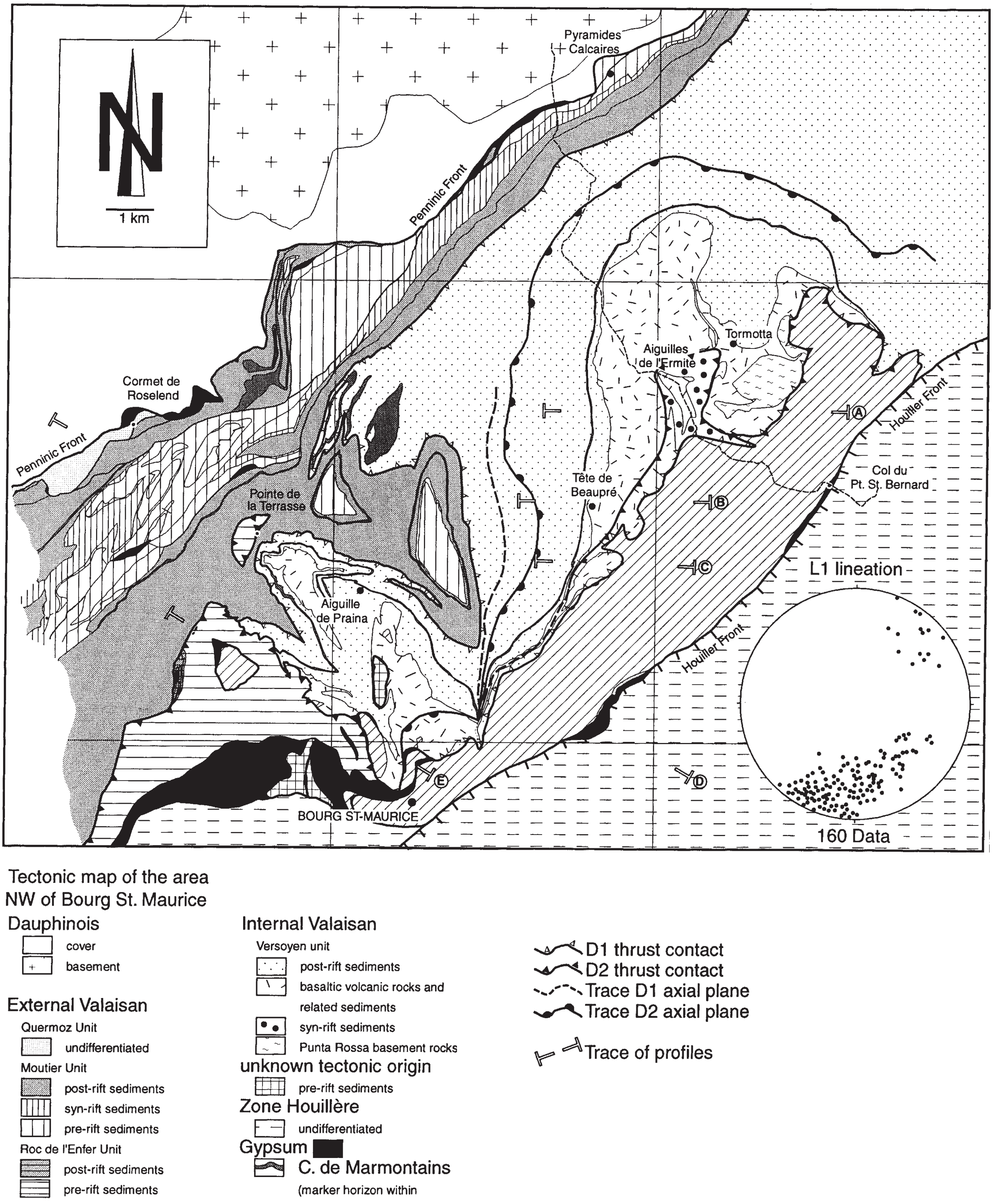


Fig. 3 Geological map of the area between the Penninic and Houiller Front, respectively, north of Bourg St. Maurice. Open triangles outline the pre-D2 thrust contact between internal and external Valaisan, *solid triangles* indicate syn- to post-D2 thrust contacts. *A–E* gives location of profiles as depicted in Fig. 4. The *stereoplot* gives the orientation of L1 stretching lineations measured within the Valaisan units

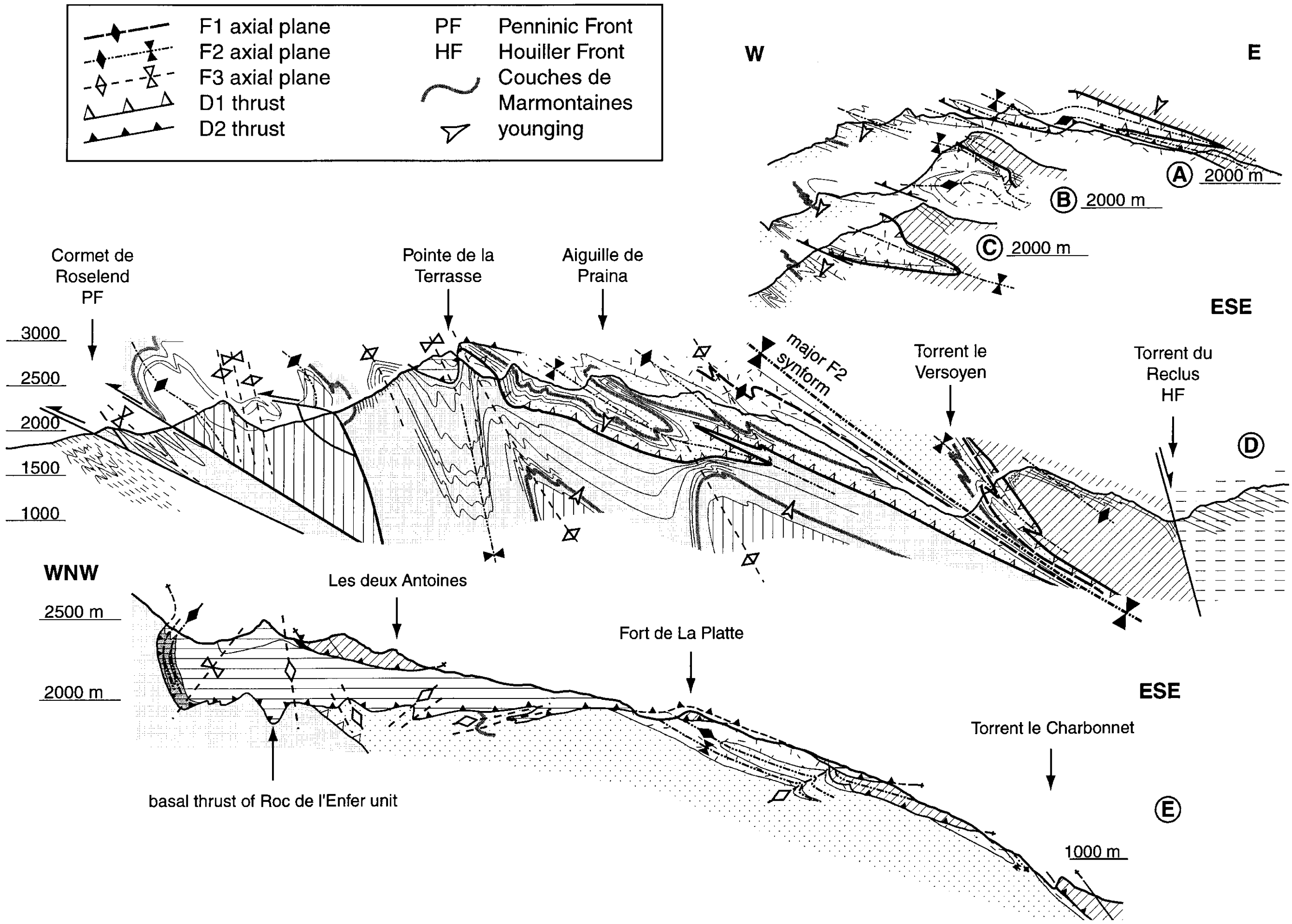


Fig. 4 Series of profiles (for location seeFig. 3) through the Valaisan; patterns are the same as in Fig. 3. No vertical exaggeration. Profiles *A–D* outline the relationship between the internal and external Valaisan. Note the increasing intensity of D3 folds towards the WNW which are cut by the Penninic Front (profile D). Profile E (note the difference in scale) illustrates primarily the basal thrust of the Roc de l'Enfer unit (late D2), cutting the D1 thrust contact between internal and external Valaisan

