A combined structural and sedimentological approach to decipher the evolution of the Valais domain in Savoy, (Western Alps)

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Andrea Loprieno
aus Pisa (Italien)

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PREFACE

The aim of this fieldbased thesis is to provide a dataset and geological evidences in order to constrain the tectonic evolution of the Valais domain in the Western Alps and to investigate its pre-deformational paleogeography. In the frame of this study several geological problems, ranging in time from the Late Jurassic-Early Cretaceous paleogeographical reconstruction of the Valaisan domain to sub-recent normal faulting are addressed. Generally speaking the rocks of the Valais domain bear information on all these problems and represent a key area for their better understanding. Deatiled mapping formed the basis of this work, aided by techniques of structural geology and basin analysis. Geological cross sections on all scales have been constructed covering the entire study area. Thus field work, interpreted in the light of up-to-date literature formed the essential tool of this work. Apart from the classical investigations carried out in the Seventies, providing a firm basis as well as an inexhaustible source of inspiration, a neat structural framework allowing for a paleogeographic reconstruction (and *vice versa*) was not available. Therefore, a solid structural database forms the main goal of this thesis with interpretations and reconstructions following there from. However the rocks of the Valaisan still bear secrets for further investigations which, most probably will have an impact on the interpretative part of this work, with the structural data presented here still surviving.

The introduction to the study area (Chapter 1) is presented in the form of a study (Fügenschuh et al. 1999) on regional geology, a publication written together with Bernard Fügenschuh, Stefano Ceriani and Stefan Schmid, who were actively involved in the Swiss National Science Foundation project termed: "The SW termination of the Valais zone and the Pennine frontal thrust near Moûtiers (Tarentaise valley, France): a combined structural and sedimentological study" carried out at the University of Basel. This thesis is a part of this project. The paper clearly focuses on the main issues of Valais domain and adjacent tectonic units (e.g. the Subbriançonnais domain) in the Penninic zone from the Pt. St. Bernard pass in the NE (Italian-French border) to the Arc Valley in the SW. It is clear that, yet, that this publication presents some results and ideas which were achieved during an early stage of the research.

Since it is rather unusual to present results in an introduction before having presented and properly analysed the data, I will briefly comment on that. The reason is simply to give an overview of the complexities of the evolution of the Valais domain, which will be extensively presented later on in the thesis.

The following two chapters deal with sedimentological observations made in the Valaisan units. Chapter 2 describes the stratigraphy of the Valaisan units introducing the new paleogeographical arrangement characterized by the subdivision into an external (continental) and an internal (oceanic) Valaisan realm. Chapter 3 is a study dealing with the syn-rift sediments. This chapter aims to test the possibility whether the external Valaisan domain (continental) bears information on two distinct rifting events. Chapter 4 is dedicated to the metamorphism with particular attention to the correlation between metamorphic stages and structural elements. This chapter introduces the reader to the structural analysis of the Valaisan units which is then furtherly investigated in the last two chapters. This part represents, probably, the most complex one of this thesis and therefore has been split into two chapters. Chapter 5 deals with the regional geometry of the nappe stack of the Valaisan units. The geometry of the main structural features associated with D1, D2 and D4 deformation phases are introduced and discussed. Finally, Chapter 6 focuses on the analysis of the structural elements related to the observed deformation phases (D1 to D6). Chapter 7 contains a summary of the main results of this study.

Abstract

The present thesis deals with the structural evolution and the paleogeography of the Valaisan units in the Penninic zone of the Western Alps (Savoy, France). A combined structural-sedimentological approach allowed to reconstruct the Late Jurassic-Early Cretaceous paleogeography and to unravel a Tertiary polyphase Alpine deformation history related to subduction and obduction.

During the late Jurassic-Early Cretaceous, rifting activity within the Subbriançonnais domain led to the individualization of an oceanic basin (the Valais oceanic domain). Yet sedimentological observations on the pre- and syn-rifting sequences infer that the Subbriançonnais domain was already affected by an earlier phase of rifting, i.e. the formation of the more internal Piemont-Ligurian oceanic domain. The younger Valais domain can be subdivided into the following paleogeographical domains: (1) The external Valaisan (Quermoz, Moûtiers, Roc de l'Enfer and Pt. St. Bernard units) and (2) the internal Valaisan (Versoyen unit). The external Valaisan is characterised by Paleozoic (Carboniferous) to Mesozoic sediments deposited onto thinned continental crust, while the internal Valaisan represents the oceanic parts of this basin. Early Cretaceous to late Cretaceous (Tertiary ?) post-rift sediments deposited onto both the external (continental) and internal Valaisan (oceanic) post-date the formation of this ocean and thus imply Earlier Cretaceous (pre-Barrêmian) opening of the Valais ocean.

During the Alpine orogeny, the Valais domain, underwent a complex evolution involving subduction and exhumation. High pressure low temperature metamorphic overprint is recorded in the internal Valaisan (Versoyen unit) and in the internal part of the external Valaisan (Pt. St. Bernard unit). The rest of the external Valaisan (Quermoz, Moûtiers and the Roc de l'Enfer units) was subjected to greenschist metamorphic condition only.

Detailed mapping revealed four main phases of deformation involving folding and thrusting (pre-D1, D1, D2 and D4) and three phases of deformation of more local impact (D3, D5 and D6). During the oldest observable deformation phase (pre-D1) thrusting affected the Versoyen and the Pt. St. Bernard units which were detached from their

substratum (oceanic and continental respectively). Microstructural observations related to this first stage of the deformation history indicates that the pre-D1 phase post-dates the HP-LT peak conditions of the Pt. St. Bernard and Versoyen units. This demonstrates that the subsequent structural evolution (D1 to D6) allows for the exhumation of the Valaisan units. Continuous thrusting and regional scale folding during D1 led to the formation of a mega-fold within the internal Valaisan (Versoyen unit) which was thrusted onto the Pt. St. Bernard and the Moûtiers units. A second phase of deformation is responsible for the formation of the regional scale Versoye synform. This structure exhibits the D1 nappe stack in its inverted limb (isoclinal refolding of the D1 nappe stack). In a late stage of the D2 deformation phase the Roc de l'Enfer unit was individualized from the Moûtiers unit along the Leisette thrust. D1 and D2 deformation phases, related to roughly N-S directed shortening, are responsible for the nappe stacking of the Valaisan units. They formed in response to sinistral transpression in a NNE/SSW-trendly corridor between the Penninic units and the European foreland. D4 oblique to the N-S trends of the D1 and D2 deformation phases, is related to a WNW-directed orogen-perpendicular shortening. During this phase the already formed Valaisan nappe stack was thrusted onto the European foreland, i.e. the Dauphinois domain along the Roselend thrust. In a late stage of the structural evolution the kinematic framework changes to an orogen-perpendicular extension.

The main conclusions of this thesis, as inferred from the combined sedimentological and structural approach can be summarized as follows:
The Valais domain marks the site of a second and a more external oceanic domain that extended all along the arc of the Western Alps. Its rather abrupt end in map view south of Moûtiers (France) is due to tectonic reasons. The Valais domain marks the site of a second and more external distinct subduction zone, separated from the internal high-pressure units of the western Alps by the Zone Houillère.

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CHAPTER 1: INTRO DUCTION

The investigated area is located SE of the Mont Blanc massif and extends from the Pt. St. Bernard pass in the NE (Italian-French border) to Bourg St. Maurice in the SW (Isere valley). This region covers approximately 150 square kilometres of mountainous terrain with altitudes varying from 500 meters to 3000 meters a.s.l. Exposure is mostly excellent. Some of the valleys are NW-SE oriented, i.e. perpendicular to the general strike of the main geological structures. These valleys (Valleé du Charbonnet and lower Valleé des Glaciers) are characterised by abrupt changes in relief and expose some spectacular (cliff) sections. However, due to their steepness large portions remains inaccessible. Smaller valleys follow geological features and are consequently NE-SW oriented (higher Valleé des Glaciers and Valleé du Versoyen). These are characterised by a more gentle topography and are of special interest regarding the three dimensionality of the Valaisan units.

For the geological setting of the study area the reader is referred to the publication included in this introduction, “Structural analysis of the Brian4onnais and Valais units in the area of Moûtiers (Savoy, western Alps): paleogeographic and tectonic consequences” (Fügenschuh et al. 1999). This paper represents a comparative study of structural evolution within the Valaisan and Subbriançonnais units. Amongst the different parts, the “Structural evolution of the Valais zone” and large parts of the “discussion” and “conclusions” are based on observations made within the area of this PhD study. Furthermore, observations related to the kinematics along the Houiller and the Penninic Front respectively, have been combined with observations from the Subbriançonnais zone further in the S. The remaining parts of Fügenschuh (et al. 1999) are based on the work of Stefano Ceriani for the “Structural evolution of the Subbriançonnais domain” and of Bernhard Fügenschuh for “Fission tracks analysis”. The “Structural analysis of the Subbriançonnais and Valais units in the area of Moûtiers (Savoy, Western Alps): paleogeographic and tectonic consequences” thus provides a useful introduction together with some first conclusions on the main problematics of the study project: “The SW termination of the Valais zone and the Penninic frontal thrust near Moûtiers (Tarentaise Valley, France): a combined structural and sedimentological study” of which this thesis is a part. However, it must be kept in mind that this publication summarised and elaborated first results of this project study.

The “Structural analysis of the Valaisan zone” exposed in Fügenschuh et al., has been refined and modified according to the new data and observations made during the following years. An exhaustive description of the structural elements (i.e. pre-D1 to D6 deformation phases) is given in

Ch. 6. It must be explicitly mentioned that that D3 of Fügenschuh et al. corresponds to D4 to this thesis. Although the main features and implications of geological profiles presented in Fügenschuh et al., are still valid they are partly modified and refined and are presented together with newly constructed cross sections in Ch. 5. Chapters 2 and 3 address both the stratigraphical and sedimentological aspects, a part, not discussed in Fügenschuh et. al. 1999.

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ORIGINAL PAPER

B. Fügenschuh · A. Loprieno · S. Ceriani
S. M. Schmid

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

B. Fügenschuh (✉) · A. Loprieno · S. Ceriani · S. M. Schmid
Geologisch-Paläontologisches Institut, Bernoullistrasse 32, CH 4056 Basel, Switzerland
e-mail: fuegenschuh@ubaclu.unibas.ch,
Tel.: +41-61-2673610,
Fax: +41-61-2673613

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

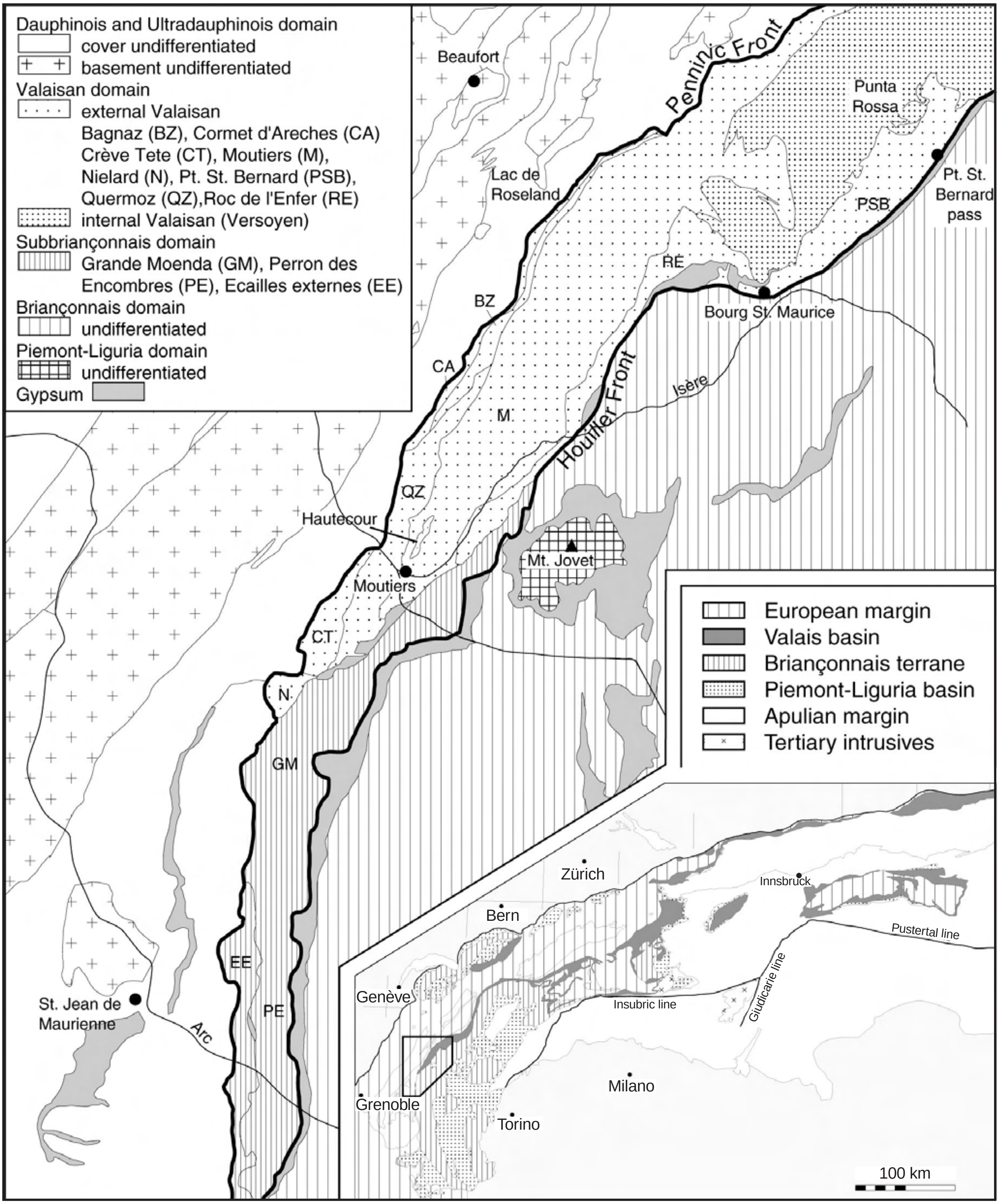


Fig. 1 Tectonic map of the area between Pt. St. Bernard pass (French–Italian border) and Arc Valley based mainly on the structural schemes of "Carte géologique de la France" feuille Bourg St. Maurice and feuille St. Jean de Maurienne. Inset shows distribution of the Valais paleogeographic domain redrawn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

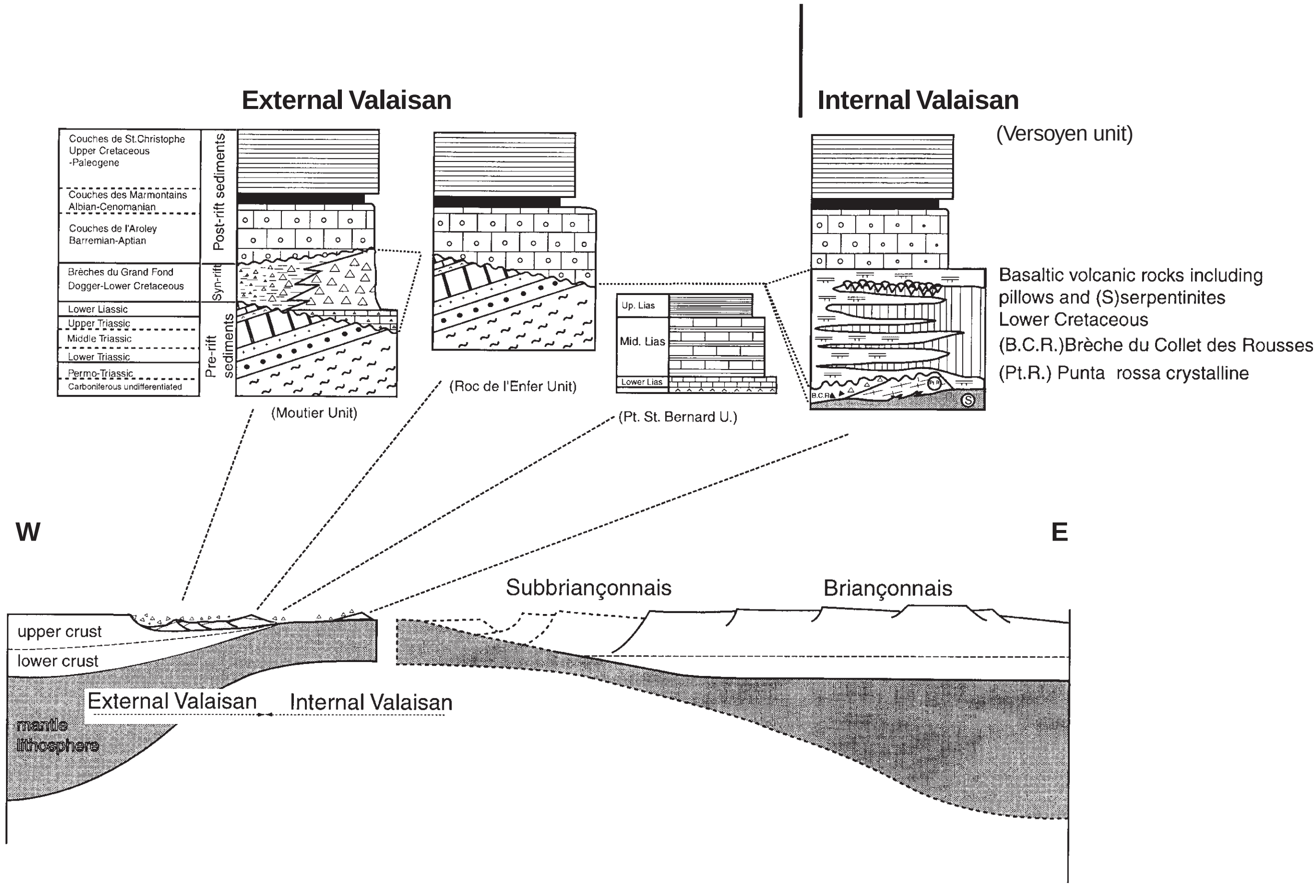


Fig. 2 Proposed palinspastic reconstruction of the different units based on our structural analysis, together with their stratigraphic record. (After Antoine et al. 1971)

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-

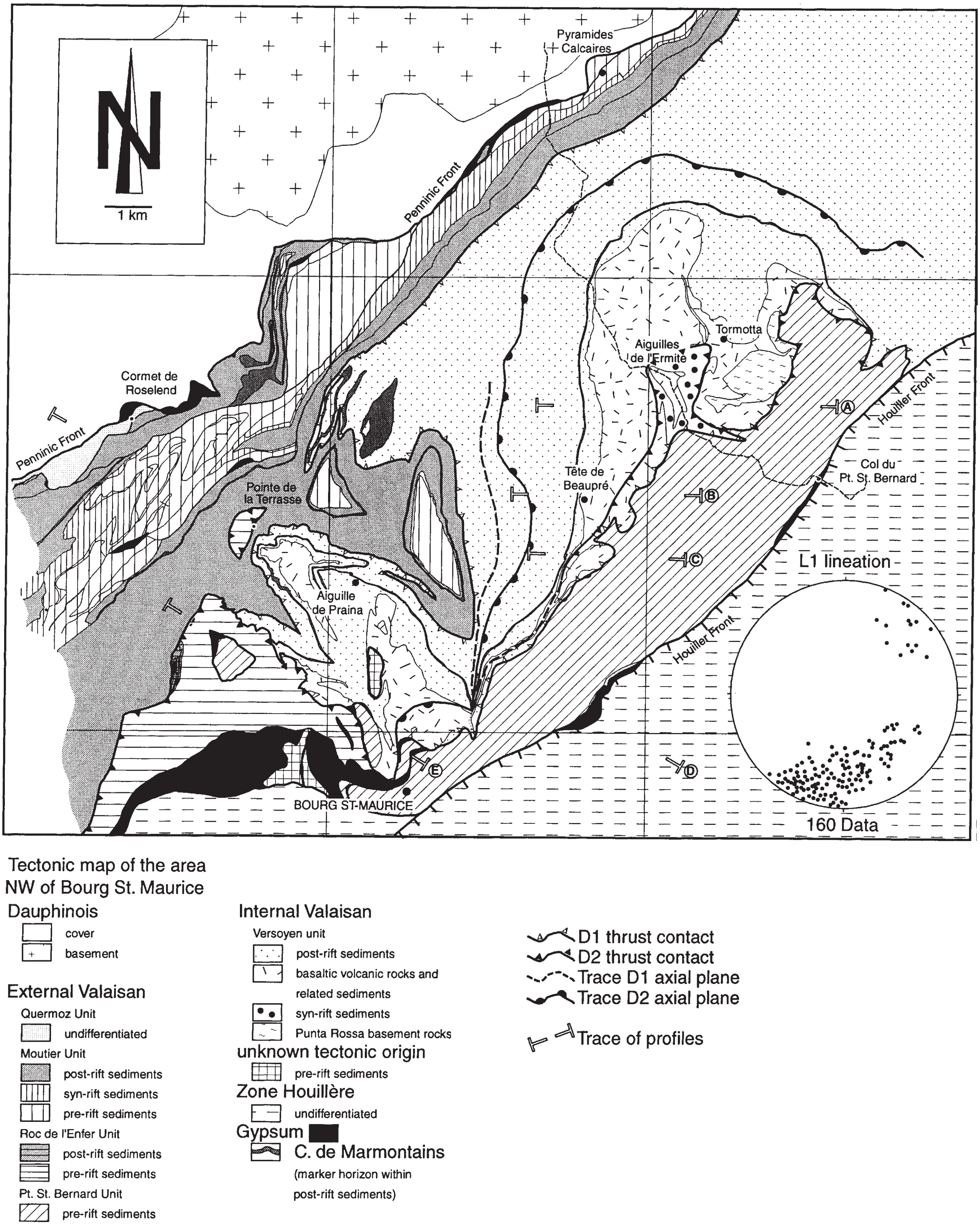


Fig. 3 Geological map of the area between the Penninic and Houiller Front, respectively, north of Bourg St. Maurice. Open triangles outline the pre-D2 thrust contact between internal and external Valaisan, *solid triangles* indicate syn- to post-D2 thrust contacts. *A–E* gives location of profiles as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

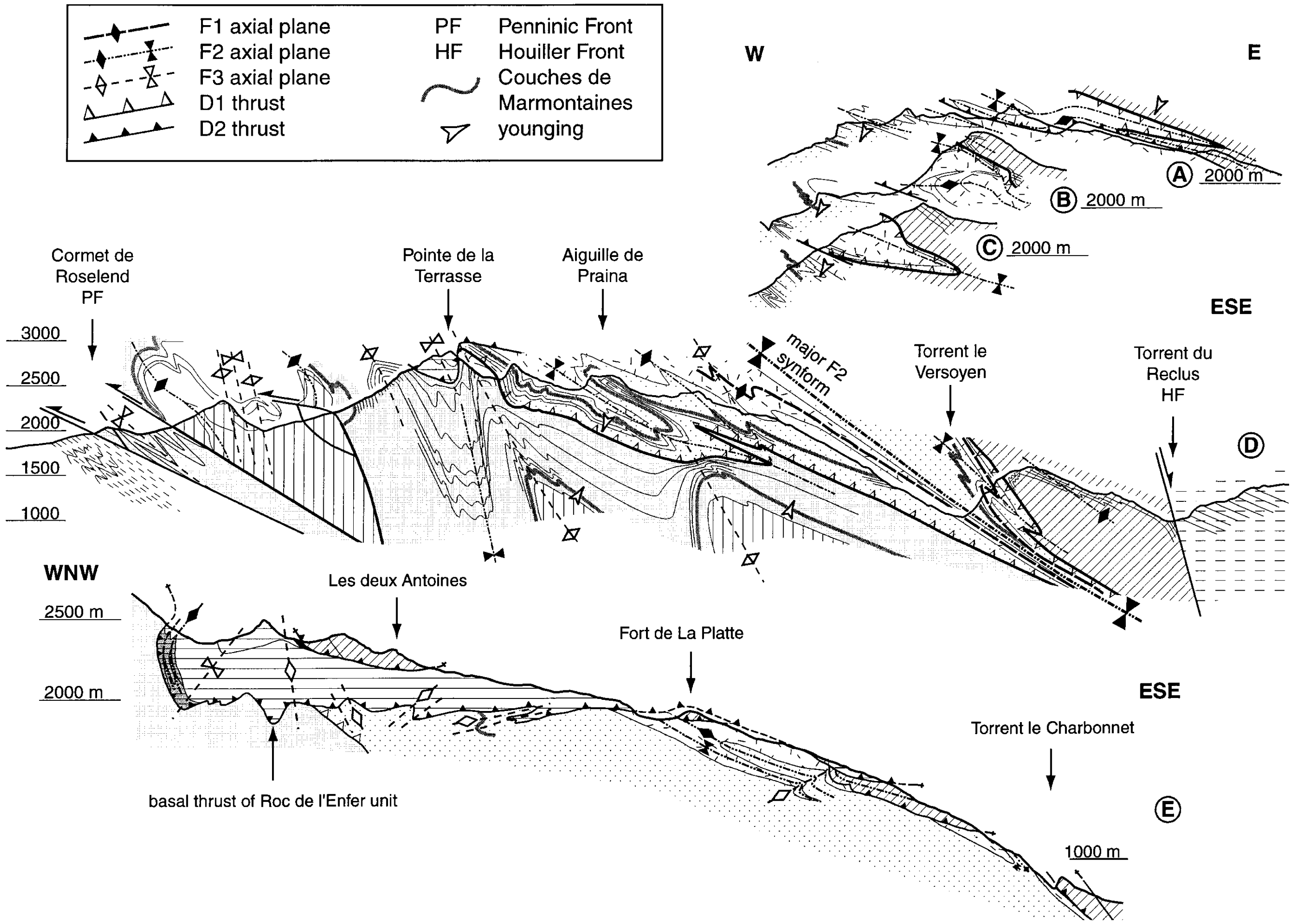


Fig. 4 Series of profiles (for location seeFig. 3) through the Valaisan; patterns are the same as in Fig. 3. No vertical exaggeration. Profiles *A–D* outline the relationship between the internal and external Valaisan. Note the increasing intensity of D3 folds towards the WNW which are cut by the Penninic 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

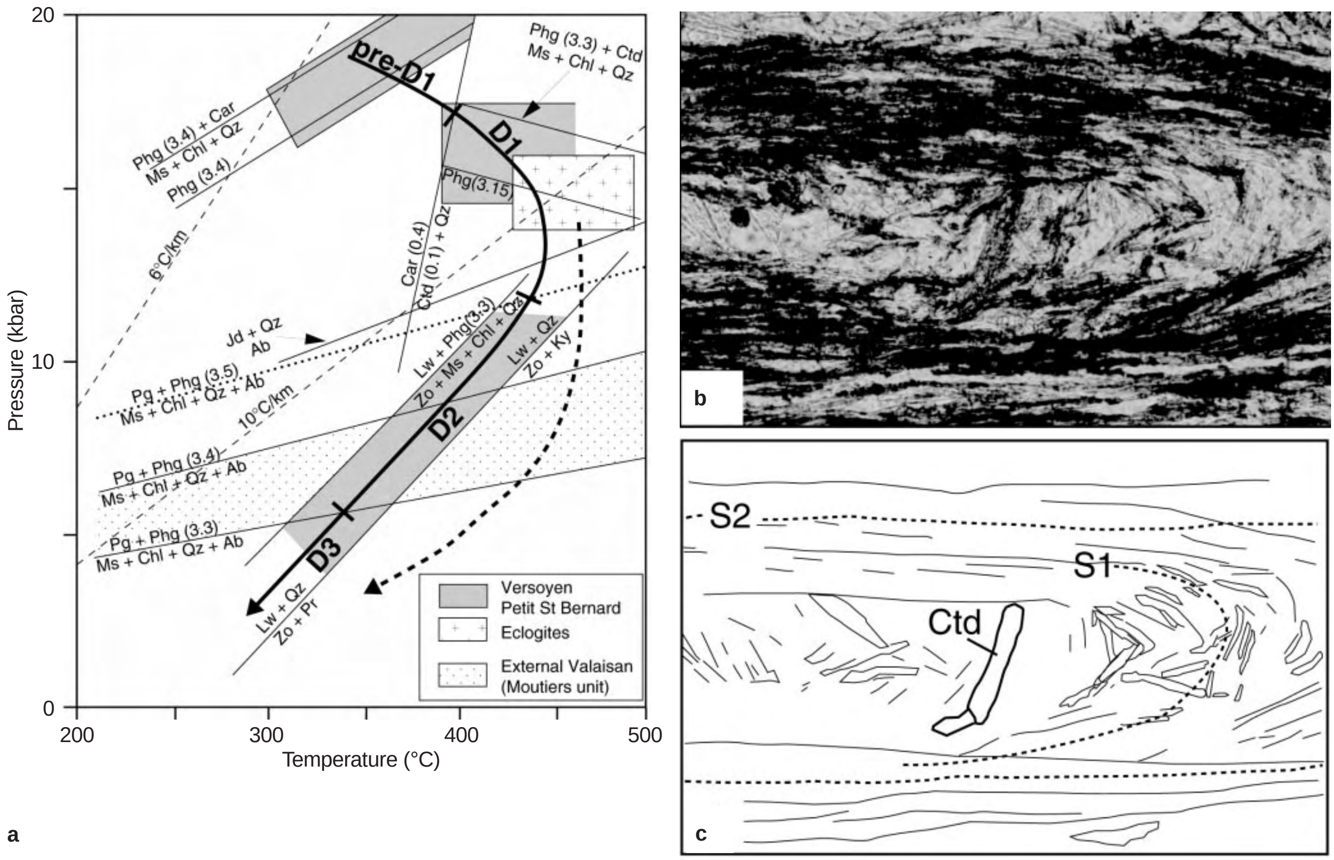
Fig. 5 Isoclinal F1 fold (axial plane indicated) in "Couches de St. Christophe" in the hinge of an F2 fold, internal Valaisan. *Arrow* indicates an F1 parasitic fold. *Camera lid* for scale

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)



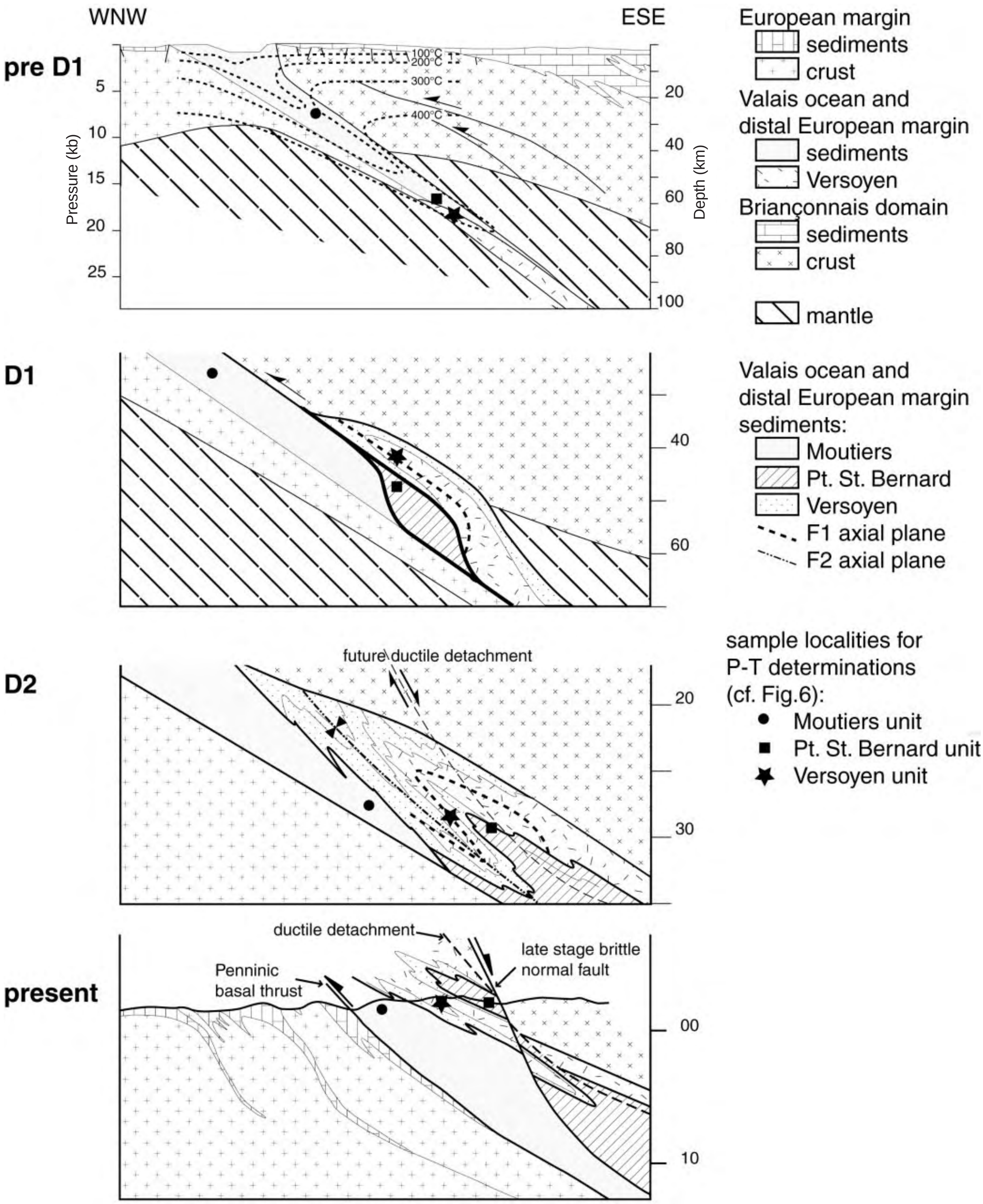


Fig. 7 Simplified sketch showing the structural evolution of the Valais units from the time of peak pressure conditions (pre-D1) to present. The P–T constraints are taken from Goffé and Bousquet (1997) and references therein (Fig. 6). Isotherms are drawn so as to fit estimated P–T conditions during the 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

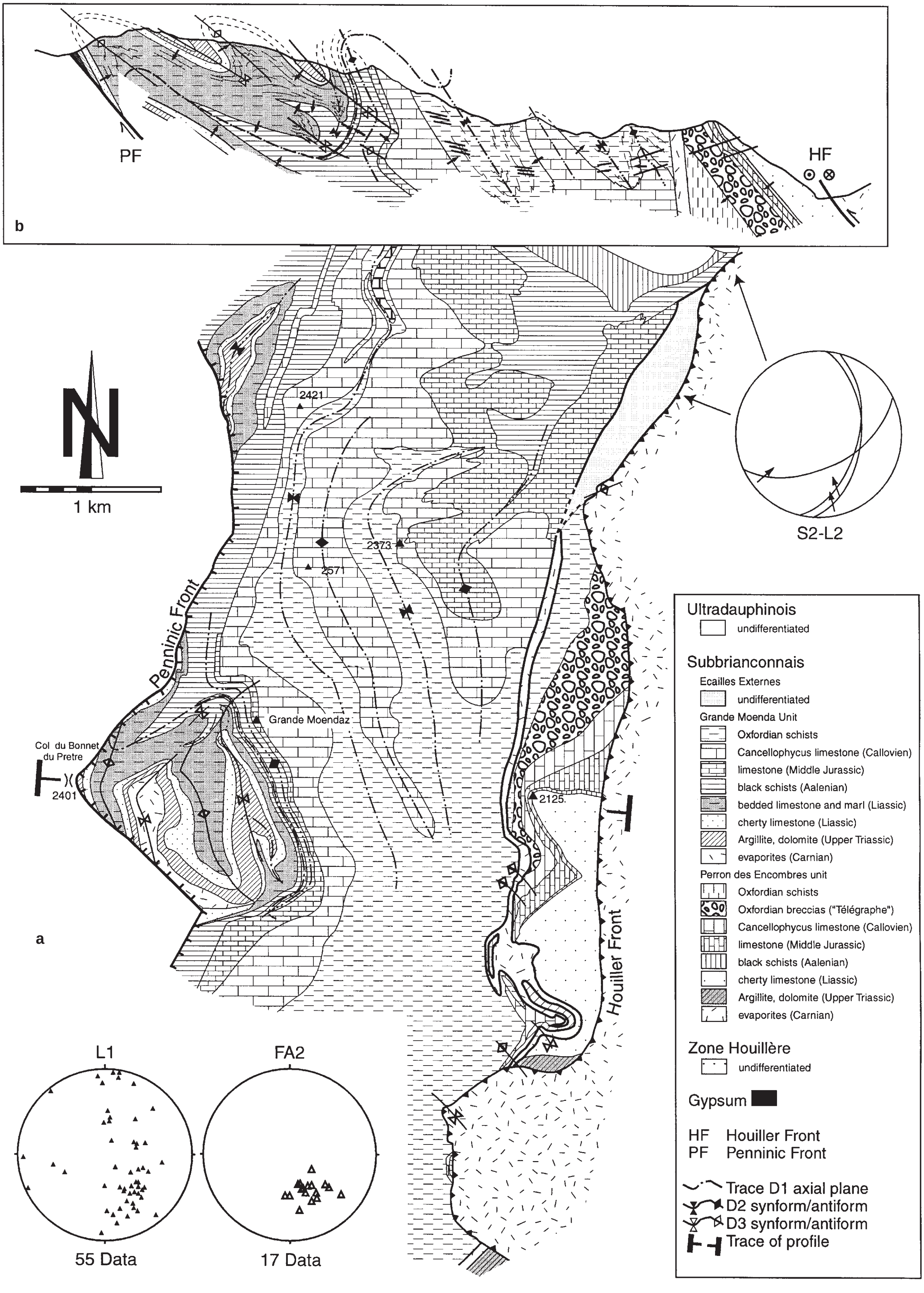

PF
HF
b
1 km
S2-L2
2421
2373
2571
Penninic Front
Grande Moendaz
Col du Bonnet du Pretre
2401
2125
Houiller Front
a
L1
55 Data
FA2
17 Data
Ultradauphinois
undifferentiated
Subbrianconnais
Ecailles Externes
undifferentiated
Grande Moenda Unit
Oxfordian schists
Cancellophycus limestone (Callovien)
limestone (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) 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-

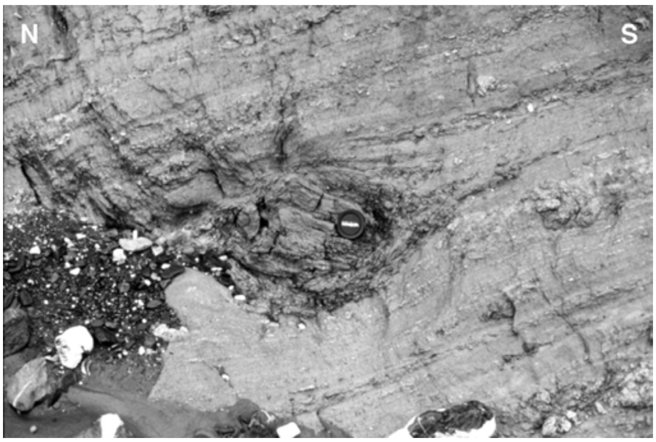


Fig. 9 σ-clast of dolomite in an anhydrite matrix indicates sinistral (i.e., top-to-the north) movement along the Houiller Front during D2. Locality: Torrent des Encombres

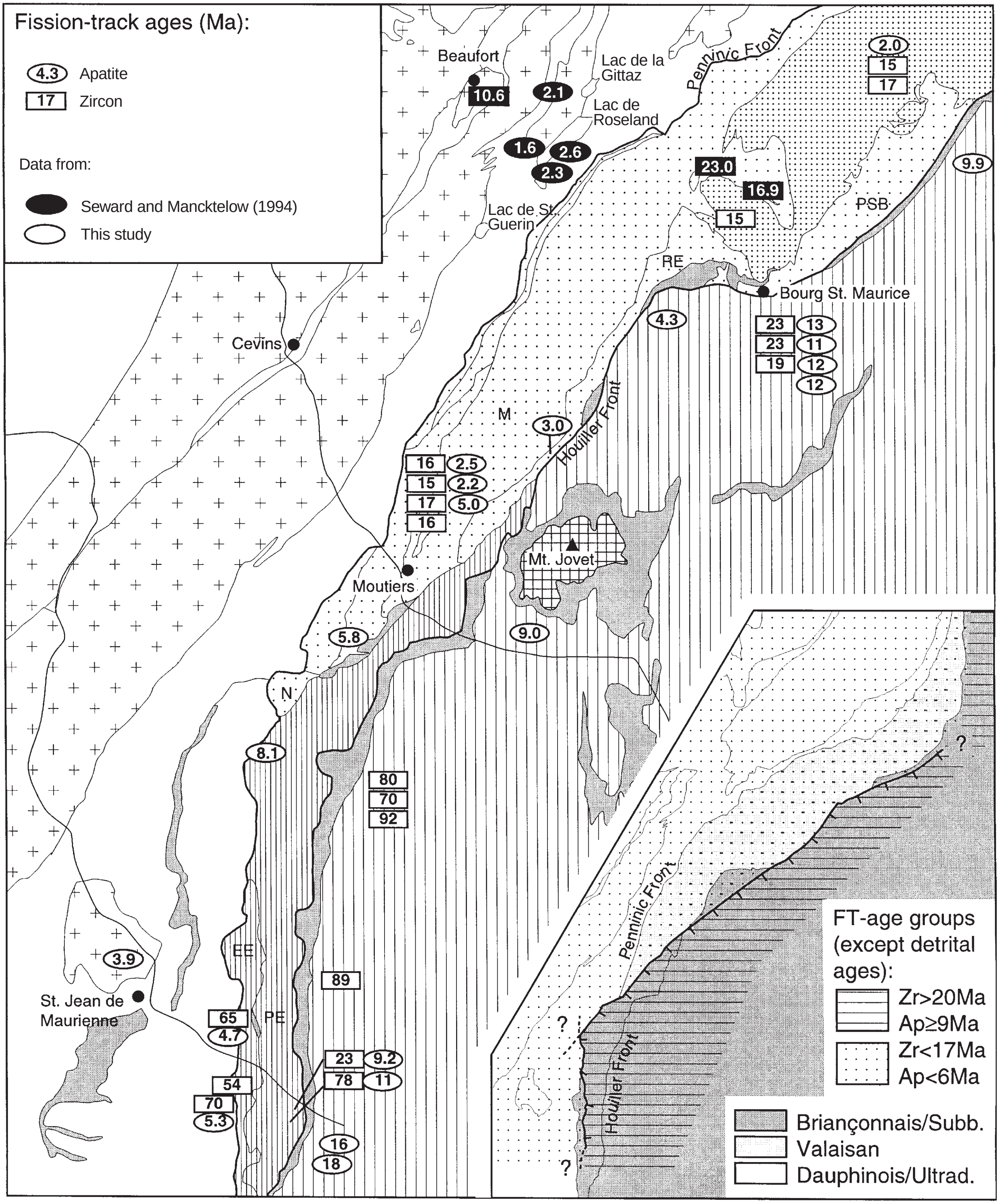


Fig. 10 Regional distribution of zircon and apatite fission-track ages from this study, together with data from Seward and Mancktelow (1994). *Inset* illustrates the map trace of late-stage normal faults, as deduced from the fission-track data

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

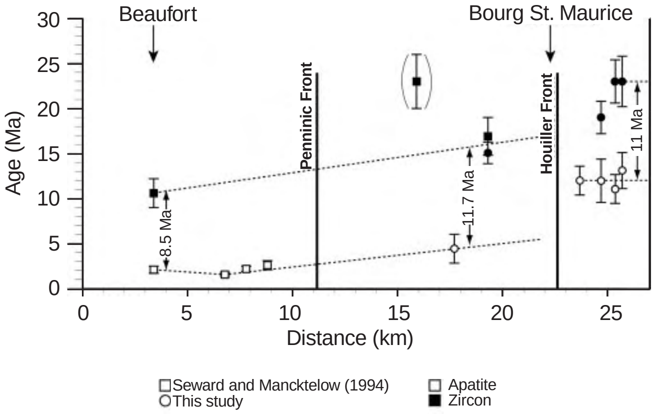


Fig. 11 Apatite and zircon fission-track ages along a WNW/ESE-trending profile through the Penninic Front and Houiller Front, respectively. The profile runs approximately parallel to the ECORS-CROP seismic line between Beaufort and Bourg St. Maurice (see Fig. 10). Note that the age difference between 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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References

Antoine P (1971) La zone des breches de Tarentaise entre Bourg-Saint-Maurice (vallée de l'Isère) et la frontière Italo-Suisse. Mém Lab Géol 9:1–367

Antoine P, Barféty JC, Vivier G, Gros Y, Fudral S, Landry P, Fabre J (1992) Carte géol France (1/50,000), feuille Bourg Saint Maurice (727). BRGM, Orléans

Antoine P, Barféty J-C, Vivier G, Debelmas J, Desmons J, Fabre J, Loubat H, Vautrelle C (1993) Notice explicative de la feuille Bourg Saint Maurice à 1/50,000, 2nd edn. BRGM, Orléans

Barbier R (1948) Les Zones Ultradauphinoise et Subbriançonnaise entre l'Arc et l'Isère. Mém Serv Carte Géol France, 291 pp

Bertrand J-M, Aillères L, Gasquet D, Macaudière J (1996) The Pennine Front zone in Savoie (Western Alps), a review and new interpretations from the Zone Houillère Briançonnaise. Eclogae Geol Helv 89:297–320

Bigi G, Cosentino D, Coli M, Parotto M, Sartori R, Scandone P (eds) (1983) Structural model of Italy, sheet no. 1, 1:500,000. C.N.R., Italy

Butler RWH, Matthews SJ, Parish M et al. (1986) The NW external Alpine thrust belt and its implications for the geometry of the western Alpine orogen. Geol Soc Spec Publ 19:245–260

Cannic S (1996) L'évolution magmatique et tectono-métamorphique du substratum du domaine valaisan (complexe du Versoyen, Alpes occidentales): implications dans l'histoire alpine. PhD thesis, Grenoble, 215 pp

Cannic S, Mugnier J-L, Lardeaux J-M (1995) Mise en évidence d'une faille normale ductile dans l'unité du Roignais-Versoyen (Alpes occidentales, France). C R Acad Sci Paris 321:513–519

Cannic S, Lardeaux J-M, Mugnier J-L, Hernandez J (1996) Tectono-metamorphic evolution of the Roignais-Versoyen unit (Valaisan domain, France). Eclogae Geol Helv 89:321–343

Debelmas K (1980) Les Alpes Franco-Italiennes. Geol Alpine 56:21–58

Elter P, Elter G (1957) Sull' existenza, nei dintorni del Piccolo San Bernardo, di un elemento tettonico riferibile al ricoprimento del Pas du Roc. Rend Acad Naz Lincei 22:181–187

Elter G, Elter P (1965) Carta geologica della regione del Piccolo San Bernardo (versante italiano). Note illustrative. Publication du centre pour l'étude géologique et pétrographique des Alpes 35:1–53

Escher A (1988) Structure de la nappe du Grand Saint-Bernard entre le Val de Bagnes at les Mischabel. Landeshydrologie und -geologie. Geol Berichte 7:1–26

Escher A, Hunziker JC, Marthaler M, Masson H, Sartori M, Steck A (1997) Geologic framework and structural evolution of the Western Swiss–Italian Alps. In: Pfiffner OA et al. (eds) Deep structure of the Swiss Alps: results of NRP 20. Birkhäuser, Basel, pp 205–222

Florineth D, Froitzheim N (1994) Transition from continental to oceanic basement in the Tasna nappe (Engadine window, Graubünden, Switzerland): evidence for Early Cretaceous opening of the Valais ocean. Schweiz Mineral Petrol Mitt 74:437–448

Freeman SR, Butler RWH, Cliff RA, Inger S, Barnicoat AC (1998) Deformation migration in an orogen-scale shear zone array: an example from the Basal Briançonnais Thrust, internal Franco-Italian Alps. Geol Mag 135:349–367

Frisch W (1979) Tectonic progradation and plate tectonic evolution of the Alps. Tectonophysics 60:121–139

Froitzheim N, Schmid SM, Frey M (1996) Mesozoic paleogeography and the timing of eclogite-facies metamorphism in the Alps: a working hypothesis. Eclogae Geol Helv 8981–110

Fudral S (1980) Une nouvelle interprétation de l'unité de Salins (zone des Brèches de Tarentaise) au nord-ouest de Bourg St. Maurice (Savoie). Consequences structurales. C R Acad Sci Paris 290:1333–1336

Fudral S (1996) Etude géologique de la suture téthysienne dans les Alpes franco-italiennes nord-occidentales de la Doire Ripaire (italie) à la région de Bourg-Saint-Maurice (France). Savoie, Chambéry, 397 pp

Goffé B, Bousquet R (1997) Ferrocarpholite, chloritoide et lawsonite dans les métapélites des unités du Versoyen et du Petit St. Bernard (zone valaisanne, Alpes occidentales). Schweiz Mineral Petrol Mitt 77:137–147

Green PF, Duddy IR, Laslett GM, Hegarty KA, Gleadow AJW, Lovering JF (1989) Thermal annealing of fission tracks in apatite. 4. Quantitative modelling techniques and extension to geological timescales. Chem Geol 79:155–182

Hurford AJ, Green PF (1983) The Zeta age calibration of fission-track dating. Isotope Geosci 1:285–317

Lancelot JC (1979) Etude tectonique de la zone des brèches de Tarentaise dans la région de Bourg-Saint-Maurice, Savoie. PhD thesis (3ème cycle), Paris

Loreau J-P, Gely J-P, Rampneaux J-P (1995) Cycles stratigraphiques dans les séries alpines du Lias et de l'Aalénien de part et d'autre du Front pennique (Savoie, France): Contrôle tectonique et rapport avec l'eustatisme. Eclogae Geol Helv 88:529–551

Maurer HR, Burkhard M, Deichmann N, Green AG (1997) Active tectonism in the central Alps: contrasting stress regimes north and south of the Rhone Valley. Terra Nova 9:91–94

Nicolas A, Polino R, Hirn A, Nicolich R, and ECORS-CROP working group (1990) ECORS-CROP traverse and deep structure of the Western Alps: a synthesis. In: Roure F, Heitzmann P, Polino R (eds) Deep structure of the Alps. Soc Géol France, Soc Géol Suisse, Soc Geol Italie

Perez-Postigo LV (1988) Contribution a l'etude géologique du Subbriançonnais entre Arc et Isère: les massifs du Perron des Encombres et de la Grande Moendaz (Alpes Occidentales-Savoie). PhD thesis (3ème cycle), Univ Savoie

Ricou LE, Siddans AWB (1986) Collision tectonics in the Western Alps. Geol Soc Lond Spec Publ 19:229–244

Schmid SM, Pfiffner OA, Froitzheim N, Schönborn G, Kissling E (1996) Geophysical–geological transect and tectonic evolution of the Swiss–Italian Alps. Tectonics 15:1036–1064

Schürch ML (1987) Les ophiolites de la zone du versoyen: témoin d'une bassin à évolution métamorphique complexe. PhD thesis, Univ Genève

Seward D (1989) Cenozoic basin histories determined by fission track dating of basement granites, South Island, New Zealand. Chem Geol 79:31–48

Seward D, Mancktelow NS (1994) Neogene kinematics of the central and Western Alps: evidence from fission-track dating. Geology 22:803–806

Soom MA (1990) Abkühlungs- und Hebungsgeschichte der Internmassive und der penninischen Decken beidseits der Simplon-Rhone Linie seit dem Oligozän: Spaltspurdatierung an Apatit/Zirkon und K/Ar datierungen an Biotit/Muskovit (Westliche Zentralalpen). PhD thesis, Univ Bern

Spencer S (1989) Nature of the North Pennine Front: French Alps. PhD thesis, Imperial College, London University

Stampfli GM (1993) Le Briançonnais, terrain exotique dans les Alpes? Eclogae Geol Helv 86:1–45

Stampfli G, Marchant R (1997) Geodynamic evolution of the Tethyan margins of the Western Alps. In: Pfiffner OA et al. (eds) Deep structure of the Swiss Alps: results of NRP 20. Birkhäuser, Basel, 223–240

Steck A (1990) Une carte de cisaillement ductile dans les Alpes Centrales. Eclogae Geol Helv 83:603–626

Trümpy R (1955) Remarques sur la corrélation des unités penniques externes antre la Savoie et le Valais et sur l'origine des nappes préalpines. Bull Soc Géol France 5:217–231

Yamada R, Tagami T, Nishimura S, Ito H (1995) Annealing kinetics of fission tracks in zircon: an experimental study. Chem Geol 122:249–258

CHAPTER 2: STRATIGRAPHY OF THE VALAISAN UNITS

A first chapter (2.1) briefly reviews earlier work and concepts on the stratigraphy of the Valais domain. It highlights the paleogeographical problems which led to the new arrangement of this realm into external and internal Valaisan domains (Fügenschuh et al., 1999). Subsequent chapters (2.2 to 2.5) deal with the description of the lithologies of the Valaisan units. These descriptions are carried out collectively, in the sense that "groups" of Valaisan formations, which represent individual tectonosedimentary stages during the evolution of the Valaisan domain, will be defined and discussed. Relevant characteristic features of the different tectonic units will be discussed in detail in Chapter 2.2.8. The last chapter (2.6) is dedicated to the ages of the Valaisan rocks, with particular emphasis on the controversy regarding the age of some of the sediments. A more detailed study of the syn-rift sediments, will be found in Chapter 3.

2.1: External and Internal Valaisan

This section, presents the new subdivision of the Valais into external and internal Valaisan realms. The external Valaisan includes, from west to east, the following units (Fig. 2.1): Quermoz, Moûtiers, Roc de l'Enfer and Pt. St. Bernard. The internal Valaisan is represented by the Versoyen unit only. Other tectonic units such as Cormet d'Arêches, Bagnaz, Crève Tête and Niélard, located outside the study area, are also ascribed to the external Valaisan based on their stratigraphic record. However, they are not discussed in detail in this study (see Fügenschuh et al., 1999 and Ceriani et al. 2000).

Two major differences emerge in respect to previous studies (Antoine et al., 1992 and 1993.) One is the attribution of the Pt. St. Bernard unit, previously ascribed to the Subbriançonnais paleogeographic domain (Elter and Elter 1957), and of the Roc de l'Enfer unit to the Valaisan (see Ch. 2.2 and 2.2.8). The Roc de l'Enfer unit was attributed to either the Valaisan or the Briançonnais domain (Antoine et al., 1992). The other matter of discussion is the subdivision of the Roignais-Versoyen unit as defined by (Antoine 1971; et al., 1992 and 1993), into the Versoyen unit, which forms the internal Valaisan, and the Roignais unit which we consider to be part of the Moûtiers unit (the external Valaisan). Both interpretations strongly influence the inferred paleogeographical arrangement of the Valaisan units.

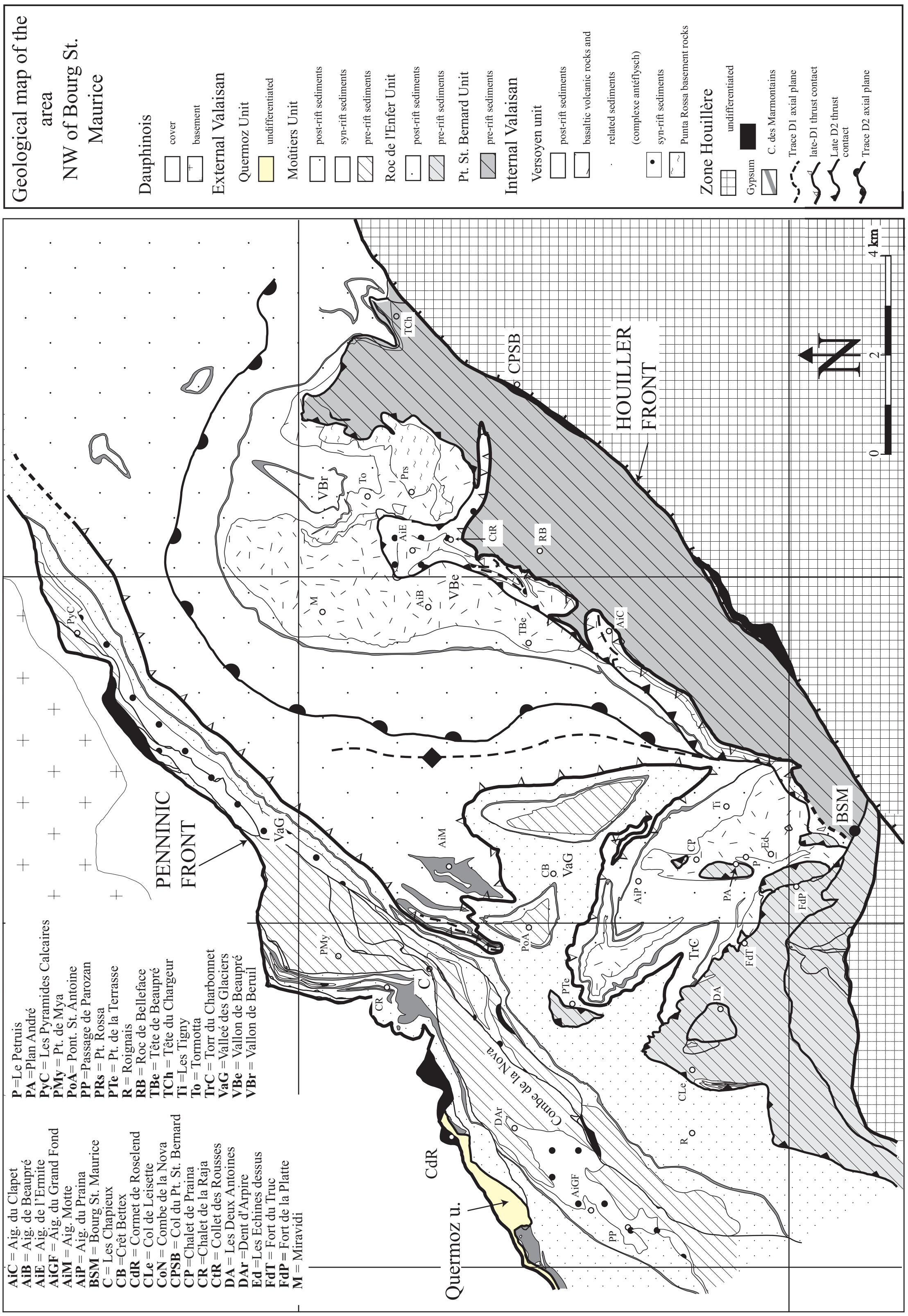


Fig. 2.1 Geological map of the study area showing the distribution of the Valaisan units. All the localities depicted here are referred to Chapter 2.

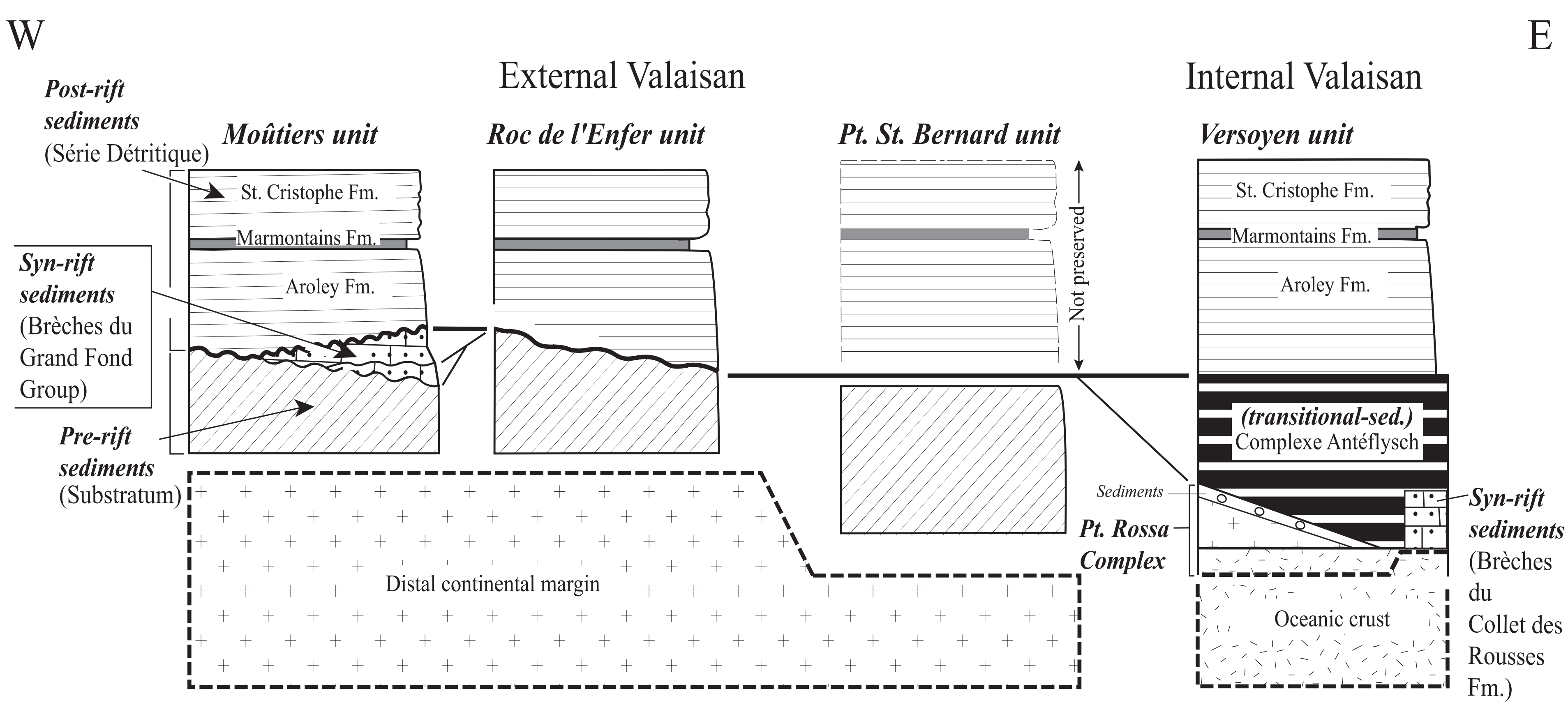
A first coherent stratigraphy for the Valaisan domain was established by the extensive work of Antoine (Antoine 1964, 1965a, b and c, 1966, 1968, 1970, 1971, 1972, Antoine et al., 1972, 1992 and 1993) This work led to a revision of the previous propositions developed by other French, Italian and Swiss workers (Barbier, 1948; Elter and Elter, 1965; Elter, 1951; Elter and Elter, 1957;

Fudral, 1973; Loubat, 1968; Schoeller, 1929; Trümpy, 1955a; Trümpy, 1955b; Zulauf, 1964). Historically, the study of the Valaisan rocks, extending from SW to NE across France, Italy, Switzerland and Austria (Froitzheim et al., 1996) was based on stimulating periods of rigorous discussions between workers belonging to different countries (a remarkable example is given in (Barbier and Trümpy, 1955) an attitude which is partially forgotten in recent times.

These studies on the stratigraphy of the Valaisan rocks were guided by the necessity to define uniform local names and definitions of the different lithostratigraphic units and, more importantly, they attempted to date the Valaisan formations, a problem which is still a matter of debate (see Ch. 2.6). Antoine in particular made a major effort to uniform previous knowledge and, based on detailed field observation, he proposed his own synthesis of the Valais domain, as presented in Antoine (1971). The present study would not have been possible without the extensive work of Antoine over 25 years (from Antoine 1965 to Antoine et al. 1993).

According to the classical view (Antoine, 1971), the Valais domain is characterised by three different groups of rocks, each corresponding to separate stages during the paleogeographic and paleotectonic development during Mesozoic times, pre-dating Alpine orogeny. These three groups are the "Substratum", the "Complexe Antéflysch" and the "Série Détritique". The corresponding episodes of the evolution of the Valais domain before the onset of the Alpine convergence are: early stages of moderate subsidence and rifting (Carboniferous – Middle Liassic); main stage of rifting and formation of the Valais domain (Late Liassic – Earliest Cretaceous) and filling of the Valais basin (Early Cretaceous – Late Cretaceous? and/or Tertiary?). This stratigraphy was further refined after the recognition of breccia formations (Fig 2.2) which are either found between the "Substratum" below and the "Série Détritique" above (i.e., the Brèches du Grand Fond of Antoine et al. 1972 and Fudral 1973), or at the base of the Complexe Antéflysch Fm. (i.e. the Brèches du Collet des Rousses of Loubat, 1968 and 1975).

In this first chapter we prefer to maintain the terms adopted by Antoine (1971) in order to be consistent with the existing literature. However, in order to emphasize the paleogeographic correlation between external and internal Valaisan, it will be necessary to use a more adequate terminology later on (Chapters 2.2 to 2.5): i.e. "pre-rift sediments" for the "Substratum", and "post-rift sediments" for the "Série Détritique".



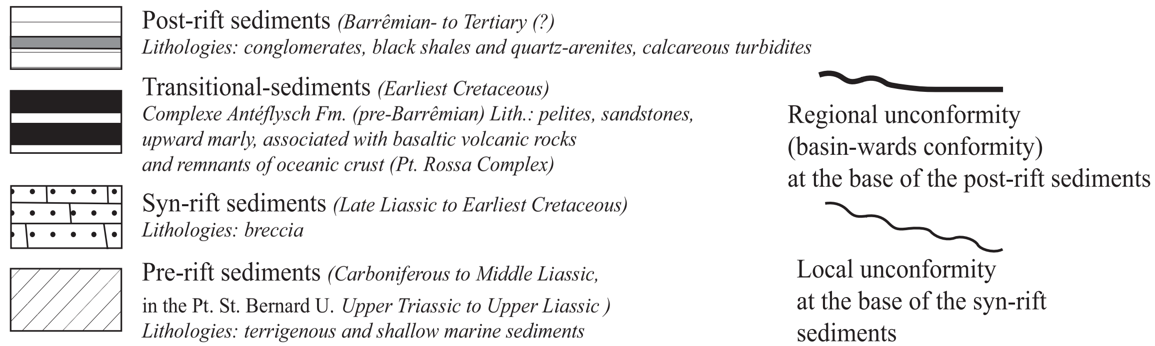


Fig. 2.2 Generalised stratigraphic scheme showing the main subdivision of the Valaisan units into: Substratum (or pre-rift sediments); Brèches du Grand Fond and Brèches du Collet des Rousses (or syn-rift sed.); Complexe Antéflysch (or transitional-sed.); Pt. Rossa Complex (remnants of oceanic crust) and Serie détritique (post-rift sed.). These groups of rocks represent two distinct realms, one deposited on thinned continental crust (external Valaisan) and the other deposited on oceanic crust (internal Valaisan).

For the intermediate breccia formations of the Brèches du Grand Fond (External Valaisan), and for the Brèches du Collet des Rousses and the volcano-sedimentary sequence (sills and black schists) of the Complexe Antéflysch Fm. of the internal Valaisan, we will use the terms syn-rift and transitional sediments. In particular we define the Brèches du Grand Fond and the Brèches du Collet des Rousses as syn-rift sediments *s.s.* (referred as syn-rift sediments) and the Complexe Antéflysch Fm. as transitional sediments.

The use of the term "transitional sediment" for the Complexe Antéflysch may appear to disturb the simple scheme regarding a subdivision into pre-, syn- and post-rift series which has been followed here to interpret the stratigraphic record of the Valais domain. The term "transitional" simply refers to distal sediments, which were deposited coevally with the formation of oceanic crust and onto oceanic crust. They are not referred as post-rift sediments because they do not unconformably overlay both external and internal Valaisan. Hence they are characteristic of the oceanic part of the Valaisan only.

According to the scheme of Fig. 2.2 the term "Complexe Antéflysch" is used for the oceanic part of the Valaisan domain only. Remnants of this oceanic crust associated with slices of continental basement are found in the "Pt. Rossa complex" (Fig. 2.2, see Ch.2.4).

The youngest group is the "Série Détritique" (or "Série Détritique de Tarentaise" of Antoine 1971). These sediments are considered to be diagnostic for the Valais domain. They are of Early Cretaceous (Barrêmian) (Trümpy 1954, Elter and Elter 1965, Sodero 1967) to probably Tertiary age. These sediments were subdivided into three formations (Trümpy 1954) which are: (1) the conglomerates of the Aroley Fm. (Early Cretaceous) at the base followed by (2) the quartz-arenites and black shales of the Marmontains Fm. (Middle Cretaceous). At the top there is an alternation of calcareous turbiditic deposits of the St. Cristophe Fm. (Late Cretaceous to probably Tertiary). The term "Flysch de Tarentaise" (Schöller 1929) was initially adopted to describe this "Série Détritique". Antoine (1971) first recognised that the term "flysch" was improperly used to describe the entire "Série Détritique" (Antoine 1971), especially concerning the conglomerates of the Aroley Fm. and the quartz-arenites of the Marmontains Fm.

Therefore, only the St. Cristophe Fm., which represents a monotonous repetition of calcareous and pelitic strata, was referred to as a flysch-type sediment by Antoine ("..le Flysch proprement dit.."). It should be remembered that the definition of C. de l'Aroley, C. des Marmontains and C. de St.

Cristophe, mentioned nowadays also by Antoine (et al. 1992) was first introduced by Trümpy (1954) in a region adjacent to our study area (Val Ferret, Switzerland). Later, the correlation between the facies in our study area and that recognised by Trümpy (1954) was discussed in two remarkable publications by Barbier and Trümpy (1955) and Elter (1960).

In the external Valaisan the ”Série Détritique” unconformably overlies the partly eroded succession of Carboniferous to Liassic terrigenous and shallow marine sediments (Fig. 2.2), i.e. the ”Substratum” of Antoine (1971). This ”Substratum” forms a second group of lithologies which represents part of the pre-Cretaceous (Carboniferous to Middle Liassic) Valaisan sedimentary succession. In the internal Valaisan, the base of the Aroley Fm. conformably and stratigraphically overlies a third group of lithologies, i.e. the Complexe Antéflysch Fm. (Antoine 1971) (Fig 2.2). Hence, the unconformity at the base of the ”Série Détritique” characteristic for the external Valaisan, passes into the correlative conformity towards the internal Valaisan (“basin-wards”).

The “Complexe Antéflysch” consists of a volcano-sedimentary sequence and of the “Pt. Rossa Complex”. The volcano-sedimentary sequence contains abundant basaltic sills (prasinites and gabbro) and pillow lavas intruded into pelitic sediments, which were collectively referred to as “black schists” (Loubat, 1968). The sills intruded “soft” and “wet” sediments (Loubat 1968, 1984; Loubat and Delaloye 1984; Schürch 1987; Cannic 1996, Cannic et al 1996) which are considered to have been deposited coevally with the intrusions. Downward, the Complexe Antéflysch, stratigraphically passes into the Brèches du Collet des Rousses (Loubat 1968) as well as into the Pt. Rossa Complex (Elter and Elter 1965, Dalla Torre 1998). Structurally both, the Brèches du Collet des Rousses and the Pt. Rossa Complex are located in the lowermost part of the Complexe Antéflysch. The Pt. Rossa complex consists of slices of granite-gneissic basement, associated with serpentinites (including ophicalcites), which are unconformably overlain by a Mesozoic conglomerate (Dalla Torre, 1998). The Pt. Rossa complex is regarded as part of the oceanic basement of the Complexe Antéflysch Fm. (Ch.2.4). The exact age of the opening of the Valaisan ocean is not known (nor that of the Complex Antéflysch and the Pt. Rossa complex). However, it is tentatively interpreted to be of the Earliest Cretaceous, since it immediately pre-dates the deposition of the Aroley Fm. of Barrêmian age (Trümpy 1954, Elter and Elter 1965, Sodero 1968).

The most important result emerging from the extensive work of Antoine is the proof that stratigraphic relationships exists between the three different groups of rocks described above which belong to a common paleogeographic domain, i.e. the Valais domain. The reconstruction of this domain, based on sedimentological criteria rather than on structural observations, shows that the ”Substratum” represents the cover sequence of a continental margin while the Complexe Antéflysch Fm. with its basaltic volcanic rocks represents a position towards or within oceanic lithosphere. Antoine (1971) was the first to demonstrate the stratigraphic transition from the Complexe Antéflysch Fm. into the “Série Détritique”. He also first discovered the coeval deposition of the “Série Détritique” onto both “Substratum” and Complexe Antéflysch. This findings indicates that from the time of deposition of the “Série Détritique” onwards all the tectonic units of the Valaisan show a common sedimentary record. This supports the idea that the Valaisan forms a coherent paleogeographical domain from the time of the deposition of the “Série Détritique” only. This domain includes a continental (Substratum) and an oceanic (Complexe Antéflysch) parts (Fig 2.2) which have an entirely different previous history.

Based on sedimentological observations, Antoine (1971) proposed a scheme for the Valaisan tectonic units which remained unchanged until recent times (Antoine et al., 1993). He divided the Valais domain into the Moûtiers and the Roignais-Versoyen units. Other tectonic units, like the Pt. St. Bernard and the Roc de l’Enfer units, according to this study also part of the Valais domain, where defined as “unités d’origine paléogeographique incertaine” (Antoine et al. 1992). However, the structure of the Valaisan nappe pile had received relatively little attention and a proper structural framework has not been clearly established yet.

There are two reasons for suggesting a redefinition of the classical tectonic subdivision of the Valais domain of Antoine (Antoine et al. 1992) as proposed in this study. A first one concerns the sedimentological record found in the “Roignais-Versoyen unit” of Antoine (et al. 1992). According to these authors, the Roignais-Versoyen unit also contains basaltic volcanic rocks of the Complexe Antéflysch Fm., as found in the sequence of the “Versoyen”, but considered to have been emplaced by volcanic activity directly onto a carbonate platform in the case of the ”Roignais-”, i.e. onto the “Substratum”. This leads to an unsatisfactory definition of the “Roignais-Versoyen” as partly oceanic and partly continental. Hence, it is more appropriate to include the “Roignais” part into the Moûtiers unit, both these parts being characterised by the same continental “Substratum”.

The second reason for redefining the classical subdivision is provided by a newly identified D1 regional scale mega-fold recognized within part of the former “Roignais-Versoyen” unit. The results of this analysis, fully treated in Chapter 5 (Chapters 5.3 and 5.4), are only briefly summarised here. A D1 megafold, formed within the Complexe Antéflysch Fm. and part of the

"Série Détritique", both formations belonging to the oceanic "-Versoyen" sequence, is thrusted onto the Pt. St. Bernard unit during a late-D1 stage, and onto what was previously considered to be the "Roignais-" part of the Roignais-Versoyen (now part of the Moûtiers unit). Both the "Roignais" and the Pt. St. Bernard unit are characterised by a continental "Substratum", overlain by the "Série Détritique". Moreover, the analysis of this late-D1 thrust reveals that the Versoyen sequence has to be rooted in a more internal position with respect to the Pt. St. Bernard and Moûtiers units.

Therefore, according to our tectonic and sedimentologic results, there is no need to postulate that the Complexe Antéflysch Fm. was directly deposited onto a continental "Substratum" (continental) and within the same tectonic unit (the Roignais-Versoyen unit of Antoine 1971). Instead we ascribe the Complexe Antéflysch Fm., including younger cover of the "Série Détritique", to the newly defined "Versoyen" unit, which represents an oceanic paleogeographic unit within the Valais domain. This oceanic Versoyen is distinct from the series characterised by a "Substratum", representing the continental part of the Valais domain. It is only the Série Détritique (post-rift) which is common to both the oceanic (internal) and continental (external) parts of the Valaisan.

This new subdivision of the Valais domain into two major tectonostratigraphic units originating from a distal continental margin (external Valaisan) and from a region floored by oceanic crust (internal Valaisan), has been systematically applied to this work. In Fig 2.3, which does not include information on the different tectonic units, the Valaisan lithologies have been mapped as pre-, syn- or post- rift sequences, or, as oceanic Complexe Antéflysch sediments (i.e. transitional sediments). Fig 2.3 also depicts the boundary between stratigraphic sequences characterised by a continental "Substratum" (pre-rift sediments deposited onto a continental margin, Fig 2.2), including its post-rift sediments (external Valaisan) and the stratigraphic sequences composed of the oceanic "Complexe Antéflysch" and its post-rift cover (internal Valaisan). In the SE part of the study area this boundary coincides with the tectonic contact between the Pt. St. Bernard unit and the Versoyen (mainly Complexe Antéflysch, compare Fig. 2.3 with Fig. 2.1). Moving westward, the trace of this same boundary must often be searched within the post-rift sediments (as in the example shown in the inset of Fig 2.3).

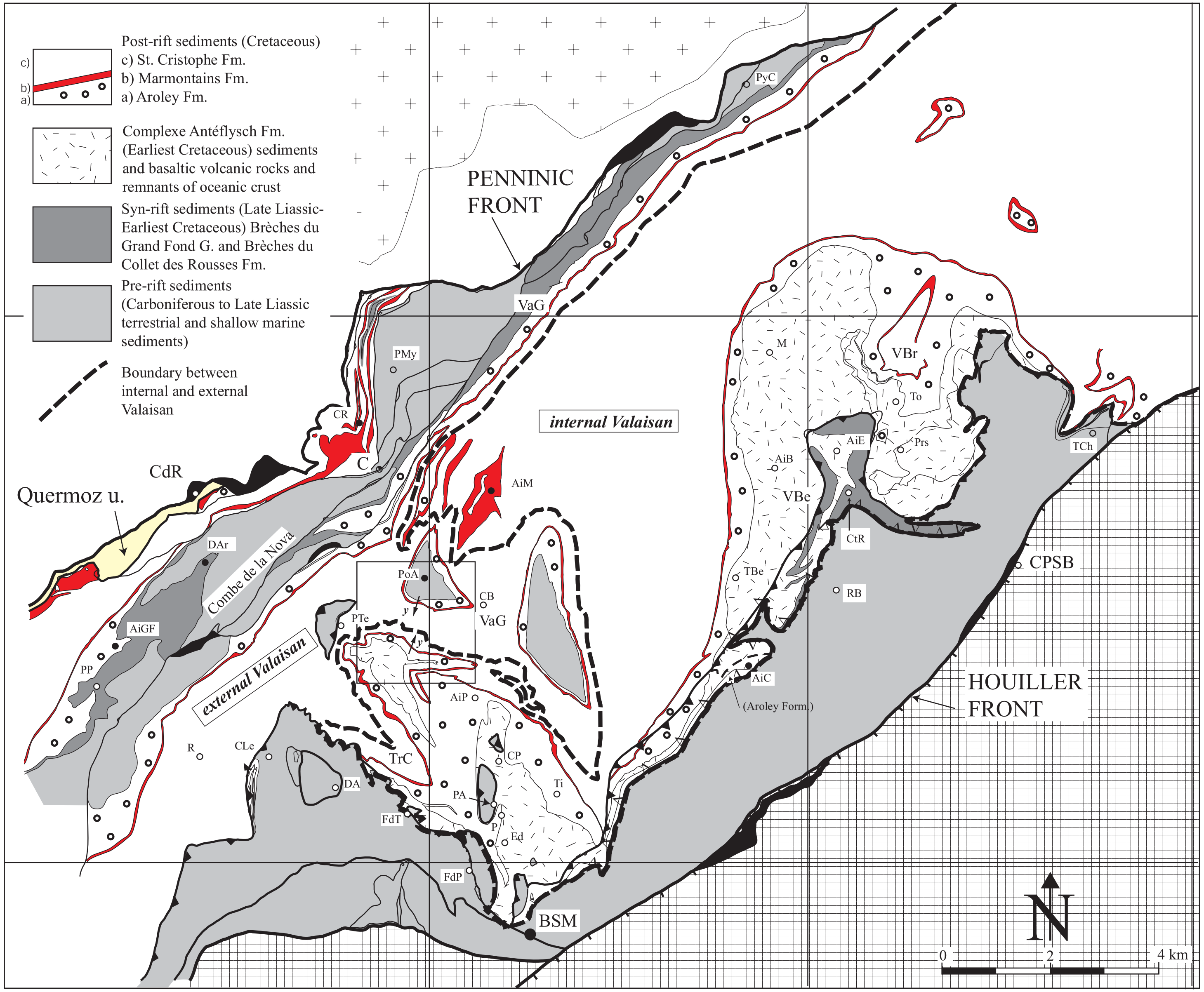


Fig. 2.3. Geological map of the Valais zone showing the outcrops of pre-, syn- and post-rift sediments and of the oceanic Complexe Antéflysch. Inset indicates a region where the post-rift sediments either rest in stratigraphic contact with pre-rift sediments (at Pt. St. Antoine, PoA) and therefore belong to the external Valaisan, or with the Complexe Antéflysch (near the Pt. de la Terrasse, PTe), and therefore belonging to the internal Valaisan.The black arrows indicate the younging direction (y). In this case the boundary between external and internal Valaisan must be searched for within the St. Christophe Formation (in white). In order to trace the boundary between the external and the internal Valaisan, the same distinction between post-rift sediments deposited onto the Complexe Antéflysch and post-rift sediments deposited onto pre-rift sediments has been applied in the entire study area.

The inset of Fig 2.3 shows post-rift sediments which are in stratigraphic continuity with pre-rift sediments at Pont St. Antoine (PoA) and in stratigraphic continuity with the Complexe Antéflysch along the ridge between the Pt. de la Terrasse (Pte) and the Aig. de Praina (AiP). In the same inset, the Marmontains Fm. (in red) indicates the presence of a duplication within the post-rift sediments. One normal sequence, together with the pre-rift sediments forms the Moûtiers unit (deposited in a continental realm), while the other overturned sequence, together with the Complexe Antéflysch (deposited in a oceanic realm) forms the Versoyen unit. The exact location of the contact between Moûtiers and Versoyen units (Fig 2.1) has been determined on the basis of a structural study (Chapter 5, compare Fig 2.3 with Fig 2.1). Retro-deforming this contact into its original position (before pre-D1) results in the attribution of the continental realm to the external Valaisan, i.e. towards the European continental margin, and of the oceanic realm to the internal Valaisan, situated between the European margin and the more internal Briançonnais domain (Figs. 2.4 and 5.5).

This simple palinspastic restoration is based on the assumption that, during the early stages of deformation (mainly D1), the structurally higher units (the Versoyen unit) were derived from more internal paleogeographic regions relative to the structurally lower units (the Pt. St. Bernard and Moûtiers units). However, this paleogeographic arrangement into external/internal (continental/oceanic) it is not only supported by structural arguments. Further evidence comes from the sedimentological analysis of the Valaisan rocks. Of particular importance is the paleogeographic facies distribution of the post-rift sediments which filled the entire Valaisan basin after the opening of the Valais ocean (see Ch. 2.5).

Finally, some remarks on the distinction of the stratigraphic record into, pre-, syn-, transitional- and post-rift sediments have to be made. The distinction into stratigraphic sequences (Fig. 2.2) is based on the recognition of unconformities, along which there is evidence of paleogeographic and paleotectonic changes mainly related to the Valaisan rifting (subaerial exposure and/or hiatus in the sedimentation and/or angular unconformity). In the external Valaisan, the Carboniferous to Middle Liassic "Substratum" is unconformably overlain by a breccia formation (the Brèches du Grand Fond Group) of Late Liassic to pre-Barrêmian age. A younger unconformity exists between the base of the Série Détritique (the Aroley Fm.) and above the Brèches du Grand Fond Group, or, directly over the substratum. Hence, the substratum is considered to correspond to the pre-rift sediments, the Brèches du Grand Fond to the syn-rift sediments and the Série Détritique to the post-rift sediments.

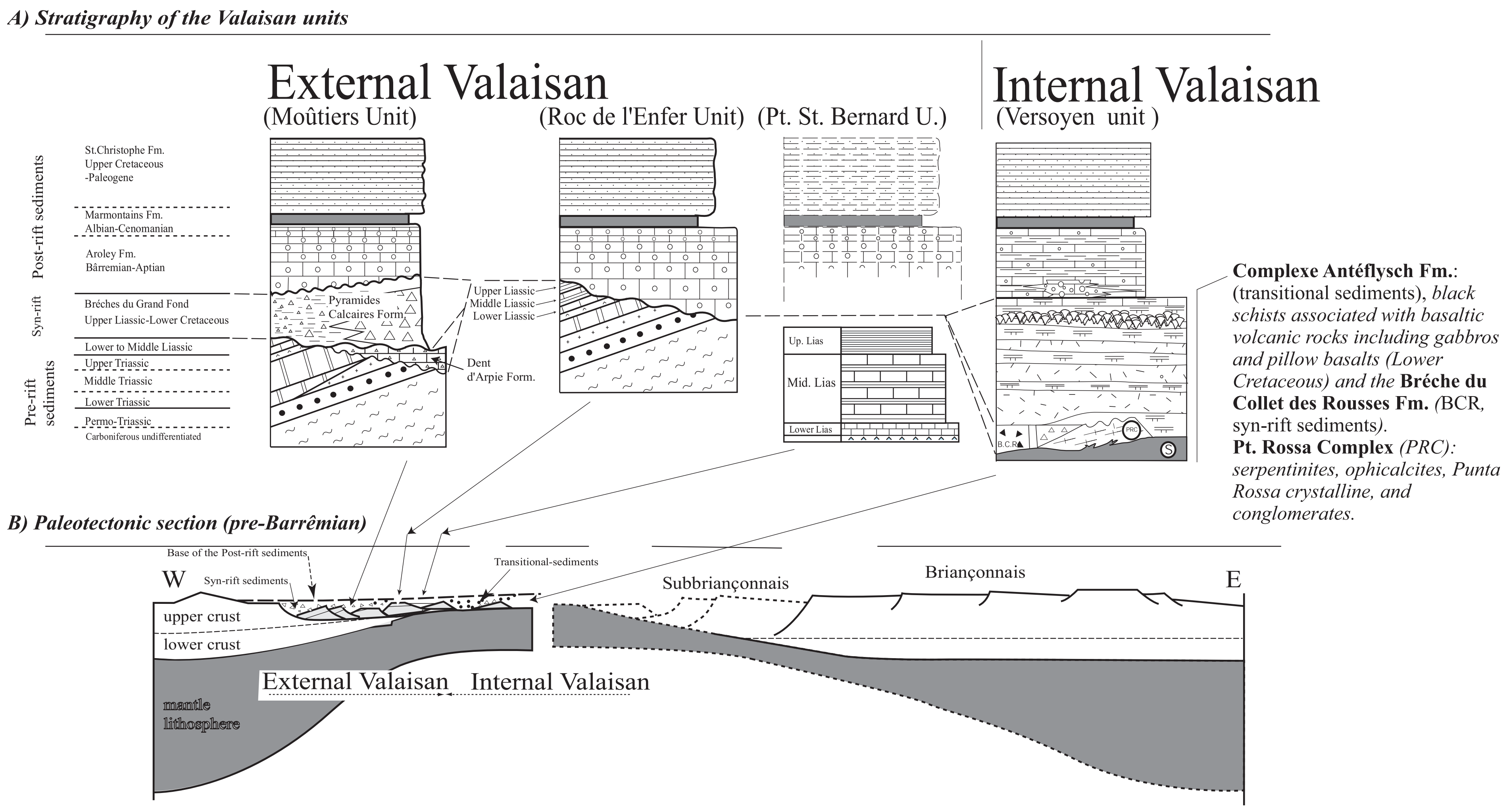


Fig. 2.4: A) Stratigraphy of the Valaisan units (modified after Fügenschuh et al. 1999). B) Paleotectonic section of the Valais domain pre-dating the deposition of the post-rift sediments.

In the internal Valaisan, the same Série Détritique overlies the Complexe Antéflysch (black schists intruded by sills) which in turn is in stratigraphic continuity downwards and into the Brèches du Collet des Rousses. The sediments of the Brèches du Collet des Rousses are tectonically detached at their base (Antoine, 1971; Dalla Torre, 1998). Therefore, their basal unconformity is not exposed. The Série Détritique of the internal Valaisan represents the basin-wards part of the post-rift sediments of the external Valaisan (Antoine 1971, see also Ch. 2.5). Therefore, the Série Détritique of the Valaisan domain unconformably overlies both, the external Valaisan (pre- and syn-rift sediments) and the internal Valaisan (Complexe Antéflysch).