sion are given by the orientation of boudin necks and deformed elongated sedimentary clasts (the latter being preferentially preserved within the post-rift Aroley breccias). The derived directions of maximum extension, although somewhat scattered due to subsequent refolding, trend roughly NNE–SSW (cf. stereoplot in Fig. 3). This intense D1 stretch very probably implies roughly north-directed D1 nappe stacking, i.e., subparallel to the present-day strike of the Western Alps. However, some reorientation due to later deformation phases could have occurred, and senses of shear related to D1 could not be found. D1 not only led to this intense large scale folding but also formed the tectonic contact between the external and internal Valaisan, as can be deduced from the fact that this thrust is folded by D2 (Figs. 3, 4). D1 detachment horizons are strongly

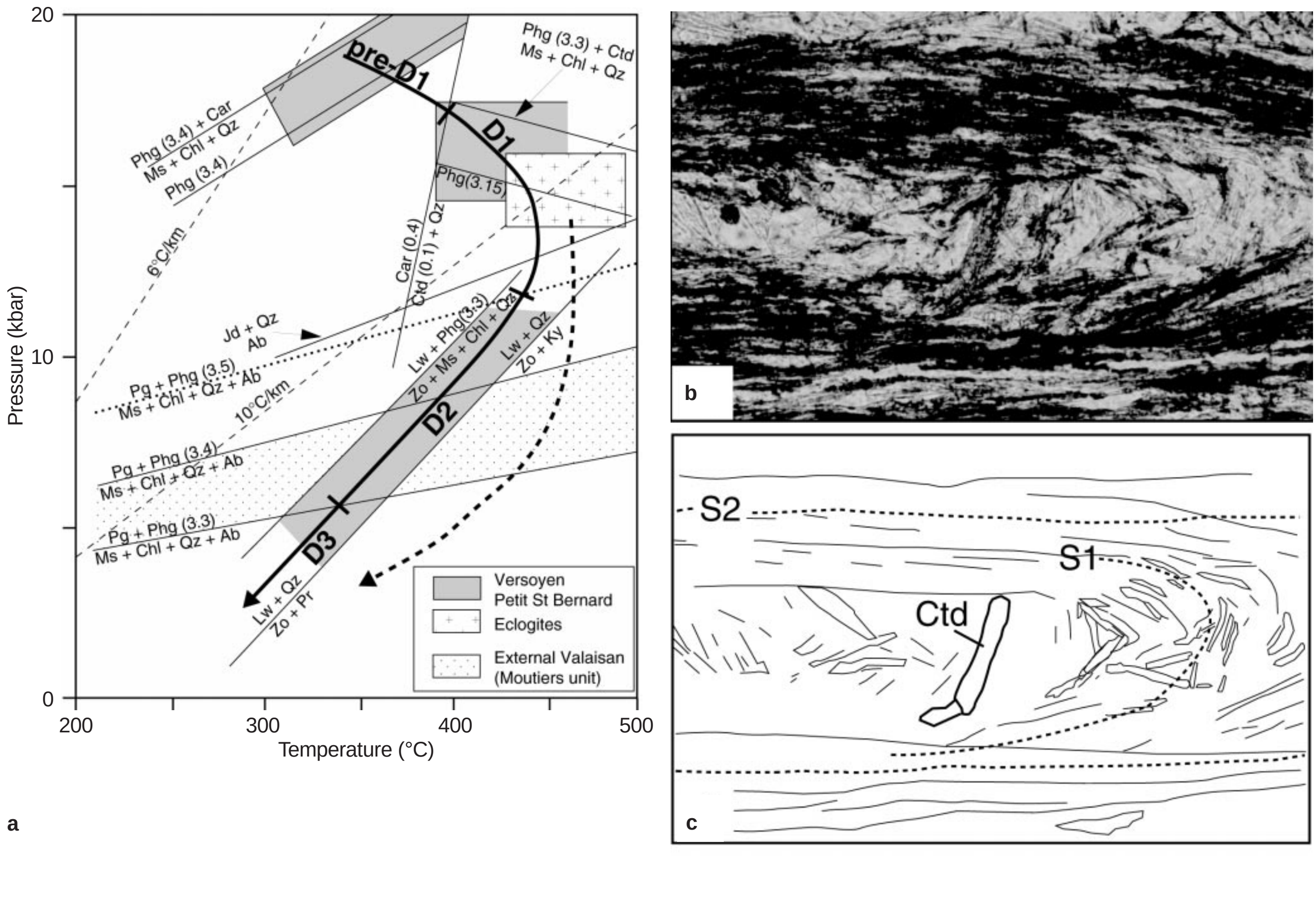
Fig. 5 Isoclinal F1 fold (axial plane indicated) in "Couches de St. Christophe" in the hinge of an F2 fold, internal Valaisan. *Arrow* indicates an F1 parasitic fold. *Camera lid* for scale

controlled by the depth of erosion of post- and syn-rift sediments (Fig. 2). The Pt. St. Bernard unit is detached along the Upper Triassic evaporites. Within the rest of the external Valaisan, however, this Upper Triassic detachment horizon had been eroded before deposition of syn- and post-rift sediments (Fig. 2), hence before the onset of D1. Consequently, most of the external Valaisan was detached along an older potential detachment horizon, namely the Carboniferous schists.

Concerning the tectonometamorphic evolution, our deformation phase D1 already postdates peak pressure conditions. Following the arguments by Goffé and Bousquet (1997) chloritoid is most likely to replace carpholite during exhumation, according to the reaction ferrocarpholite ⇒ chloritoid + quartz + water. Whereas Goffé and Bousquet (1997) state that chloritoid commonly is aligned within the main foliation, without specifying this foliation, our own observations allow us to observe chloritoid parallel to, i.e., grown within the first foliation, and refolded by, D2 (Fig. 6). Hence, no penetrative deformational features (schistosity, lineation) related to subduction and associated with the growth of carpholite during peak pressure conditions (pre-D1) have been detected so far within the Valaisan. Presumably this is due to the pervasive D1 overprint of all pre-D1 subduction-related structures.