This regional scale unconformity is considered as characteristic for post-rift sediments (i.e. the Série Détritique). Hence, the Complexe Antéflysch and the underlying Brèches du Collet des Rousses, which are older than the unconformity at the base of the post-rift sediments and are considered as syn-rift sediments in a wider sense: the syn-rift sediments s.s. of the Brèches du Collet des Rousses and the transitional sediments of the Complexe Antéflysch. The basement underlying the Brèches du Collet des Rousses and/or the Complexe Antéflysch is considered oceanic and it is in part represented by the Pt. Rossa Complex.

2.2: Lithological characteristics of the external Valaisan: the pre-rift sediments of the Moûtiers and Roc de l'Enfer units

The external Valaisan, characterised by the presence of pre-rift sediments, is subdivided into the Moûtiers, Roc de l'Enfer and the Pt. St. Bernard units (Fig 2.4a). The Moûtiers unit shows the most complete sequence, including pre- syn- and post-rift sediments (Fig. 2.5). The Roc de l'Enfer unit, now tectonically juxtaposed onto the Moûtiers unit (Ch. 6.3.2 to 6.3.5), contains mainly pre- and post-rift sediments identical to those of the Moûtiers unit (Fig. 2.6). The Pt. St. Bernard unit only contains pre-rift sediments (Fig. 2.7, see also Ch 2.2.8). The attribution of all these three units to the external Valaisan is justified by the fact that all of them contain pre-rift sediments representing an eroded carbonate platform which was deposited onto continental crust.

The observed dismembering of the external Valaisan into separate tectonic units is controlled by the presence or absence of detachment horizons within the pre-rift sediment series: the Upper Triassic evaporites and the Upper Liassic black schists.

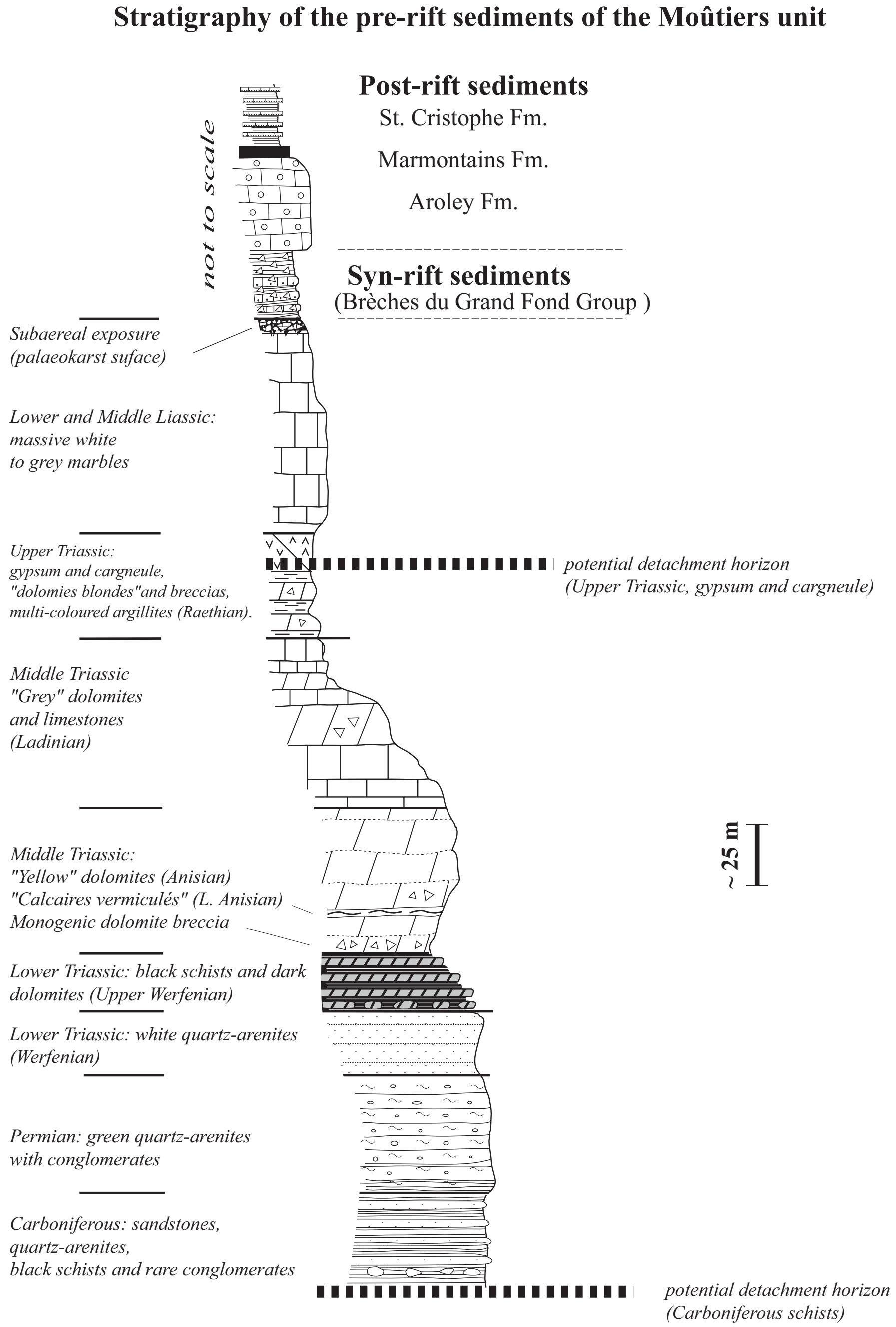


Fig. 2.5 Schematic stratigraphic section of the pre-rift sediments of the Moûtiers unit. After Antoine (1971, *et al.* 1992), Fudral (1973, 1998).

The Upper Liassic horizon is only preserved within the Roc de l'Enfer and the Pt. St. Bernard units. In the Moûtiers and Roc de l'Enfer units, due to the unconformity below syn- and or post- rift sediments (Fig 2.4a) two different situations can be recognised. Where these evaporites were preserved from subsequent erosion they acted as detachment horizons. In this case a stratigraphic sequence starting with the Upper Triassic has been separated from older formations of the continental margin. Where the Upper Triassic evaporite horizon is missing due to subsequent erosion below the post-rift sediments, Lower and Middle Triassic carbonates as well as Permo-Triassic quartz-arenites were detached as one coherent series along Carboniferous schists. The contact between pre- and syn-rift sediments, or that between pre- and post-rift sediments, has never been observed to act as a detachment horizon. This is demonstrated by the abundance of primary stratigraphic contacts between these formations in the study area (Fig 2.1). In the case of the Moûtiers and Roc de l'Enfer units, which together constitute the bulk of the external Valaisan, the evaporite horizon is generally missing and these units have their stratigraphic base in the Carboniferous schists (Figs. 2.5 and 2.6). However, where the evaporites are present, they pre-determine a fragmentation of the stratigraphic sequence. In the Moûtiers unit an upper stratigraphic sequence consisting of Upper Triassic to Cretaceous sediments can be locally separated from a lower sequence which comprises Middle Triassic carbonates, Permotriassic quartz-arenites and Carboniferous schists. Field examples of this can be observed in the Pyramides Calcaires and in the Pt. de Mya regions, respectively. Within the Roc de l'Enfer unit the partitioning into sub-units becomes more pronounced compared to the Moûtiers unit. This led to the formation of a complex system of thrust sheets (the Deux Antoine, the Forclaz and the Plan Andre thrusts sheets all together referred as to as the Roc de l'Enfer unit, see Ch. 6.3.2 to 6.3.5). This particular character of the Roc de l'Enfer unit is related to the presence of the Upper Triassic evaporites and of another detachment horizon located within the Upper Liassic schists (Fig. 2.6).The same detachment horizons (Upper Triassic and Upper Liassic) also characterise the Pt. St. Bernard unit (Fig. 2.7, see also Ch. 2.2.8). The sedimentary record of the Moûtiers unit shows a complete basin evolution during the Mesozoic when this region evolved in a platform environment (pre-rift sediments) followed by the coarse clastics of the syn-rift stage (mainly breccias).

The age of deposition of these breccias is unknown. Sedimentation may have taken place between Late Liassic and Lower Cretaceous. A thick series of post-rift sediments completes the stratigraphy of the Moûtiers units. Contrary to the Moûtiers unit, the Roc de l'Enfer unit represents a region where the same carbonate platform series (pre-rift sediments) is in direct stratigraphic contact to the post-rift sediments and where the syn-rift sediments are only locally preserved. These differences can be interpreted in terms of syn-sedimentary formation of *horst* and *graben* structures, mainly related to the opening of the Valaisan ocean. The transition from pre-rift sediments to post-rift sediments, without the presence of the intermediate syn-rift sediments, characterises parts of the Moûtiers unit (at the Pont St. Antoine, Fig 2.3) and the entire Roc de l'Enfer unit (at the Col de Leisette, Fig 2.3). These situations can be interpreted as evidence for horst structures within the external Valaisan. The different lithologies of pre-rift sediments of the external Valaisan will be summarized below and complemented with new observations (Fig. 2.5 and 2.6).

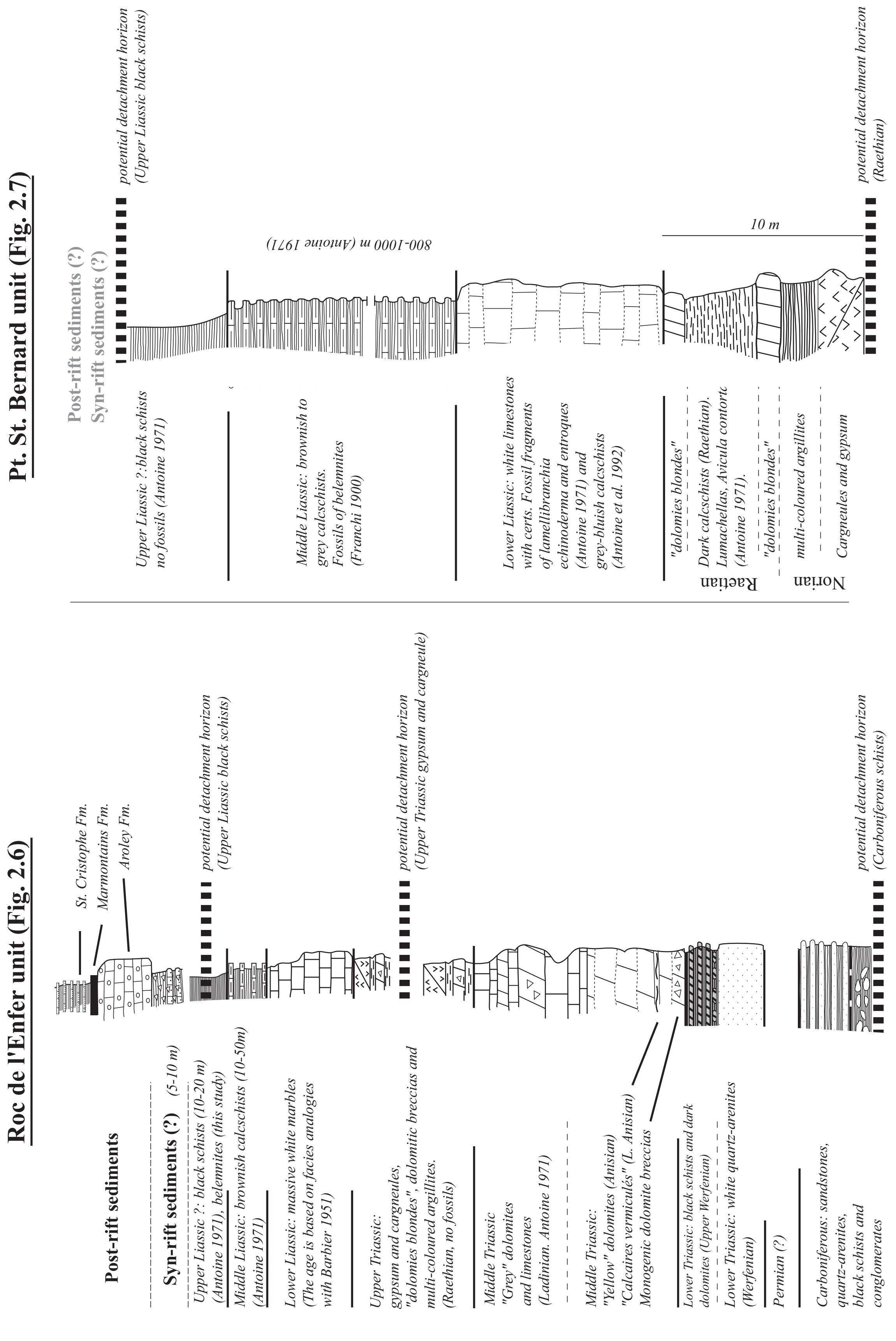


Fig. 2.6: Schematic stratigraphic section of the pre-rift sediments of the Roc de L'Enfer unit. Modified after Antoine (1971, *et al.* 1992). Fig. 2.7: Schematic stratigraphic section of the Pt. St. Bernard unit after Antoine (1971, *et al* 1992) and Bucher 1999.

2.2.1: Carboniferous.

Except for small relics of continental basement preserved near Moûtiers (S of the study area), namely the Hautecour crystalline (Antoine 1971), or at the Pt. Rossa massif, the Carboniferous rocks are the oldest lithologies exposed within the study area. They consist of light grey sandstones, quartz-arenites, grey to black schists and rare conglomerates. All together, these lithologies are commonly referred as the "Houiller". The repetition of grey to black schists and sandstones represents the most common facies of the Carboniferous. In the field the Carboniferous black schists may be confused with similar sediments in the Marmontains Fm. or parts of the Complexe Antéflysch at a first glance. In these cases the attribution to the Carboniferous is based on the following criteria: (1) the presence of a millimetric (1-2 mm) detrital white mica in the black schists, (2) the repetition between black schists and sandstones and (3) the absence of carbonates. Carboniferous rocks generally do not react to hydrochloric acid. In the Moûtiers and Roc de l'Enfer units, the Carboniferous represents a preferred detachment level, and therefore its thickness is unknown. Also, the Carboniferous is locally seen in direct stratigraphic contact with the post-rift formations, such as the Roc de l'Enfer unit (Pt. de la Terrasse, see Ch. 6).

2.2.2: Permian.

Typical Permian rocks can be observed in the Combe de la Nova, Pt. de Mya and Pyramides Calcaires areas. In these regions the Permian is seen in stratigraphic continuity with the older Carboniferous strata and the younger Lower Triassic white quartz-arenite. Generally, the lithologies of the Permian show a massive aspect and greenish colour. The Permian contains red and green conglomerates and quartz-arenites. Pinkish to white (rarely green) quartz clasts, millimetre to centimetre sized, constitute the main grains in sandstones and conglomerates. They are enclosed in a groundmass of similar compositional origin: mainly fine-grained quartz, albite and sericite. Conglomerates are found at different levels in the Permian beds. These levels are good marker horizons for the regional mapping of bedding (S0). Commonly, the Permian rocks form the core of antiformal anticlines, surrounded by light-coloured younger formations (especially the yellowish dolomite of the Triassic). Today, the thickness of this formation is hundreds of meters (500 m at the Combe de la Nova and 200-300 m at the Pt. de Mya). However, considering intense folding, the original thickness should not have exceed 100-150 m. The boundary between Permian and Carboniferous is gradational everywhere. In some places (Combe de la Nova) the base of the Permian consists of some intensely foliated green to white quartz-arenites which gradually pass into the Carboniferous black schists. In such cases, the boundary is difficult to map (Fig. 2.5). Notably within the Roc de l'Enfer unit, the Permian is reduced to only 1 m of thickness (Antoine 1971).

2.2.3: Lower-Triassic.

Complete stratigraphic sections from the Permian to the Middle Triassic limestones and dolomites can be observed within both the Moûtiers and Roc de l'Enfer units (Antoine 1971). In the Moûtiers unit, significant examples are located at the Combe de la Nova (SW of Les Chapieux, x 939700 m; y 2085100 m; z 2180 m, see Fig. 2.8), N of the Pt. de Mya (Antoine 1971 and at x 942675 m; y 2089950; z 2400) and at the Pyramides Calcaires (Elter G. 1960, Elter G. and P. 1965, Antoine 1971). The Lower Triassic formation starts with homogeneous massive quartz-arenites (the basal quartz-arenites), light grey to white in colour. They are between a few meters (2-10 m) and some tens (30-50 m) of meters in thickness. The basal white quartz-arenites show strong affinities with similar late-Permian lithologies and are classically considered Werfenian in age (Antoine 1971). The overlying sediments, limited in thickness to a few meters (5-15 m), clearly represent the transition from the quartz-arenites to Middle Triassic limestones and dolomites. The transition starts with black schists interbedded with rare strata of a dark dolomite with a reddish weathered surface. Further up section a distinct sedimentary bedding (an alternation of dark dolomites and black schists), is well defined and completes the transition to the dolomitic Middle Triassic lithologies (Fig. 2.8). Based on the stratigraphic continuity between quartz-arenites and Middle Triassic carbonates, a late-Werfenian age may be attributed to the black schists, including the dark dolomites ("schistes suprawerféniens" of Antoine 1971). However the white quartz-arenites at the base and the black schists, show lateral variation in thickness. In some cases the Middle Triassic carbonates have been observed in direct sedimentary contact with the Permian, suggesting local gaps in sedimentation during the Lower Trias (Fig. 2.9).

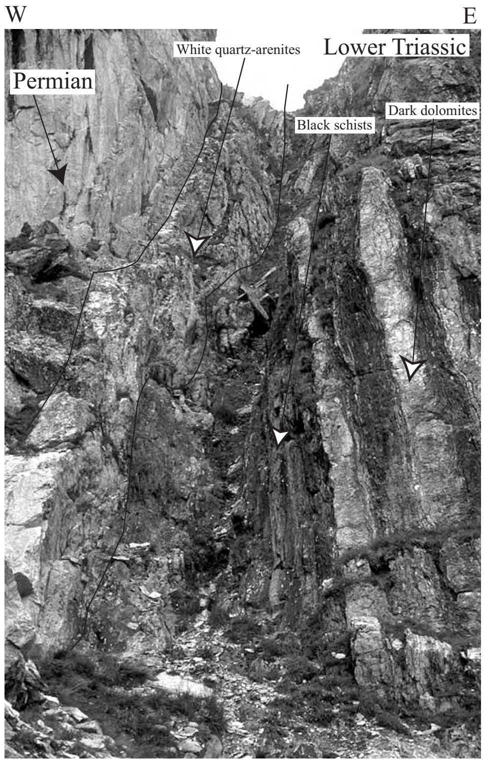


Fig. 2.8: Complete transition from Permian conglomerates to Lower Triassic quartz-arenites. In the same outcrop, Middle Triassic dolomite follows in continuity eastward and outside the photograph (Combe de la Nova, Moûtiers unit, external Valaisan, coordinates in Ch.2.2).

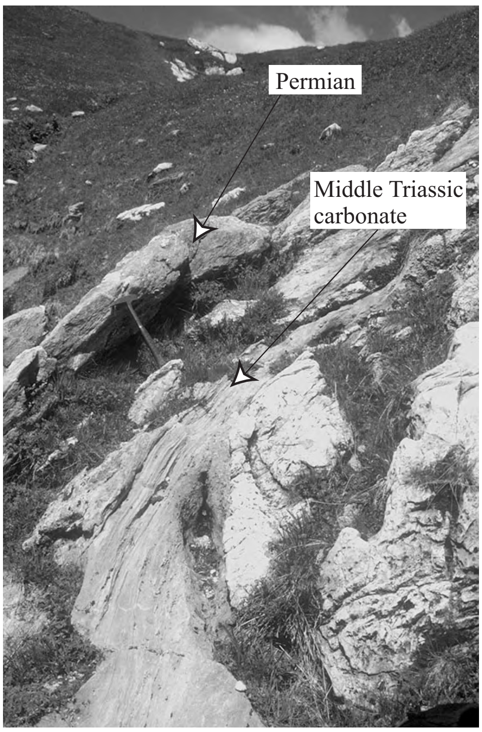


Fig. 2.9: Stratigraphic contact between Permian conglomerates and Middle Triassic carbonates. In this example the transition is incomplete and the Lower Triassic is missing (Pt. de Mya, Moûtiers unit, external Valaisan, coordinates in Ch.2.2).

2.2.4: Middle Triassic.

Two different members characterise the Middle Triassic carbonates. The lower member ("yellow" member) consists of brownish to yellowish weathered dolomites, and it comprises a series of well-bedded limestones. The "Calcaires vermiculés" are a distinctive bed within this member (Antoine 1971) constituted by a dark grey limestone locally showing abundant traces of bioturbation. This facies is also found in the Briançonnais series where it has been dated as Lower Anisian (Debelmas and Lemoine, 1961; Ellemberger, 1958). Therefore this lower "yellow" member may be of Anisian age. A second member of the Middle Triassic formation consists of greyish weathering limestones and dolomites ("grey member"). Bedding is thicker and well expressed, compared to the lower "yellow" member. Characteristic facies types of this "grey" member are monogenic breccia horizons (Fig. 2.10) which exclusively contain angular dolomite clasts in a matrix of similar origin. The dolomitic matrix is diagnostic for distinguishing this facies from younger syn-rift sediments, which are characterised by a calcareous matrix. Based on a lithostratigraphic correlation with the Briançonnais domain this younger "grey" member within the Middle Triassic may be correlated with the Ladinian stage (Antoine, 1971; Ellemberger, 1958).

Fig. 2.10 Middle Triassic (Ladinian) monogenic breccia. The suite of the clasts ranges between Lower (dark grey to black) to Middle Triassic dolomites (light grey). The occurrence of this facies within the Middle Triassic sequence of the external Valaisan can be interpreted as the result of a moderate rifting activity. (Moûtiers unit, external Valaisan)

2.2.5: Upper Triassic.

A complete stratigraphic succession from Middle Triassic to Upper Triassic was observed rarely (Antoine, 1971). This is due to the fact that the Upper Triassic, mainly formed by gypsum and cargneule, represents a potential detachment horizon exposed along tectonic contacts. However, in a few cases Upper Triassic lithologies were observed in stratigraphic continuity with Lower Liassic limestones. These stratigraphic sections can be found along the W side of the Crêt Baudin (x 942250 m; y 2086550; z 1620, see Fig 2.11) and at the pass between the two Pyramides Calcaires (Val Veni, Italy). At the Pyramides Calcaires, the uppermost lithology attributable to the Upper Triassic is a series of yellow dolomitic breccias contained in a matrix of similar origin (Antoine, 1971). Zulauf (1964) and Elter and Elter (1965) assumed a stratigraphic continuity with the well dated Lower Liassic limestone, and they interpreted these breccias as Rhaetian in age. At the Crêt Baudin, a series of multi-coloured argillites ("pélites vertes") and yellow to brownish weathering dolomite ("dolomites blondes") is found to belong to the Upper

Fig. 2.11 Rare occurrence of a stratigraphic contact between a massif grey Lower Liassic limestone (left) and the Upper Triassic evaporites and yellow dolomite (right). The general absence of the Upper Triassic within the external Valaisan is due to downcutting erosion related to the later opening of the Valais ocean.

Triassic. At this outcrop the uppermost Upper Triassic beds consist, as in the case of the Pyramides Calcaires, of yellow dolomitic breccias (Fudral, 1998). In both examples, the attribution of the uppermost part of the Upper Triassic to the Rhaetian can be put into doubt because of the absence of mollusc fossils diagnostic for the Rhaetian (i.e. Avicula contorta is the most typical). Antoine (1971) interprets the dolomitic breccia as representative for periods of condensed sedimentation (emersion?) during the Upper Triassic, responsible for the absence of Rhaetian fossils. Within the paleogeographically more internal Roc de l'Enfer unit, in the valley of Charbonnet, Antoine (1971) described a continuous succession from the Middle Triassic to the Lower Liassic (Fig. 2.6). In this example, the Upper Triassic comprises: black schists, breccia with dark dolomitic components, "dolomites blondes" and grey calcschists Antoine 1971. According to Antoine the presence of breccia with components of dark dolomite confirms the Rhaetian age for the uppermost Upper Triassic. Classical, Rhaetian facies can be observed at the base of the Pt. St. Bernard unit only (see Ch. 2.2.8).

2.2.6: Liassic.

Within the Moûtiers unit, there are two major key outcrops exposing the Liassic lithologies: one located at the Crêt Baudin near Les Chapieux, and another one at the Pyramides Calcaires. These outcrops have been studied in detail by Schöller (1929), Barbier (1951a), Elter (1951,1954), Elter and Elter (1965), Antoine (1971) and Fudral (1973, 1998). At the Pyramides Calcaires a grey to bluish marble, overlying the Upper Triassic dolomitic breccias, shows a light grey to white patina. This limestone displays considerable lateral variations in thickness, ranging from more than 100 m in its central portion to a few meters towards SW. Also at the Crêt Baudin the same marble stratigraphically overlies Upper Triassic lithologies. The attribution to the Liassic epoch is based on the following criteria: 1) both outcrops show evidence for a stratigraphic transition to Upper Triassic sequences beneath; 2) at the Pyramides Calcaires, Barbier (Barbier, 1951b) described the occurrences of belemnites and ammonites, 3) a similar lithology in the region of Etroits du Saix near Moûtiers, has been attributed to Lower and Middle Liassic by the discovery of a large ammonite (*Arietites* ?) by Barbier (Barbier, 1951a). In general the white to grey marbles lack diagnostic fossils and are lithostratigraphically correlated with the limestones at Etroits du Saix (also referred to as the "Lias de Tarentaise", Antoine *et al* 1992). Hence, the limestone at the Pyramides Calcaires and at the Crêt Baudin (Moûtiers unit) is attributed to the Lower and Middle Liassic (Antoine, 1971).

Within the Roc de l'Enfer unit (Fig. 2.6) an almost complete Liassic sequence is found in the Valleé des Charbonnet. This sequence comprises, above some remnants of Upper-Triassic evaporites, the same grey-bluish marble (Lower Liassic, Antoine 1971) observed in the Moûtiers unit. However in this unit the limestone does not show evidence for emersion. Upwards the limestone is overlain by a 10-20 m thick sequence of calcschists (alternation of impure limestones and pelitic levels) which contain belemnites (Middle Liassic, Antoine 1971). This lithology passes upwards into a sequence of black schists whose thickness is unknown, due to tectonic shearing. These black schists are indicative for the Upper Liassic. A complete section of the Liassic sequence within the Roc de l'Enfer unit can also be observed along the Torrent du Charbonnet near Bourg St. Maurice. In this locality the new finding of a fossiliferous horizon within the black schists, revealing abundant rests of belemnites and ammonites probably correlates this formation to the Liassic epoch (Fig. 2.12).

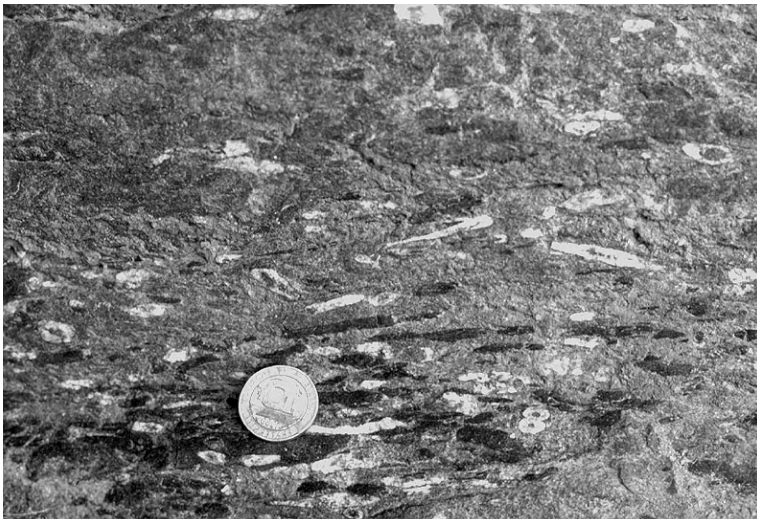


Fig. 2.12 Fossiliferous horizon at the base of Upper Liassic black schists, Roc de l'Enfer unit, external Valaisan (Plan Andre thrust sheet, Ch. 6), Torrent le Charbonnet, loc. Le Bourgeat. (945225 x; 279900 y; 900 z). In this example several sections of belemnites are recognizable.

2.2.7: Dogger.

The existence of lithologies of Middle Jurassic age within the external Valaisan has long been a matter of discussion. Dogger (such as Malm) was generally considered to be absent. The only evidence for the presence of an oolitic limestone referred to as Mid-Jurassic in age, was observed in the form of rare clasts in the Early Cretaceous Aroley Fm. (Antoine 1971). Unfortunately, during this study no evidence for the existence of such clasts was found. We cannot confirm nor reject the observation of Antoine (1971).

2.2.8: The Pt. St. Bernard unit and the paleogeography of the pre-rift sediments of the external Valaisan

The Pt. St. Bernard unit is composed of an incomplete stratigraphic succession, which consists mainly of Middle Liassic calcschists (Fig. 2.7).

An important detachment horizon is present at the base of the unit, often cutting out parts of the Upper Triassic series. Antoine (1971) describes the occurrence of cargneule and gypsum associated with Upper Triassic dolomite ("dolomies blondes")

at the base of the Pt. St. Bernard unit. Other lithologies belonging to the Upper Triassic are black schists with fossils of *Avicula contorta* of Rhaetian age (Antoine 1971) and multi-coloured argillites (Bucher 1999). The Upper Triassic sequence is of limited thickness (10 m). Upwards section, the Lower Liassic starts with a bluish limestone which passes towards a light grey limestone with cherts nodules (Lower Liassic). In stratigraphic continuity follows a thick sequence of calcschists (800-1000 m, Antoine 1971) which consists in a regular centimetre-scale alternation of dark grey to brownish impure limestones and pelitic levels. This sequence is interpreted to be of Middle Liassic age because of the occurrence of belemnites (Antoine 1971, Franchi 1900). The youngest lithology of the Pt. St. Bernard unit is a series of black schists. Their attribution to the Upper Liassic (Antoine et al. 1992) is based on lithostratigraphic evidence only. The Liassic stratigraphic record of the Pt. St. Bernard unit may be correlated with the onset of fault-controlled basin development pre-dating the Valaisan rifting. The great thickness and monotonous lithology of the Middle Liassic series suggest that it was deposited in a slowly subsiding outer shelf and slope environment.

Due to the lack of sediments younger than the Late Liassic black schists, previous workers did not attempt to make direct facies correlations between the Pt. St. Bernard unit and the Valaisan units. Therefore the paleogeographic attribution of this unit has been derived from its structural position. In all previous works (based on Elter and Elter 1957) the Pt. St. Bernard unit was considered as a slice of sedimentary rocks pinched between the Valaisan units (the Versoyen unit) below and the Houillèr Zone above. This tectonic position has been considered to indicate that the Pt. St. Bernard unit is part from the Subbriançonnais domain.

However, this structural argument is invalidated by the new findings presented in Chapter 5.5 (and schematically depicted in Fig 5.5) which show that the present–day position of the Pt. St. Bernard unit is mainly due to regional scale folding (D1 and D2). These structural arguments strongly argue for a paleogeographic derivation of the Pt. St. Bernard unit from the Valais domain. Hence, the stratigraphic record of the Pt. St. Bernard unit, only containing pre-rift sediments deposited onto a continental margin, is ascribed to the external Valaisan. In this case, however, and in analogy with all other Valaisan units, a former presence of a post-rift sequence on top of the preserved Liassic needs to be postulated. This younger cover must have been detached along a potential horizon located within the Upper Liassic black schist formation (Fig 2.7).

In reality, the question after the paleogeographic origin of the Pt. St. Bernard unit may largely be a semantic one. The Subbriançonnais paleogeographic domain individualized on the external margin of the Briançonnais domain and from Middle to Upper Liassic times onwards associated with the mid-Jurassic opening of the Piemont-Ligurian ocean. This domain has been defined by Barbier (1948) in a region located far S of our study area (see also Fügenschuh et al. 1999). The Valais paleogeographic domain (external and internal Valaisan) is also situated in a region located on the external margin of the Briançonnais domain, i.e. between the Briançonnais and the European margin. However, the Valaisan as s distinct paleogeographical domain individualized only later and in response to the Early Cretaceous opening of the Valais ocean. Therefore, in Middle to Late Liassic times no Valais paleogeographic domain existed yet. In other words, the stratigraphic information coming from the Liassic sediments of the Pt. St. Bernard unit alone is not diagnostic for its attribution to either the Valais or Subbriançonnais domain.

Although the structural arguments discussed in the work of Elter and Elter (1957) are invalidated, we agree with these authors on the observation that the Pt. St. Bernard unit shows a stratigraphic record which can easily be correlated with Subbriançonnais units (Gran Moëndaz and Perron des Encombres units of Barbier 1948). The thick Liassic sequence of the Pt. St. Bernard unit, and the presence of Upper Triassic and Upper Liassic detachment horizons are indeed characteristic features of the Subbriançonnais of the Gran Moëndaz unit (Ceriani et al., 2000). However, it is postulated, is that the Valais ocean opened later (in the Cretaceous) and within a paleogeographic region characteristic for the previously (Jurassic) individualized Subbriançonnais domain. We assume that the Pt. St. Bernard unit preserved its Liassic sequence from the syn- and post- rift erosion characteristic for the rest of the Valaisan because it already represented a graben structure where erosion could not affect the Liassic beds. Facies variations in the Liassic of the external Valaisan, can be observed from W to E. In fact, the Liassic pre-rift sediments (pre-rift is in respect to the Valaisan rifting) of the entire external Valaisan (Moûtiers, Roc de l'Enfer and Pt. St. Bernard units) show a thickening trend from more external (W) to more internal (E) units. This trend is related to the opening of the Piemont-Ligurian ocean.

In conclusion, in the external Valaisan domain we observe the superposition of two rifting phases (a rift in rift!). The pre-rift sediments of the external Valaisan were deposited in the Subbriançonnais paleogeographic domain, the term “pre-rift” relating to the opening of the Valais ocean. During a second rifting event (Upper Jurassic-Lower Cretaceous) the Subbriançonnais domain individualized into internal Valaisan (the Versoyen unit) and external Valaisan (Moûtiers, Roc de l’Enfer and Pt. St. Bernard unit). The external Valaisan experienced extension and erosion, together with the deposition of the reworked material (syn-rift formation). Later on, (post-Barrêmian), external and internal Valais basins were filled with post-rift sediments. In consequence, only the Versoyen unit represents a “true” Valaisan unit in the sense of an oceanic domain, while the external Valaisan may be interpreted either as Subbriançonnais or Valaisan, depending on whether one refers to the opening of the Piemont-Liguria ocean or to the opening of the Valaisan ocean. According to this model, the external Valaisan may be considered as a part of the Subbriançonnais which is found in a more external position with respect to the subsequently formed Valaisan ocean.

It is interesting to note that the Subbriançonnais defined by Barbier south of our study area is non-metamorphic, while the external Valaisan was metamorphosed with an increasing grade from the Moûtiers to the Pt. St. Bernard unit (see Ch.4). The present analysis (Ch. 4 to 6) indicates out a subduction scenario which is common to both external and internal Valaisan units. The non-metamorphic character of the Subbriançonnais domain S of the study area indicates that this domain remained at shallower structural levels during the subduction of the Valais domain. Therefore, one may attribute all units affected by subduction to the Valaisan, while the Subbriançonnais (together with the Briançonnais) remained within an upper plate position.

2.3: The syn-rift sediments

While syn rift sediments will be extensively discussed in Ch. 3, this section (Ch.2.3) only briefly discusses the main subdivisions and differences in interpretation in respect to previous work. The rifting phase in the external Valaisan, preceding the opening of the Valais ocean, is recorded by sedimentation of the “Brèches du Grand Fond” (Antoine et al., 1972). These syn-rift sediments are bracketed between two unconformities: one at the base of the syn-rift sediments, dated as post-Middle Liassic, and, one at the top and in contact to the post-rift sediments dated as pre-Barrêmian. Hence, the age of these syn-rift sediments is bracketed between Upper Liassic and Barrêmian.

Therefore, it is possible that the "Brèches du Grand Fond" record both rifting events: the "Piemont-Liguria" and the "Valaisan" rifting events. Within the Brèches du Grand Fond Group two distinct lithostratigraphic formations can be defined (Fig. 2.13).

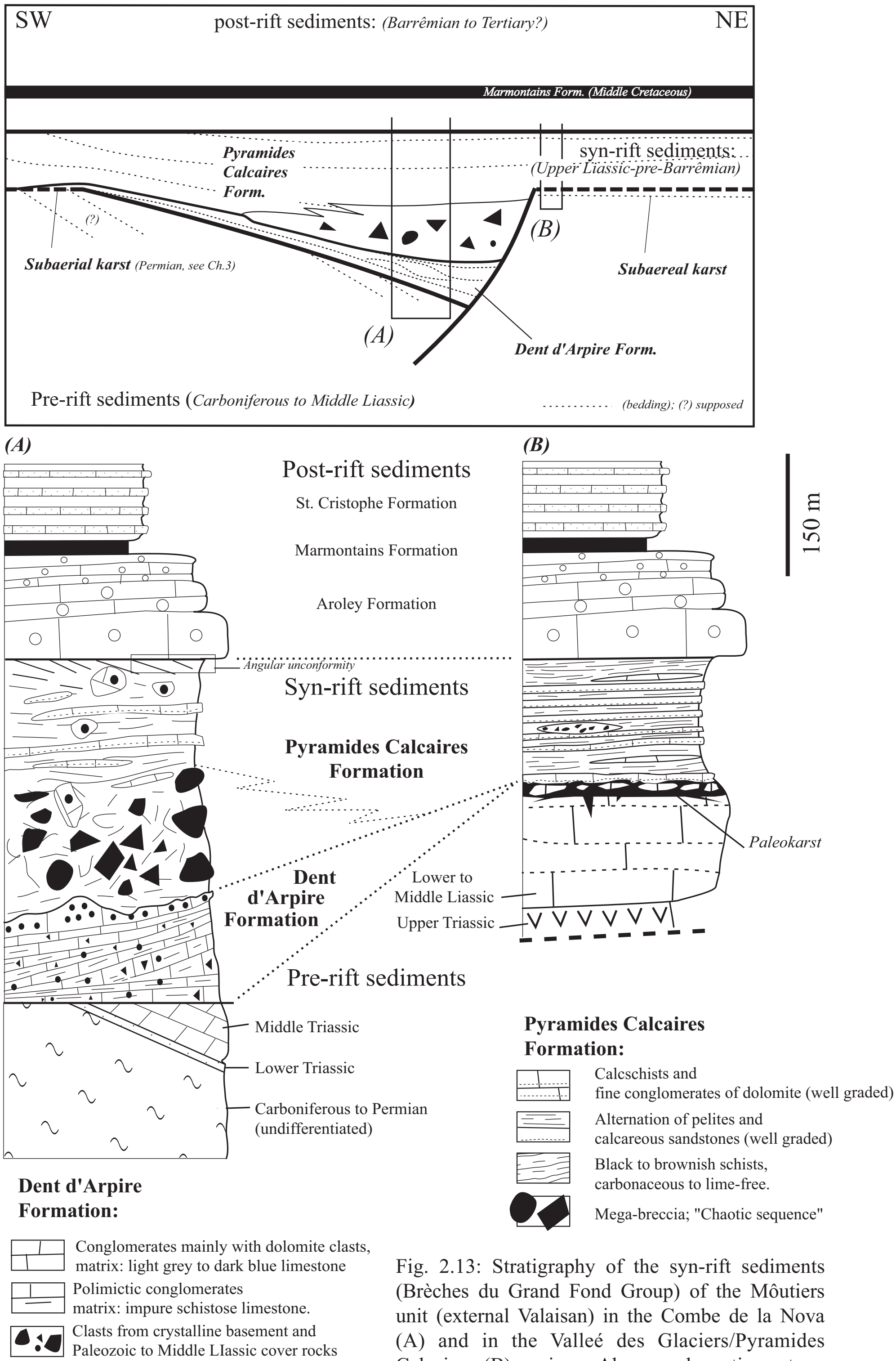
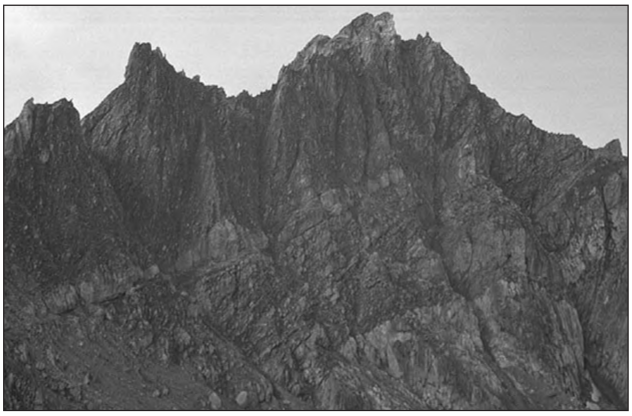
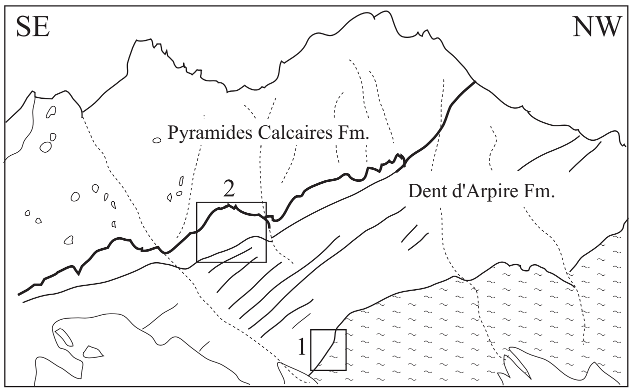


Fig. 2.13: Stratigraphy of the syn-rift sediments (Brèches du Grand Fond Group) of the Môutiers unit (external Valaisan) in the Combe de la Nova (A) and in the Valleé des Glaciers/Pyramides Calcaires (B) regions. Above: schematic cartoon showing the correlation between the Dent d'Arpire and the Pyramides Calcaires Formations and the possible geometry of the depositional setting. Further explanations in Ch. 3.

A basal sequence, referred to as the "Dent d'Arpire Fm.", discordantly overlies Permian to Middle Triassic pre-rift sediments. The second and younger one is the "Pyramides Calcaires Fm." which either unconformably overlies the Dent d'Arpire Fm. (such as in the Combe de la Nova area) or discordantly rests on Lower to Middle Liassic pre-rift sediments (such as in the high Valleé des Glaciers and Pyramides Calcaires areas). The Dent d'Arpire Fm. consists of series of well bedded matrix-supported polymictic breccias. The lowermost beds show a marly matrix, while upwards the matrix gradually becomes pure limestone. The uppermost beds of the Dent d'Arpire Fm. consist of a massive thick conglomerate with mainly dolomite clasts in a light grey to bluish limestone matrix. The top surface is irregular, probably due to the erosion preceding (or accompanying) the deposition of the younger Pyramides Calcaires Fm. (Fig. 2.14).



A



B

Fig. 2.14 Detail of the W side of the peak quoted 2620 m in the Combe de la Nova area which shows the subdivision of the Brèche du Grand Fond Group (syn-rift sediments, of the Moûtiers unit external Valaisan) into two lithostratigraphic formations: the Dent d'Arpire Fm. (at the base) and the Pyramides Calcaires Fm. (above). In this example the Dent d'Arpire Fm. is in depositional contact with the Permian conglomerates (inset 1). The contact between the two formations (inset 2) is interpreted as an erosional surface produced at the expense of the (old) Dent d'Arpire Fm. during the deposition of the (younger) Pyramides Calcaires Formation. Further explanations in text (Ch. 3).

The Pyramides Calcaires Fm. shows a markedly schistose character. The most abundant facies types are brownish calcschists and a characteristic alternance between black shales and fine calcareous sandstones. The calcschists contain fine-conglomerates in which Middle Triassic dolomite clasts are predominant. Another typical facies is a "chaotic" sequence found in the Combe de la Nova area and in which the original bedding is characteristically not preserved, the size of clasts ranging from a few centimetres to several meters (10-30m). In the Combe de la Nova area, where both formations are exposed, the base of the Pyramides Calcaires Fm. contains large well-rounded clasts of metric size (1-2 m), resulting from the reworking of the lower Dent d'Arpire Fm.. In general, the Brèches du Grand Fond Group is characterised by clasts, which are closely comparable with the lithologies forming the pre-rift sediments (Carboniferous to Middle Liassic) and, additionally, crystalline basement rocks.