The second phase of deformation D2 formed tight to isoclinal folds on all scales, among them spectacular folds in the "Vallée des glaciers" (e.g., Antoine 1971; Fudral 1996).Other examples of F2 folds are described by Cannic et al. (1996) who deduced their existence from the distribution of igneous textures within mafic laccoliths (area of the Aiguilles de l'Ermite). F2 axial planes and associated intense axial plane schistosity (S2) gently dip toward the east to SE. F2 fold axes are subparallel to the D1 principal stretch, gently (5–20°) plunging toward 190–220° in the western part and toward 180–150° in the eastern part of the working area. The dip angle increases southwards to values between 35 and 40°. The axial trace of a major D2 synform, refolding the D1 nappe stack, can be mapped north of Bourg St. Maurice (Fig. 3). Most of the internal Valaisan, forming the core of this D2 synform, rests in its inverted limb. Toward the south, near Bourg St. Maurice, this D2 structure is related to a complex D2 thrust system, which is responsible for the detach-

Fig. 6 **a** Pressure–temperature paths determined for the Valaisan domain redrawn after Goffé and Bousquet (1997), indicating the P–T conditions which prevailed during pre-D1, D1, and D2/D3 deformation, respectively, based on microstructural observations. Data from Goffé and Bousquet (1997) and Cannic et al. (1996). *Solid line* P–T path proposed by Goffé and Bousquet (1997). *Broken line* P–T path proposed by Cannic et al. (1996). Line indicating the reaction Pg + Phg (3.5) → Ms + Chl + Qz + Ab, valid for the internal Valaisan, is taken from Cannic (1996), all other reactions are taken from Goffé and Bousquet (1997). **b, c** Thin section and line drawing, respectively, of chloritoid-bearing black schists ("complexe antéflysch," internal Valaisan)



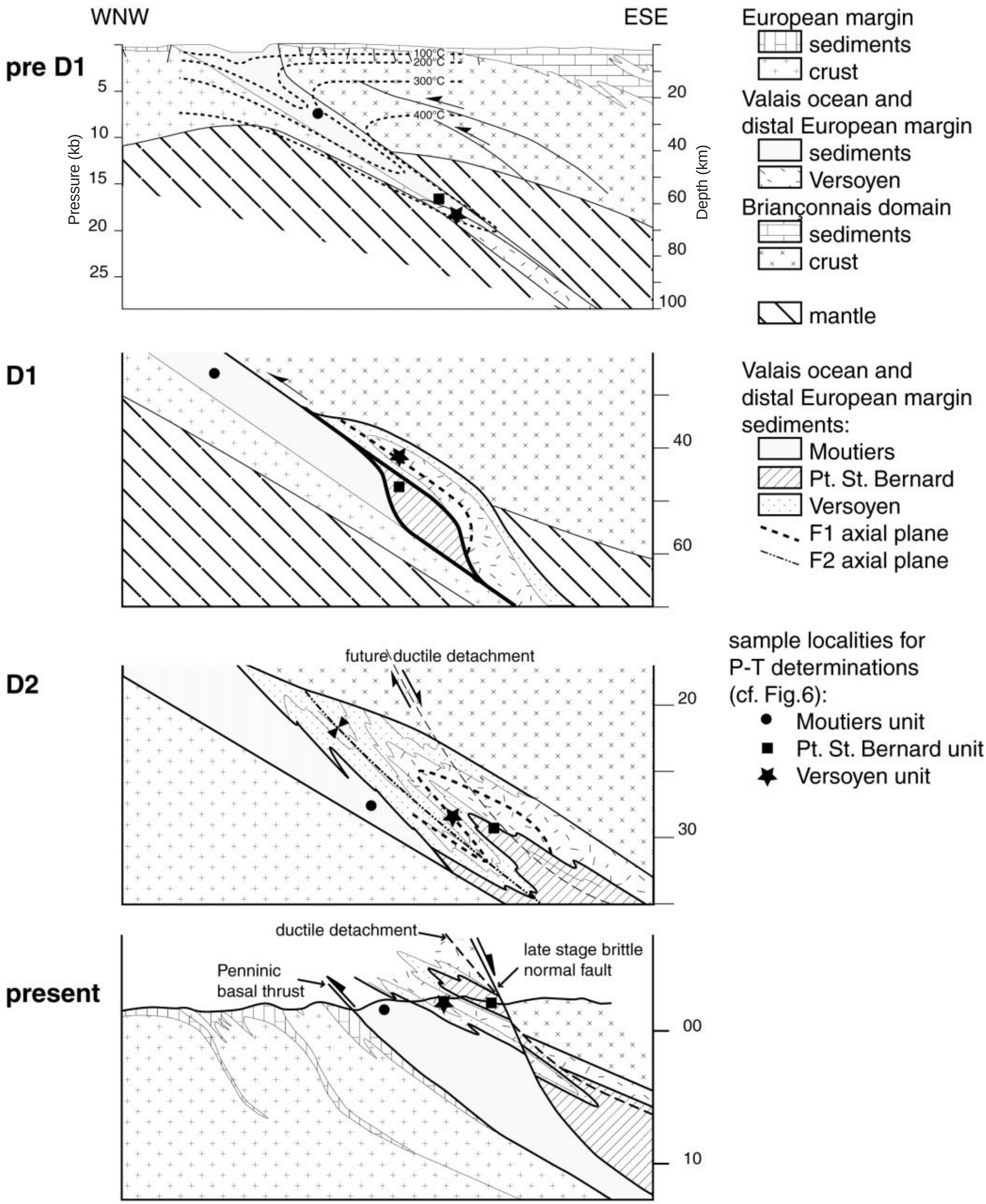


Fig. 7 Simplified sketch showing the structural evolution of the Valais units from the time of peak pressure conditions (pre-D1) to present. The P–T constraints are taken from Goffé and Bousquet (1997) and references therein (Fig. 6). Isotherms are drawn so as to fit estimated P–T conditions during the pre-D1 stage at peak pressure conditions, i.e., 17–18 kbar, 350 °C (Goffé and Bousquet 1997). Depth during D1 and D2 deformation is according to Fig. 6. The present-day situation is modified after Schmid and Kissling (submitted)

ment of the Roc de l'Enfer unit from the rest of the External Valaisan. Along this thrust L2 stretching lineations, defined by the principal elongation direction of the rock-forming minerals within pelitic levels of post-rift sediments, strikes parallel to the F2 fold axes. Shear sense indicators, although rarely observable, consistently indicate top-to-the-north-directed transport during D2. The geometry of the basal D2 thrust of the Roc de l'Enfer unit is illustrated in Fig. 4 (profile E). Based on the available data we infer that tectonic transport during D2 is also roughly top to the north, i.e., almost parallel to the present-day strike of the orogen. The simplified sketch given in Fig. 7 outlines the main features during the D1 and D2 phases.

The third phase of deformation (D3) also involves regional scale folding (Fig. 4). Dominantly NW-verging F3 folds display a strain-gradient within the investigated area. Away from the Penninic Front a spaced cleavage (S3) is observed, dipping steeply toward SE or, occasionally, toward NW. In the central region of profile D in Fig. 4 relatively open F3 folds exhibit a steeply SE-dipping axial plane, whereas further south (profile E in Fig. 4) box-like folds with SE- and NW-dipping axial planes are observed. Toward the Penninic Front, however, F3 folds gradually become tighter and perfectly isoclinal close to the Penninic basal thrust. They asymptotically merge into parallelism with this coevally active thrust, clearly indicating that here the Penninic Front is the map trace of a syn-D3 WNW-directed thrust (see section Penninic Front).

Herein the profiles are discussed in more detail.

Profiles A–D (Fig. 4) aim to document the structural evidence supporting the arguments which lead to the proposed distinction between External and Internal Valaisan. The southernmost outcrops of volcanic rocks together with their post-rift sediments (internal Valaisan) can be found along the ridge between Pte. de la Terrasse and Aig. de Praina (cf. Figs. 3, 4). From Aig. de Praina down section, i.e., normal to the F2 axial plane, one crosses volcanic rocks of the internal Valaisan. They form the core of a parasitic D2 east-facing synformal anticline, (folding an already inverted sequence). Hence, this part of the internal Valaisan had been inverted prior to D2 folding. Further down one crosses a 1000-m-thick sequence of intensely folded post-rift sediments to finally reach pre-rift sediments of the external Valaisan displaying an upward younging sequence (Fig. 4, profile D).

The rocks exhibiting the opposing younging directions described above, found on the same (i.e., inverted) limb of the major F2 synform (Fig. 4, profile D), are significantly different with respect to their stratigraphic record. The normal (upward younging) part at the base exhibits post-rift sediments deposited onto an eroded carbonate platform (external Valaisan). The inverted (downward younging) part has identical post-rift sediments in direct sedimentary contact with an oceanic substratum, i.e., pre-rift sediments containing basaltic rocks of the internal Valaisan or Versoyen ("complexe antéflysch" of Antoine 1971). Hence, along the same limb of a major F2 fold two distinct paleogeographic domains can be recognized, juxtaposed by D1 folding and thrusting. It becomes evident that parts of the internal Valaisan have been overturned and thrusted onto the external Valaisan during D1 folding and thrusting, associated with fold nappe formation above a D1 thrust (Figs. 3, 4) running within the Couches de St. Christophe (i.e., the youngest formation of the post-rift sediments; see Fig. 7). North-directed thrusting immediately followed D1 folding since this thrust cuts the F1 axial plane (D1 reconstruction in Fig. 7) but is clearly folded by F2 (Fig. 4).

These observations highlight the fact that hitherto undectected D1 folding and thrusting juxtaposed oceanic over continental realms of the Valaisan. They are crucial for the paleogeographic reconstruction depicted in Fig. 2, which shows that the Valaisan in Savoy comprises a former ocean–continent transition, situated at the external margin of the former Valais oceanic domain.

Further to the SW the internal Valaisan, and hence also the tectonic contact with the external Valaisan is no longer exposed and buried beneath the D2 thrust of the Roc de l'Enfer unit (Fig. 3). Yet, further to the NE, the contact between internal and external Valaisan is proposed to be represented by the tectonic contact between the Pt. St. Bernard and Versoyen units. Profiles through this area, depicted in Fig. 4 (profiles A–C) reveal the following:

1. A late stage syn-D2 thrust (Fig. 3) cuts previously formed F2 folds and reactivates former D1 thrusts.
2. F2 folds affecting the internal Valaisan exhibit WNW-facing antiformal anticlines and synformal synclines, respectively. This contrasts with observations made in the area between Pte de la Terrasse and Aig. de Praina (profile D) where anticlines are synformal and synclines antiformal. This observation demands an eastward closing F2 major syncline refolding an F1 axial trace.
3. Younging directions in the Pt. St. Bernard unit, found in the upper limb of this major F2 fold, imply that this unit was in an upright position before D2, as is the case for the External Valaisan depicted in profile D (Fig. 4).

The structural evolution of the Valaisan units is summarized in Fig. 7. Note that movements during stages pre-D1, D1, and D2 took place under sinistral compression, i.e., out of plane with respect to the sketches and toward north. From top (pre-D1) to bottom (present) the figures zoom in more and more and thus the scale gradually decreases (see y-axis for scale). Pressure and temperature constraints are taken from Goffé and Bousquet (1997). The frame regarding pre-D1 shows the proposed situation during peak pressure conditions together with the shape of the isotherms in a subduction scenario. The shape of these isotherms qualitatively corresponds to what is expected in a subduction scenario and drawn such as to satisfy the petrological constraints given by Goffé and Bousquet (1997) regarding peak pressure conditions (Fig. 6). There are no direct constraints on the age of peak pressure conditions. In analogy with the subduction of the North Penninic ocean in eastern Switzerland, we assume a Late Eocene age (Schmid et al. 1996). Also unknown is the relative timing of this event in the Pt. St. Bernard and the Versoyen units, respectively. We assume that both these units were subjected to peak pressure at about the same time which keeps the distance between the Moutiers unit and the Versoyen (and thus the width of the Valais domain) as small as possible. During D1 the Pt. St. Bernard unit individualized from the rest of the external Valaisan. Furthermore, the Versoyen unit, i.e., the internal Valaisan, was thrusted as a D1 fold nappe over the external Valaisan and the Pt. St. Bernard unit. As outlined previously and depicted in Fig. 4 (profile D), the D1 thrust beneath the inverted limb of the D1 fold nappe runs inside the Couches de St. Christophe. During D1 deformation pressure decreased (see Fig. 6) while the temperature increased: Material points crossed the strongly bent isotherms toward higher temperatures prevailing above the subduction channel. Hence, decompression to more moderate pressures during D1 took place in a compressive scenario. D2, which took place under greenschist facies conditions, led to the formation of a regional-scale synform–antiform pair, the synform of which is still preserved below the present day erosional surface. Two more steps follow and are not represented separately in Fig. 7:

1. Post D2 and pre D3 ductile normal faulting allows for the exhumation of the D2 structures in the Valais units relative to the Briançonnais. The proposed future position of this top SE detachment, situated beneath the Houiller Front, and mapped out NE of the Pt. S. Bernard pass, is indicated as a stippled line in the representation of D2.
2. Thrusting of the whole nappe stack onto the European foreland (Dauphinois) along the Penninic basal thrust and during D3. The representation of "the present" (redrawn from Schmid and Kissling, submitted) aims to illustrate the present-day situation and also depicts a late brittle normal fault revealed by fission-track dating and overprinting the former ductile detachment mentioned previously (see Fission-track dating). Removal of the effects of normal faulting reveals that the internal Valaisan has to be rooted between the Pt. St. Bernard unit and the Zone Houillère.

Profile E (Fig. 4) illustrates mainly the individualization of the Roc de l'Enfer unit from the rest of the External Valaisan during D2. The Roc de L'Enfer unit (Fig. 3) has already been investigated both in terms of its stratigraphic record (Barbier 1948; Antoine 1971) as well as its structural evolution (Fudral 1980, 1996; Antoine et al. 1993). It extends from Moûtiers to Bourg St. Maurice, forming a thin sliver of mainly Carboniferous schists and conglomerates following the Houiller Front. Within the area studied this subunit occasionally also contains small relics of Mesozoic pre-rift sediments directly overlain by post-rift sediments above an angular unconformity (see Fig. 2). The northernmost outcrops can be found along the ridge of the Pte. de la Terrasse, forming the peak itself (Figs. 3, 4; Fudral 1980). At this locality the Roc de l'Enfer unit is mostly made up of Carboniferous sandstone but also contains small occurences of Cretaceous post-rift sediments which unconformably and stratigraphically overlie the Carboniferous and form the core of a D3 synform (Fig. 4, profile D). The basal D2 thrust of this klippe is folded by D3. A comparable situation, depicted in Fig. 4, profile E, can be found further to the SW. There, however, the post-rift sediments rest on Lower Triassic quartzites. The internal deformation of the Roc de L'Enfer unit is characterized by the same three deformation phases described for the rest of the Valaisan. Whereas D3 structures deform the D2 basal thrust of the Roc de l'Enfer unit, this thrust cuts D2 fold axial traces in its footwall (e.g., near Fort de la Platte; Fig. 4, profile E). Shear sense criteria observed along this D2 thrust unequivocally indicate top-to-NNE-directed transport. Since this thrust is folded by D3 and displays a roughly north-directed transport direction, we interpret this thrust as a D2 structure, active during the last stages of this deformation phase. Profile E in Fig. 4, in combination with Fig. 3, clearly shows that this late-stage D2 thrust is responsible for the disappearance of the Versoyen unit toward SW, where the Versoyen is buried underneath the Roc de l'Enfer unit.