Previous work, in the Combe de la Nova region referred to the Dent d'Arpire and the Pyramides Calcaires Formations as "ensemble conglomératique inférieur et supérieur" (Antoine et al. 1972, Fudral 1973). The Pyramides Calcaires Fm. in the Valleé des Glaciers and Pyramides Calcaires area was previously described as "Complexe Antéflysch and attributed to the Roignais-Versoyen unit by Antoine (1971). Antoine et al. (1972) suggested a Callovo-Oxfordian age for the "ensemble conglomératique inférieur et supérieur". New ammonite fossils found in the lower part of the Dent d'Arpire Fm. indicate an Late Liassic (Toarcian) age for this formation (see Chapter 3.1.2) while the Pyramides Calcaires Formation remained undated.

The syn-rift sediments of the internal Valaisan (Versoyen unit) are represented by the Brèches du Collet des Rousses Fm. and by the Complexe Antéflysch Fm. (see Fig. 2.2) .The Brèches du Collet des Rousses Fm. is made up by coarse conglomerates containing Lower Liassic clasts within a calcareous matrix. Upwards, this formation passes into a regular sequence of calcschists (see Ch. 2.4.4). The age of this formation was proposed to be Late Liassic to Dogger (Antoine, 1971), however the upper age limit is inferred without palaeontological data. Hence, its age is between Late Liassic and -Barrêmian.

In the sense of Antoine (1971) the "Complexe Antéflysch" comprises all the lithologies found between the pre-rift-sediments and the base of the post-rift Aroley Fm. Later, Antoine et al. (1972) treated a part of this formation as an independent lithostratigraphic unit referred to as the "Brèches du Grand Fond" of the Moûtiers unit. Consequently, the Complexe Antéflysch was then restricted to the "Roignais-Versoyen" unit (Antoine, 1972). Within this unit, the Complexe Antéflysch was subdivided into two distinct facies intervals. One corresponds to the "black schists" associated with the basaltic volcanic rocks (our Complexe Antéflysch), while the other consists of brownish calcareous schists. The calcareous schists contain predominantly fine conglomerates with dolomitic clasts, but they never contain basaltic rocks or serpentinites.

Both these facies types have never been observed together within the same outcrop. On the contrary, according to Antoine (1971), their distinct facies suggest that they were deposited in two different basin settings. The black schists and associated basaltic volcanic rocks are restricted to the SE part of the study area where they form the bulk of the Versoyen unit.

The calcareous schists with fine conglomerates form a continuous narrow band which can be followed from Les Chapieux to the Pyramides Calcaires in the higher Valleé des Glaciers, (Antoine, 1971; Mennessier et al., 1976) situated in the NW part of the study area. Moreover, the calcareous schists with fine conglomerates stratigraphically overlie pre-rift sediments, while the black schists and volcanic rocks show a stratigraphic contact with the Brèches du Collet des Rousses Fm. or with the Pt. Rossa Complex. According to the distinction between external and internal Valaisan, we interpret that part of Antoine's Complexe Antéflysch containing basaltic volcanic rocks to have been deposited onto newly formed oceanic crust (internal Valaisan) while the calcareous schists containing fine conglomerates, overlying the pre-rift sediments, are interpreted to have been deposited on continental crust (external Valaisan). For this reason we redefined Antoine's "Complexe Antéflysch" by restricting this term to the sediments containing basaltic volcanic rocks (including the Pt. Rossa Complex). The brownish calcareous schists, however, are considered as part of the syn-rift Pyramides Calcaires Fm. (see also Ch. 2.3).

2.4: Lithologies characteristic for the internal Valaisan

2.4.1: The Complexe Antéflysch Fm. of the Versoyen unit

The Complexe Antéflysch Fm., as redefined in the previous chapters (Ch. 2.2 and 2.3), has two main occurrences in the SE part of the study area (Figs. 2.1 and 2.3): (1) NW of Bourg St. Maurice and (2) between Bourg St. Maurice in the SW, all the way up to near La Thuille village (I) in the NE. Further SE the Complexe Antéflysch Fm. wedges out under the Roc de l'Enfer unit, for tectonic reasons (together with the Versoyen unit).

The lithologic contrast between the dark green to black coloured rocks of the Complexe Antéflysch with the light coloured post-rift sediments is probably one of the most evident optical features of the study area (Fig. 2.15). The volcano-sedimentary sequence is characterised by a sub-horizontal alternation of basaltic volcanic rocks (sills), intruded into dark coloured sediments (black schists). The apparent thickness of the Complexe Antéflysch is estimated to range between 200 and 450 m. However, due to intense Alpine deformation and isoclinal folding the original thickness is unknown. Stratigraphically, this sequence is included between the Brèches du Collet de Rousses Fm. beneath and the post-rift sediment (Aroley Fm) at the top. The Complex Antéflysch Fm. in its present day position is inverted in most places due to large scale folding (D2) and moderately dips towards SE. Loubat (1968) and Antoine (1971) were the first to recognise inverted bedding in the Complexe Antéflysch. Their polarity criteria were based on different macroscopic and microscopic trends recorded in the sills and the black schists. Moreover, Antoine (1971) also demonstrated that the same younging direction was also observable in the post-rift sediments (Aroley Fm).

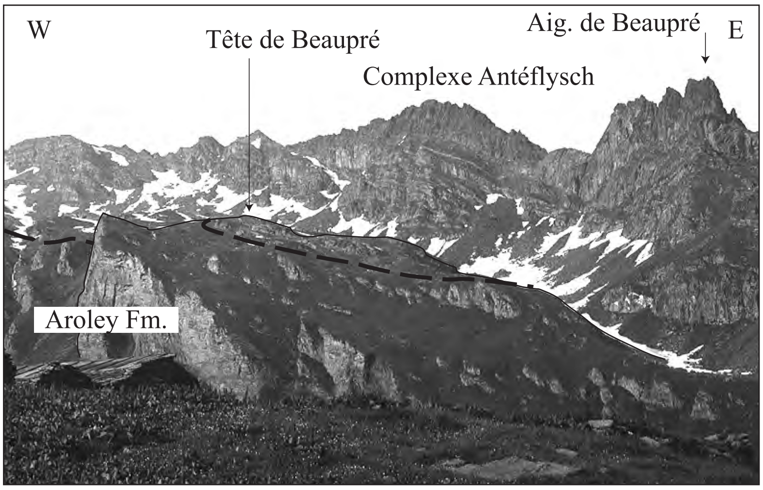


Fig. 2.15: Panoramic view of the overturned series W of the Aiguilles de Beaupré, Valleé du Versoyen (Versoyen unit internal Valaisan). The dark coloured strata in the background are black schists while the interbedded massive lighter levels are sills. Lower part of photo is occupied by the Aroley Formation (post rift sediments).

Despite of HP-LT metamorphic overprint and intense Alpine deformation, it is still possible to distinguish different types of basaltic rocks. However, in the field they are difficult to be recognised. Based on observed magmatic and metamorphic mineral assemblages, several varieties of volcanic rocks and sediments were described by detailed petrographic studies (Loubat 1968, Schürch 1987). From the stratigraphic bottom to the top of the Complexe Antéflysch Fm. the following volcanic rocks were recognized: meta-gabbros, meta-basalts, pillow lavas and volcanic breccias (the Brèches du Miravidi of Loubat 1968, Antoine 1971). For a detailed petrographic study of the volcano-sedimentary sequence of the Versoyen unit the reader is also referred to Lasserre and Lavergne (1976), Cannic (1996), Bousquet (1998) and Dalla Torre (1998). A review of the Alpine metamorphic overprint is presented in Ch. 4. By the help of magmatic textures within the Complexe Antéflysch, Cannic (1996) and Cannic et al. (1996) were able to carry out polarity criteria within the sills in order to decipher the complex folding. However, in the field isoclinal folding is evident and "classical" geological cross sections can easily be constructed (Dalla Torre 1998, see also Ch. 5, Fig. 5.14).

The following characteristics of the Complexe Antéflysch Fm. (Fig. 2.16) can be observed along a profile from the Aig. de l'Ermite (E) to the Aig. de Beaupré (W) (Fig. 2.17).

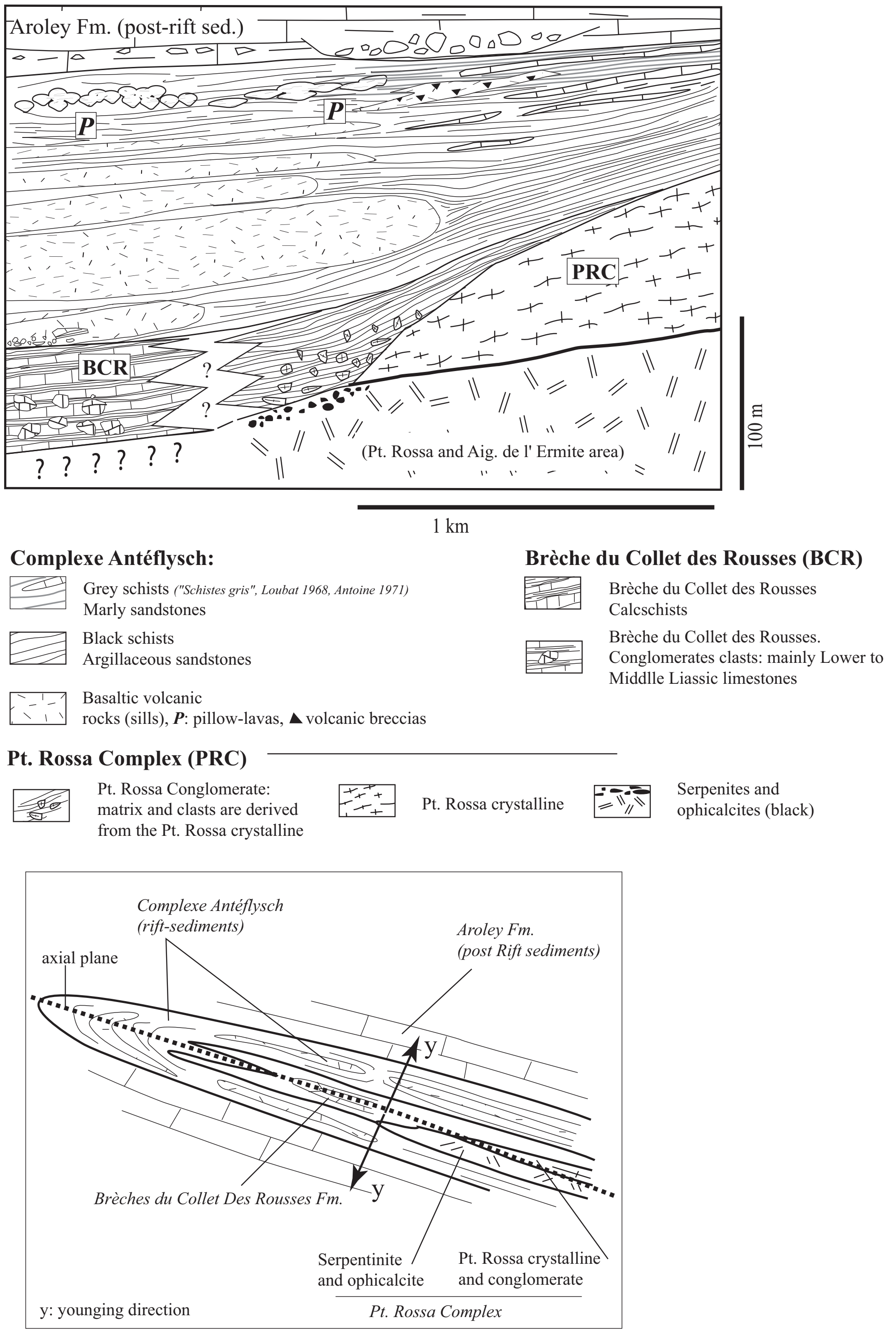


Fig. 2.16 Above: Schematic cartoon showing the inferred primary relationschips between the Complexe Antéflysch Fm. (including the Brèches du Collet des Rousses Fm.) and the Pt. Rossa Complex. B below: Based on this scheme, when large scale isoclinal folding develops, the Brèches du Collet des Rousses and the Pt. Rossa Complex form the core of this structure (see Ch. 5). For sedimentological features within the Aroley Fm. see Ch. 2.5.

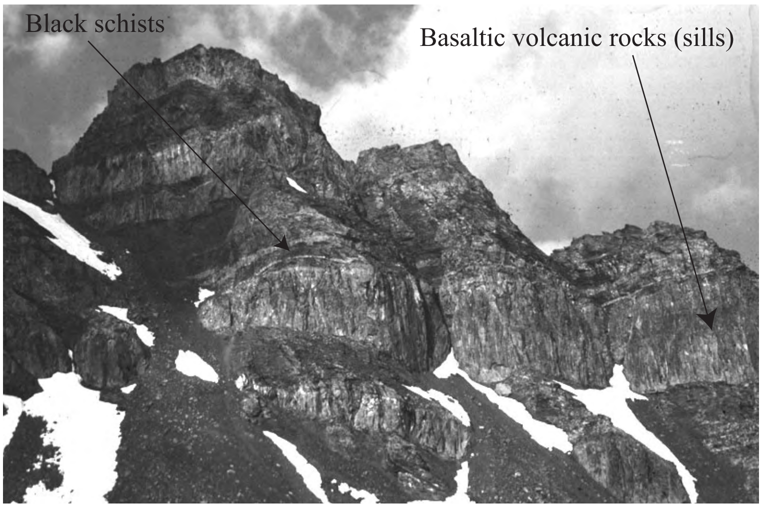


Fig. 2.17) The Aig. de l'Ermite, high Vallon de Beaupré, Complexe Antéflysch Fm. (Versoyen unit, internal Valaisan). The lowermost sills (lower right corner of photograph) are close to the contact with the Brèches du Collet des Rousses formation.

The sills, about 20 in numbers (Loubat 1984), decrease in thickness from the stratigraphic bottom (more than 50 m thick) toward the top of the sequence (1-10 m) (Loubat, 1984; Schürch, 1987). Laterally, the thickness of the basaltic sills remains constant for individual sills and they can be traced over long distances (1 km by 2 km), offering useful markers for the three dimensional geometry of the folding. Their shape is tabular and sub-parallel to bedding planes. Examples of discordant dykes were not observed.

Meta-gabbro can be observed at the Aig. de l'Ermite in the core of the thickest sills (Cannic et al., 1996; Schürch, 1987). Macroscopically, they are dark green to dark grey, moderately coarse-grained and extremely massive rocks. A reddish weathering is also typical (Schürch, 1987). Microscopically they show relicts of a magmatic assemblage made by pyroxene (augite), hornblende and apatite. Upwards in the sequence, and in correspondence with sills of reduced thickness, the Alpine metamorphic overprint, i.e. mainly the greenschist retrograde stage (D2), dominates over the primary magmatic texture. These sills lack an internal differentiation between fine and coarse grained textures and can be referred as meta-basalts or prasinites. They show a light green colour and generally have a finer grained texture than the meta-gabbro. In the thinnest sills (0.50 to 1 m) a penetrative Alpine schistosity can be recognised. The light green colour is mainly due to chlorite and green amphibole, epidote and albite. Examples of these prasinites are widespread through the whole Complexe Antéflysch. However, they are best exposed in the Aig. the Praina area. Pillow lavas systematically occur close to the contact with the post-rift sediments (Antoine, 1971). Good examples can be studied in the Tormotta area and at the Aig. du Praina further SW (Fig. 2.18) and in the area between the Tête de Beaupré and the Pt. Rossa (Antoine, 1971; Loubat, 1965; Loubat and Antoine, 1965). Schürch (1987) mentions a 60 m. thick succession of pillow-basalts (Pointe Fornet, I). In association with the pillow lavas volcanic breccias are also observed (the Brèches du Miravidi, Fig. 2.19).

Fig. 2.18: Pillow lava from the uppermost sills of the Complexe Antéflysch Fm. (Versoyen unit, internal Valaisan). This outcrop is located at the Chalet de Praina at the base of a well visible cross (1970m).

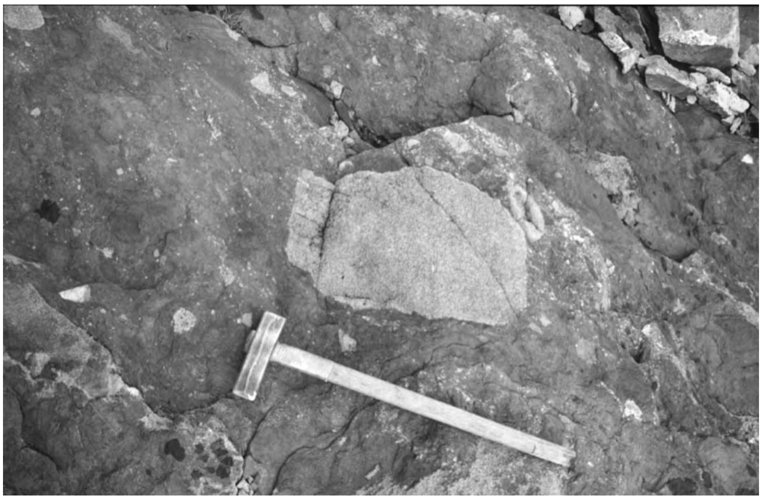


Fig. 2.19: Brèche du Miravidi (Loubat 1968, Antoine 1971). Fragments of basaltic volcanic rocks are enclosed in a matrix of similar composition (NW of the Aig. de Beaupré, qt. 2700).

The best examples of this particular rock type are located SW of the Miravidi peak and in the Pt. Rossa area. Within a fine grained massive dark matrix, green to light-green clasts with a marked angular shape are recognized. These clasts represent fragments of sills and pillow lavas and their largest size ranges between 30 cm and 1 m. In some occurrences the matrix strongly recalls the texture of the sills, while in other cases its dark colour makes it similar to the black schists.

The "black schists" represent the sediments of the Complexe Antéflysch Fm. , intercalated between sills. From stratigraphic bottom to top of the sequence the sediment horizons found between the sills increase in thickness (from 1-5 m to 15-20m), a trend which is the reverse of that observed for the sills. The term "black schists comprises several macroscopically different types. Except for their Alpine metamorphic overprint (see Ch. 4), their original mineralogical composition was formed by white mica, quartz, calcite, chlorite, tourmaline, oxides and organic material. The varying proportions of these minerals control the different facies of the "black schists". The most common facies is a black to dark blue intensely foliated rock. However, in some instances, this rocks is more massive and does not split into thin laminae. Characteristic is a typical metallic glint and a local rust-coloured alteration due to the high content in opaque minerals. This rusty facies can be seen between the lowest sills at the base of the Aig. de l'Ermite. In many places, a regular millimetric to centimetric alternation of light coloured carbonaceous and dark pelitic strata can be observed. Microscopically, the carbonaceous levels are dominated by calcite with subordinate quartz and white mica. The pelitic strata are formed by millimetric alternances of quartz and white mica. However, in the pelitic strata fine grained black to reddish oxides predominate. In other occurrences this alternation is reduced to thin carbonaceous or quartzarenitic lens shape layers, embedded within predominantly pelitic beds.

Up sequence the black schists show a gradual trend towards more carbonaceous schists which, referred to as to "Les schistes gris" (grey schists) by Loubat (1968) and Antoine (1971). However, Loubat (1968) notes that the mineralogical composition of the grey schists does only differ that of the black schists regarding the amount of calcite (Antoine 1971). The grey schists are only several meters thick (10-20 m) and consistently located near the top of the sequence. This observation is interpreted by Antoine as one of several evidences that there is a gradual stratigraphic transition from the Complexe Antéflysch Fm. to the overlying calcareous Aroley Fm. The sedimentary protolith of the black schists and of the grey schists can be defined as an immature silty sandstone and calcareous sandstone respectively. From a geochemical analysis, Antoine (1971) proposes that the sediments of the Complexe Antéflysch Fm. were produced by the erosion of a subaerial domain located far from the depositional site. Hence, these sediments possibly represent a distal facies of the proximal syn-rift sediments in the External Valaisan (Pyramides Calcaires Fm. of the Moûtiers unit, see Ch. 3).

The mode of intrusion of the sills into the sediments was subject to various interpretations (Kelts, 1981; Loubat and Delaloye, 1984; Schürch, 1987). These authors agreed that the volcano-sedimentary sequence is sub-horizontally layered and that dykes have not been observed. The sediments are considered to have been unconsolidated at the time of intrusion (Kelts, 1981; Loubat, 1968; Loubat, 1984; Loubat and Delaloye, 1984; Schürch, 1987). This can be confirmed by additional arguments. Firstly, there is no evidence for xenoliths consisting of black schists within the sills. Secondly, as long as the magmas intrude laterally into soft sediments (as it appear to be the case of the Complexe Antéflysch), the forming of dykes is not necessary. Thirdly, the presence of water within the sediments is documented by the occurrence of hydrothermal alteration and contact metamorphism (Loubat and Delaloye, 1984; Schürch, 1987).

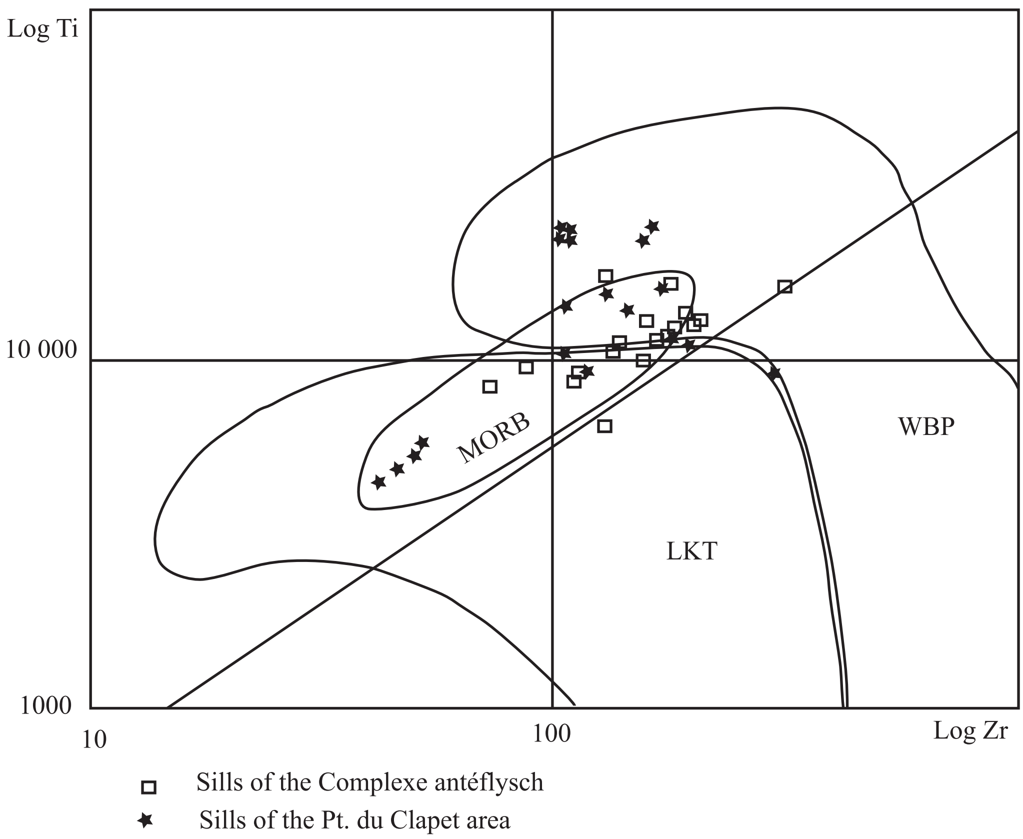


Fig. 2.20. Zr-Ti diagram of Pearce (1980) showing a differentiation of the basaltic sills from the MORB type to the WPB type (after Schürch 1987). Note that the majority of the points cluster are in the MORB field. LKT, Low K-Tholeites; WPB, Within Plate Basalts; MORB Middle Oceanic ridge Basalts.

The last argument, concerns the formation of “adinoles” which form at chilled margins. Such porcelain-like seams consist of albite which formed within the pelitic rocks at the contact with the sills (Loubat and Delaloye, 1984). The sills of the Complexe Antéflysch Fm. show a tholeiitic character (Fig. 2.20) with affinity to middle oceanic ridge basalts (MORB) and with a minor continental (alkaline) affinity which can be explained by assimilation of sedimentary material during their intrusion (Cannic, 1996; Schürch, 1987).

2.4.2: The Pt. Rossa Complex

The Pt. Rossa Complex is situated on the SW side of the Vallon du Breuil, NW of the Pt. St. Bernard pass. The term “Pt. Rossa Complex” comprises different lithologies, which are structurally below the Complexe Antéflysch. This complex contains a mixed continental and oceanic lithological association. In Fig. 2.16 the Pt. Rossa Complex is only very schematically presented. For a more detailed description of the rock types, the reader is referred to Dalla Torre (1998).

The Pt. Rossa Complex contains serpentinites, ophicalcites and meta-gabbros, as well as granites and granite-gneisses. Conglomerates and arkoses which stratigraphically overlie this complex are referred as to as the Pt. Rossa sediments. The serpentinites are dark green with a coarse grained texture. A reddish colour due to weathering is typical. A variety of serpentinite consists of a light green rock made up by chrysotile and talk. The serpentinites were interpreted to have originated from upper mantle material (Schürch, 1987). The ophicalcites consist of a dark green, (white on fresh surfaces), micritic carbonate with fragments of serpentinite. Meta-gabbros often occur in direct contact with serpentinites. The mineral constituents are plagioclase and Mg-Fe chlorite, minor actinolite-tremolite, clinozoisite, epidote and titanite. Lens shaped light-coloured rocks in a chlorite matrix are also interpreted as meta-gabbros. The gneiss and granite rocks are characterised by reddish surface weathering (Pt. Rossa). Microscopically, the granite consists of quartz, plagioclase, white mica and minor chlorite. The gneisses show increased amounts of white mica and chlorite. In the Pt. Rossa massif the granite occupies the core of the Pt. Rossa massif. The gneiss is strongly sheared and forms the outer part of the Pt. Rossa massif (Dalla Torre, 1998). The Pt. Rossa sediments are represented by conglomerate and arkose collectively referred. The conglomerates contain pebbles derived from the Pt. Rossa crystalline (Antoine, 1971; Elter and Elter, 1965). The intervening matrix is an arkosic sandstone. In the upper part of the Pt. Rossa sediments a trend to arkosic sandstones is observed. Hence, from base to top a fining upward trend is evident. The Pt. Rossa sediments were correlated with the Permian conglomerate by (Antoine, 1971). However a Mesozoic age appears more reasonable for two reasons: (1) These sediments seem to represent the lateral equivalent of the Brèches du Collet des Rousses; (2) these sediments grade upwards into the complexe Antéflysch Fm. of which they probably represent the stratigraphic base.

The contact between serpentinites and granites or gneisses is always of tectonic nature (Dalla Torre, 1998). Along such a contact, locally preserved N of the "Lacs de la Tormotta", the granite-gneiss shows a transition into cataclasites, as they approach the serpentinites. The Pt. Rossa sediments (with a thickness of about 50 m), unconformably overlie both serpentinites and the granites-gneisses (Dalla Torre, 1998). This is good evidence for deposition of these sediments after the formation of tectonic contacts between continental and oceanic part of the Pt. Rossa Complex.
The contact between the Pt. Rossa Complex and the overlying Complexe Antéflysch Fm. is difficult to be properly interpreted. Previous work (Antoine, 1971; Dalla Torre, 1998) could not clearly decide on the sedimentary versus tectonic nature of this contact. This ambiguity is caused by the strong Alpine overprint and high competence contrast between the rock-types involved. However, at many localities a stratigraphic transition from the Pt. Rossa Conglomerate into the black schists of the Complexe Antéflysch Fm. could be inferred (Elter and Elter, 1965). The present-day position of the Pt. Rossa Complex is clearly within the Complex Antéflysch.

In conclusion, the tectonic assembly of granites and gneisses with serpentinites and ophicalcites postdates the deposition of the Pt. Rossa sediments. This situation is very similar to that described for the opening of the Valais ocean in the Engadine window by (Florineth and Froitzheim, 1994). The assemblage of continental basement rocks and serpentinites could be interpreted to have occurred along a rift-related extensional detachment which predates the deposition of the Pt. Rossa sediments (Florineth and Froitzheim, 1994; Froitzheim et al., 1996). It is proposed that the Pt. Rossa Complex represents the transition between the continental margin (external Valaisan) and the oceanic domain of the Valaisan ocean. However, it must be noted that further evidence discussed in the following chapter suggests a different depositional setting for the internal Valaisan from that described in Florineth and Froitzheim (1994).

2.4.3: The nature of the contact of the Complexe Antéflysch Fm. with the post-rift sediments

The understanding of the nature of the contact between the Complexe Antéflysch Fm. and the overlying post-rift sediments is of fundamental importance for paleogeographic reconstructions of the Valaisan domain. This contact is well exposed in the study area. It can be easily observed in the Aig. de Praina region (NW of Bourg St. Maurice) and northwards, along the eastern side of Valleé du Versoyen (F) up to the Tête du Chargeur (I), situated N of the Col du Pt. St. Bernard (Fig. 2.1). In this area the post-rift sequence starts with the Aroley Fm. which is composed of "schistose limestones", with local occurrences of fine conglomerates (mainly dolomite clasts). This facies of the Aroley Fm. may easily be confused with the St. Cristophe Fm. (the youngest post-rift formation) and, consequent, incorrect interpretations of this contact (Elter and Elter, 1965). However, the fine conglomerates within the calcschists are very diagnostic for their attribution to the Aroley Fm., and their occurrence was systematically mapped (see Ch. 2.5.1).

The relationships between Complexe Antéflysch Fm. and post-rift sediments were studied in great detail by Antoine (1971) who postulated a stratigraphic transition. Recently, this interpretation was questioned by Cannic (et al. 1996) who postulated a tectonic separation between the "black schists" of the Complexe Antéflysch Fm. and the Aroley Fm. We found no evidences (such as fault rocks, cargneule or gypsum) for such a tectonic contact, and neither did other workers (Fudral, 1998; Fudral and Guillot, 1988).

In the following, we describe a typical stratigraphic section (Fig. 2.21) and summarise sedimentological arguments for a stratigraphic transition such as proposed by Antoine (1971). In the region of the Aig. de Praina, the best example of this contact can easily be observed a few meters N and above the point 1331m below the road between Les Echines dessus (1328 m) and Les Tigny (1450 m) (Fig 2.1). At this locality the Complexe Antéflysch Fm. overlies the Aroley formation in an overturned sequence. Here the observed total thickness of the Complexe Antéflysch Fm. is estimated between 30 and 60 m., but due to lack of outcrops the contact with older formations is not preserved. The uppermost sill is massive and very thick (30 m). It represents a light-green to yellow-green, fine-grained volcanic rock which is interpreted as a basalt (Schürch 1987). Near this locality the same sill also shows pillow structures (Loubat 1965, Loubat and Antoine 1965, Schürch 1987 see also Fig. 2.18). At its top the sill is in stratigraphic contact with the sediments of the Complexe Antéflysch Fm. which directly passes into the Aroley Fm. (Fig. 2.21).

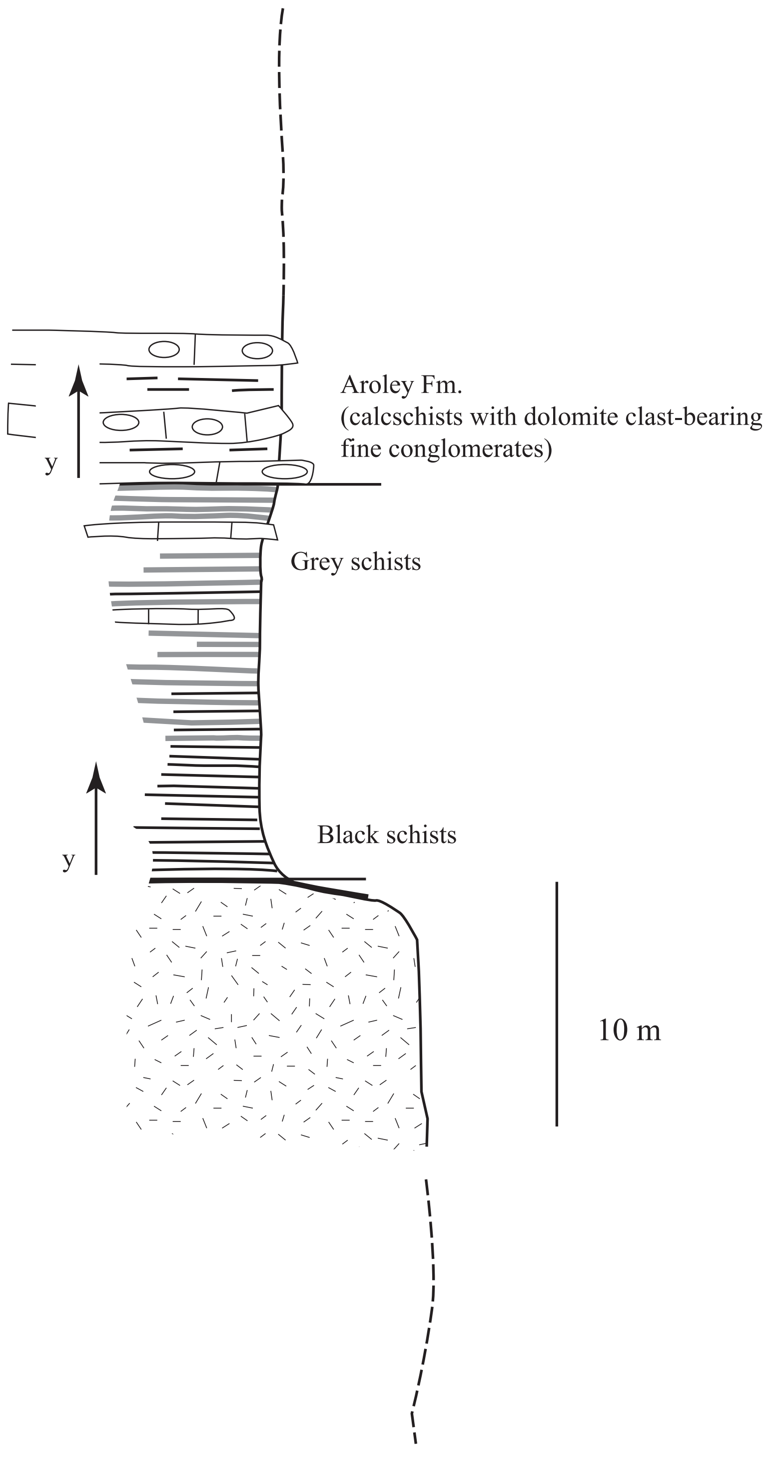


Fig. 2.21: Stratigraphic section depicting the depositional contact between the Complex Antéflysch Fm. and the younger Aroley Formation (post-rift sediments) of the Versoyen unit, internal Valaisan (Aig. du Praina area, for the coordinate see Ch. 2.3). In the Complexe Antéflysch Fm. a progressive increase of calcareous material from the black schists to the grey schists is observed. The boundary with the Aroley Fm. is gradual within 1 metre.

The contact between sill and black schists is sharp. The uppermost 7-10 meters of the Complexe Antéflysch Fm. are characterised by an irregular alternation of calcareous and pelitic beds. The calcareous beds consist of a grey limestone. The thicknesses of the beds range between 2 and 5 cm, locally thicker beds are also recognized (10-15 cm). Conglomerates are not yet observed. The schists, mainly pelitic, are grey in colour, monotonous and calcite-bearing.

This lithology is typical for the "grey schist" of the uppermost part of the Complexe Antéflysch. Over a short distance they pass into impure limestone and shales which contain fine conglomerates. The lowermost of these fine conglomerates is correlated with the base of the Aroley Fm. However, the contact is a gradual one over a thickness of 1 m. The fine conglomerates are frequent in the lowermost 5-7 meters where they show a matrix supported fabric. The matrix consists of a yellowish weathering limestone with a light-grey to bluish colour along fresh fractures. Within the lowermost meters of the Aroley Fm. the thickness of the limestone beds is constant (about 7-10 cm). Internally, mica-rich lamina are found parallel to bedding. The clasts of the conglomerates exclusively consist of yellowish weathering dolomite. On broken surfaces the clasts are dark grey to yellow and strongly recall the Middle Triassic dolomite of the pre-rift sequence. Their size is constant at a scale between 0.5 to 1-2 cm. The form is generally subrounded to rounded with a marked tabular character. The clasts show strong flattening parallel to bedding due to tectonic streaning. Measured long axes of clasts (mean azimuth 176°/ dip angle 35°) are in agreement with other structural data coming from the area (see Fig. 6.3 L1 lineation map). Up-section the clasts disappear and the rocks develops the aspect of ordinary calcschists. Further to the N (500 m N of le Tigny), the Aroley Fm. is observed in continuity with the black shales of the younger Marmontains Fm. which in turn gradually passes into the St. Cristophe Fm.. Here, the estimated thickness of the Aroley formation is in the order of one hundred meters. It is interesting to mention that along the road between Plan André and Le Petruis (x: 2080650,y: 945000,z:1590) the lowermost meters of the Aroley Fm. consist of a mica-bearing pelite, containing a metric clast of carbonate rocks and a decimetre clast of basement rock (micaschists). The matrix-supported texture suggest that this sediment was deposited by debris-flows. These observations confirm earlier descriptions made by Antoine (1971) in other regions of the study area.

The arguments supporting a stratigraphic contact between the Complexe Antéflysch Fm. and the post-rift sediments can be summarised as follows (see also Antoine, 1971). Firstly, the Complexe Antéflysch Fm. is always in contact with one and the same formation, i.e. the post-rift sediments of the Aroley Fm.

Secondly, the stratigraphic younging directions in the post-rift sediments indicate that they occur in an overturned sequence (Antoine, 1971), and the same younging direction was observed in the Complex Antéflysch Fm. (Loubat 1968, see Ch. 2.4). This means that the Complexe Antéflysch Fm. and the Aroley Fm. form a coherent stratigraphic sequence from older to younger. This coherence is also proven by the gradual transition from the black schists to the grey schists in the Complexe Antéflysch. These simple, but crucial observations contradict the idea of tectonic contact. Additional arguments, proposed in this study, come from the structural data (Ch. 5 and 6) and the reconstructed metamorphic evolution (Ch. 4) of the Valaisan units.

2.4.4: The nature of the contact of the Complexe Antéflysch Fm. with the Brèches du Collet des Rousses

The contact between the Brèches du Collet des Rousses Fm. and the base of Complexe Antéflysch Fm. is exposed at the Collet des Rousses (2854 m) and in the Pt. Rossa area where the Collet des Rousses Fm. is in tectonic contact with the structurally lower Pt. St. Bernard unit within an overturned sequence. The contact with the Complexe Antéflysch Fm. can be particularly well observed between the Collet des Rousses and the foot of the southern slope of the Aig. de l'Ermite (2670 m). The lateral extension of the Brèches du Collet des Rousses was generally underestimated (Antoine 1971, Antoine et al. 1993). New mapping revealed the occurrence of this formation from N of the Aig. du Clapet up to 2 km NW of the Pt. St. Bernard pass (Lancebrannette) and westward to the foot of the Aig. de l'Ermite.

Lithologically, the Brèches du Collet des Rousses consist of matrix-supported poorly sorted conglomerates, and of calcschists (referred to as the *"calcschistes du glacier d'Arguerey*" by Antoine 1971) which generally do not contain clasts. The suite of clast types is dominated by Lower to Middle Liassic limestones, with minor occurrences of Middle Triassic dolomite clasts. Crystalline basement pebbles were not observed, but they were depicted in Fudral (1998, Fig. 132.4). The clasts are centimetric to metric (1m) in size, with a well rounded to subangular shape. They are enclosed in a schistose to calcareous matrix (Fig. 2.22a).

The sedimentary nature of the contact between Complexe Antéflysch Fm. and Brèches du Collet des Rousses was already demonstrated by Loubat (1968). He was also the first to attribute the Brèches du Collet des Rousses Fm. to the Versoyen unit. Previously, this formation was thought to be part of the Pt. St. Bernard unit (Elter and Elter, 1965). Recently, Cannic et al.(1996) again postulated the presence of a tectonic contact between Brèches du Collet des Rousses and Complexe Antéflysch. However, this assumption is invalidated by the observations of Goffé and Bousquet (1997) and by the present work.

A) B)

Fig. 2.22: A) Conglomerate of the Brèche du Collet des Rousses (southern slope of the Aig. de l'Ermite, qt. 2760). B) Stratigraphic contact between the Brèches du Collet des Rousses (above, with conglomerate) and the black schists of the Complexe Antéflysch (below) observed in the overturned sequence at the Collet des Rousses (qt. 2770 m). C) Xenoliths of Middle Liassic limestone in a sill of the Complexe Antéflysch (near Collet des Rousses, F). The xenoliths are the same as the sedimentary clasts in A).

C)

The contact of the Brèches du Collet des Rousses with the Complexe Antéflysch Fm. shows a variety of features. The contact between conglomerates and calcschists of the Collet des Rousses Fm. and the black schists of the Complexe Antéflysch Fm. can be observed best at the Collet des Rousses or in the Pt. Rossa area (Antoine, 1971; Dalla Torre, 1998; Loubat, 1968; Loubat, 1984). At the Collet des Rousses the transition from the conglomerates to the black schists is clearly stratigraphic (Fig. 2.22b). The contact surface, although sharp, does not reveal any indication of tectonic movements. In the Pt. Rossa area (Lancebrannette) the contact between conglomerates and black schists is a gradual one over several meters (10m). Of particular interest is the occurrence of a direct magmatic contact of the conglomerates with the sills, without intervening black schists. At this locality (N of the Aig. du Clapet, Fig. 2.22c) fragments of the Brèches du Collet des Rousses are observed as xenoliths within the sill which directly overlies the conglomerate (Loubat, 1968, Dalla Torre 1998). This implies that the sills post-dates the deposition of the conglomerates and additionally proves that the contact between Collet des Rousses Fm. and Complexe Antéflysch Fm. is a stratigraphic one.

The significance of the Brèches du Collet des Rousses Fm. is not completely clear yet. This is due to the following reasons: (1) According to Schöller (1929) and Antoine (1971) the Brèches du Collet des Rousses Fm. is Upper Liassic to Dogger in age. This age is based only on occurrence of belemnites in the matrix of the conglomerates (Schoeller 1929). (2) The Brèches du Collet des Rousses shows a gradational upward transition into the Complexe Antéflysch and therefore they occupy the same stratigraphic position as the sediments and the other rocks of the Pt. Rossa Complex (Fig. 2.16). (3) They contain mainly clasts of pre-rift sediments (Middle Triassic and Lower to Middle Liassic) which are characteristic of the external Valaisan. Moreover, the Brèches du Collet des Rousses is more a conglomerate than a breccia due to the rounded shape of its components.

Thus, the Brèches du Collet des Rousses Fm. displays features characteristic for the internal Valaisan (structural and stratigraphic position) as well as for the external Valaisan (lithologies, pre-rift sediments). A possible explanation for these Jurassic conglomerates could be that the Brèches du Collet des Rousses formed at the transition between the continental (external Valaisan) and oceanic (internal Valaisan) domains.

2.4.5: The nature of the basement of the internal Valaisan

The recognition of the nature of the basement of the Internal Valaisan is hampered by the fact that, the Versoyen sediments were detached at the base of the Complexe Antéflysch during the Alpine convergence (Ch. 5). The nature of the basement can be reconstructed only based on circumstantial arguments. The lithologic composition of parts of the Pt. Rossa Complex and the Complexe Antéflysch, however, suggests that the Versoyen unit was at least partly floored by oceanic crust. Additional arguments may be provided by looking for the feeder dykes which supplied the magma which fed the sills. Previous work (Loubat, 1984; Schürch, 1987) postulates that such a feeder dyke may have been located in or near the Pt. du Clapet area (Fig. 2.23).