Structural evolution of the Subbriançonnais domain

Detailed structural mapping in the area of Grande Moenda and Perron des Encombres also revealed three phases of deformation associated with folding. The first phase (D1) is characterized by isoclinal folds (F1) of kilometric scale. A penetrative axial planar cleavage (S1), subparallel to bedding, is associated with the development of these folds. F1 axial planes and fold axes roughly strike NNW–SSE, displaying a big scatter due to refolding by D2 and D3. Since F1 fold hinges can hardly ever be directly observed, their presence is inferred from repetitions of stratigraphic formations and opposing younging directions. Previously, repetitions in the stratigraphic record were generally interpreted in terms of sedimentary facies changes (e.g., Barbier 1948; Perez-Postigo 1988). Thrusting of the Perron des Encombres subunit onto the Gde. Moenda subunit is also attributed to D1 since this thrust is folded by D2. Yet, the contact between the two subunits has been partly overprinted by late normal faulting, obliterating its previous history. L1 lineations (Fig. 8a) predominantly strike north–south with a great variation in plunge due to subsequent refolding; hence, no direct inferences on the transport direction during D1 can be made.

The second phase (D2) involves large-scale folding (F2). Major tight-to-isoclinal F2 folds have typical wavelenghts of the order of 200–300 m. F2 folds have steeply east- to ESE-dipping axial planes and south-plunging axes (Fig. 8a). A penetrative axial-planar cleavage (S2) is present throughout. Especially the incompetent lithologies, such as the Cancellophycus formation and the Oxfordian schists, are intensely folded by spectacular parasitic F2 folds on all scales. Elongated clasts, stretched belemnites, as well as boudinaged competent layers allow determination of a NNW/SSE-oriented direction of maximum extension (L2), oriented parallel to the F2 fold axes. Top NNW D2 shear sense indicators are found only near the Houiller Front and are discussed later.

In contrast to the previous phases of deformation, D3 is much less penetrative. It causes a spaced cleavage associated with fairly open, occasionally chevron-type folds that die out rapidly along their axial planes. In the Grande Moëndaz area (i.e., close to the Penninic Front; Fig. 8a) F3 folds exhibit fold axes plunging toward the south to SE with their axial planes dipping 30–40° toward the SE. Further to the east, on the other hand, near the Houiller Front, the same chevron-type folds have axial planes dipping toward the SW (axes plunging toward SE). Since no overprinting relations have been observed between F3 fore- and backfolds, they are both attributed to D3.

In summary, the structures observed within the Subbriançonnais units exhibit strong similarities with those described for the Valaisan domain further north. In analogy the D1 and D2 deformation phases are assumed to have formed during top north- to NNW nappe stacking. However, the metamorphic evolution before D1 (no high-pressure overprint) and during D1 (no evidence for temperatures >300 °C) was different from that of the Valais units.

Penninic Front

As previously mentioned, the tectonic significance of the Penninic Front (thrust vs normal fault) is still a matter of debate. This is the main reason for using the term “front” which simply denotes the map trace of a complex fault zone. Our observations reveal significant differences along the Penninic Front both in terms of kinematics as well as concerning the temperature conditions during deformation.

In the north, where the Penninic Front separates the Dauphinois from the Valais paleogeographic domain, this fault zone is characterized by a zone of SL tectonites several tens to 100 m wide. From SW to NE, the NE/SW-striking foliation steepens continuously from values of 30° in the Roseland pass area into a subvertical position at Pyramides Calcaires (Fig. 1). Thus, the orientation of the stretching lineation changes from 35° toward 147° (SW) to values of 74° toward 085° (NE). This steepening is regarded as a late feature in response to the exhumation of the external massifs, since the Penninic Front is steepest close to the culmination of the Mont Blanc massif. In the axial depression between the Mont Blanc and Belledonne massifs (Roseland area), the moderate dip of the Penninic Front fault zone has been determined from the ECORS-CROP seismic profile which imaged this fault zone down to great depth (Nicolas et al. 1990). Hence, the 30° dip is inferred to approximate the original orientation of a thrust which was active during late stages of D3, since D3 displays increasing strain toward the Penninic Front. Shear sense indicators such as shear bands and asymmetric clasts consistently indicate top-to-the-WNW-directed thrusting. The temperature during deformation was sufficient for ductile deformation of calcite and led to brittle–ductile transitional behavior in quartz. In summary, the Penninic Front in the area between the Roselend pass and the Pyramides Calcaires represents the map trace of the syn-D3 basal thrust of the Penninic units which took place under lowermost greenschist facies conditions.

Further to the south the Penninic Front separates Ultradauphinois from Subbriançonnais units, crosscuts F3 folds, and is typically marked by the presence of a thick layer of anhydrite, transformed to gypsum near the surface. This evaporite layer exhibits an intense, steeply (54°) east-dipping foliation and a stretching lineation dipping with 52° toward azimuth 110°. Close to the evaporite the limestones of the Subbriançonnais unit show brittle overprint. East-side down displacement can be deduced on faults and slickensides dipping 70° toward 112°. All these observations indicate post-D3 normal faulting across the Penninic Front fault zone under sub-greenschist facies conditions, overprinting a former D3-thrust.

Houiller Front

The Houiller Front fault zone separates Alpine low-grade metamorphosed rocks of the Zone Houillère from (a) Valaisan units subjected to high-pressure/low-temperature metamorphism in the north, and (b) low-grade metamorphosed Subbriançonnais units in the south. In addition, the kinematics of movement also change from north to south.

In the north, from Bourg St. Maurice across the French–Italian border until La Thuile, the Houiller Front is outlined by a several-meters-thick layer of upper Triassic evaporites (Gypsum and Cargneules) dipping steeply (80°) toward the SE and crosscutting all previous structures. Although no direct evidence for relative displacement has been found thus far within these evaporites, the steepening of all the structures within the footwall (i.e., Valaisan units) toward the Houiller Front is compatible with SE-directed normal faulting. This effect can best be seen in the “Torrent de Reclus” NE of Bourg St. Maurice. Late normal faulting taking place under brittle conditions is supported by

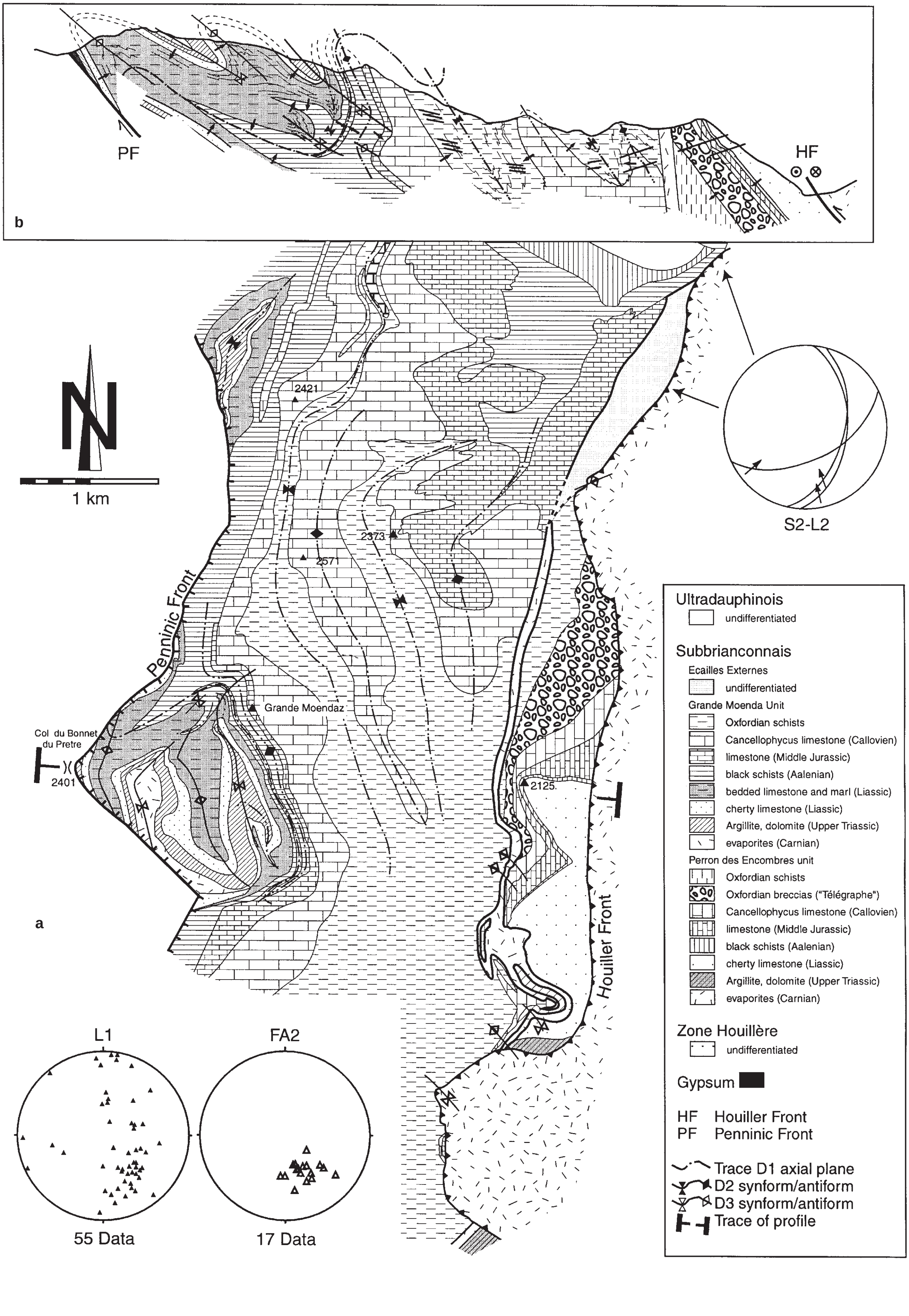

b
PF
HF
a
1 km
2421
2373
2571
Penninic Front
Grande Moendaz
Col du Bonnet du Pretre
2401
2125
Houiller Front
S2-L2
L1
55 Data
FA2
17 Data
Ultradauphinois
undifferentiated
Subbrianconnais
Ecailles Externes
undifferentiated
Grande Moenda Unit
Oxfordian schists
Cancellophycus limestone (Callovien)
limestone (Middle Jurassic)
black schists (Aalenian)
bedded limestone and marl (Liassic)
cherty limestone (Liassic)
Argillite, dolomite (Upper Triassic)
evaporites (Carnian)
Perron des Encombres unit
Oxfordian schists
Oxfordian breccias ("Télégraphe")
Cancellophycus limestone (Callovien)
limestone (Middle Jurassic)
black schists (Aalenian)
cherty limestone (Liassic)
Argillite, dolomite (Upper Triassic)
evaporites (Carnian)
Zone Houillère
undifferentiated
Gypsum
HF Houiller Front
PF Penninic Front
Trace D1 axial plane
D2 synform/antiform
D3 synform/antiform
Trace of profile

Fig. 8 Geological **a** map and **b** profile of the Subbriançonnais in the area of Grande Moendaz. Also shown are stereoplots (equal area, lower hemisphere) of L1 stretching lineations and F2 fold axes. *Stereoplot in the upper right corner* gives the orientation of the D2 mylonitic foliation and the D2 stretching lineation (*arrows*) along the Houiller Front, refolded by D3

fission-track data discussed later. Northeast of the Pt. St. Bernard pass this late brittle normal fault is seen to overprint a ductile post-D2 top SE extensional detachment mentioned previously, also situated at the base of the zone Houillère.

South of Moûtiers the thickness of the upper Triassic evaporites outlining the external limit of the Zone Houillère increases significantly. Near St. Martin de Belleville in the "Torrent des Encombres" (Fig. 8) the massive anhydrite layer altered to gypsum displays an eastward (45°) dipping mylonitic foliation together with an approximately north/south-trending stretching lineation. The mylonites postdate D1, cutting the tectonic contact between the Perron des Encombres and the Grande Moenda units. However they are refolded by open F3 folds implying an activity of the Houiller Front in that area during D2. Decimeter- to meter-size clasts of dolomite allow establishment of top-to-the-north-directed transport of the Zone Houillère (Fig. 9). This implies sinistral transpression across the syn-D2 Houiller Front fault zone in this southern area.

Fission-track dating

This chapter discusses the first results of still ongoing thermochronological work in light of previously published fission-track data (Seward and Mancktelow 1994). The aim is to extend the work of Seward and Mancktelow (1994) toward the SW (Maurienne Valley)

Fig. 9 σ-clast of dolomite in an anhydrite matrix indicates sinistral (i.e., top-to-the north) movement along the Houiller Front during D2. Locality: Torrent des Encombres

as well as toward the SE into the Zone Houillère in order to obtain additional constraints on the thermal and structural evolution. Sample localities, together with zircon and apatite fission-track ages, are shown in Fig. 10. Mineral separation, mounting, polishing, etching, and counting was carried out using standard techniques as described by Seward (1989). All samples were treated using the external detector method with a zeta value (Hurford and Green 1983) of 348 (FCT, SRM 612) for zircon and 357 (Dur, CN5) for apatite.

The 40 new age data obtained so far (30 samples analyzed yielded 21 apatite and 19 zircon ages) allow distinction between two areas. A more external area is characterized by apatite ages ≤ 5 Ma and zircon ages ≤ 17 Ma. A second and more internal area revealed apatite ages ≥ 9 Ma and zircon ages ≥ 20 Ma. With the exception of one sample (i.e., the topographically lowest sample yielding a zircon age of 23 Ma), temperatures for all other samples to the south of Moûtiers have not been sufficient for fully resetting the zircons. These samples are supposed to yield detrital zircon ages, rather than cooling ages of samples subjected to temperatures in excess of 300 °C. This is also suggested by the strong internal variation of the single grain ages which fail the chi-square test. Inspection of Fig. 2 from Seward and Mancktelow (1994), containing their data together with data from Soom (1990), reveals that the same age pattern is also observed further towards the NNE in the area of Visp in Switzerland.

Within the studied area (Fig. 10) the boundary separating these two age domains strikes approximately 050°, making an angle of $\pm 30°$ with the general strike of the units (020). In the north it coincides with the Houiller Front (Pt. St. Bernard pass area, where normal faulting was postulated across the Houiller Front). Southwest of Bourg St. Maurice it first follows the Isère valley and coincides with the contact between the Valais and Subbriançonnais units NE of Moûtiers. South of Moûtiers it presumably follows the contact between the Subbriançonnais and Ultradauphinois units. Finally, in the Maurienne Valley, a similar offset in apatite fission-track ages is observed between the Ecailles externes and the Perron des Encombres subunits, i.e., within Subbriançonnais units.