In this area the basaltic sills in the Complexe Antéflysch Fm. are thickest (600 m, Schürch 1987) and massive (compare Fig. 2.23 to Fig. 2.17). The structure of the feeder was assumed to be sub-horizontal (feeder-sill) by Loubat (1984) or tree-like (the "Cèdre" of Schürch 1987) (Fig. 2.24). The validity of the two models is difficult to be tested in the Aig. du Clapet area. However, the petrographic analysis of Schürch (1987) provides two interesting hints. (1) Within the massive sill in the Pt. de Clapet area, the occurrence of a serpentinite rocks similar to those in the Pt. Rossa Complex.

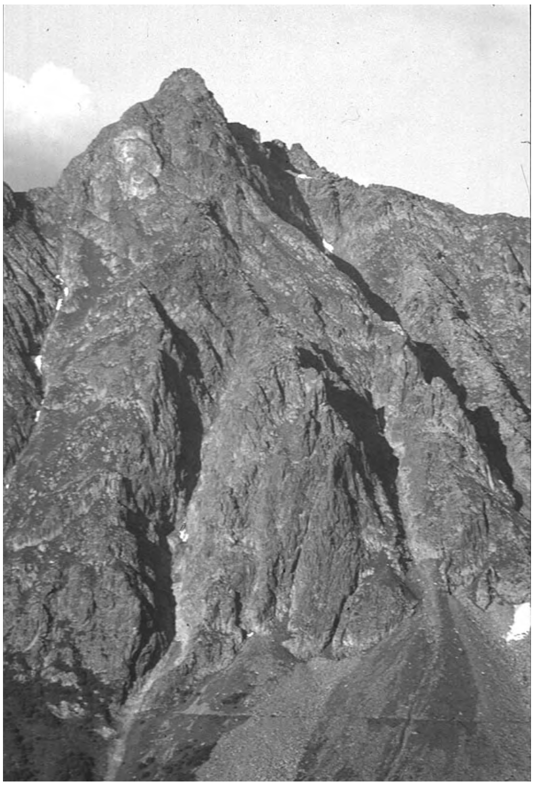


Fig. 2.23 Massive gabbroic volcanic rocks on the NW side of the Aig. du Clapet. This region was interpreted as the feeder dike (Loubat 1968, 1984) or as massive sills (Schürch 1987).

(2) In the same area, Schürch (1987) describes the occurrence of "*eléments bréchiques leucocrates*" (i.e. crystalline clasts) in a matrix formed essentially by chlorite and minor pyrite (Schürch 1987, p.32). According to Schürch, the presence of serpentinite rocks in the basement of the Pt. du Clapet area and the occurrence of clasts suggest the occurrence of our Pt. Rossa Complex rocks underneath the Pt. du Clapet area (i.e. at the base of the Complexe Antéflysch).

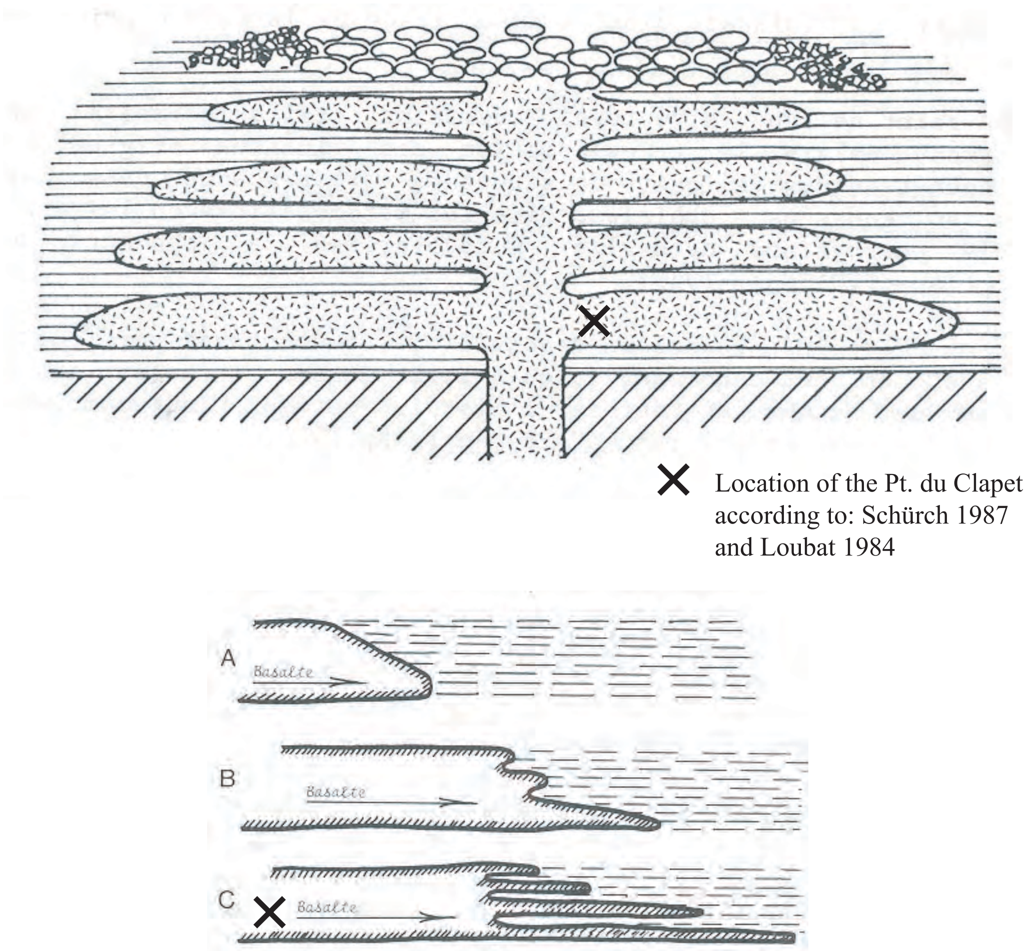


Fig. 2.24: Previous models for the location of the feeder for the sills of Versoyen unit. It is to note that the feeder is supposed to be sub-vertical (feeder dyke, Schürch 1987) or sub-horizontal (feeder sill, Loubat 1984).

According to these observations, it seems reasonable to assume that the feeder dyke was located in a region which was floored by (or situated close to) exhumed mantle material.

The MORB character of the basalts, the occurrence of pillow basalts and the presence of ophicalcites point to the exposure of serpentinites at the sea-floor and are good arguments for the existence of oceanic basement in the internal Valaisan. Another argument in favour for an oceanic nature of this basement comes from large scale correlations with other regions of the Valaisan domain. Evidence confirming the fact that oceanic crust existed in the Valais basin was discussed in Florineth and Froitzheim (1994) in the Engadine window of Eastern Switzerland and Austria (see Ch. 2.4.2). Recently, and in a portion of the Valais domain situated closer to the study area, Steinmann and Stille (1999) provided geochemical evidence which indicates that the Misox zone, attributed to the Valaisan, was underlain by oceanic crust.

The exact geometric relationships between Pt. Rossa Complex, Brèches du Collet des Rousses and sills, as well as the location of the feeder dyke are far from being resolved. Nevertheless all these lithologies exhibit close depositional relationships with the black schists of the Complexe Antéflysch Fm. Assuming that the sediments of the Complexe Antéflysch were unconsolidated during the intrusion of the sills, their sedimentation must have taken place synchronous with the final rifting or early drifting stages in the Valais ocean which included: exhumation of continental mantle rocks and the intrusion of the sills and pillow basalts. If we only consider the stratigraphic relationship between black schists and Pt. Rossa Complex, we could clearly infer a post-rift age for the Complexe Antéflysch Fm. Based on its lithostratigraphic position (below the Aroley Fm.) and on its lithological composition the Pyramides Calcaires Fm. can be considered as a more proximal equivalent of the Complexe Antéflysch Fm. (see also Ch. 3.3). However, this Pyramides Calcaires Fm. shows a clear unconformity to the overlying Aroley Fm. and is interpreted as a syn-rift formation. It follows from this that the Complexe Antéflysch Fm. is transitional between syn- and post-rift sediments.

The first sediments found on both continental and oceanic domains, defining the base of the post-rift sediments in the strict sense belong to the Barrêmian Aroley Fm. (discussed in the next Chapter). As a consequence, the term “transitional” implies a late syn-rift stage for the deposition of the Complexe Antéflysch which we correlate with the upper part of the Pyramides Calcaires Fm. If this is correct the age of the opening of the Valais ocean would have to be indirectly inferred as Early Cretaceous (pre-Barrêmian).

2.5: The post-rift sediments

General remarks

Before the deposition of the post-rift sediments (Fig. 2.4), the Valais basin consisted of a thinned continental margin, which exhibited both a subaerial relief and faulting related to the syn-rift stage of sedimentation (the external Valaisan), and of an adjacent region floored by oceanic crust which represented the basin-ward continuation (the internal Valaisan). After the initiation of opening of the Valaisan ocean, extensional faulting within the continental margin died out and thermal subsidence drove the basin-wide deposition of post-rift sediments.

The post-rift sediments consist of a continuous, less than 1 km thick sequence of conglomerates, black shales, schists and calcschists which correspond to the Aroley, Marmontains and St. Cristophe Formations respectively. On a regional scale Aroley Fm. onlaps the pre- and syn-rift sediments of the external Valaisan. Basin-wards it concordantly overlies the transitional sediments of the Complex Antéflysch Fm. of the internal Valaisan (Fig. 2.25). Particularly, in the external Valaisan the lower part of the post-rift sequence (the Aroley Fm. and part of the Marmontains Fm.) is dominated by gravity-controlled sedimentation (gravity flow sedimentation, Lomas 1992). Towards the internal Valaisan a general thinning and fining trend of the Aroley as well as of the Marmontains Formations (Antoine 1971) suggests the presence of a more distal depositional setting for the lower part of the post-rift sediments (Chapters 2.5.1 and 2.5.2). At the same time the stratigraphic sections indicate an upward thinning and fining trend in the post-rift sediments of both the external and the internal Valaisan. This is generally recorded by the shift from a conglomerate facies in the Aroley Fm. and part of the Marmontains Fm. towards a monotonous alternation of arenaceous and argillaceous beds in the St. Cristophe Fm. model for the Valais basin during the deposition of the post-rift sequence.

However, we are confronted with the problem that the sediment series are strongly deformed (see Ch. 4 to 6). For this reason the reconstruction of the depositional setting can only be inferred from a few but reliable localities (Ch. 2.5.1).

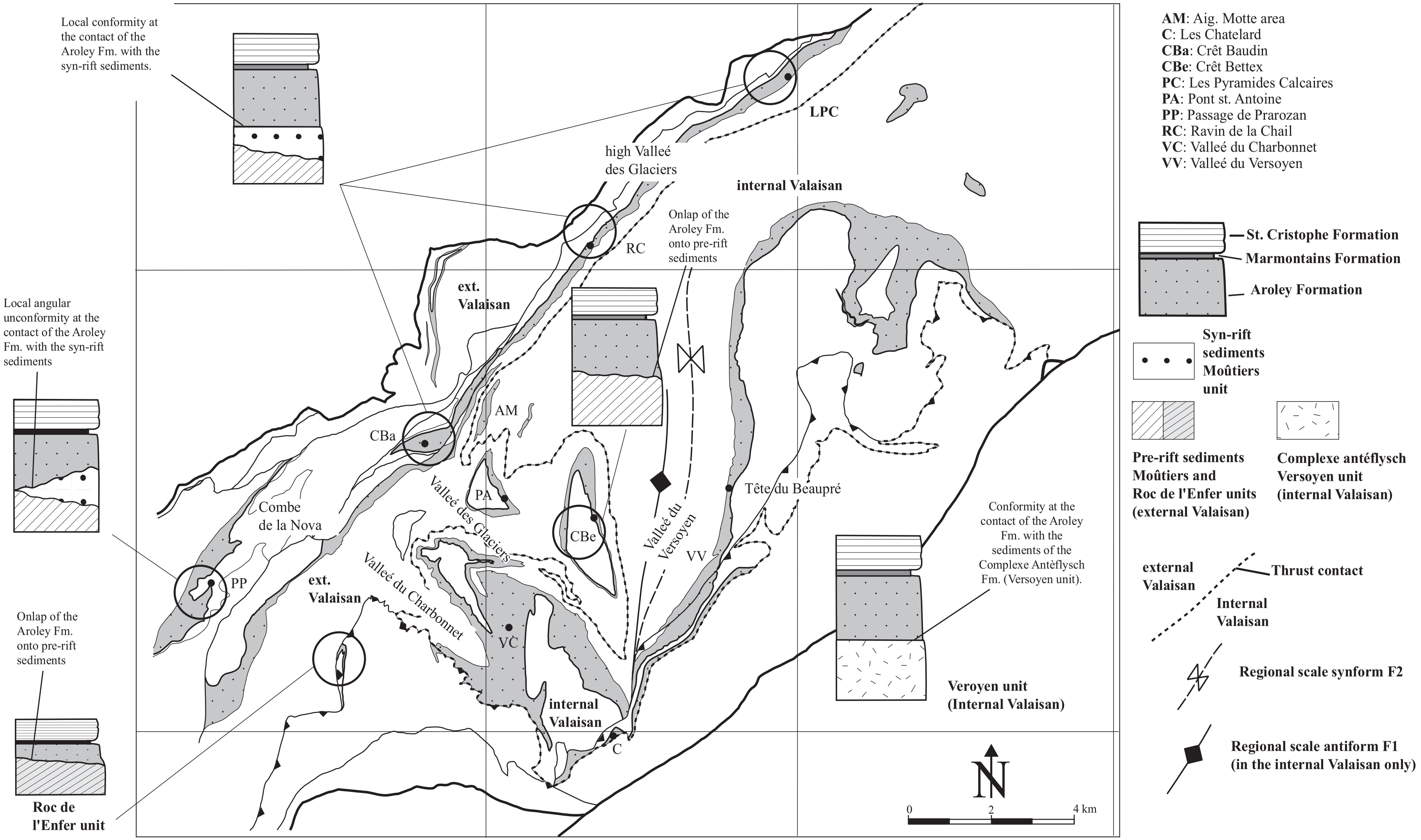


Fig. 2.25 Outcrops of the Aroley Formation within the Valaisan units (external and internal Valaisan). Schematic stratigraphic columns show the relationships of the Aroley Formation with older pre- and syn-rift sediments as well as transitional formations. note the difference occurrence of the unconformity at the base of the aroley Formation, discordant within the external Valaisan and concordant within the internal Valaisan. All the localities referred in Chapter 2.5 are indicated.

2.5.1: The Aroley Formation

It is remarkable that everywhere in the study area the post-rift sediment start with the Aroley Fm., However, the Aroley Fm. onlapping pre- and syn-rift sediments, may become younger towards the external Valaisan. Primary stratigraphic contacts between the Marmontains Fm. or the uppermost St. Cristophe Fm. with pre- or syn-rift sediments have never been observed so far. Fig. 2.25 illustrates distribution of outcrops of the Aroley Formation and its sedimentary relationships with older stratigraphic sequences (pre-syn- and rift-sediments). Regarding the external Valaisan, the Aroley Formation is best developed within the Moûtiers unit. In the Roc de l'Enfer unit, according to our findings (see Chapter 6.3.2), it is only locally preserved (Col de Leisette Fig. 2.25) ranging in thickness from a few meters (less than 2m in some cases) to 20-30 m. Except for its reduced thickness the Aroley Formation within the Roc de l'Enfer unit is identical to that of the Moûtiers unit.

Within the Moûtiers unit the Aroley formation outcrops near Les Chapieux, from the Combe de la Nova up to Les Pyramides Calcaires and along the lower part of the Valleé des Glaciers in the localities Pont. St. Antoine and Crêt Bettex (Fig 2.25). The overall thickness ranges between 50 and 150m. Where the beds are intensely folded (Crêt Baudin) they can reach an apparent maximum thickness of more than 400m. In the internal Valaisan (Versoyen unit) the Aroley Formation is exposed in the Valleé du Charbonnet, as well as along the SE side of the Valleé du Versoyen (F) and up to the Valleé de Breuil (I). Other important outcrops are located between Bourg St. Maurice and the Pt. Du Clapet (see Ch.5). Apart from a very few outcrops located close to the Pennine Front and within a complex system of thrust sheets, each occurrence of the Aroley Fm. is still attached to pre- or syn-rift sediments, and in stratigraphic continuity with the younger Marmontains Fm.. This allows to make sedimentological observations along vertical profiles located above a variety of pre-Aroley formations.

In the external Valaisan (Moûtiers unit) the most common facies are massive conglomerates with a grey to blue calcareous matrix. Generally bedding is well developed, beds range between 0.3 and 15-20 m. Interbedded pelitic levels are usually reduced to a few centimetres and rarely reach more than 1 m. Often they are even absent. The clast size ranges between 1cm to 3m. However, most clast sizes are between 1 and 10 cm (Lomas, 1992).

The clasts are composed of lithologies comparable to those of the pre-rift sediments (Carboniferous to Middle Liassic) and also include crystalline rocks (mainly micaschists and granite). Clasts of Middle Triassic dolomite and Lower to Middle Liassic limestones are predominant (Lomas, 1992).

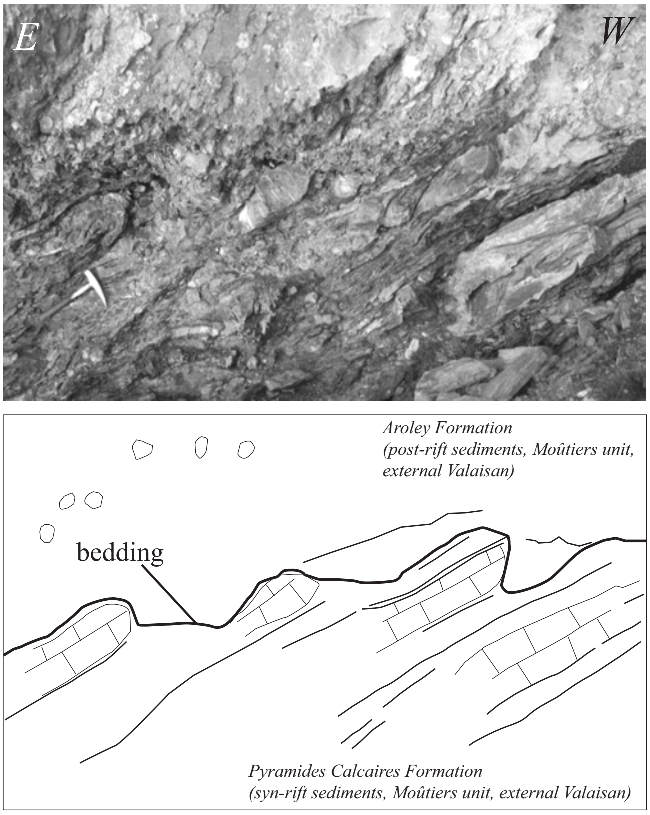
In the internal Valaisan (Versoyen unit), the Aroley Formation shows remarkable sedimentological differences in respect to the external Valaisan. The dominant facies consists of bluish impure limestones, interbedded with pelitic levels, forming a calcschist which contains conglomerates. In a few isolated occurrences a massive grey limestone can be observed. Both facies (calcschists and massive limestones) were observed at the base of the Aroley Formation (i.e. close to the contact with the Complex Antéflysch Fm.) and are interpreted as lateral equivalents of the Aroley Fm. in the external Valaisan. The sedimentation of the Aroley Fm in the internal Valaisan is characterised by an upward fining trend of the components and by a decreasing occurrence of conglomerate beds. The clast-size ranges from millimetres to 5 cm. and the clast composition is dominated by Middle Triassic dolomite. The thickness of conglomeratic beds is constant and ranges around 10-30 cm.

The sedimentological differences of the Aroley Formation between external and internal Valaisan were first observed by Antoine (1971) who interpreted them in terms of a pronounced proximal to distal facies variation. Using low-strain exposures from different localities, Lomas (1992) proposed a sedimentological reconstruction of the depositional system of the external Valaisan. However, both authors did not take into account the geometry of the Valaisan nappe stack. This work aims to integrate both the detailed work of Lomas (1992) on the proximal Aroley Formation, and the large scale observations of Antoine (1971) into the detailed structural framework of the Valaisan units.

Our observations focus on the unconformity at the base of the Aroley Formation from the external to the internal Valaisan (Fig. 2.25). In some crucial localities vertical stratigraphic sections are briefly discussed. Due to tectonic overprint, single outcrops cannot be directly correlated from one locality to another. Such outcrops are discussed in Chapters 2.5.1.1 and 2.5.1.2, and integrated in chapter 2.5.1.3 along a geological cross section at the scale of the study area, qualitatively retro-deformed to the original position before the onset of the Alpine deformation.

2.5.1.1:The Aroley Formation of the external Valaisan

Fig. 2.26 illustrates the angular unconformity at the base of the Aroley Formation, above the Pyramides Calcaires Formation (syn-rift sediments of the Moûtiers unit), which can be observed at the Passage the Prarozan, near the Col du Grand Fond (Combe de la Nova area x: 937’300 y: 2082’425 z: 2660).



A)

B)

Fig. 2.26 a: Photograph and line drawing of the angular unconformity between the Aroley Formation (above) and the Pyramides Calcaire Formation (below) observed at the Passage de Prarozan. Fig. 2.26b: The first bed of the Aroley Fm consists of a coarse clast-supported 5m thick bed of conglomerates. (scale is 27 cm)

The bedding of the calcschists of the Pyramides Calcaires Formation is clearly truncated beneath the conglomerates. The base of the Aroley Fm. is generally planar to weakly undulating. This unconformity can also be observed at a large scale by a panoramic view of the NW side of the Aig. Du Grand Fond (Antoine et al., 1972). Locally, moderate late tectonic movements (in the range of few meters of displacement) may have taken place along this surface.

However, due to the absence of consistent evidence for larger movements along this contact, we ascribed this unconformity be original. Assume that the deposition of the Aroley Fm. occurred along a sub-horizontal plane the calcschists of the Pyramides Calcaires Fm. must have gently dipped towards ENE (069/12). In this example the Aroley Formation starts with a coarse clast-supported 5 m thick conglomerate bed, which is moderately to poorly sorted. Clasts are generally well rounded and their sizes range between 5 to 18 cm. However, larger components up to 35 cm can also be observed. Up-section the bedding within the Aroley Formation remains thick (5-10 m).

In a more internal position, the basal contact of the Aroley Formation can also be observed along the higher part of the Valleé des Glaciers (Fig. 2.27). In these exposures, evidence for an angular unconformity could not be observed. The contact is concordant and transitional over a few meters (2-3 m) and (see Ch. 3.4). The Aroley Fm. starts with thick (2-10 m) conglomerate beds showing an upward thinning trend (Fig. 2.28). Conglomerate beds are generally sheet-like with marked planar surfaces. The components of the Aroley Formation show vertical variations in their composition. In one bed they are almost completely composed of crystalline rocks (Fig. 2.28). However, in general, dolomite represents the most abundant clast type. This character, together with a sub-angular clast shape suggests a derivation from local sources.

Moving eastwards, the Aroley Formation can also be observed along the lower Valleé des Glaciers from the Crêt Baudin up to Les Châtelard (near Bourg St. Maurice). This part of the Valleé des Glaciers offers the possibility to link observations on the Aroley Fm. over a distance of more than 5 km (not retro-deformed). The Aroley Fm. can be observed along the bottom of this valley (Pt. St. Antoine and Crêt Bettex) or along the ridge between the Pt. de la Terrasse and the Aig. du Praina up to Les Châtelard (Fig. 2.25). The upper Aroley Formation also finds a continuation in the Aig. Motte area (i.e. on the other side of the Valleé des Glaciers in respect to the Pt. De la Terrasse). The lower Aroley onlaps the pre-rift sediments (of the Moûtiers unit) and represents the proximal Aroley Formation. The higher Aroley Fm. is concordant with the Complex Antéflysch (i.e. it belongs to the Versoyen unit of the internal Valaisan) and represents the distal Aroley Formation (see Ch. 2.5.1.2). These two occurrences are separated by about 600 m of post-rift sediments, mainly the St. Cristophe Formation, which belongs to the Moûtiers as well as to the Versoyen unit (see Fig. 2.3).

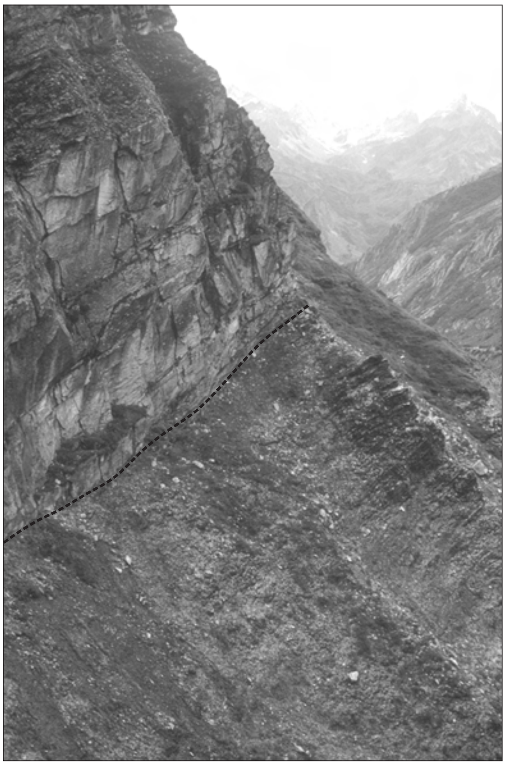


Fig. 2.27 Concordant contact between the Aroley Formation (post-rift sediments) and the Pyramides calcaires Formation (syn-rift sediments), external Valaisan, Ravin de la Chail, Valleé des Glaciers (x:945'500, y:2'090'450, z:2200).

From NW (at the Pt. St. Antoine) towards SE (at the Crêt Bettex) the Aroley Formation onlaps Middle Triassic dolomite and Lower Triassic quartz-arenites, respectively (Fig. 2.29). The Aroley Formation starts with coarse clast-supported conglomerates interbedded with thin (0.1-1 m) litharenites. This facies is the same as observed at the Passage the Prarozan. A detailed stratigraphic study of the outcrop at the Pt. St. Antoine was carried out by Lomas (1992) and his main observations are briefly summarised in the following (see also Fig. 2.30). The first 40 m of the Aroley Formation consist of coarse, clast-supported thick (1-8m) beds characterised by the absence of internal organisation (e.g. absence of normal grading). This part of the Aroley Formation also represents the coarsest in terms of maximum clast size and bedding thickness. Up section (between 40 and 80 m), the dominant facies consists of normally graded clast-supported conglomerates interbedded with thin matrix-rich beds. Bed thickness generally decreases (0.3 to 2 m) and scoured bedding surfaces can oft be observed. In both facies, normal gradual grading occurs.

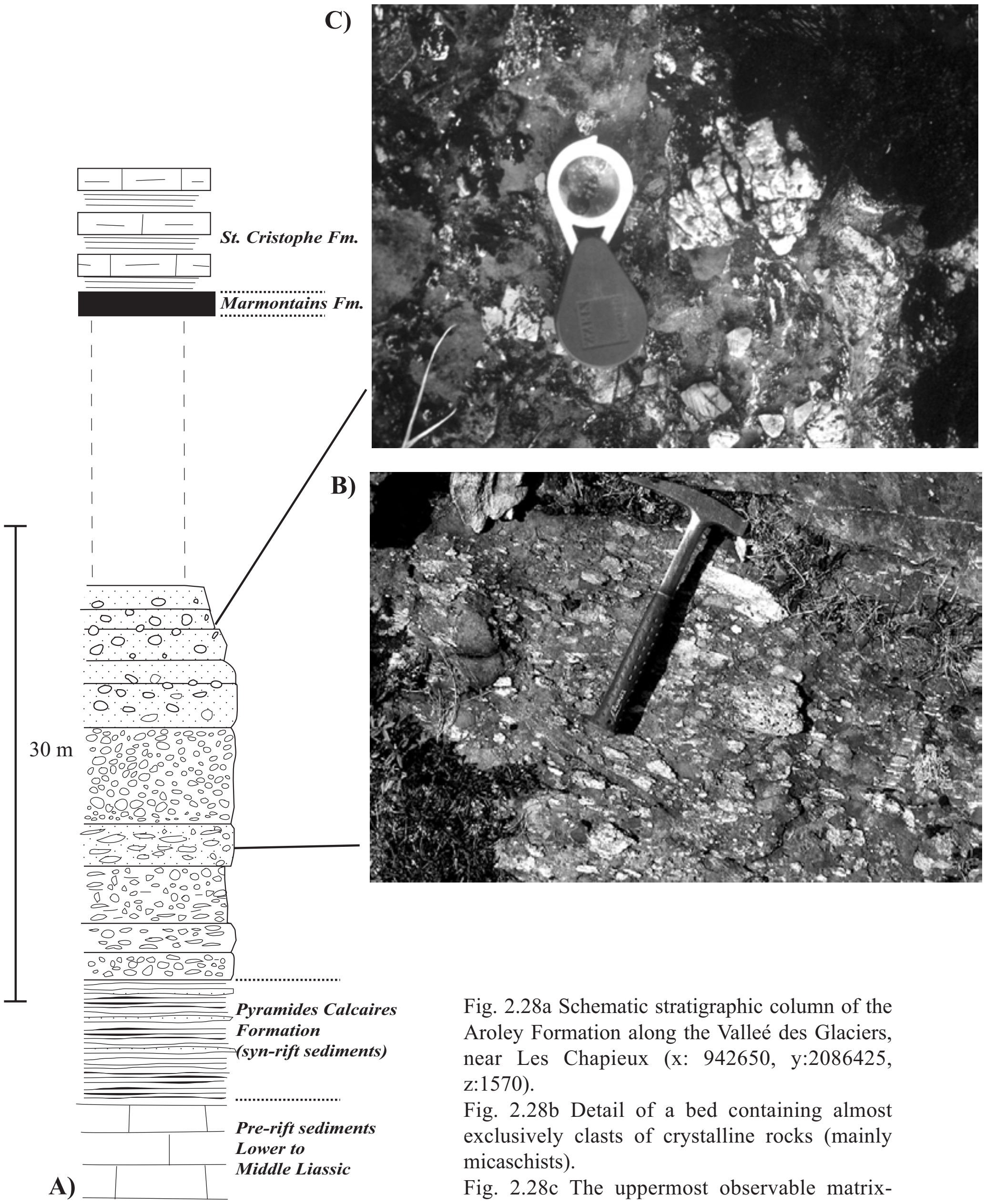


Fig. 2.28a Schematic stratigraphic column of the Aroley Formation along the Valleé des Glaciers, near Les Chapieux (x: 942650, y:2086425, z:1570).
Fig. 2.28b Detail of a bed containing almost exclusively clasts of crystalline rocks (mainly micaschists).
Fig. 2.28c The uppermost observable matrix-supported conglomerate bed.

Finally, the uppermost part of the Aroley Formation (between 80 and 120 m) shows thin conglomerate beds (0.4 to 2 m) and well developed high-angle cross-bedding. The thinnest beds (0.25 to 1.5 m) comprise a coarser clast-supported zone at the base which gradually passes into a matrix supported facies in the upper part of the bed. In these occurrences the matrix is represented by a calcareous sandstone. According to Lomas (1992) the vertical transition between the different facies types is always gradual and can be considered as a continuous sedimentary event. The vertical succession of clast types indicates the predominance of dolomite and limestone, associated with a slight upward increase of non-carbonate components (schist, sandstone and granite).

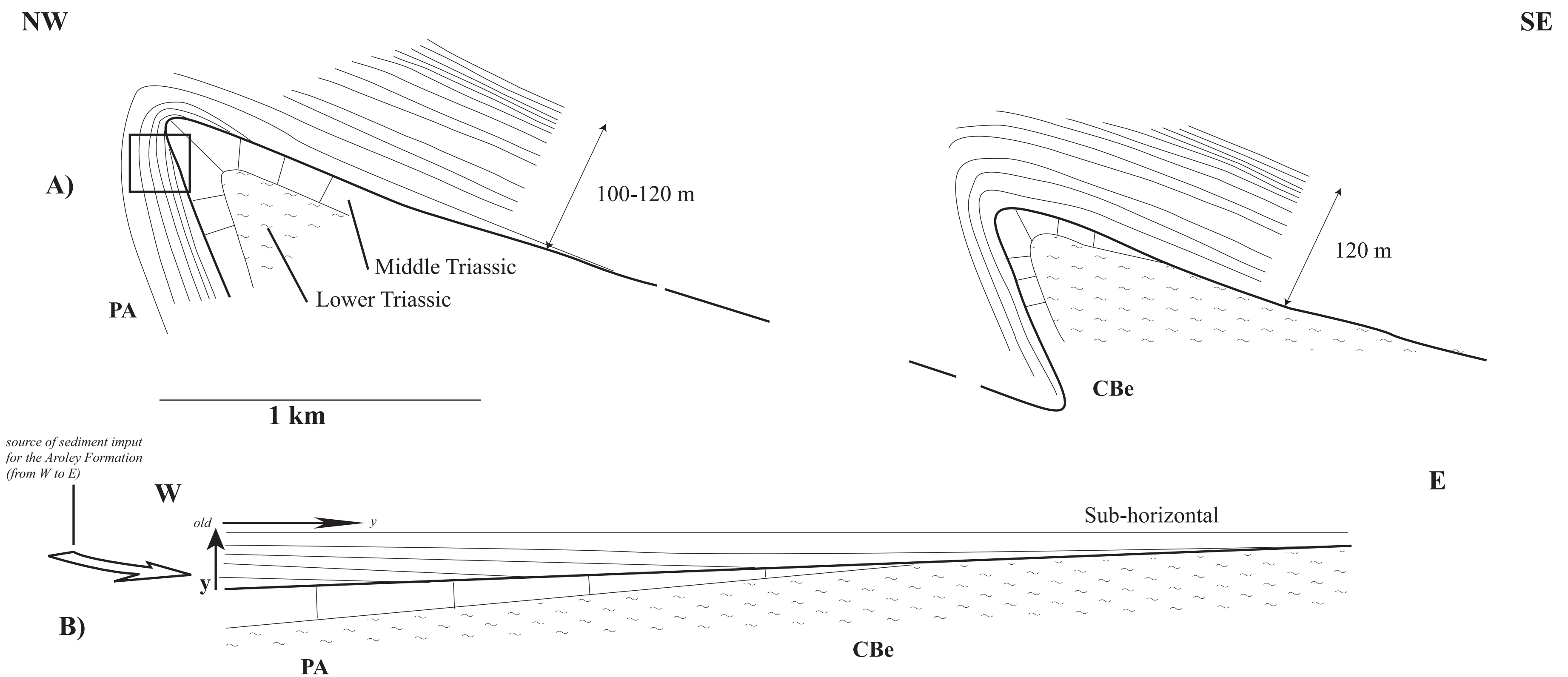
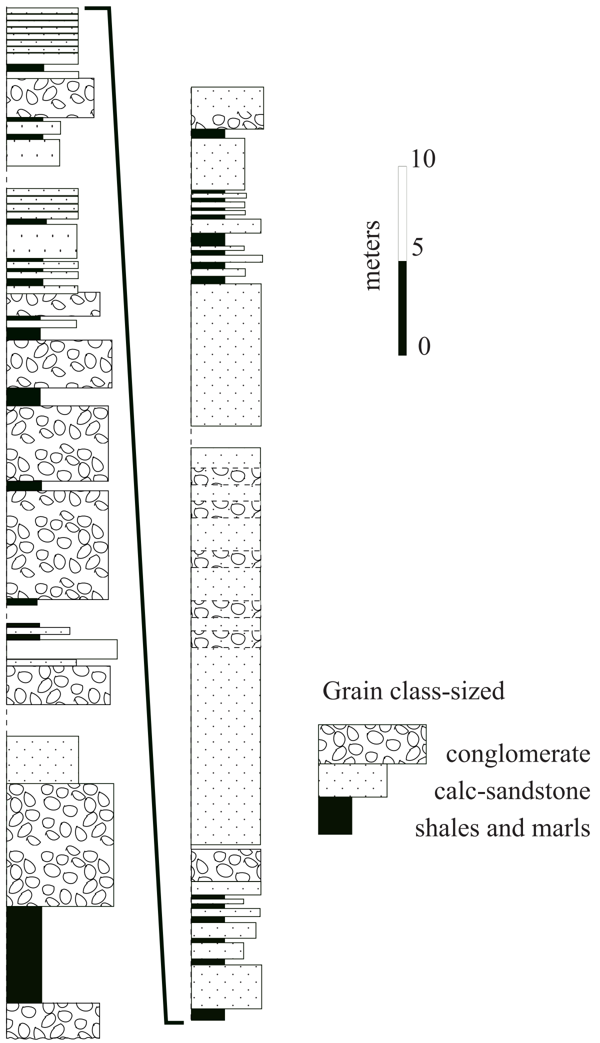
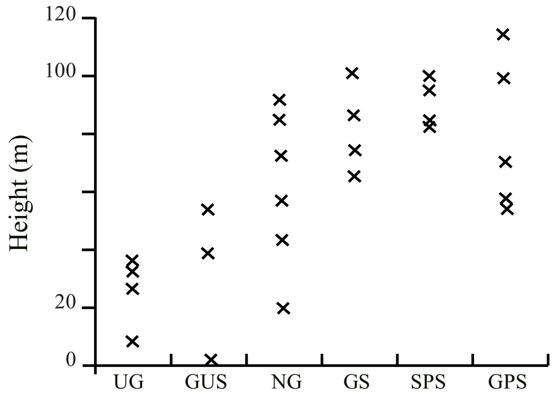


Fig. 2.29a Schematic cross section along the lower Valleé des Glaciers, from the Pt. St. Antoine (PA) to the Crêt Bettex (CBe) showing the upward thinning trend of the Aroley Formation. Inset at the Pt. St. Antoine locality shows the onlap of the first beds of the Aroley Formation onto Middle Triassic dolomite (x: 942'975, y: 2084'650, z: 1650 +/- 20 m). At the Crêt Bettex locality the Aroley beds are concordant with Middle and Lower Triassic. Fig. 2.29b shows the same profile after folding is removed. The beds of the Aroley Formation exhibit an onlapping direction from W toward E. This direction suggests that the Aroley Formation was sourced from the W (from the continental margin of the external Valaisan) rather than from the E (i.e. the oceanic domain of the internal Valaisan). **y**: younging direction



A)



B)

UG: very coarse, matrix-poor, disorganised conglomerate. Lack of vertical grain-size segregation. Clast-supported. Beds typically 2-8m thick. Very coarse grain-size

GUS: Coarse, clast-supported, crudely graded conglomerate. Normally graded, sometimes with a thin basal zone of inverse grading. No discernible stratification.

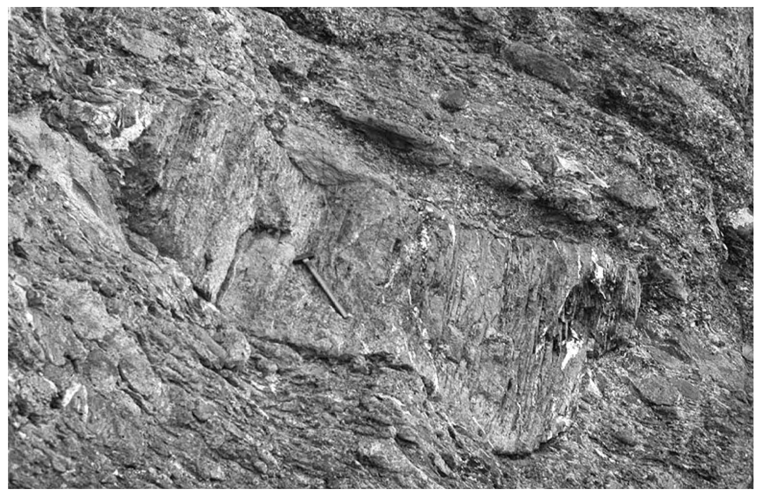
NG: Normally graded conglomerate. Graded often progressive rather than abrupt. Crude low-angle stratification occasionally present. Clast-supported, with a marked paucity of arenaceous material.

GS: Crudely stratified conglomerate. Normal grading confined to basal portion of bed. Grading typical abrupt rather than gradational. Gently inclined (< 20° to bed base) near planar stratification transitional to multiply graded profile.

SPS: Cross-stratified pebbly sandstone or fine conglomerate. Cross-bedding typically of symptotically- or planar- based tabular morphology, less commonly through-like. Basal part may show weak normal grading.

GPS: Graded pebbly sandstone. Normally graded through most of the bed. Matrix supported except in lower portion. Crude low angle stratification common in higher parts.

(simplified after Lomas 1992)



C)

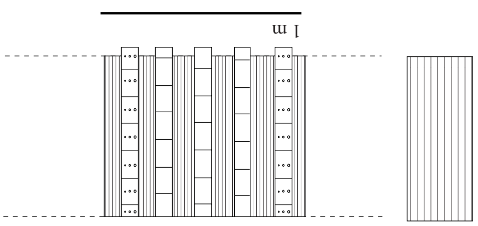
Fig. 2.30 Figure showing the main results of the sedimentological study of the Aroley Formation at the Pt. St. Antoine locality performed by Lomas (1992). A: Simplified stratigraphic log and B: graph showing the stratigraphic height at which beds of various facies are encountered. C: coarse grain supported facies of the Aroley Formation (Facies UG of Lomas 1992), observed at the Pont St. Antoine, Valleé des Glaciers with large clast of Middle dolomite is 3m long.

The presence of crystalline basement clasts, such as micaschists and gneiss, should also be pointed out. They have been observed within the stratigraphically lower strata with a slight increase toward the higher strata.

2.5.1.2: The Aroley Formation of the Internal Valaisan

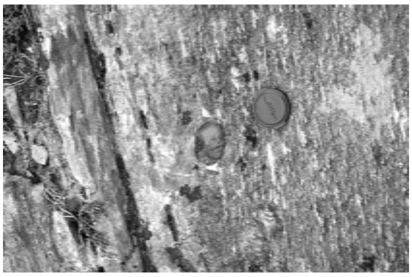
The occurrences of the Aroley Fm. of the internal Valaisan in the Valleé des Glaciers (around the Pt. de la Terrasse) show a monotonous repetition of calcschists. Except for rare occurrences of fine-conglomerates, the attribution to the Aroley Formation is inferred from the stratigraphic transition into the Marmontains Formation (quartz-arenites and black shales) which also helps to define the younging direction within the post-rift sediments.

Lithologically, the Aroley Formation of the Valleé des Glaciers is characterised by a regular alternance between decimetre (8-25 cm) thick bluish limestones and argillaceous beds. The rare occurrence of fine-conglomerate is restricted to the base of the limestone beds. In these cases, a normal grading is evident (Fig. 2.31, x: 942'525, y: 2083'250, z: 2450). In the Aig. Motte area (along its western slope above the Les Chapieux village) the distal Aroley Formation displays a massive well bedded bluish limestone which lacks large pebbles. Further SE at Les Châtelard, the Aroley Formation consists of a massive dark grey to bluish limestone with a well developed bedding outlined by thin (0 to 2cm) pelitic schists (Fig. 2.32). The calcareous beds are generally sheet-like, with a planar base and bed the thickness varies between 1-2 m. Locally, abundant occurrences of fine conglomerate are evident. These fine-conglomerates are characterised by limited matrix-supported domains which are laterally not consistent. The clasts, always of small dimensions, from a few millimetres (0.1-0.5 cm) to 2 cm, are homogeneously dispersed in the matrix. The lithology of the clasts is dominated by yellow dolomites with a sub-rounded shape and strongly flattened within the foliation plane (S1). Conglomerates of similar composition sporadically comprise isolated meter-scale clasts at Plan André. long the Valleé du Versoyen the most interesting outcrops of the Aroley Formation show the occurrence of an isolated coarse conglomerate interbedded with calcschists. In this area the overall thickness of the Aroley Formation ranges between 100 and 200 m. It is interesting to note a direct correspondence between well developed bedding and the occurrence of coarse conglomerates. This can be observed at the Tête du Beaupré where coarse clast-supported conglomerate beds with a total thickness of 30 m occur towards the top of the Aroley Formation (Fig. 2.33). This particular facies is laterally not consistent and passes into the calcschist.



(Complexe antéflysch)

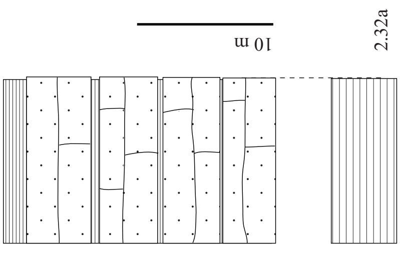
2.31a)



2.31b)

Fig. 2.31a schematic stratigraphic column of the Aroley Formation in the Valleé des Charbonnet showing the regular alternance between calcareous and argillaceous beds.
Fig. 2.31b detail of the same outcrop showing a bed of the Aroley Formation which exhibits internal grading. Due to folding the Aroley formation is overturned and this feature is interpreted as normal grading (x:942'525, y:2083'250, z:2450).

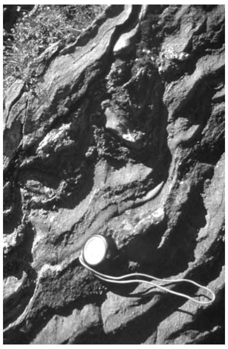
2.32b



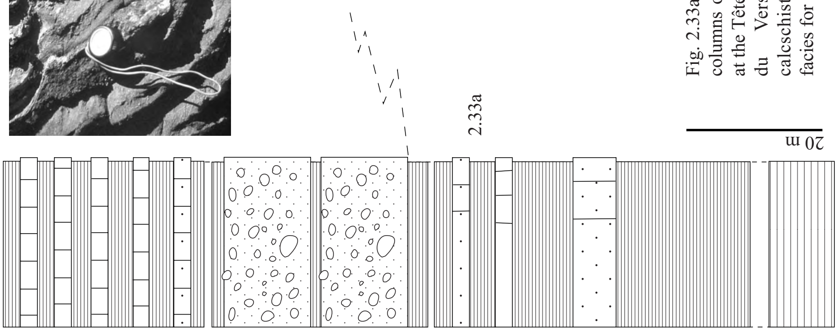
2.32a

(Complexe antéflysch)

Fig. 2.32a Schematic stratigraphic column of the Aroley at the Chatelard.
Fig. 2.32b Matrix-supported fine conglomerate dispersed within a limestone matrix containing millimetre size clasts of dolomite (large clast is 5 cm long).



2.33b



2.33a

(Complexe antéflysch)

Fig. 2.33a Schematic stratigraphic columns of the Aroley Formation at the Tête de Beaupré area, valleé du Versoyen. Schists and calcschists are the dominant facies for the Aroley Formation

Fig. 2.33b In this area also thick conglomerate beds with coarse clasts be observed. These beds are laterally not continuous and pass into the calcschists.

2.5.1.3: The depositional setting of the Aroley Formation

Information concerning the depositional setting of the Aroley Formation can be inferred from sedimentological observations (Antoine, 1971; Fudral, 1973; Lomas, 1992) and by the reconstruction of the geometry of the Valaisan basin before the onset of the Alpine deformation. Fig. 2.34 shows the simplest way to achieve the original geometric configuration. This approach is geometrically correct but dimensions are qualitative. Fig. 2.35 shows the location of the outcrops of the Aroley Fm. described before. Two facies associations can be distinguished, based on bed thickness, grain size, and on the proportion of grading and bedding style (Walker, 1967; Walker, 1976). Other geometrical (bed shapes) or depositional marks (scour marks) have not been observed.

The Aroley Formation of the external Valaisan (outcrops: PP, Cba, PA and Cbe, see Fig. 2.25) starts with thick beds (Figs 2.26, 2.27) ranging between 2 and 10 m. The beds are generally clast-supported conglomerates. They show no grain-size sorting and a very coarse mean and maximum grain size (Lomas, 1992). Upwards a thinning and fining trend corresponds to a gradual increase of normal grading and a trend towards matrix supported clast fabrics.

The Aroley Formation of the internal Valaisan (outcrops C, VC, AM and VV, see Fig. 2.25) displays thin beds (Figs. 2.31, 2.32, 2.33 and 2.35), ranging between 8 cm and 2m. Conglomerates are intercalated within a regular alternation of calcarenitic and argillaceous beds. Generally, the occurrence of coarse conglomerates (Fig. 2.33) correlates with a large bed thickness (10 m). Normal grading and a matrix supported clast fabrics with a relatively small clast size (1 mm to 5 cm) are characteristic features of these beds. Due to intense folding, the study of variation of bed thickness along vertical profiles is made impossible. However, an almost regular bed thickness through out the whole formation is suggested from field mapping. Interestingly, conglomerate beds (Figs. 2.32 and 2.33) are generally found at the base of the formation. In the internal Valaisan, conglomerates are never been observed close to the contact with the Marmontains Fm. Generally, the occurrence of coarse conglomerates (Fig. 2.33) correlates with a large bed thickness (10 m). Comparing the sedimentological features of the Aroley Formation the lateral thinning and fining trend suggests a proximal (external Valaisan) and distal (internal Valaisan) depositional setting.

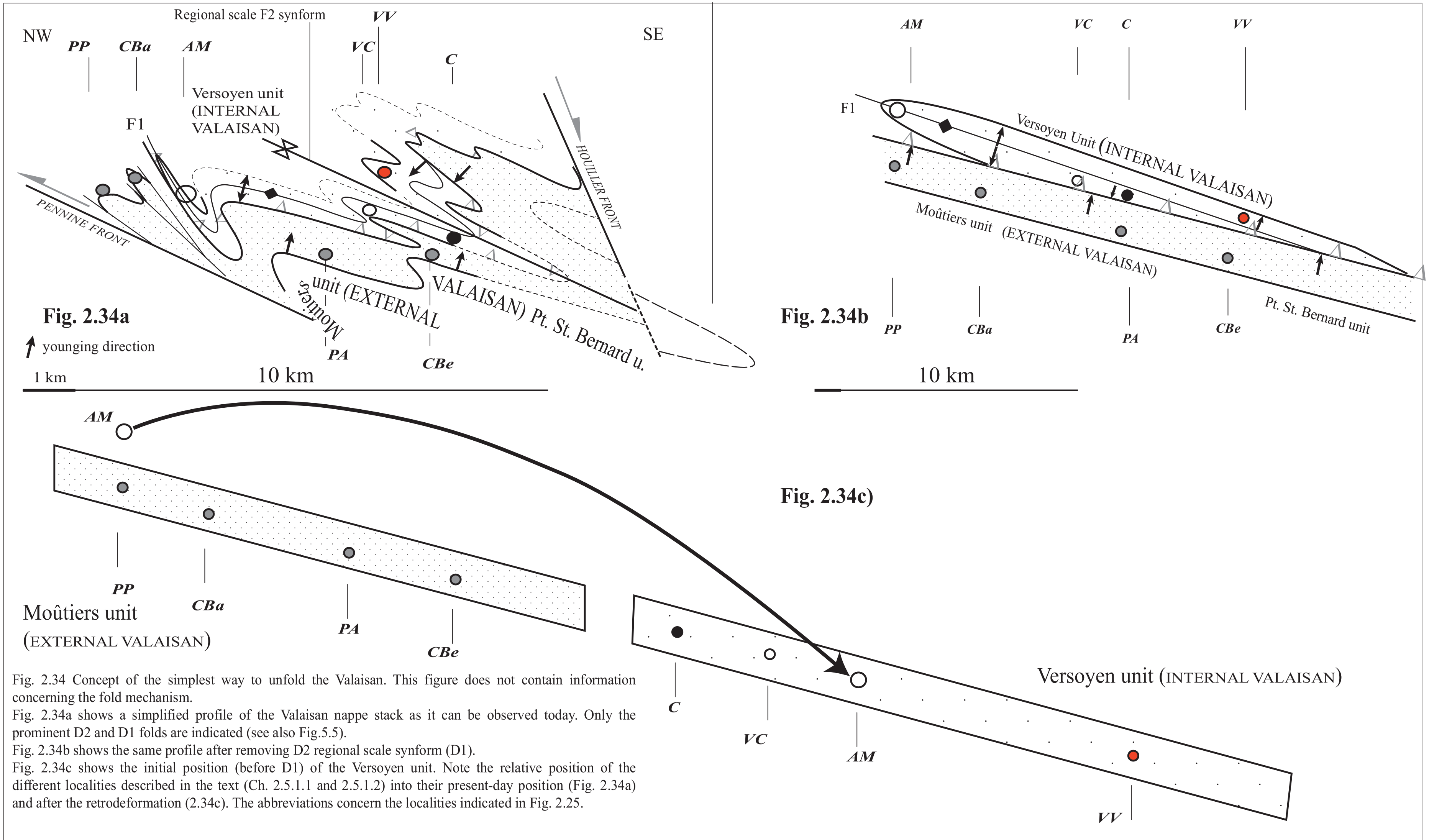


Fig. 2.34 Concept of the simplest way to unfold the Valaisan. This figure does not contain information concerning the fold mechanism.
Fig. 2.34a shows a simplified profile of the Valaisan nappe stack as it can be observed today. Only the prominent D2 and D1 folds are indicated (see also Fig.5.5).
Fig. 2.34b shows the same profile after removing D2 regional scale synform (D1).
Fig. 2.34c shows the initial position (before D1) of the Versoyen unit. Note the relative position of the different localities described in the text (Ch. 2.5.1.1 and 2.5.1.2) into their present-day position (Fig. 2.34a) and after the retrodeformation (2.34c). The abbreviations concern the localities indicated in Fig. 2.25.

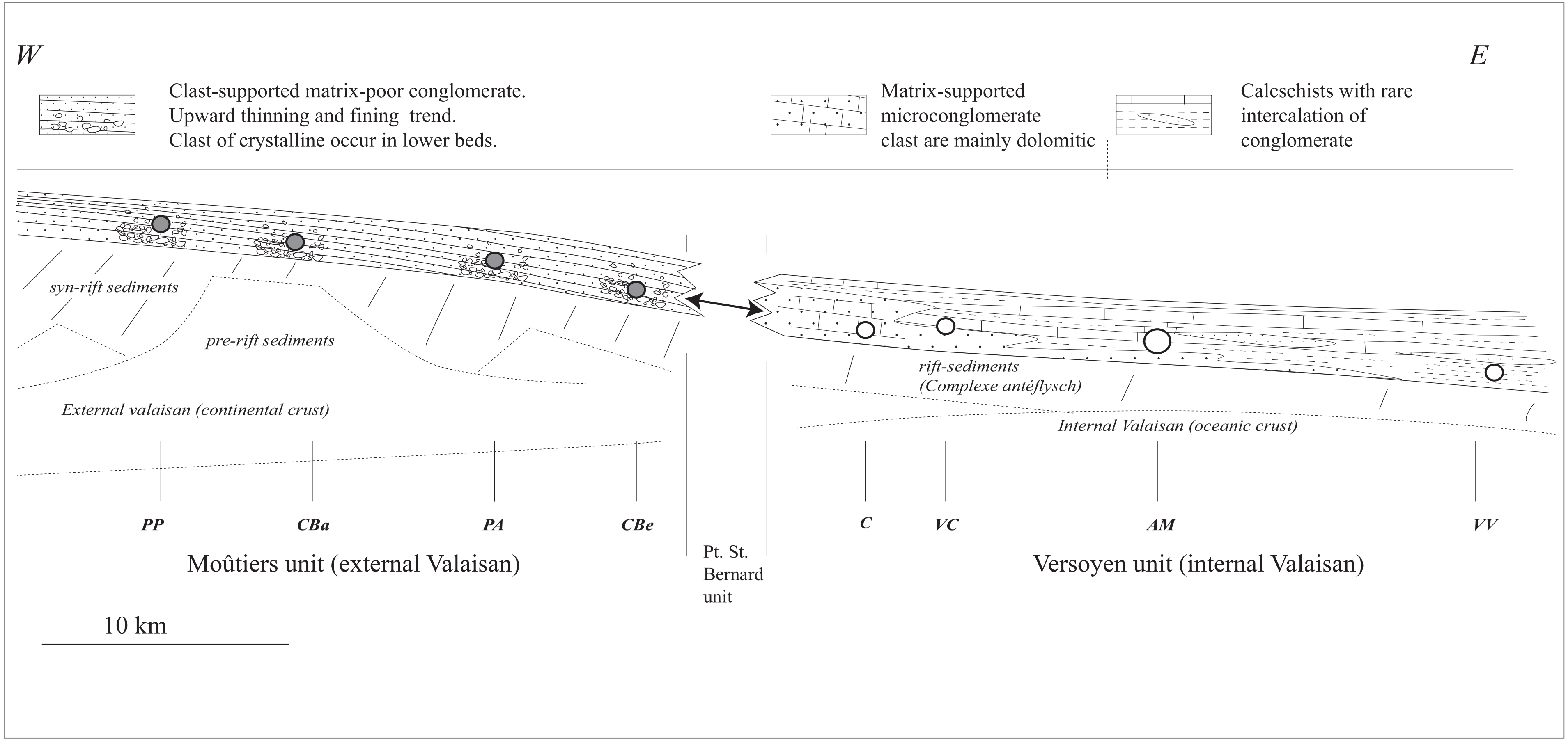


Fig. 2.35 Shows the facies variation of the Aroley Formation within the Valaisan basin. Orientation is based on present day coordinates and on paleocurrent directions (Lomas 1992). Modified after Antoine (1971).

The variation in the composition of the clast-types of the Aroley Formation from the external to the internal Valaisan is also indicative for this same arrangement. In the external Valaisan the Aroley Formation exhibits clasts which include crystalline rocks and sediments from Carboniferous to Middle Liassic sources. The clasts from the Aroley Formation in the internal Valaisan clast rarely comprise granite and micaschists, the clast source being mainly limited to Triassic dolomite (mainly Middle Triassic). However, clasts of granite and micaschists are occasionally present at different proportions, in both proximal and distal Aroley Formations. These clasts were also observed even in the stratigraphically lowest strata. Lomas (1992), describes only a slight upward increase in the proportion of non-carbonate components.

The deposition of the proximal Aroley Formation was driven by gravity-controlled mass-flow processes (Lomas, 1992) in a submarine setting. The sub-marine setting is evidenced by the few poorly preserved benthic foraminifera (Sodero, 1968; Trümpy, 1952; Trümpy, 1955a). The lack of evidence for subaereal exposure (hematite pigmentation, aeolian deposits, dessiccation cracks, paleosols, etc) additionally point to a subaqueous depositional environment. The water depth, however, remains unknown. It is assumed to be generally greater than storm wave base (a few 100 m, Lomas 1992). This is suggested by the absence of wave reworking of the sediments. For the internal Valaisan the most likely setting is one which is deeper and quieter than that of the external Valaisan. A lower depositional energy of the Aroley formation of the internal Valaisan is suggested by the decrease in the coarse-clast portion of the sequence and by the occurrence of graded bedding. Another argument in favour for a quiet and distal depositional setting for the Aroley Formation is suggested by the absence of evidence of reworking or erosional surfaces at the contact with the lowermost Complex Antéflysch Fm. These observations can be interpreted in terms of decrease of the depositional energy, compatible with a distal depositional region. The sporadic occurrence of laterally discontinuous beds with coarse clasts can be interpreted in terms of channels.

The detrital source area of the Aroley Formation was located somewhere in the W, i.e. continent-ward in respect to the internal oceanic Valaisan. The, palaeocurrent directions were proposed to be towards NE by Antoine (1971). This was mainly based on the presence of a slice of crystalline rocks (Hautcour crystalline near Moûtiers) in the SW part of the Valais Zone, in direct stratigraphic contact with the post-rift sediments. The commonly observed SW-NE directed orientation of the long axes of the clasts, interpreted by Antoine as a paleocurrent feature, however representsa tectonic feature (see Ch. 6). The same interpretation, from a sedimentological perspective, was also proposed by Lomas (1992) who noted a reorientation of the grain orientation distribution towards parallelism with local fold axes even in the least strained beds (Fig. 14 of Lomas 1992). Lomas (1992) proposes a WNW-ESE (Mean Lineation Vector 119°, after removing tectonic rotation, Lomas 1992) oriented palaeocurrent direction. Another argument in favour of a western origin of the clasts source is suggested by Fig.2.29b. This reconstruction shows that the Aroley Formation, over a distance of 6-7 km (after retro deformation), onlaps pre-rift series toward the E. Finally, the lack of a particular trend in the range of clast types along vertical sections and the presence of crystalline-type clasts also in lowermost beds indicate that basement rocks were exposed before as well as during the deposition of the post-rift sediments. This point to a very pronounced stretching of the passive margin leading to tectonic unroofing of basement rocks during the rifting period.

2.5.2: The Marmontains Formation

The Marmontains Formation can easily be distinguished from the underlying Aroley Formation and the overlaying Cristophe Formation. The Marmontains Formation generally consists of a regular alternation of black shales and quartz-arenites. Macroscopically, the quartz-arenites form continuous massive beds showing a characteristic green to brownish colour. The thickness remains laterally constant over great distances and ranges between 0.1 and 0.5 m. The black shales consist of a regular intercalation of black to dark grey fine grained schists. The black shales are generally carbonate-free. In this case the distinction of the black shales from the Carboniferous black schists is based on the presence or absence of a detrital white mica. The black shales of the Marmontains Fm. never contain white mica which are diagnostic for the Carboniferous schists. The thickness of single beds of black shales is comparable to that of the sandstones (Fig 2.36). It has to be noticed that the quartz-arenites are characteristic of the external Valaisan only. In the internal Valaisan the Marmontains Fm. consist of black shales only. Therefore the thickness of this formation ranges between a few meters (5-8 m) in the internal Valaisan and up to 30 m in the external Valaisan. Despite its moderate thickness, this formation is crucial for deriving the younging direction within the post-rift sequence. The formation is extremely useful in those regions where the Aroley Formation is represented in a distal facies and therefore closely resembles the St. Cristophe Formation (e.g. within the Versoyen unit of the internal Valaisan). Within the external Valaisan the most complete sequence of the Marmontains Formation can be observed in the Combe de la Nova (x 949675; y 2085700; z 2320) as illustrated in (Fig 2.37a).

Fig. 2.36: Alternation between quartzit-arenite and black shales strata, Marmontains Formation, Moûtiers unit, external Valaisan.

The first sediments above the Aroley Formation are black shales. The top of the Aroley Fm. consists of an alternation of black shales (same as those of the Marmontains Fm.) and conglomerates of the Aroley Formation. The Marmontains Fm. starts with the first bed of sandstone. Up section the alternance between black shales and sandstones becomes more regular and the sandstone beds can reach a thickness of the 10-15m. This profile illustrate the common facies of the Marmontains Formation in the external Valaisan. Within the thickest beds of sandstone a coarse conglomerate facies can often be recognised (Fig 2.37b). This coarse conglomerate facies is always found observed towards the top of the formation (Fudral, 1973). The clasts are exclusively made up of dolomite. This facies was referred by Antoine (Antoine et al., 1992) as the “Grès grossiers et conglomeratiques” and is found in the Moûtiers unit. The Marmontains Formation of the Versoyen unit lacks quartz-arenites, the black shales are the only lithology present. The Marmontains Fm. also shows a marked thinning character from external to internal Valaisan. In the internal Valaisan, the thickness ranges from a few meters (5m) up to 20 m. However, due to the lack of quartzites, the exact limits of the formation are difficult to be traced. The transition to older and younger formations is extremely gradual.

In conclusion chances in the facies distribution of the Marmontains Fm. do occur between external and internal Valaisan. Towards the internal Valaisan the coarse conglomerates and the quartz-arenites progressively disappear. This indicates a fining trend towards the basin, in agreement with trend discussed for the Aroley Fm (see Ch.2.5.1).

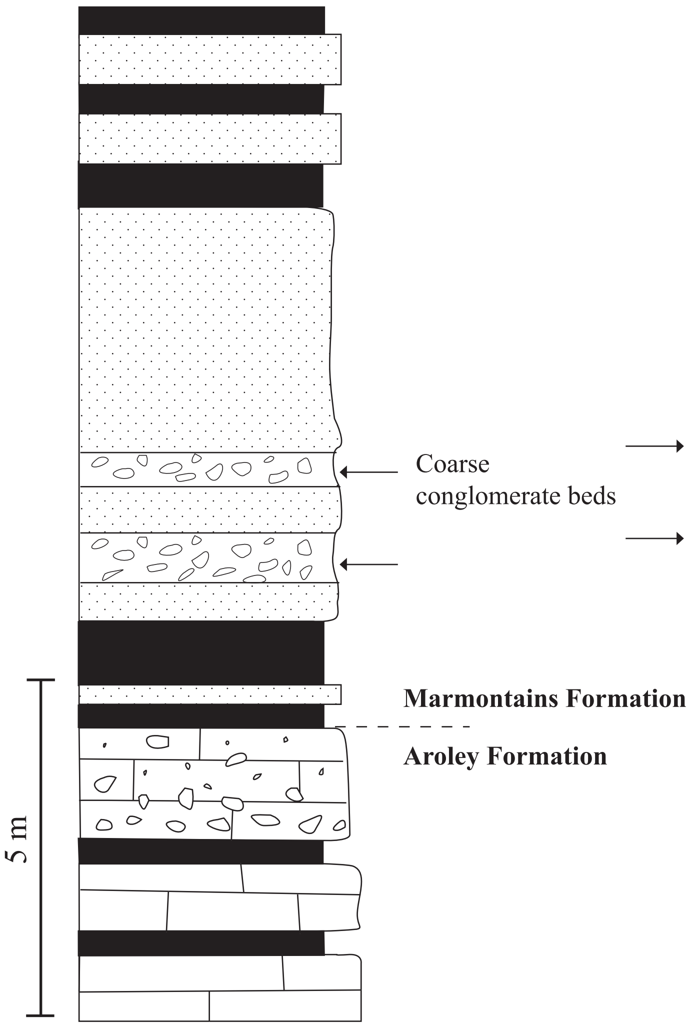


Fig. 2.37a

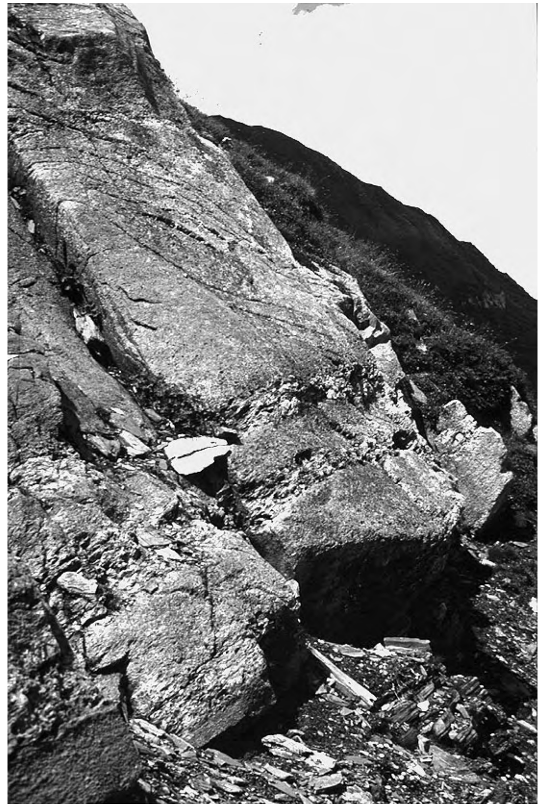


Fig. 2.37b

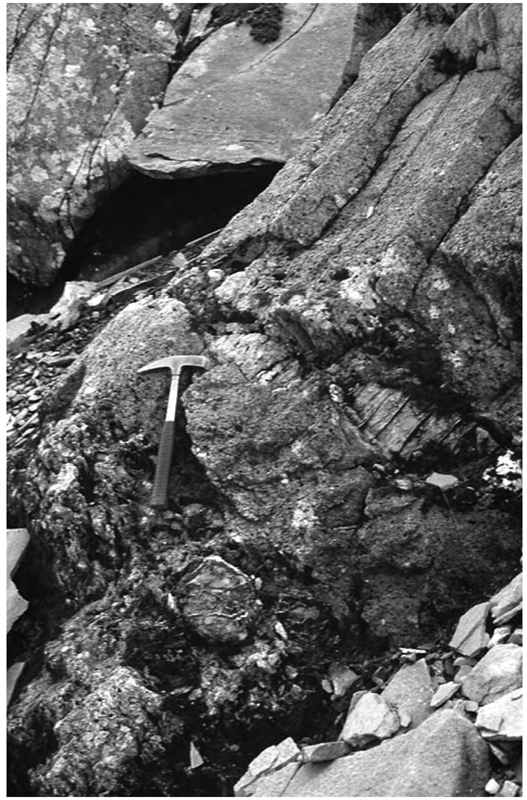


Fig. 2.37c

Fig. 2.37a Stratigraphic column of the Marmontains Formation.
Fig. 2.37b Coarse conglomerate facies in the Marmontains Formation. Black shales are at base of a massive sandstone bed. Combe de la Nova, Moûtiers unit, external Valaisan.
Fig. 2.37c detail of Fig. 2.37b showing dolomite clasts.

2.5.3: The St. Cristophe Formation

The youngest formation amongst the post-rift sediments is represented by the St. Cristophe Formation which forms the majority of the post-rift sediments. This very monotonous formation shows an alternation of calcarenaceous strata and black schists which can be interpreted in terms of a turbiditic sequence (Fig. 2.38). In some places the rhythmic character is not so pronounced and the St. Cristophe Fm. shows a calcschist facies. This formations is easily observable at the bottom of the cliff exposed in the Valleé des Glaciers. The thickness of the formation has been estimated to be in the order of several hundreds of meters (600 to 900 m by Antoine 1971). However, according to the results of our structural study, a mean value around 500 m seems more likely. Notably, the St. Cristophe Formation is very similarly developed in the internal and external Valaisan. The St. Cristophe Formation best corresponds to what is referred to as “Bündnerschiefer” in the Valaisan of eastern Switzerland.

Fig. 2.38 St. Cristophe Formation, Valleé du Versoyen, Versoyen unit, internal Valaisan. Calcareous centimetric beds with argillaceous and arenaceous interbeds.

2.6: The controversy on the age of the Valaisan rocks (post-rift sediments and Complexe Antéflysch).

The question after the age of the post-rift sediments is controversial. Much of the effort to determine the age of the Valaisan rocks concentrated on the post-rift sediments, probably because they are the most abundant and typical lithologies (Antoine, 1971; Barbier, 1948; Barbier and Trümpy, 1955; Fudral, 1998; Schoeller, 1929). In the following a historic review of this subject is presented. In previous work, the concept of pre-, syn- and post-rift sedimentary sequences was not adopted, because these definitions were developed only later on (Mitchum *et al.* 1977). Therefore we will follow the original nomenclature (referred in *italics*).

Since the Valaisan rocks only delivered few fossils, early attempts (Barbier and Trümpy, 1955) to date them consisted in lithostratigraphic correlations on a large scale. The approach of Barbier and Trümpy (1955) was based on earlier work. However, it represents the first essay towards a regional synthesis of the *zone des Brèches de Tarentaise* (Barbier, 1942). The zone des Brèches de Tarentaise was subdivided into a more external *Digitation de Niélard* and a more internal *Digitation de Moûtiers* (Barbier, 1948), the latter corresponding to our study area. Both zones consist of a *Flysch,* overlying a substratum. In case of the "Digitation du Niélard" Schöller (1929) and Barbier (1948) proposed a Tertiary age for the "Flysch" while the "Digitation du Moûtiers" represented a region where the age of the "Flysch" remained unknown to these authors. In a region located further N the Val Ferret (CH), and in a more external position in respect to the "Digitation du Moûtiers", Trümpy (1952, 1955) defined the *zone de Ferret*. This zone is characterised by a "Flysch" which is detached from its original substratum. According to this author, the base of this "Flysch" is formed by the "Couches de l'Aroley", which was dated Barrêmian-Aptian after the discovering of benthic foraminifera (*Orbitolina* sp.) by Trümpy (1954).Therefore, Barbier and Trümpy (1955) concluded that the age of the "Flysch" of the "Digitation du Moûtiers" should be "*intermédiaire*" between the Tertiary age inferred from the "Digitation du Niélard" in the SW, and the Cretaceous age inferred for the "zone de Ferret".

A Cretaceous age for the "Flysch" of the "Digitation de Moûtiers" was also proposed by Elter (1954). At the Pyramides Calcaires, he discovered small fragments of *Orbitolina* within the first strata of a series situated between the Lower to Middle Liassic limestone and the base of "Couches des Aroley". The exact attribution of these strata was controversial. According to Elter (1954) and Trümpy (in: Barbier and Trümpy 1955) they represented the base of the "Couches de l'Aroley", while Antoine (1971) ascribed this horizon to the "Complexe Antéflysch" (see below). Sodero (1967) described abundant and well preserved examples of *Orbitolina* and *Dasycladacee* discovered in the "Couches de l'Aroley" in the Sapin Valley (Aosta Valley, I). This locality, located outside the study area, clearly represents a part of the "Digitation de Moûtiers"(it is probably part of the internal Valaisan).

All this evidences which consistently points to a Barrêmian-Aptian age for the base of the "Couches de l'Aroley", was questioned by findings of Antoine (1965, 1971). Antoine (1965) discovered a single pyritised *Globotruncana lapparenti coronata* fossil within the same strata where Elter (1954) discovered Orbitolina sp.(i.e. underlying the Couches des Aroley). The inferred age for these strata (part of the Complexe Antéflysch for Antoine) was Turonian-Santonian. Consequently, he considered the overlying Couches des Aroley (referred by Antoine as to the "formation basal du Flysch") as Campanian in age.

At the end of the Sixties, the unique finding of Antoine led to two different solutions of the problem concerning the age of the Valaisan rocks. According to a first solution (Trümpy 1954, Sodero 1967, Elter and Elter 1965) the Couches de l'Aroley are regarded as Barrêmian-Aptian in age, the Couches des Marmontains as Aptian-Cenomanian ("middle" Cretaceous) and the Couches de St. Cristophe as Late Cretaceous to possibly Paleogene. Consequently, the Complexe Antéflysch was indirectly dated as Early Cretaceous. According to a second solution (Antoine 1971, Antoine et al 1993) the Complexe Antéflysch is regarded as Late Cretaceous (Turonian-Santonian) in age. Hence, the Couches de l'Aroley would be Campanian while the Couches de Marmontains and the younger Couches de St. Cristophe are tentatively dated as Maastrictian-Paleocene (Antoine 1971, Antoine et al. 1993). This solution suggests that the Orbitolina fossils found by Elter (1954) are reworked. Fragments of Orbitolina were also observed during this study in the Pyramides Calcaires Fm. However, the question if they where reworked remains.

This study cannot presents new fossil findings which would resolve this problem. A single fossiliferous horizon found in the Complexe Antéflysch Fm. yielded pyritised microfossils which proved to be both unidentifiable (probably Radiolarians) and of little use for dating these rocks (Fig. 2.39). Unfortunately, the examination of several thin sections from the locality where Antoine discovered his Globotruncana (Ravin de la Chail, high Valleé des Glaciers) did not reveal any fossils providing additional arguments for the age proposed by Antoine (1965, 1971). Unfortunately, the thin section described by Antoine (1965) is lost (pers. com. Antoine, January 2000) and a confirmation of the attribution of its fossil was no more possible.

However, lithofacies correlations with other Valaisan sequences on a large scale furnish alternative arguments for deciding on the age of the post-rift sediments. The north-eastern continuation of the Valais domain is found in the zone of Sion-Courmayeur of Switzerland (Jeanbourquin and Burri, 1991; Trümpy, 1988) and further east in the Misox zone (Steinmann, 1994). It is widely assumed

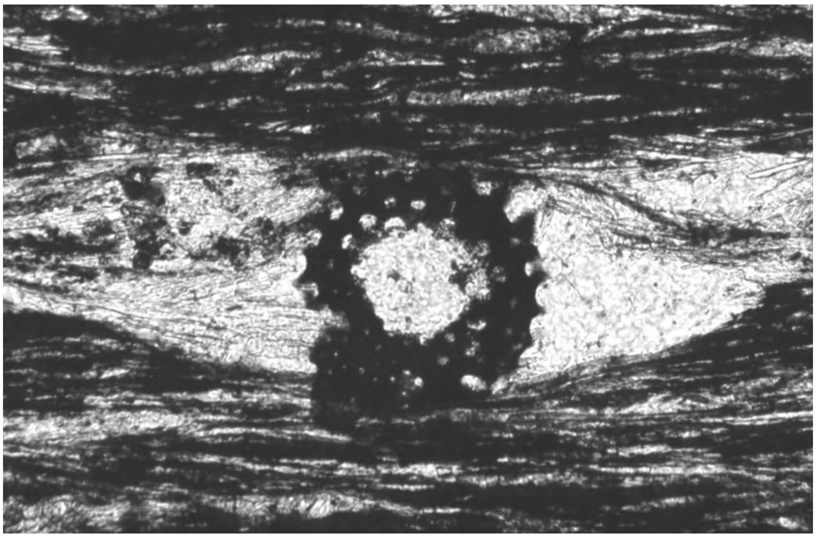


Fig. 2.39 Photograph of thin section of the black schists of the Complexe Antéflysch Fm. Probable radiolarian fossil. (A. du Clapet region, AL 9617b; x: 948125, y: 2081950, z: 1810).

that the easternmost prolongation of the Valais domain is exposed in the Engadine and Tauern windows (Froitzheim et al., 1996; Trümpy, 1988). Due to change in the strike of the Alpine chain, the term “Valaisan domain” is generally replaced with the equivalent term “ North Pennine Basin” (NPB, Froitzheim et al. 1996). The low-grade Misox zone (lower greenschist facies Steinmann and Stille 1999) consists of large masses of Bündnerschiefer sediments and basaltic rocks (Steinmann, 1994). According to Steinmann (1994) this zone represents the nappe stack of two tectonic units called Tomül and Grava nappes, intercalated between three melanges zone and one “Schuppenzone” (Valser Melange, Aul Schuppenzone, Grava melange and Tomül melange). The Tomül and the Grava units contain the best preserved sedimentary sequence of Bündnerschiefer which, in the Tomül units stratigraphically overlay basaltic rocks. The thick (2.5 km) monotonous sedimentary sequence of the Bündnerschiefer of the Tomül and Grava units, several formations were lithostratigraphically defined such as: Bärenhorn Fm., Nollaton Fm. and Nollakalk Fm. The Nollaton Fm. consists of a composite thick (400-600m) succession of black shales to schists, alternating with calcareous sandstones and green to grey quartz sandstones. The black shales are generally carbonate-free. The overlying Nollakalk Fm. is dominated by a monotonous (350-450 m) repetition of calcareous sandstones alternating with black schists to form turbiditic sequences.

When the Nollaton and the Nollakalk Fms. are compared with the post-rift sediments of the study area, they show remarkable lithostratigraphic similarities with the Marmontains and the St.

Cristophe Fms. In particular, the Nollaton Fm. displays a characteristic facies association of black shales and green quartz sandstones which is lithologically directly comparable with the quartz sandstones of the Marmontains Fm. Another argument consists in the presence, in both formations (the Nollaton and the Marmontains Fms.) of a characteristic carbonate-free black shales interval. This particular stratigraphic feature is correlated with the anoxic event which in the middle Cretaceous characterised the entire Tethyan realm (middle Cretaceous 'global anoxic events' Allen and Allen 1992). It is therefore likely that the Nollaton Fm. and the Marmontains Fm. are of the same age. The overlying Nollakalk Fm. with its monotonous repetition of calcareous sandstones and black schists can therefore be lithologically correlated with the St. Cristophe Fm. However, the Nollaton and the Nollakalk Fms. represent an almost completely fossil-free series. Based on stratigraphic comparisons and geochemical investigations (Neodymium isotopes, organic carbon) Steinmann (1994) proposed an Aptian-Albian age for the Nollaton Fm. and a Cenomanian to younger age for the Nollakalk Fm. Hence, an Aptian-Albian age for the Marmontains Fm. and a Cenomanian age for the St. Cristophe Fm. can reasonably be assumed for the Valaisan domain of the study area. Moreover the Aroley and Marmontains Formations of the Valaisan can be lithostratigraphically correlated to the Tristel and Gault Formations in the Falknis and Tasna nappes, respectively. The Barrêmian to Aptian Tristel Fm. (Schwizer, 1983) is made up by calciturbitites with breccia horizons with a gradational upward transition into the Aptian to Cenomanian sandstones and black shales of the Gault Fm. (Hesse, 1973, Allemann, 1957).

In conclusion, correlation of the Marmontains Fm. with the middle Cretaceous anoxic event and with equivalent formations, strongly pointing to a middle Cretaceous age for this formation, which is in contrast with the Late Cretaceous age proposed for this formation by Antoine (1971). Therefore, ages suggested by Trümpy (1954) seem more reasonable and are also proposed in the present study. According to several authors (Frisch, 1979; Froitzheim et al., 1996; Stampfli, 1993; Stampfli and Marthaler, 1990; Stampfli et al., 1998; Steinmann and Stille, 1999; Trümpy, 1976) the Valais basin evolved into an oceanic domain which is now incorporated, as an ophiolitic suture zone, within the Pennine nappe pile all the way from the western to the eastern Alps. The opening of the Valais ocean has been connected with the opening of the Atlantic, i.e. the Bay of Biscay during the Late-Jurassic-Early Cretaceous period. In our study area this period corresponds to the deposition of the syn-rift sediments (Late- Liassic-pre-Barrêmian) and formation of the Complexe Antéflysch (Neocomian?). This is in accordance with a Barrêmian to Aptian age for the first post rift sediments, as proposed by Trümpy (1954).

CHAPTER 3: THE B RÈCHES DU GRAND FOND GROUP

Introduction

It was proposed in Chapter 2 that the external Valaisan suffered two subsequent rifting events. An older one related to the deposition of a thick deep-water Middle to Late Liassic basin in the Pt. St. Bernard unit is ascribed to the Piemont-Ligurian rifting (Middle Liassic to Dogger). The younger and second rifting event was responsible for the main stratigraphic features which characterise the Valaisan units (e.g. the syn- and post-rift sequence) and is related to the Valaisan rifting (Upper Jurassic to Lower Cretaceous). This chapter focuses the attention to the syn-rift sediments of the external Valaisan (Brèches du Grand Fond Group) which theoretically encompasses a wide age bracket between Upper Liassic and Barrêmian (Fig. 3.1). This very long time interval makes it possible that both rifting stages may be recorded by the conglomerate of the Brèches du Grand Fond Group. However, because of dating problems it is hard to decide how much of the Brèches du Grand Fond Group may have formed during each rifting stage. Based on detailed field work, the first part of this chapter (Ch. 3.1 and 3.2) will describe the facies of the Dent d'Arpire and Pyramides Calcaires Formations (Brèches du Grand Fond Group, Moûtiers unit) in more detail. In the second part (Ch. 3.3) timing and depositional setting of the Brèches du Grand Fond Group will be discussed.

The syn-rift sediments consist of the Brèches du Grand Fond Group of the Moûtiers unit (external Valaisan) and the Brèches du Collet des Rousses Fm. of the Versoyen unit (internal Valaisan) while the Complexe Antéflysch Fm. of the Versoyen unit is considered as "transitional" between syn- and post-rift sediments. The Brèches du Grand Fond Group, occurring in the NW part of the study area, is subdivided into two formations: the lower Dent d'Arpire Fm. and the younger Pyramides Calcaires Fm.

A complete section of the syn-rift sediments of the external Valaisan can be studied in the Combe de la Nova region, SW of Les Chapieux (Fig. 3.2). Toward NE the Brèches du Grand Fond Group outcrops along the higher parts of the Valleé des Glaciers. In this area the syn-rift sediments are represented by the younger Pyramides Calcaires Fm. only. In Fig. 3.2 two main unconformities, which define the depositional interval of the syn-rift deposits, are mapped.

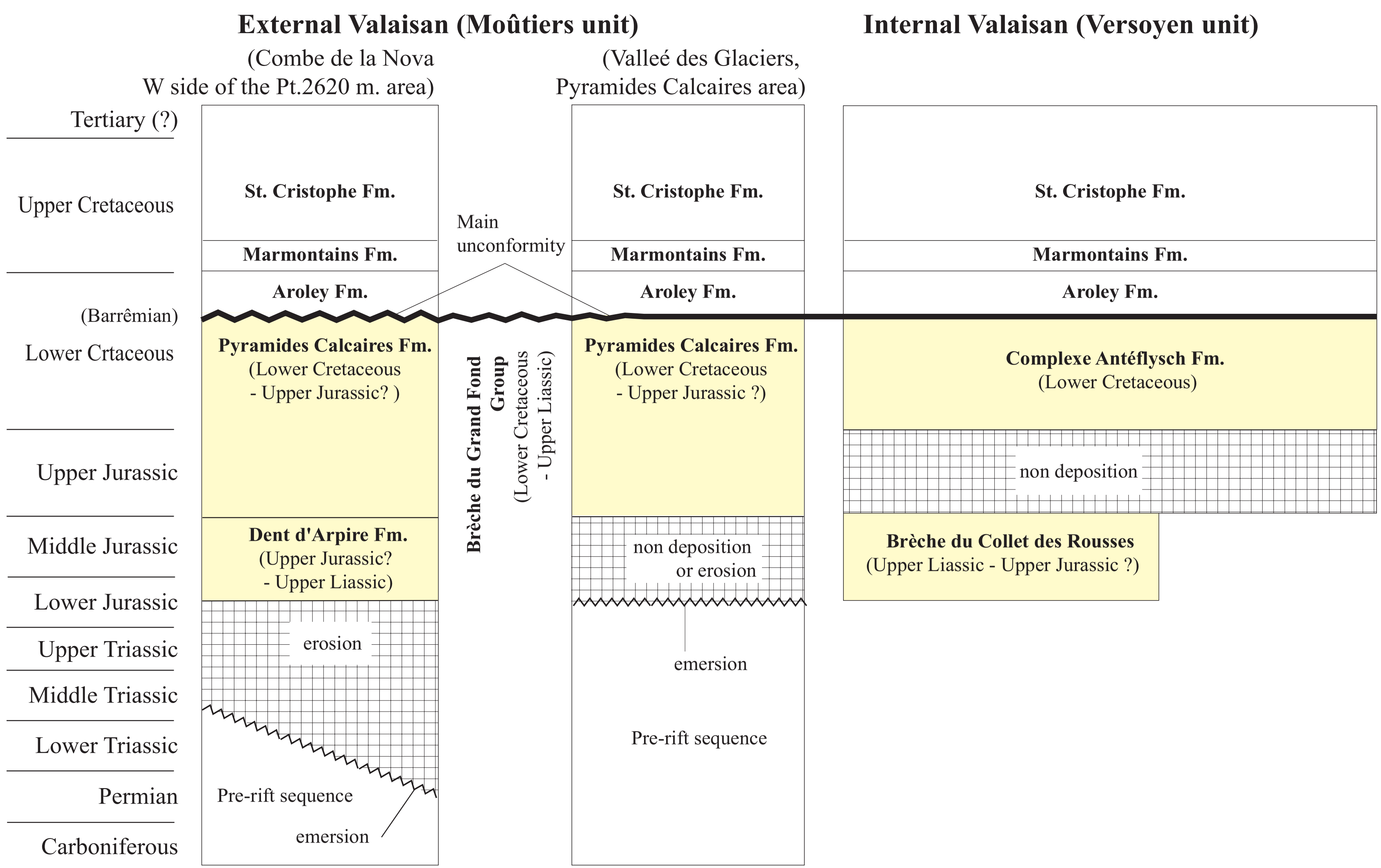
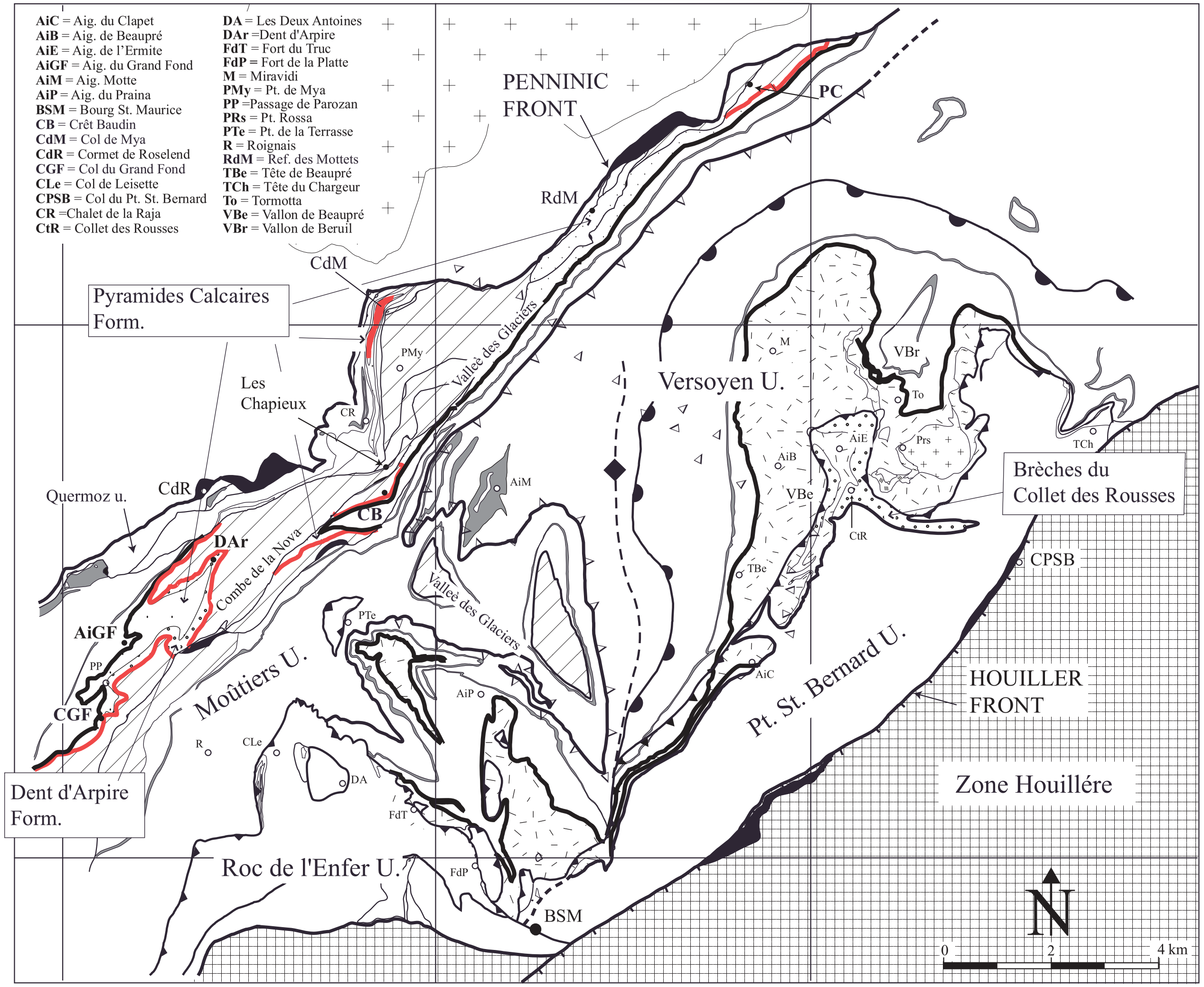


Fig.3.1 Lithostratigrapy of the Moûtiers and Versoyen units. In both units the formations indicated in yellow were deposited sometimes between Upper Liassic and Barrêmian. All of them rest below a regional scale unconformity at the base of the Aroley Fm. (post-rift) which is expressed as an angular unconformity in the outer basin (external Valaisan), passing basinwards (internal Valaisan) to a correlative conformity (Barrêmian).