A plot of fission-track data projected onto a NW/SE-trending profile approximately parallel to the ECORS-CROP seismic line (Fig. 11) reveals that ages get increasingly older towards the Houiller Front where a jump of the order of 7 Ma occurs. Furthermore, it can be seen that the cooling rates as determined from the difference between zircon and apatite fission-track ages are roughly the same across the profile. This excludes the possibility that the jump in ages has been caused by differential cooling (exhumation) of the different units during cooling below the annealing temperatures for zircon (240 ± 60 °C; Yamada et al. 1995) and apatite (90 ± 30 °C; Green et al. 1989). The age offset is best explained in terms of very late normal faulting with the internal (or southeastern) part forming the hanging-

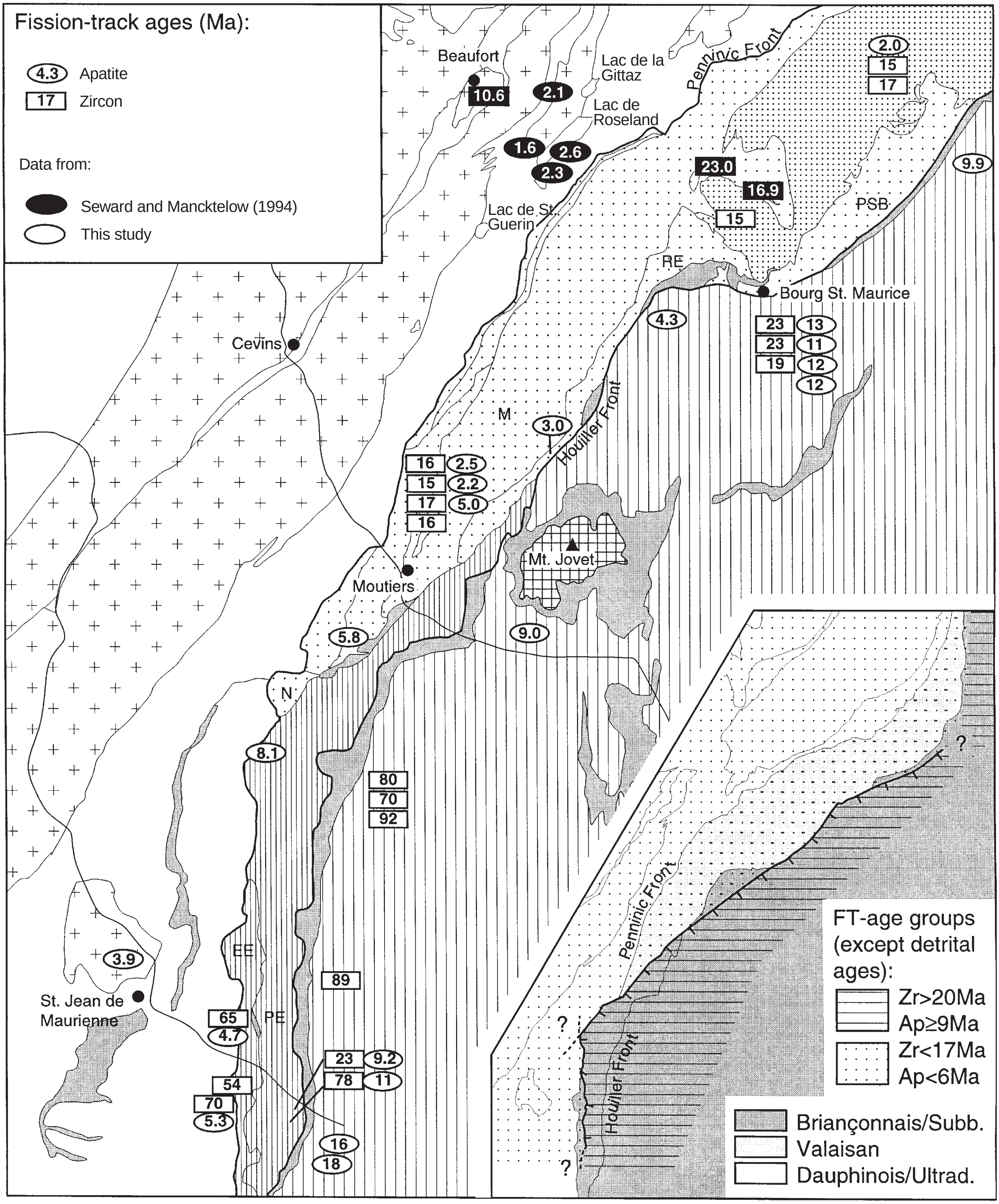


Fig. 10 Regional distribution of zircon and apatite fission-track ages from this study, together with data from Seward and Mancktelow (1994). *Inset* illustrates the map trace of late-stage normal faults, as deduced from the fission-track data

wall. The oldest apatite fission-track ages from the footwall (± 5 Ma) thus give a maximum age for the activity of this normal fault. Vertical offset across this normal fault can roughly be estimated to be of the order of 3–4 km, depending on the assumed geothermal

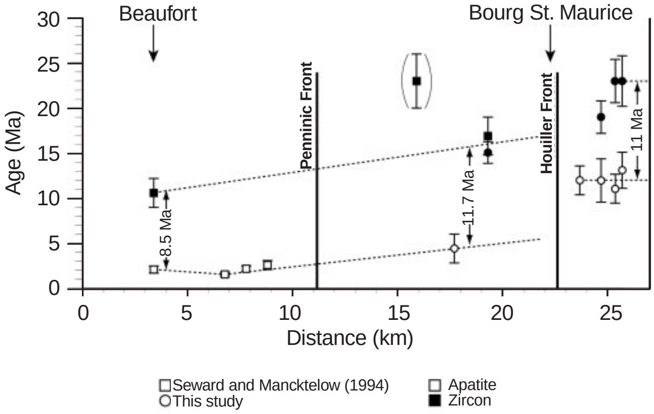


Fig. 11 Apatite and zircon fission-track ages along a WNW/ESE-trending profile through the Penninic Front and Houiller Front, respectively. The profile runs approximately parallel to the ECORS-CROP seismic line between Beaufort and Bourg St. Maurice (see Fig. 10). Note that the age difference between zircon and apatite is almost the same for the Valaisan domain (11.7 Ma) and the Zone Houillère (11 Ma), suggesting that normal faulting post-dates the cooling history as determined for the Valais units and the Zone Houillère

gradient. It is interesting to note that the very young age of normal faulting at the Houiller Front is in agreement with the present-day stress field deduced for the internal zones of the Western Alps, indicating orogen-perpendicular horizontal extension within the internal zones of the Western Alps (Maurer et al. 1997).

Discussion

The wedging out of the Valaisan paleogeographic units in map view is likely to be closely related to the fact that only those tectonic units which are derived from the Valais domain have been subjected to high-pressure metamorphism. In addition to the difference in metamorphism between the Valais and Subbriançonnais units, the Valaisan units themselves display a metamorphic gradient, as recently demonstrated by Goffé and Bousquet (1997). According to these authors the external Valais unit was never subjected to pressures greater than 8 kbar, whereas for the internal Valais unit maximum pressures were around 18 kbar. The need for an important normal fault separating the Versoyen from the Pt. St. Bernard unit, as proposed by Cannic et al. (1995, 1996), became invalid after relics of high-pressure metamorphism were also found within the Pt. St. Bernard unit (Goffé and Bousquet 1997), which is part of the Valais paleogeographic domain according to our findings. Thus, the internal boundary of the exhumed high-pressure rocks has been shifted further towards the east, namely to the Houiller Front which marks the boundary between the various Valais units and the low-grade metamorphosed rocks of the Zone Houillère, situated in the hangingwall of the North Penninic subduction zone. Paleogeographically, the Zone Houillère occupied a position similar to that of the Subbriançonnais south of Moûtiers. This suggests that the Briançonnais s.l. formed the upper plate with respect to a North Penninic subduction zone within the Valais units, a subduction zone which is extremely unlikely to have laterally ended south of Moûtiers.