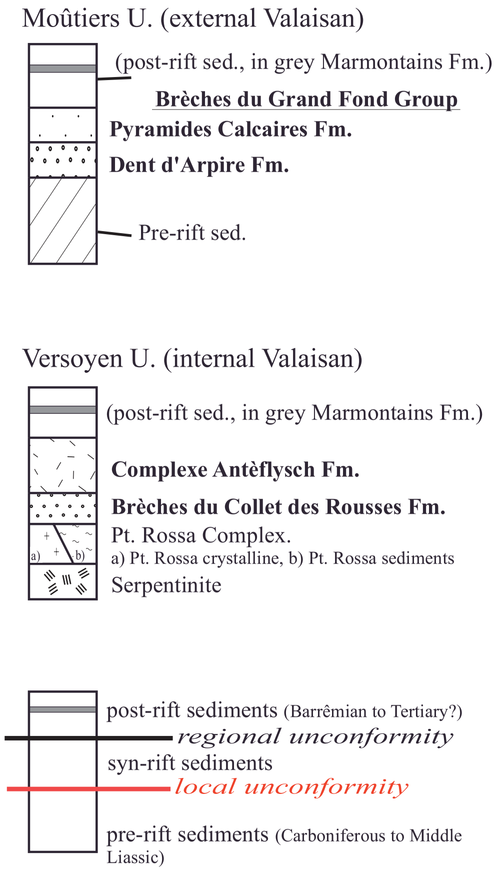


Fig. 3.2 Outcrop distribution of the syn-rift sediments within the study area.

3.1: The Brèches du Grand Fond Group in the Combe de la Nova region: the Dent d'Arpire and the Pyramides Calcaires Formations

Fig. 3.3 presents a detailed geological map of the Combe de la Nova region and the location of the outcrops described in this chapter. The Brèches du Grand Fond Group is described in two selected areas. One, located to the N (area A), is situated on the western slope of the peak quoted 2620 m. The second one in the S (area B), covers a larger area near the Col du Grand Fond (Fig. 3.3).

3.1.1: The Dent d'Arpire area

The area A of Fig. 3.3 is part of an NE-SW oriented ridge formed by the Dent d'Arpire-Aig. du Grand Fond-Pointe du Presset peaks, limiting the Combe de la Nova from the meadows of the Cormet de Roselend pass. The steep NW side of this ridge offers cliff sections where the basal unconformity of the Brèches du Grand Fond Group can be well observed (Fig. 3.4). Two outcrops are situated on the normal and on the inverted limbs of a large scale W-facing F2 antiformal anticline, respectively (Fig. 3.5). Therefore, this area offers the possibility to correlate (after retro-deformation to their original position before D2) these two outcrops directly. At the topographically higher outcrop (i.e. located on the normal limb of the F2 fold, see Fig. 3.4) Permian rocks (below) are in contact with a thick sequence of the Brèches du Grand Fond Group (above).

For a schematic stratigraphic section representative for this outcrop the reader is referred back to Fig. 2.13a. The lowermost part of the Brèches du Grand Fond Group, the Dent d'Arpire Fm., is characterised by a good stratification and by a massive conglomeratic bed at the top of the formation. In this section, the thickness of the Dent d'Arpire Fm. reaches its maximum value of around 40-60 m. The rest of the exposed Brèches du Grand Fond Group consists of the younger Pyramides Calcaires Fm. This is clearly recognized by means of the presence of large (0,5 to 20m) breccia components and by the absence of a clearly defined stratification (Fig. 3.4).

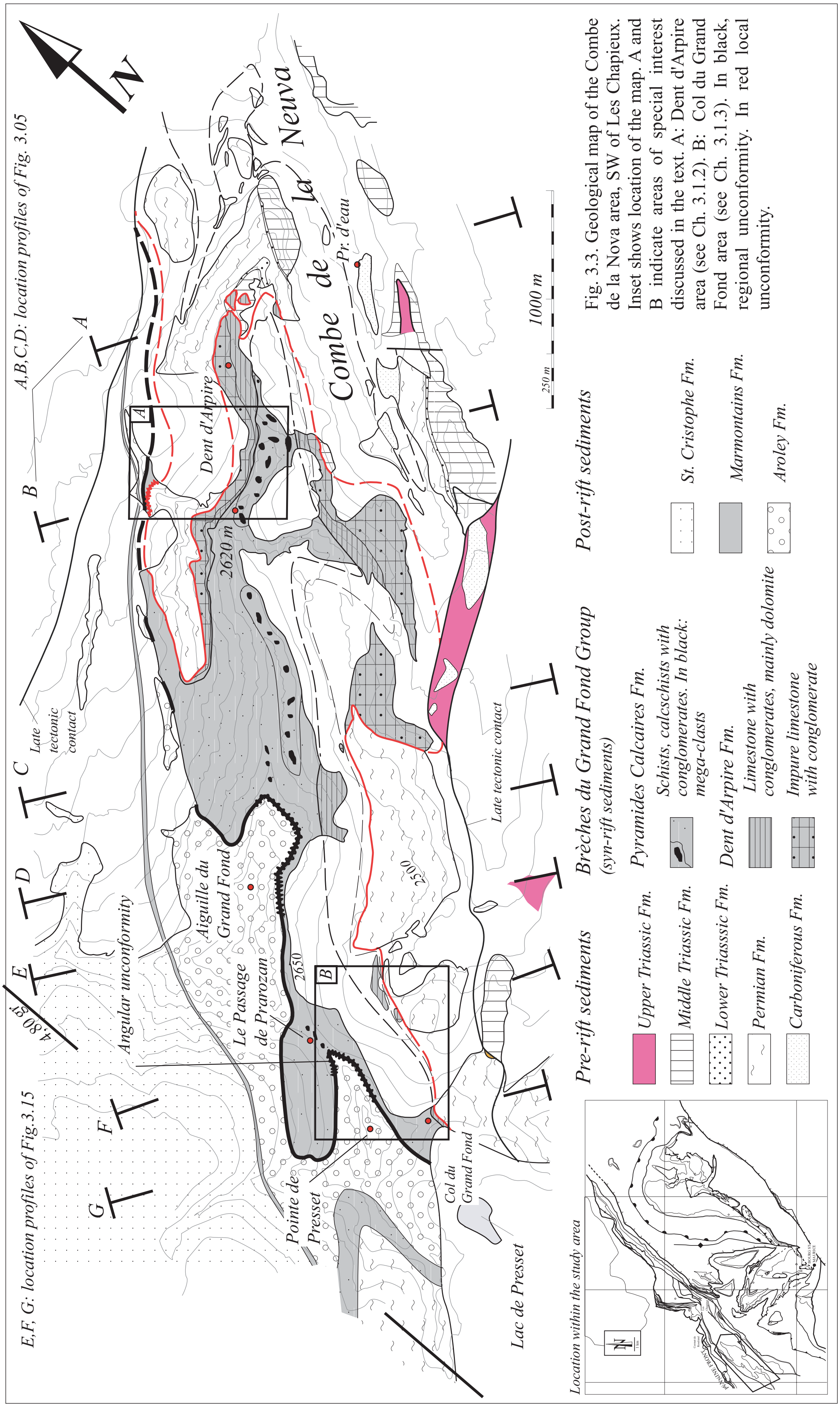


Fig. 3.3. Geological map of the Combe de la Nova area, SW of Les Chapieux. Inset shows location of the map. A and B indicate areas of special interest discussed in the text. A: Dent d'Arpire area (see Ch. 3.1.2). B: Col du Grand Fond area (see Ch. 3.1.3). In black, regional unconformity. In red local unconformity.

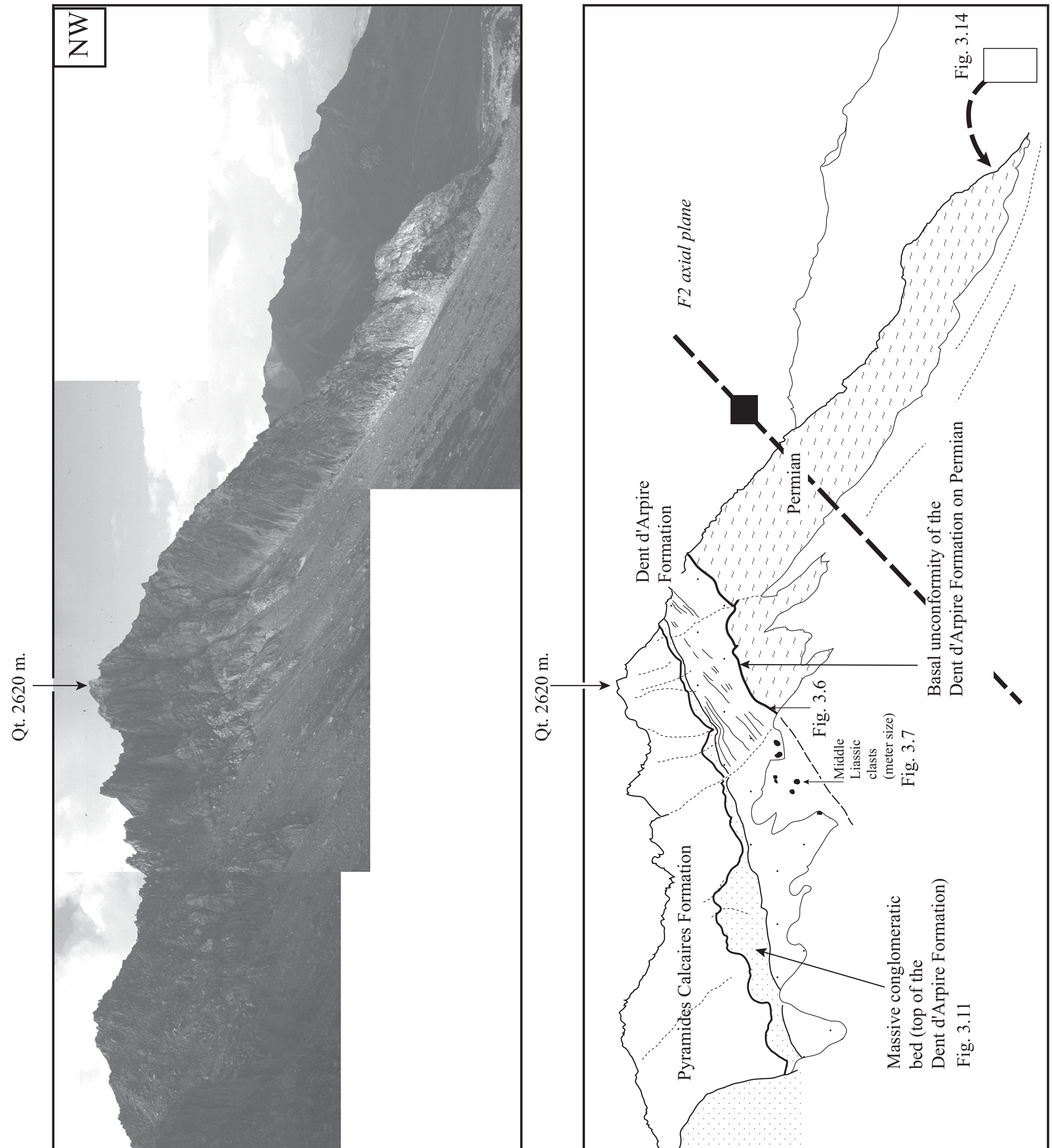


Fig. 3.4. The Brèches du Grand Fond Group at the slope W of Pt. 2620 m. Insets show location of the outcrops discussed in Ch. 3.1.1.1. See also Figs. 2.13 and 2.14 of Ch.2.

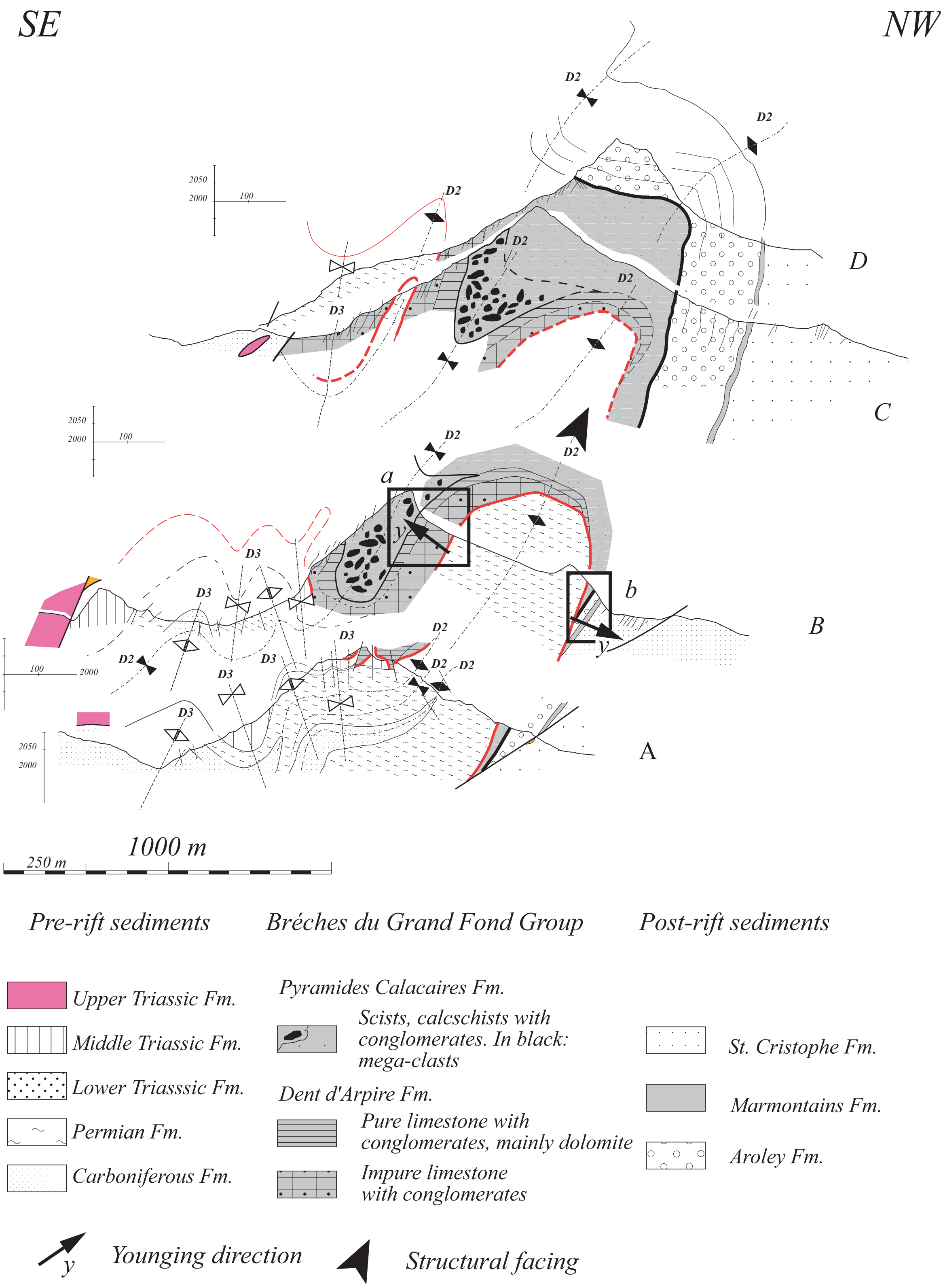


Fig. 3.5. Structural framework of the outcrops described in text. Frames discussed outcrops. Location of the profiles in Fig. 3.3. The two outcrops areas, a and b, located on the western slope of the point quoted 262 0m., rest on the normal and inverted limb of a NW-facing F2 antiformal anticline.

At this outcrop, the contact which separates the Dent d'Arpire Fm. from the underlying Permian quartz-arenites and conglomerates is locally concordant (Fig. 3.6a), but clearly discordant at a larger scale. The lack of evidence for subaerial deposits (e.g. hematite pigmentation) along the contact and within the first conglomerates, suggests that the deposition most likely took place in an aquatic environment.

Fig. 3.6a. Contact between the base of the Dent d'Arpire Fm. and the Permian Fm. (open arrow), Moûtiers unit external Valaisan (see also Fig. 3.7b). The firsts beds consist of a coarse clast-supported conglomerate (see Fig. 3.6b) intercalated with dark argillaceous layers (black arrow). Dog for scale bar approximately 70 cm high. Coord. x:938750, y:208650, z: 2350.

Fig. 3.6b. Characteristic aspect of the lower part of the Dent d'Arpire Fm., Moûtiers unit, external Valaisan. In this example the clasts (mainly dolomites) are embedded in an impure schistose lightly-coloured limestone. This exposure is located a few meters above the contact shown in Fig. 3.6a.

These first beds consist of coarse, clast-supported, 0.5 to 2 m thick beds of conglomerates which display a marked lateral variation in the amount and the nature of the matrix, even over a distance of a few meters. Often, the matrix is an impure schistose light-coloured limestone (Fig. 3.6b). In other cases the clasts are enclosed in dark argillaceous sediments which only occur in the lowermost part of the Dent d'Arpire Fm. Up section, the proportion of calcareous matrix increases towards a light-grey to dark-blue pure limestone composition.

Despite the change of composition of the matrix, the clast composition remains constant throughout the whole Dent d'Arpire Fm. Up section the pebbles mainly consist of dolomites and limestones with strong affinities to the lithologies of the pre-rift sediments. Of particular interest is the suite of clast-types found in the first 10-20 m above the Permian. In the first 5 m. above the Permian formation massive white limestone clasts (Fig. 3.7) were observed. Based on facies analogy, this lithology is interpreted as Lower to Middle Liassic limestone and it probably represents the youngest extrabasinal lithology (i.e. the youngest pre-rift sediment).

Fig. 3.7a

Fig. 3.7b

Fig. 3.7a,b. Details of the clast suite in the first meters of the Dent d'Arpire Fm. above the Permian Fm. Fig. 3.7a: large meter size boulder of Lower to Middle Liassic limestone in the lower part of the Dent d'Arpire Fm., Moûtiers unit external Valaisan. This lithology is interpreted as the youngest extrabasinal component. For location see Fig. 3.4. Fig. 3.7b, the same white limestone (black arrow) occurs in the first bed of the Dent d'Arpire Fm. Open arrow indicates the contact with the Permian Fm..

The other pebbles are grey to dark blue limestones (Fig. 3.8) and brownish coarse calcareous sandstones (Fig. 3.9). Both kind of pebbles have a decimetre to metric size and characteristically a tabular shape. The grey to bluish limestone is interpreted as an intrabasinal clast reworked during the deposition of the Dent d'Arpire Fm. The calcareous sandstone is characterised by a light grey to white colour in fresh fracture and a brownish alteration. The sand grains consist of quartz and rock fragments (dolomite). Typically, the sand-sized quartz grains stand out on weathered surfaces. In thin section, the interstitial space is occupied by calcite and fine grained white mica. Quartz forms polycrystalline millimetre-sized grains characterised by sutured internal grain boundaries. This feature is indicative for a metamorphic source of this quartz.

Fig. 3.8 Detail of the clast suite of the basal part of the Dent d'Arpire Fm., Moûtiers unit external Valaisan. In this example decimetre-sized tabular shape clast of grey limestone was observed. This lithology corresponds to part of the matrix of the Dent d'Arpire Fm. Hence, it is interpreted to have been reworked soon after its deposition (intrabasinal clasts). It has to be noticed that the matrix is massive and consists of an almost pure limestone. Compare the aspect of this exposure to the matrix in Fig. 3.7b which is characteristic for the lowermost part of the Dent d'Arpire Fm..

The coarse calcareous sandstones are also interpreted to represent intrabasinal clasts (i.e. they were also deposited at the same time as sedimentation of the Dent d'Arpire Fm.). The coarse calcareous sandstone is of particular interest, because it has never been observed in the stratigraphic succession of the pre-rift sediments, nor as a component in the post-rift sequence (Aroley Fm.). Large parts of the Dent d'Arpire Fm. exhibit a crude stratification, gently inclined in respect to both the base and the top of the formation (Fig. 3.4).

Fig. 3.9. Detail of a clast of calcareous coarse sandstone referring millimetre sized, well rounded quartz sand grains set in a calcareous matrix. In this lithology the coarse sand grains tend to stand out relative to the calcareous matrix which is easily dissolved out at the surface. This kind of component occurs only in the Brèche du Grand Fond Group and was never observed at the outcrop within the pre-rift sediments sequence nor as clast within the conglomerates of the Aroley Fm.. Therefore this lithology is interpreted as a intrabasinal lithology.

This feature, which is absent within individual beds, is interpreted as a large scale cross stratification (clinoform) affecting the entire formation. This cross bedding can be better observed along a NE-SW profile (roughly perpendicular to the view of Fig. 3.4). The dip to the NE of the clinoform suggests that the depositional system propagated from SW to NE. This orientation is obtained after retro deformation (see below). The uppermost bed of the Dent d'Arpire Fm. also part of the clinoforms is a particular one and consists of a 10-20 m massive conglomerate. It outlines the contact with the overlying Pyramides Calcaires Fm. This thick bed contains abundant small and well rounded clasts (mainly dolomitic), regularly dispersed in a pure limestone matrix (Fig. 3.10).

Fig. 3.10. The uppermost bed of the Dent d'Arpire Fm., consisting of a massive limestone matrix with centimetre sized pebbels.

Moreover, it displays a greater lateral facies continuity, when compared to the other facies of the Dent d'Arpire Fm. For this reason it was mapped separately in Fig. 3.3 as "limestone with conglomerate". However, it is interesting to note that the thickness of this bed (Fig. 3.11) varies laterally ranging from a few meters to 10-20 m. The base of this bed is planar and concordant with the lower formation, its top shows a strongly irregular shape. This observation suggest that the top of the Dent d'Arpire Fm. corresponds to an erosional surface formed before the deposition of the younger Pyramides Calcaires Fm.

The younger Pyramides Calcaires Fm. starts with an highly variable set of polymictic beds, comprising large boulders embedded in a calcareous to pelitic matrix. In this area, the Pyramides Calcaires Fm. consists of a clast-supported poorly sorted conglomerate. A clear stratification is totally absent, only axial plane surfaces (mainly S4) can be observed. Eastwards, the proportion of large clasts increases. Their long axes may reach up to 20-30 m. (e.g. mega clasts of Lower to Middle Liassic limestones Fig. 3.12). Blocks of Permian and Middle Triassic dolomite are also commonly observed.

The large clasts are embedded in a clast-supported matrix of decimetre sized conglomerate. This particular coarse grained facies of the Pyramides Calcaires Fm., which contains clasts larger than any other facies, reaches a thickness of about 100m.

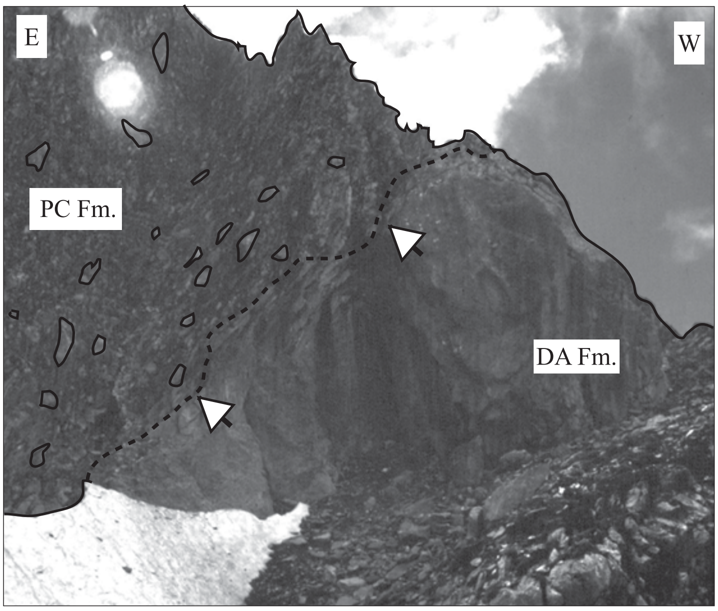


Fig. 3.11 The contact of the Dent d'Arpire Fm. (DA Fm.) with the younger Pyramides Calcaires Fm. (PC Fm). Here, the Pyramides Calcaires Fm. consists of a mega breccia. The uppermost thick conglomerate bed of the Dent d'Arpire Fm. shows an upper surface characterised by an irregular topography (open arrows). This is interpreted as an erosional surface, pre-dating the deposition of the Pyramides Calcaires Fm. (see Ch. 3.1.2).

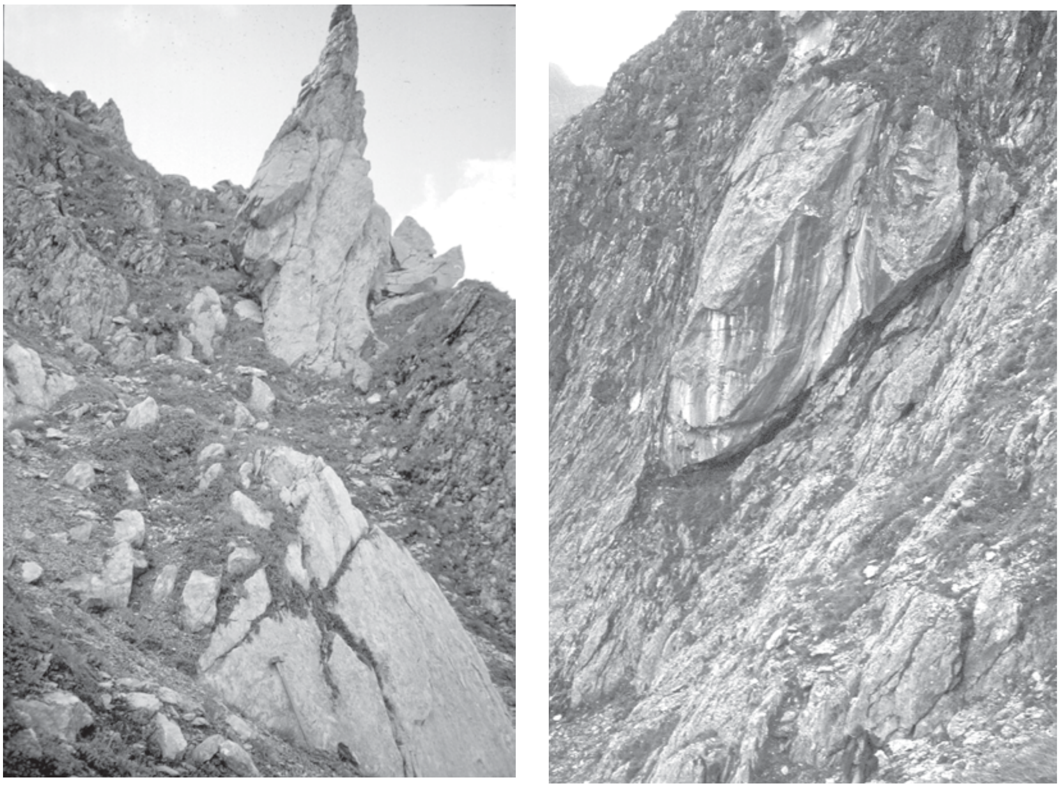


Fig. 3.12a

Fig. 3.12b

Fig. 3.12a and b. Examples of Lower to Middle Liassic mega-clasts within the Pyramides Calcaires Fm on the E slope of the Pt. 2690, Combe de la Nova (Moûtiers unit, external Valaisan).

The second outcrop, located at the lowermost base of the cliff section (for location see Fig. 3.4; cord. x 938500; y 2085125; z 2230), offers the possibility to study a stratigraphic series enclosed between the Permian series and the Aroley Fm. This interesting exposure in the Combe de la Nova region was subject to previous detailed descriptions (Antoine et al. 1972, Fudral 1973, Fudral 1998). However, due to dating problems different authors presented different lithostratigraphic interpretations (for a synthesis see Fudral 1998). Note that this place is located on the inverted limb of the F2 fold. At this outcrop, a 5 to 10 m thick white to bluish massive limestone (Fig. 3.13a) directly overlies the Permian series. Up section follow grey to brown schists and calcschists, containing conglomerates, which stratigraphically pass into the Aroley Fm. The contact of the limestone with the Permian series displays a centimetre thick reddish deposit (Fig. 3.13b) which results from sub-aerial exposure of the Permian (paleokarst surface), pre-dating the deposition of the massive limestone (Antoine et al. 1972). Another unconformity can be observed at the contact between the limestone and the overlying calcschists (see Fig. 6 of Antoine et al. 1972). In contrast to previous work, which considered the first beds above the paleokarst to be Liassic in age (Antoine et al. 1972), our observation indicates that this limestone belongs to the Brèches du Grand Fond Group.

Fig. 3.13a Detail of the same contact further illustrated in Fig. 3.13b. The first meters of the Dents d'Arpire Fm. consist of a bluish massive limestone.

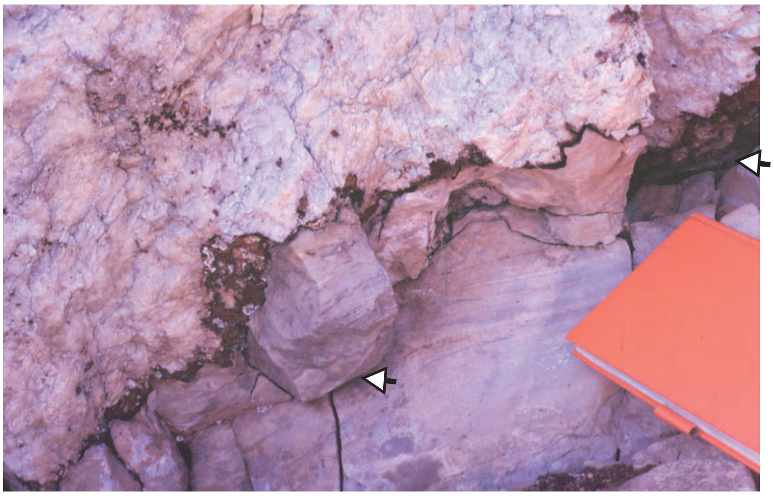
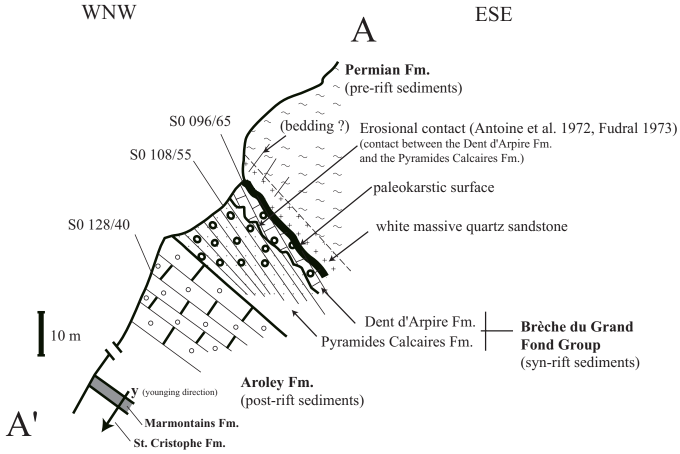


Fig. 3.13b Millimetre to centimetre thick paleokarstic surface (open arrow) at the contact between the Dent d'Arpire Fm. (below) and the Permian (above). The uppermost meters of the Permian consist of a massive white quartz sandstone. This outcrop is located on the inverted limb of an F2 fold (see Fig. 3.4).

This correlation is supported by the observation that centimetre-sized clasts, mainly dolomites, are embedded within this limestone (Fig. 3.14) which in fact represents a fine conglomerate.



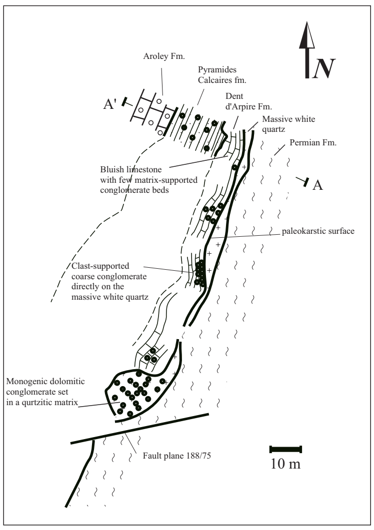


Fig. 3.14. Overview of the outcrop area located at the base of the Pt. 2620 (see Ch. 3.1.1 and Fig. 3.4 for location). Above: simplified section modified after Antoine et al. 1972 and Fudral 1973,1998. Left, sketched geologic map of the outcrop area, showing the occurrence of conglomerate within the first meters of the Dent d'Arpire Fm., above the Permian Fm.

In analogy with the outcrops described before, this conglomerate is included between the Permian series and an erosional contact with schists and calcschists at its top. This allows to correlate this horizon with the Dent d'Arpire Fm. The overlying schists and calcschists represent the Pyramides Calcaires Fm. according to our interpretation.

The erosional contact of the Dent d'Arpire Fm. with the Pyramides Calcaires Fm. indicates erosion of the Dent d'Arpire Fm. before the deposition of the Pyramides Calcaires Fm. In comparison to previous outcrop, the Dent d'Arpire Fm. shows a very reduced thickness (5 to 10 m), the same for the Pyramides Calcaires Fm. (30 m, see Fig.3.14).

3.1.2: The Col du Grand Fond area

In the Col du Grand Fond area the thin Dent d'Arpire Fm. unconformably overlies the Permian series (Figs. 3.3 and 3.15) which do not show evidence for pre-depositional emergence (e.g. hematite pigmentation). The contact surface is planar and subparallel to bedding of the first meters of the Dent d'Arpire Fm. This formation starts with a massive bluish limestone. This limestone is followed by a thick section of the Pyramides Calcaires Fm., which is in turn overlain by the Aroley Fm. (at Passage de Prarozan, see Ch. 2.5.1.1). The following description particularly refers to the Col du Grand Fond area situated on the northern slope of this pass. This region, despite less perfect outcrops in respect to the other formations, offers good exposures of the Dent d'Arpire Fm.

The thin Dent d'Arpire Fm. contains coarse clast-supported horizons which are irregularly interbedded with matrix-supported intervals (Fig. 3.16). Bedding is characteristically 0.5 to 2 m thick and displays a well-developed stratification. The irregular contact surface between clast-supported and underlying matrix-supported beds can be explained by the formation of erosional channels. Locally, coarse clast-supported 3 to 5 m. thick well sorted conglomerates can be observed, as well as normal grading (Fig. 3.17). The matrix-supported beds consist of a massive blue limestone with small, dispersed conglomerate pebbles. This matrix contains fossils of crinoids, belemnites and large ammonites (*Lithoceras sp*., L. Hottinger, pers. comm.) (Fig.3.18).

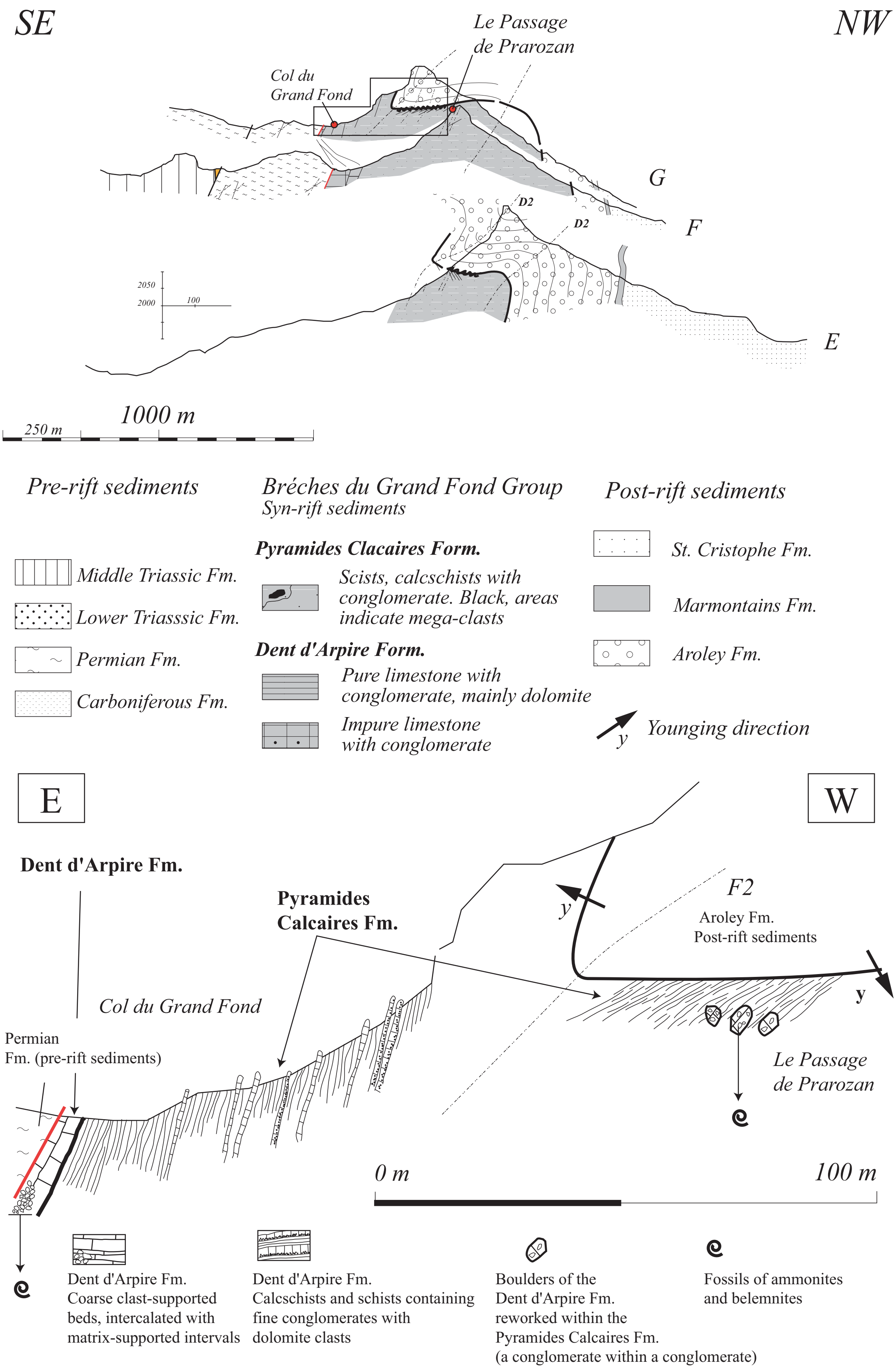


Fig. 3.15 Structural framework (above) of the area near the Col du Grand Fond. Below: enlargement of the inset of profile G and schematic stratigraphic section. In this area the Dent d'Arpire Fm is reduced to a few meters. Location of the profiles is indicated in Fig. 3.3. The local unconformity at the base of the Dent d'Arpire Fm. is shown in red.

Fig. 3.16. The Dent d'Arpire Fm. in the Col du Grand Fond area (Moûtiers unit, external Valaisan). Generally this formation consists of coarse clast-supported and very well sorted intervals, intercalated with matrix-supported beds. This example shows a normal graded coarse interval (thin black arrow) which is presently overturned. The top of this horizon is concordant with bedding (open arrows) while its base shows an irregular shape (black arrow) due to local channeling.

The ammonites suggest an Upper Liassic (Toarcian) age for this part of the Dent d'Arpire Fm. The clasts show the same spectrum as in other occurrences of the Dent d'Arpire Fm. (Fig. 3.19).But there are several differences regarding the fragment sorting of clasts: (1) a general fining of the clasts size, which ranges between a decimetre and 1 m, (2) the lack of boulders of Lower to Middle Liassic limestones near the contact to the Permian, and, (3) the occurrence of clasts of granite, micaschists, Permian and Carboniferous lithologies.

However, the brownish coarse calcareous sandstones were also found here. In some cases it was observed that this calcareous coarse sandstone is conglomeratic, containing clasts of Middle Triassic dolomites.

Fig. 3.17. Dent d'Arpire Fm. in the Col du Grand Fond are (Moûtiers unit, external Valaisan). Coarse clasts supported conglomerate, which in some cases, displays normal grading. Black arrows indicate repeated depositional sequences indicating normal grading.

In the Col du Grand Fond area, the facies of the overlying Pyramides Calcaires Fm. is finer grained. The matrix is more variable, and everywhere shows a markedly schistose character. The most typical facies are brownish calcschists with thin clast-rich layers and dark grey to black schists, bearing rusty iron nodules. In the clast-rich layers Middle Triassic dolomite pebbles predominate.

Locally, large clasts of Carboniferous schists, Permian and Middle Triassic dolomite can be observed.

Fig. 3.18 Fossil of ammonite (*Lithoceras sp.*, L. Hottinger, pers. comm.) Dent d'Arpire Fm., Brèches du Grand Foud Group, Moûtiers unit, external Valaisan. Scale is 18 cm.