We discuss a model for the structural and metamorphic evolution of the studied area, dealing with the Valaisan and Subbriançonnais paleogeographic domains, respectively. This model attempts to integrate the presented structural and fission-track data as well as the published petrological data (Schürch 1987; Cannic et al. 1996; Goffé and Bousquet 1997).

The suggested relative positions of the different units before subduction/collision occurred is shown in the palinspastic reconstruction in Fig. 2. The units are, from external to internal: (a) the external Valaisan, forming the European margin s.s. and comprising the Moûtiers and Pt. St. Bernard units; (b) the internal Valaisan, representing the relics of the Valais ocean, identical to the Versoyen unit s.str. of the French authors; and (c) the Subbriançonnais which represents the external margin of the Briançonnais microplate (comprising parts of the Zone Houillère and the Subbriançonnais cover nappes south of Moûtiers).

Since it is likely that the age of the high-pressure overprint is the same for both the Pt. St. Bernard unit and the internal Valaisan, they are supposed to lie close to each other during peak pressure conditions in a subduction scenario. Whereas these two units have been subducted to pressures as high as 18 kbar, the Moûtiers unit was only subducted to depths corresponding to 8 kbar (Goffé and Bousquet 1997). During our deformation stage D1, which postdates peak pressure conditions, the internal Valaisan, which must have been partially exhumed before D1, was thrust northwards onto the external Valaisan as a major recumbent fold nappe (Fig. 7). In the southern section (i.e., the area south of Moûtiers where the Valaisan units are missing), on the other hand, subduction of the Valaisan continued without exhumation while intense D1 and D2 folding and/or thrusting occurred in the upper plate (Subbriançonnais and Zone Houillère). A possible reason why obduction of the high-pressure rocks occurred only in the north is the angle between the subduction direction (roughly NNW–SSE) and the strike of the colliding chain. In the south, the north/south-trending units enclose a small angle with the assumed subduction direction giving rise to a large component of sinistral strike-slip movement. Toward the northeast the overall orientation of the mountain chain gradually changed and units strike NNE–SSW in the area of Bourg St. Maurice and ENE–WSW in Switzerland, thus forming a high angle with the inferred subduction direction.

During D2 refolding of the nappe stack gave rise to the formation of the kilometer-scale synform as observed in the profile through the northern area (Figs. 4, 7). The high-pressure rocks of the internal Valaisan thus form part of the core of this synform, whereas the Pt. St. Bernard nappe comes to lie on the upper limb. In the southern section refolding by D2 involved the previously formed tectonic contact between the Gde. Moenda and the Perron des Encombres subunits, respectively, and north-vergent thrusting of the Briançonnais continued.

During the third stage (D3) the tectonic scenario changed dramatically: north–south convergence, roughly parallel to the strike of the Western Alps, was now followed by WNW–ESE shortening perpendicular to the strike of the Western Alps. F3 folding is linked to WNW-vergent thrusting of the internal units onto the European foreland along the Penninic Front. The basal D3 thrust (Penninic frontal thrust) is still preserved and observable in the northern section in the Roseland pass area.

Finally, late (post 5 Ma) top-to-the-SE-directed normal faulting affected the entire area (Fig. 10). The normal fault is essentially parallel to the Houiller Front in the northern part of the working area (i.e., north of Moûtiers) and partly reactivated and overprinted the Penninic Front to the south. This late normal fault plays an important role with respect to the structural observations available along the Houiller Front and Penninic Front fault zones, respectively, and is partly responsible for the abrupt character of the SW termination of the Valaisan. However, it has to be clearly stated that the overall wedging out of the Valaisan has to do primarily with the D1 and D2 deformation phases, and particularly with WNW-vergent thrusting during D3. It is obvious that the relatively minor vertical offset of the order of 2–3 km is not at all sufficient to exhume the Versoyen and Pt. St. Bernard high-pressure rocks.

Thus, it is clear that we miss a major detachment allowing for the exhumation of the Valais units relative to the Zone Houillère which had to be active prior to D3. The need for such ductile detachment has already been discussed by Cannic et al. (1995, 1996). Our proposed location of this detachment is indicated in Fig. 7 (“future ductile detachment” in the D2 frame). The late brittle normal fault discussed previously plays a major role with respect to the detachment: The ductile detachment is presently hidden below the Zone Houillère in the hangingwall of this brittle normal fault, whereas it is above the present-day topography in its footwall. Therefore, the only chance to observe the detachment is in the area of the Pt. St. Bernard pass. In fact, greenschist facies mylonites indicating top-to-the-SE-directed normal faulting and cutting D2 structures have recently been found in this area (S. Bucher, pers. commun.). However, it has to be kept in mind that early exhumation of the high-pressure units of the Valais units to an intermediate depth around 30 km is related to D1 and D2 deformation which took place in a compressive scenario above the subduction channel (see Fig. 7), predating final exhumation by post-D2 extension.

Conclusion

Despite striking differences concerning metamorphic evolution, the present study reveals strong similarities concerning the internal deformation within the Subbriançonnais and the Valais units, respectively. Both these units are characterized by three phases of deformation. The first two phases are presently characterized by tight to isoclinal folds. Fold axes and stretching lineations strike north–south and shear sense criteria indicate top-to-the-north-directed transport during D1 and D2. The third phase led to open folding during WNW-directed thrusting. This can be inferred from the increasing strain toward the Penninic Front in the Valais domain and the coeval activity of D3 folding and WNW-directed thrusting along the Penninic Front in the Cormet de Roseland area.

North-directed transport along the segment of the Western Alps considered in this study implies sinistral transpression on the scale of the entire Alpine chain, as postulated by Ricou and Siddans (1986). During D1 and D2 the Western Alps were situated at the western edge of the Adriatic promontory which moved north relative to stable Europe. The changeover to D3 deformation postdates final collision in the Eocene and is probably linked to the westward movement of the Adriatic promontory, linked to the activity along the dextral Insubric line during the Oligocene (Steck 1990; Schmid and Kissling, submitted). In conclusion, our data definitely favor a tectonic origin for the disappearance of the Valais units south of Moûtiers, suggesting that the Valais paleogeographic domain formerly extended all along the arc of the Western Alps.

Several kinematic steps during a complex and truly three-dimensional structural evolution contributed to this wedging out. Late-stage D2 top-to-the-north thrusting is held responsible for the burial of the Versoyen unit beneath the external Valaisan (Roc de l’Enfer unit), and hence its wedging out in map view near Bourg St. Maurice (Fig. 3). The final wedging out of the rest of the Valaisan SE of Moûtiers is due partly to tectonic omission related to post-D3 normal faulting, the locus of which changes from a more internal position in the north, i.e., at the Houiller Front, to a more external position in the south, i.e., at the Penninic Front (inset of Fig. 10). It is probably due partly to previous top-to-the-north transport of the Subbriançonnais units and the zone Houillère over the Valaisan units during D1 and D2 in a scenario of sinistral transpression. However, due to the later overprint by normal faulting, there is no direct evidence for such top-to-the-north transport in the area around Moûtiers. Consequently, it is hard to decide which of these two factors played a predominant role for the final wedging out of the Valaisan units. However, we firmly postulate that the

Valaisan units are presently buried underneath the Subbriançonnais units and the Zone Houillère south of Moûtiers.

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