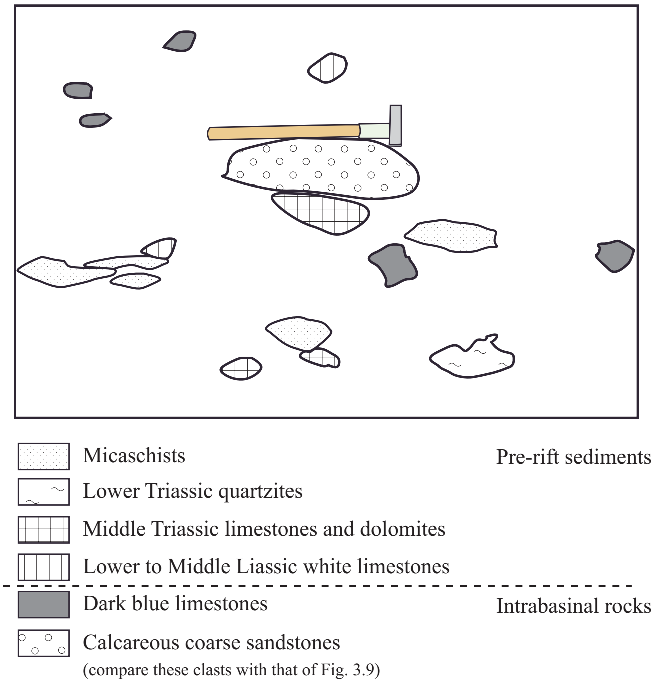


Fig. 3.19 Coarse-clast supported and moderately sorted conglomerate facies of the Dent d'Arpire Fm. in the Col du Grand Fond area (Moûtiers unit, external Valaisan).
The clasts include:

Interesting is the presence of large boulders of a reworked conglomerates (Fig. 3.20). Such clasts can be observed along the crest between the Aig. du Grand Fond peak and the point quoted 2692 m (cord.: x 938075, y 2084400, z 2670), as well as at the Passage de Prarozan close to the Col du Grand Fond (Antoine et al., 1972).

Fig. 3.20. Large well rounded boulders of conglomerate (open arrows) within the Pyramides Calcaires Fm. (Moûtiers unit, external Valaisan)

The large (2-3m) well rounded boulders bear a light grey to blue limestone matrix with centimetric clasts of dolomite. The matrix of one of these clasts revealed well preserved belemnites (Fig. 3.21). Based on facies analogies and the belemnite fossils, this clast-type may be interpreted as being reworked from the Dent d'Arpire Fm.

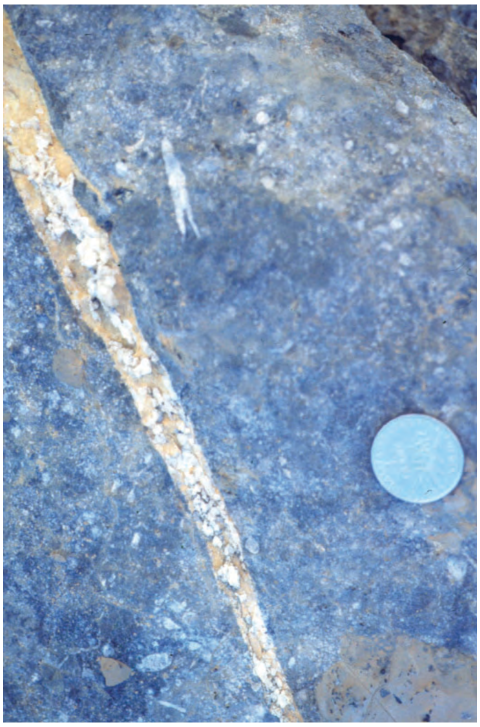


Fig. 3.21. Detail of one of the boulder shown in Fig. 3.20.
A belemnite is contained in the matrix of a dolimite-bearing conglomerate. This conglomerate represents a boulder in a younger conglomerate. Thus large boulder was derived from the Dent d'Arpire Fm.

It is proposed that it may have been derived from the uppermost thick massive conglomerate bed below the contact with the Pyramides Calcaires Fm. The rounding of the boulders suggests that these clasts were eroded and transported after complete lithification.

3.2: The Brèches du Grand Fond Group in the Valleé des Glaciers: the Pyramides Calcaires Formation

3.2.1: Description of the various lithologies

From the Pyramides Calcaires (NE), across the Col de la Seigne and all the way to the Crêt Baudin (SW) the massive light coloured Aroley Fm. is continuously exposed as part of a normal sequence (Fig. 3.22). The lithologic contrast with the underlying dark rocks, the Pyramides Calcaires Fm., is striking. At the Pyramides Calcaires as well as at the Crêt Baudin, the Pyramides Calcaires Fm. unconformably overlies Lower to Middle Liassic limestones. At these localities, a complete succession of pre- ,syn- and post-rift sediments can be observed. A similar situation is also found at the Col de Mya (Fig. 3.23). In the Valleé des Glaciers (between Les Pyramides Calcaires and the Crêt Baudin), detailed structural mapping revealed that late stage thrusting (D4) is responsible a duplication of the Pyramides Calcaires Fm. or for the superposition of syn-rift sediments onto pre-rift sediments (e.g. at the Pt. de Mya massif, Fig. 3.24). Due to the new interpretation of the lithologies, the geologic map of Fig. 3.24 shows strong differences in respect to previous investigations (Antoine 1971, Antoine et al 1992, 1993, Mennessier et al. 1976). A number of relatively good exposures along small creeks on the left side of the Valleé des Glaciers, allow the construction of five synthetic stratigraphic sections of the Pyramides Calcaires Fm. presented in Fig. 3.25.

Several lithologies (Fig. 3.26) are found, such as: black shales, calcareous sandstones, calcschists, grey limestones and lenses of coarse conglomerates. The different lithologies contain a variable amount of rock fragments of mainly dolomitic and calcareous composition. The close alternation of different lithologies makes it difficult to trace clear stratigraphic limits. However, the logs of Fig. 3.25, show that portions of the sections are dominated by a particular facies.

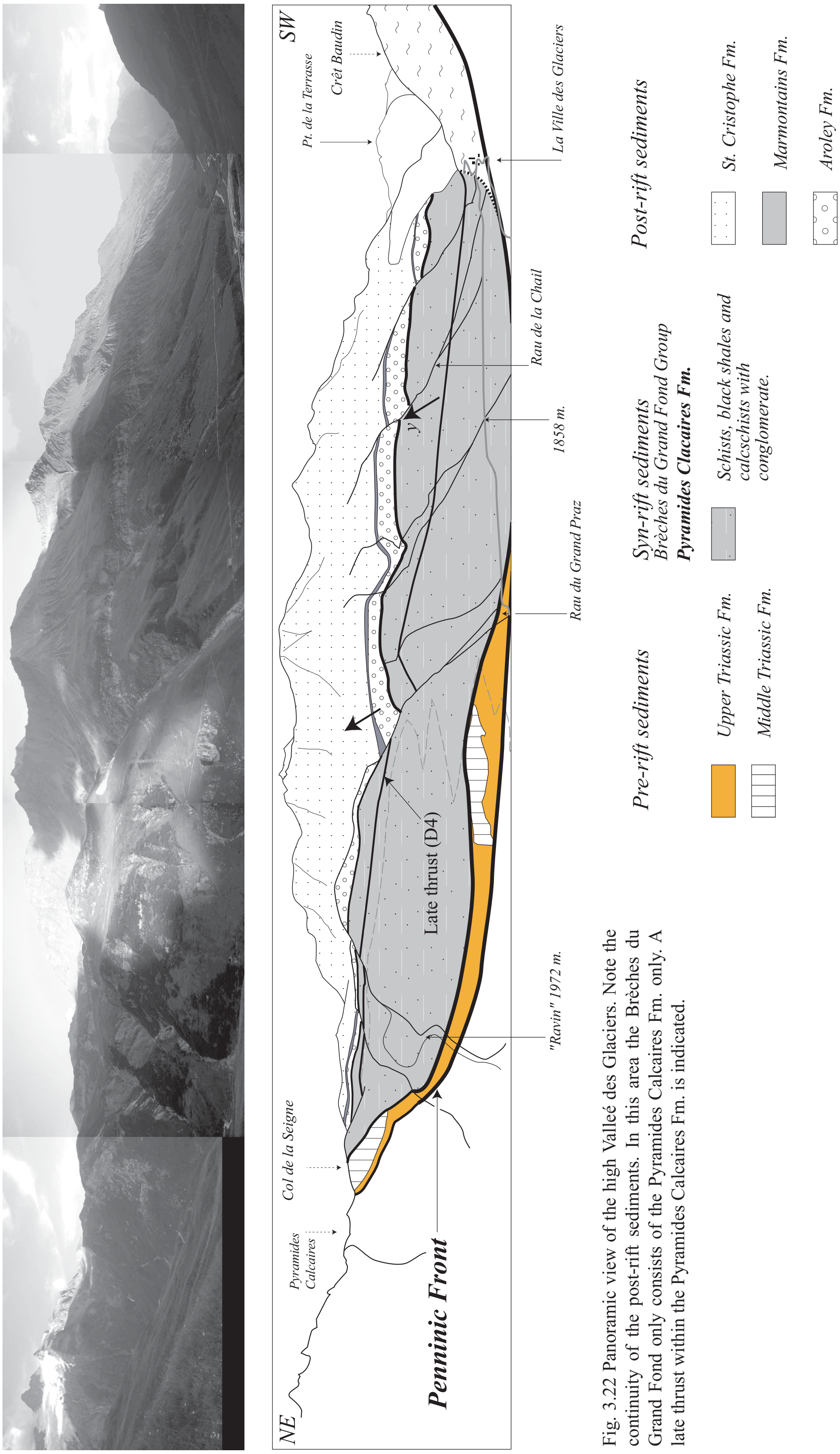


Fig. 3.22 Panoramic view of the high Valleé des Glaciers. Note the continuity of the post-rift sediments. In this area the Brèches du Grand Fond only consists of the Pyramides Calcaires Fm. only. A late thrust within the Pyramides Calcaires Fm. is indicated.

The dark shales represent the most abundant lithology. Microscopically, they show an association of fine organic material, dark iron-oxides and quartz. Subordinate constituents are plagioclase (albite) and white mica. The millimetric lamination with alternating light and dark layers is a sedimentary structure typical for this formation. Fine organic material with clay is concentrated in thin dark laminae which are separated by quartz-rich levels (light laminae). This facies is typically carbonate-free. In a few examples a coloured banding of black and greenish-white very thin beds (1 to 2 cm) is observed (Fig. 2.26d).

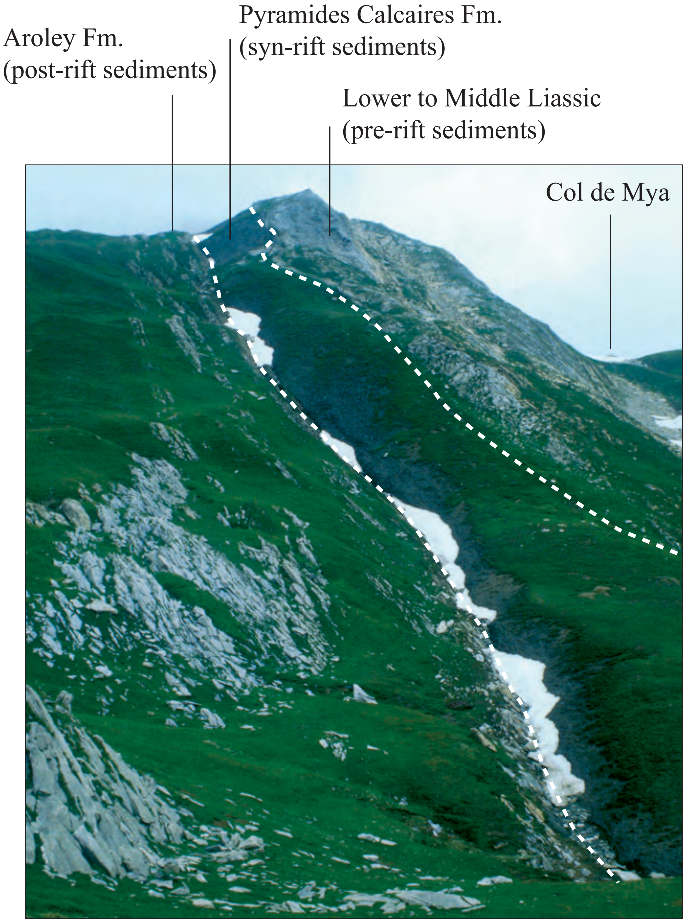
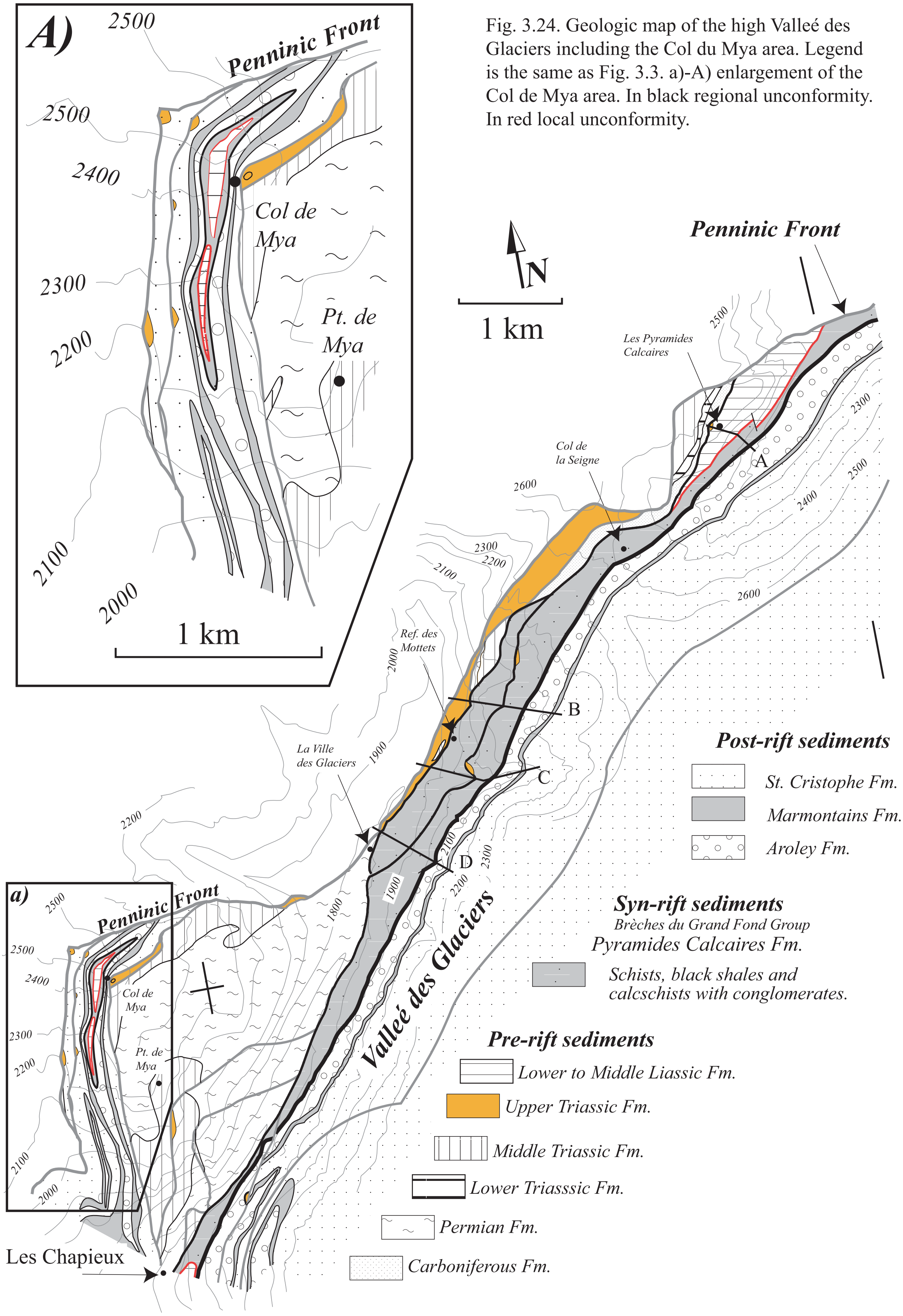


Fig. 3.23. View of the Col de Mya, N of les Chapieux, Moûtiers unit, external Valaisan.

An increase in iron-oxide within laminae rich in organic material seems to be correlated with the occurrence of greenish-white beds. Less frequent are dark shales with a massive texture and grey colour which bear calcite. This variety was not distinguished because it gradually passes to more typical carbonate-free black shales.

Calcareous sandstones are interbedded with black shales and form thin (3-10 cm) beds. The constituents are quartz, calcite, plagioclase (albite) and lithoclasts. Minor white mica and iron oxides mark the bedding.

Fig. 3.24. Geologic map of the high Valleé des Glaciers including the Col du Mya area. Legend is the same as Fig. 3.3. a)-A) enlargement of the Col de Mya area. In black regional unconformity. In red local unconformity.



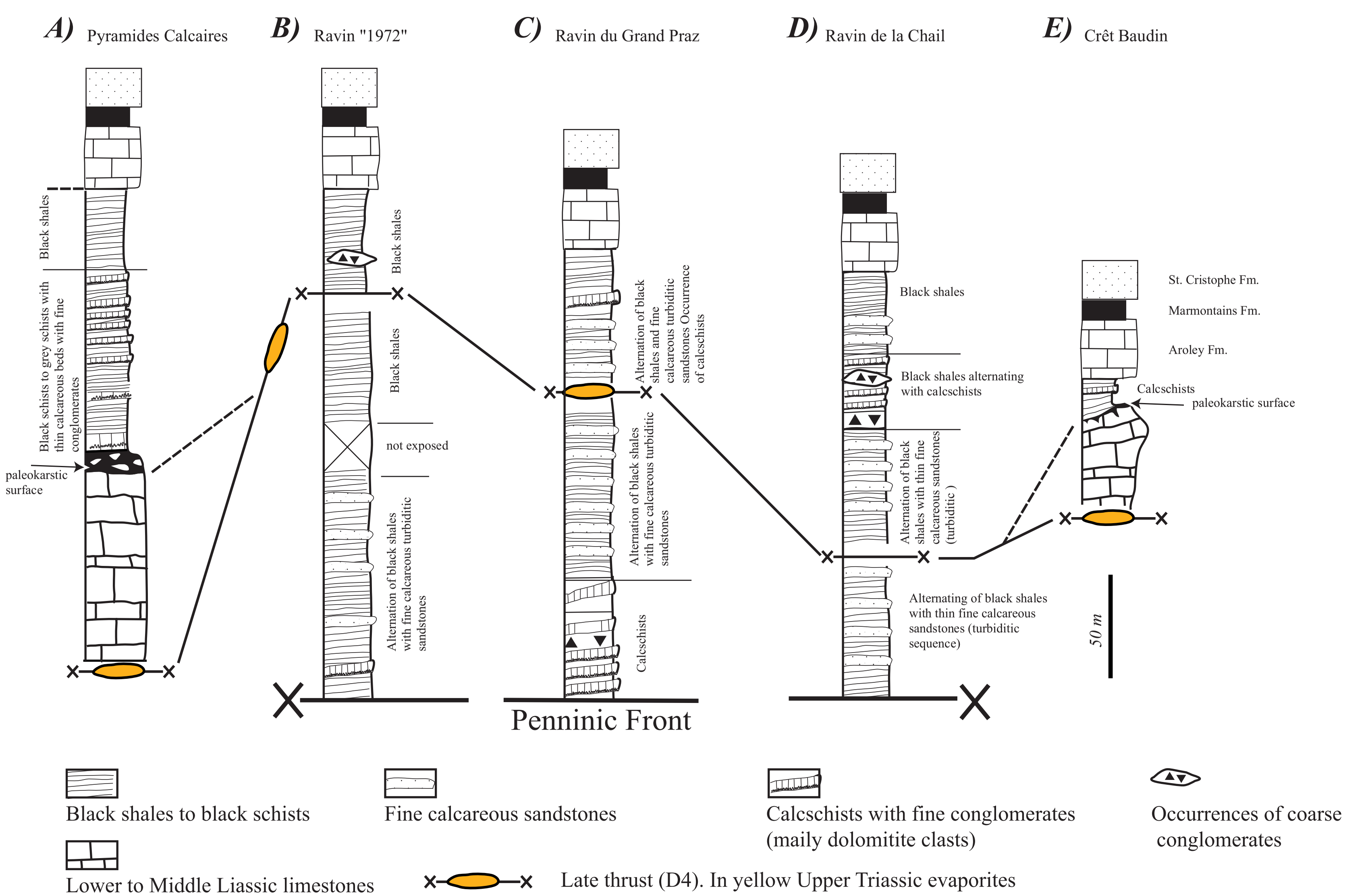
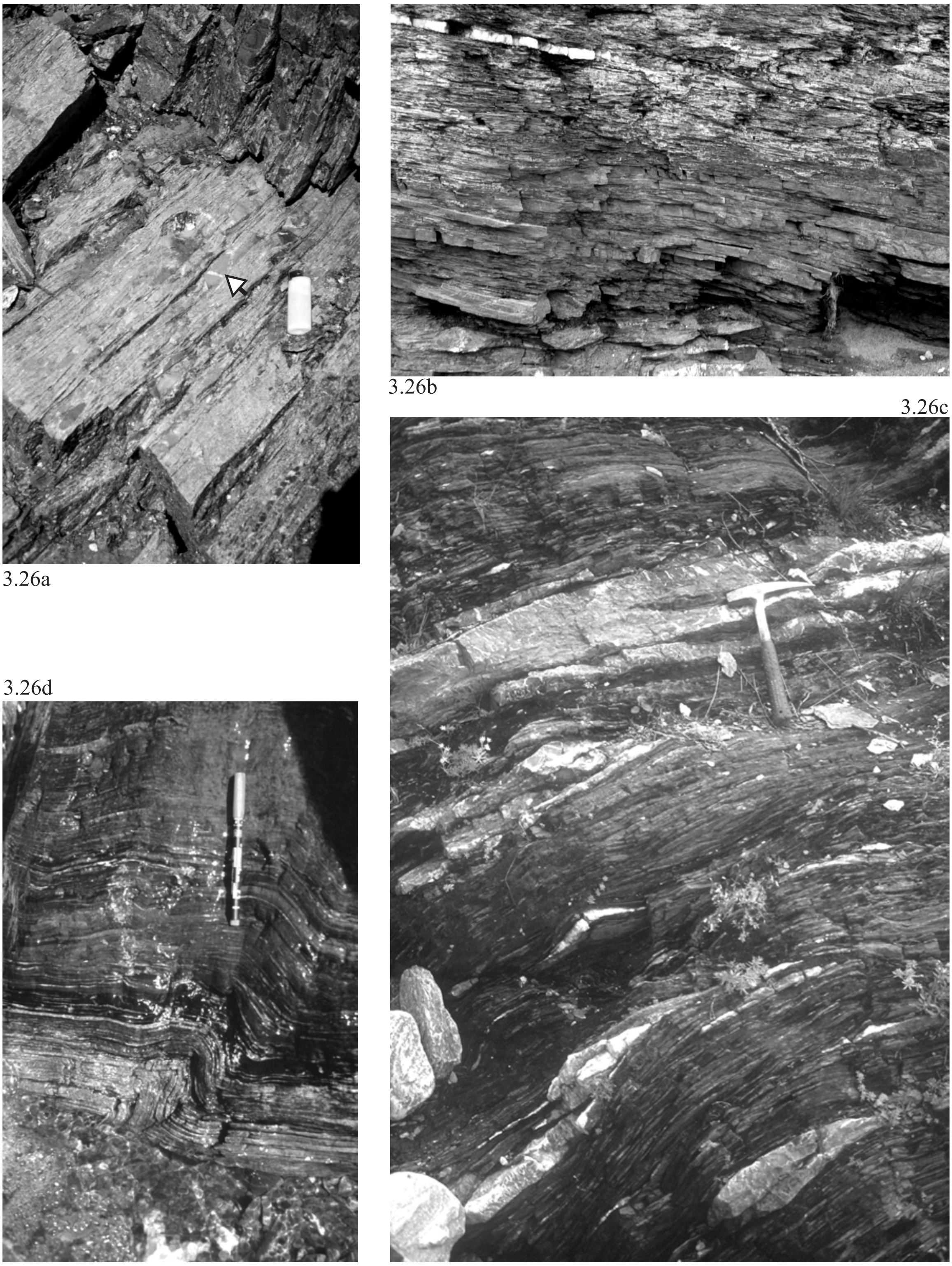


Fig. 3.25. Schematic stratigraphic sections of the Pyramides Calcaires F. in the high Valleé des Glaciers. For locations of the sections A,B,C and D see Fig. 3.24. Sections A and E are modified after: Antoine 1971 (a) and Fudral 1972 (B). In sections B,C and D late thrusting (D4) is resonsible for a duplication of the Pyramides Calcaires Fm.



3.26a 3.26b 3.26c 3.26d

Fig. 3.26 Compilation of different lithologies of the Pyramides Calcaires Fm. in the Valleé des Glaciers (Moûtiers unit, external Valaisan). Fig. 3.26a Black shales referring clasts of dolomite (open arrow). This lithology was previously interpreted as Carboniferous schists. Fig. 3.26b Carbonaceous variety of the black shales (calcschists). Fig. 3.26c Alternation of black shales and calcareous beds with millimetre scale clasts of dolomite. Fig. 3.26d Example of black and greenish-white varieties of black shales.

Most of the quartz is monocrystalline (0.1mm). Polycrystalline larger quartz grains are rarely recognizable (0.5 mm) and are generally of the same dimensions as plagioclase grains (0.4 to 0.7 mm). Macroscopically the calcareous sandstones and conglomerates are white to pale grey and show normal grading.

The rock fragments are exclusively fine-grained sedimentary rocks, in particular sub-rounded millimetric to centimetric clasts of dolomites (Middle and Lower Triassic).
The calcschists form thin calcareous beds with a typical yellow to brownish colour (Fig. 2.26b). These rocks in fact represents fine conglomerates whereby the clasts represent at least 50% of the total volume. The clasts have a sub-rounded shape and a yellowish weathering colour. In fresh fractures and in thin sections different varieties of the Lower to Middle Triassic dolomites can be recognised. In the thickest beds a white limestone (Lower to Middle Liassic ?) can also be observed. In all the observed beds the fragments sorting is graded. The calcschists in fact represent thin debris-flows.

Several lens shaped conglomerates occur, dispersed in the lower to middle parts of the studied sections. They wedge out over some tens of meters (Fig. 3.27).

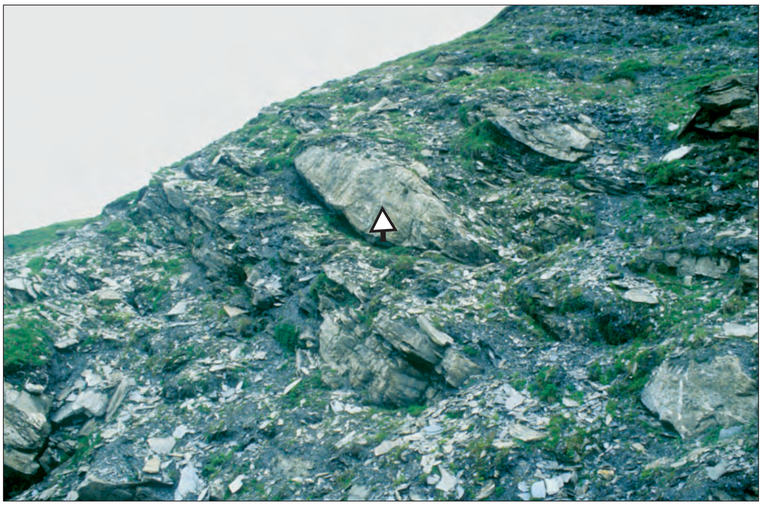


Fig. 3.27 Lens shaped occurrence of a coarse conglomerates (open arrow) within a sequence of black shales and calcschists. Near the Ravin de la Chail, Valleé des Glaciers (Moûtiers unit, external Valaisan)

Their mean thickness is between 2 to 8 m and they contain large (up to 1 m) clasts of limestone and smaller (millimetre to 10 cm) clasts of dolomite which are comparable to clasts observed in the calcareous sandstones or in the calcschists.

In the conglomerate lenses bedding is still observable by interbedded black shales similar to those observed in the rest of the section. Most probably, these conglomerate bodies represent debris flow channel fills.

The prominent rhythmic bedding and grading suggests that the Pyramides Calcaires Fm. here represents a turbiditic series. The presence of thick intervals dominated by black shales led Antoine (1965, 1971) to interpreted a great part of this lithology as Carboniferous schists. However, the attribution of these black shales to the Pyramides Calcaires Fm. is based on the following observations: 1) rhythmic intercalation of black shales and calcareous sandstones, 2) the occurrence of fine conglomerates with Triassic dolomite clasts (Fig. 3.26a) and 3) the absence of other lithologies characteristic for the Carboniferous Fm. (i.e. quartz-arenites). These arguments strongly suggest a Mesozoic age of these sediments. The occurrence of white mica in the black shales may be due to reworking of Carboniferous series.

3.2.2: Detailed description of sections of the Pyramides Calcaires Formation in the Valleé des Glaciers

Detailed descriptions of vertical sections through the Pyramides Calcaires Fm. in the Valleé des Glaciers focus on those exposures where a complete sequence, i.e. pre- syn- and post-rift sediments, is exposed. These are found at the Pyramides Calcaires and at Crêt Baudin (see Fig. 3.25), as well as at the Col de Mya. All other sections, “Ravin 1972”, Ravin du Grand Praz and Ravin de la Chail show syn- and post-rift sediments only. However, they offer the possibility to observe the upper parts of the Pyramides Calcaires Fm. below the Aroley Fm. which are common to all the sections. Amongst the latter, only the Ravin de la Chail section will be described.

The section at the Pyramides Calcaires was earlier studied in detail by Schöller (1929), Barbier (1951), Elter P. (1951, 1954) and Elter and Elter (1965). According to our observations (Fig. 3.25) the section comprises light grey marbles (Lower to Middle Liassic pre-rift sediments) at the base and the conglomerates of the Aroley Fm. at the top of the Pyramides Calcaires Fm. The 150 m thick Pyramides Calcaires Fm. comprises black shales alternating with fine-grained calcareous sandstone beds. The black shales have a high calcite content interbedded with quartz-rich laminae. Different outcrops along the SE slope of the Pyramides Calcaires (SE peak) offer the possibility to observe the uppermost part of the Liassic limestone and the lower 10-15 m of the Pyramides Calcaires Fm.

The uppermost meters of the massive Liassic limestone are fractured and overlain by dismembered meter to centimetre sized boulders (Fig. 3.28). The shape of the boulders is rounded in case of the large boulders and subangular to subrounded in case of the smaller clasts. The space between the boulders is filled by a fine-conglomerate, containing almost exclusively dolomitic clasts (Fig. 3.29). The fine-conglomerate are interpreted to represent the base of the Pyramides Calcaires Fm.

Fig. 3.28. The uppermost part of the pre-rift sediments at the Pyramides Calcaires is formed by large boulders of limestone, separated by a fine conglomerate filling the space between the clasts (Moûtiers unit, external Valaisan). The boulders consist of Lower or Middle Liassic and are interpreted as part of an *in situ* breccia.

The blue to grey calcareous matrix of this fine-conglomerate contains rests of belemnites, gastropods, crinoids and small fragments of Orbitolina (Elter 1954, Elter and Elter 1965). The top of the Liassic limestone shows a reddish to yellowish hematitic to limonitic pigmentation (Fig. 3.30) which strongly suggests the presence of a karst surface (Antoine, 1971; Elter and Elter, 1965), unconformably overlain by the Pyramides Calcaires Fm. Above the fine conglomerates, the Pyramides Calcaires Fm. consists of a monotonous alternation of black shales with calcareous sandstones and dolomitic fine conglomerates. The uppermost 20-30 m of the Pyramides Calcaires Fm. shows a predominance of black shales in respect to the calcarenaceous beds.

The section at the Crêt Baudin is similar to that of the Pyramides Calcaires (Fig. 3.25). However, the thickness is strongly reduced. This outcrop shows Lower to Middle Liassic limestone, discordantly overlain by schists and fine conglomerates which are topped by the Aroley Formation.

Fig. 3.29 Details of the fine conglomerates containing almost exclusively dolomitic clasts

Fig. 3.30a

Fig. 3.30b

Fig. 3.30a. Examples of the karstic breccia encrusting the surface of the Lower to Middle Liassic limestone at the Pyramides Calcaires (Valleé des Glaciers). Colour variation (Fig. 3.30b) are transitional and reflect the different content of hematite (red) and limonite (yellow) which is probably due to climatic changes.

At the interface between the Lower to Middle Liassic limestone and the schists, a yellowish crust on the limestone (Fig. 3.31) was described by Fudral (1973). This limonitic layer again represents a karst surface as seen in the Pyramides Calcaires section.

The section at the Ravin de la Chail (Fig. 3.32) can be studied in continuous outcrops along the creek and in discontinuous outcrops in the meadows around the creek. Below the base of this outcrops, the lower part of the Pyramides Calcaires Fm. is truncated by a late thrust. Our description starts with outcrops above this thrust which are continuous up to the Aroley Fm. The lower part of the section (between 0 and 70m, from the first observable outcrops) consists of carbonate-free black shales with fine grained calcareous sandstone (cm scale) intercalations.

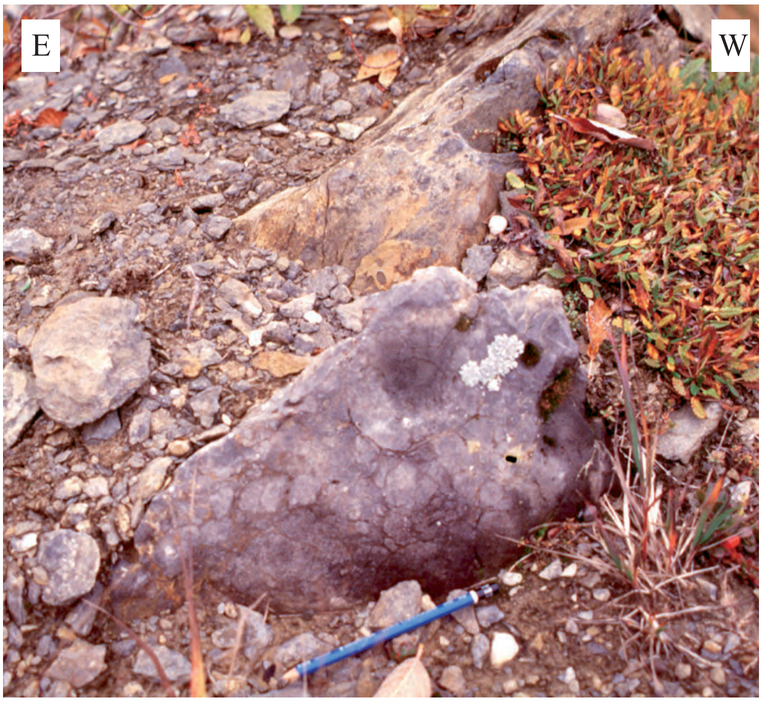


Fig. 3.31. The Lower to Middle Liassic limestone at the Crêt Baudin. The contact with the Pyramides Calcaires Fm. is not showed in this photograph and can be observed a few meters to the W according to an inverted sequence. The Liassic limestone shows a yellowish matrix (on the back ground) suggesting the presence of a paleokarstic surface at the contact between Liassic and the Pyramides Calcaires Fm. (Moûtiers unit, external Valaisan).

The shales locally contain detrital white mica. For this reason this part of the section was erroneously interpreted to belong to the Carboniferous schists (Antoine, 1971). The fine grained calcareous sandstones form a rhythmic alternation with the black shales, suggesting a turbiditic process of deposition. Between 20 and 30 m of this section the black shales are dominating. A single black shale interval, lacking in the calcareous-sandstone fraction is 10 m thick. The first coarse conglomerate horizon occurs between 70 and 80 m.

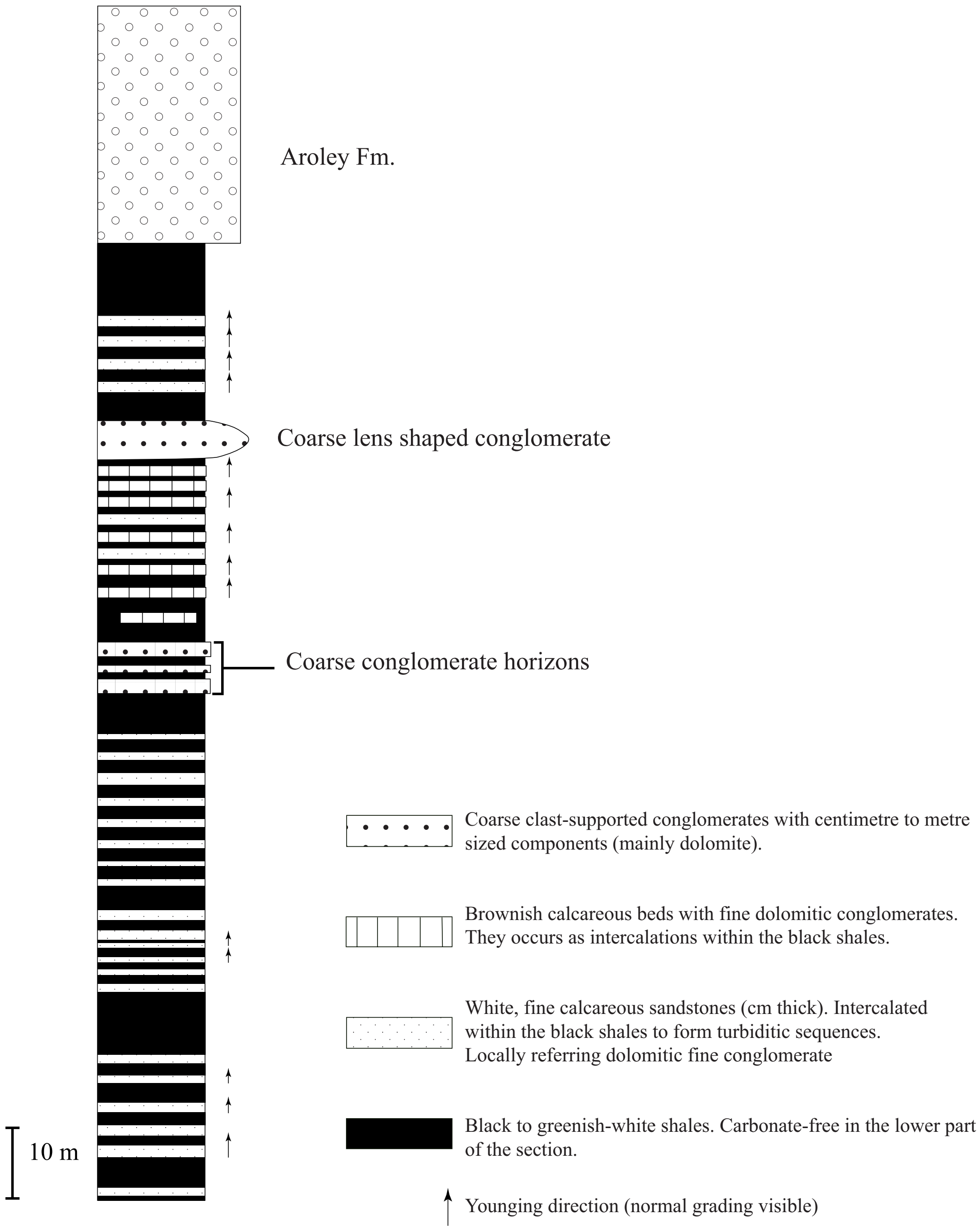


Fig. 3.32 Stratigrafic section of the Pyramides Calcaires Fm. in the Valleé des Glaciers (Ravin de la Chail, Moûtiers unit, external Valaisan). Coord. x:945500, y:2090450, z:2070-2200

This level is laterally continuous and can be observed on both sides of the creek. It is 5-10 m thick and composed of three individual conglomerate beds, intercalated by decimetre thick intervals of sandstone-shale couplets. The matrix is a grey limestone with a yellowish weathering colour. The clasts consist of grey limestone (Liassic ?) and black to yellow Triassic dolomites. These conglomerates are overlain by a 20-25 m of black shales with calcareous sandstones and fine conglomerates (between 80 and 105m) with yellow-brown dolomite clasts. Within the lens-shaped uppermost coarse conglomerate (between 105 and 110 m) bedding is underlined by black shales. The components are coarser compared to the lower conglomerate beds, a large-size boulder of white limestone (Lower-to Middle Liassic ?) has a diameter of 1 m. The uppermost part of this section (between 110 and 135m) is formed by a homogeneous repetition of black-shales, calcareous sandstone and dolomitic fine-conglomerate. In the upper part of this interval (10 m below the Aroley Fm.), shale predominates. The transition to the Aroley Fm. can best be observed on the right side of the creek. There over a distance of 0.5 m few lenses of conglomerates are interbedded with black shales representing the stratigraphic transition from the Pyramides Calcaires Fm to the Aroley Fm.

When recording this section, particular attention was paid to the recognition of sedimentologic polarity in order to detect possible tectonic complications (mainly folds and tectonic contacts). Polarity criteria are well represented by grading in calcareous sandstones. In all the observed beds the younging direction pointed upwards towards the Aroley Fm. The absence of major sheared horizons as well as Upper Triassic cargneule or evaporites excludes that this section is affected by strong tectonic deformation (tectonic contact). However, a tectonic contact is present at its base (Fig. 3.24).

At the Col de Mya another occurrence of the Pyramides Calcaires Fm. has been newly recognized. Here, conglomerates are observed in the lowermost 3 m of the Pyramides Calcaires Fm. (Fig. 3.33). Otherwise, the formation mainly consists of calcschists with discontinuous beds of coarse conglomerates. They contain centimetre-sized limestone and dolomitic clasts embedded in a yellowish calcareous matrix (Fig. 3.33a). Up section only a few beds of fine conglomerate with dolomitic clasts can be observed. Important is the occurrence of decimetre-sized clasts of a calcareous sandstone, identical to that observed in the Combe de la Nova area (Fig. 3.33b, compare with Fig. 3.19b). The section shows a general fining upward trend with an upward decrease in fine conglomerates. The topmost part of the Pyramides Calcaires Fm. entirely consists of black shales and minor calcareous sandstones. The contact with the underlying limestone appears concordant over the whole observed area. The contact surface do not show karst features. Even if without palaeontological evidences, it is attributed to the Lower to Middle Liassic in analogy to the sections observed at the Pyramides Calcaires and at the Crêt Baudin.

Fig. 3.33a

Fig. 3.33 Outcrops of the Pyramides Calcaires Fm. in the Col de Mya area (Moûtiers unit, external Valaisan). Fig. 3.33a shows the first meters above the Liassic limestone (located on the left and not visible in this photograph). The first sediment (cm thick) consists of a laterally discontinuous coarse conglomerate covering the surface of the Liassic limestone. Upwards there are calcschists with a few fine-conglomerate beds (not in this outcrop). Fig. 3.33b decimetre sized clast of coarse calcareous sandstone identical to those observed within the Dent d'Arpire Fm. (compare to Fig. 3.19b).

Fig. 3.33b

3.3: Timing and depositional setting of the syn-rift formations of the external Valaisan (Brèches du Grand Fond Group, Moûtiers unit)

The age of most of the syn-rift Formations (Brèches du Grand Fond Group) as well as the formation boundary is badly constrained. Nevertheless, in order to reconstruct the relative timing of deposition we attempt to correlate the different outcrops of the Dent d’Arpire and the Pyramides Calcaires Formations. The correlation horizons are the formation boundaries. Evidence for two superimposed rifting events (pulses) is found in the Combe de la Nova region where both formations (the Dent d’Arpire and the Pyramides Calcaires Fms.) are exposed. In order to constrain the structural setting of this area the geological cross sections shown in Figs. 3.5 and 3.15 (for location of the profiles see also Fig. 3.3) have been constructed.

Even though these profiles are not discussed in detail it is easy to observe that folding (D2 and D4) affected this portion of the external Valaisan which, before the onset of D2 folding, represented a normal sequence. This simplifies the retro-deformation, which in its simplest form is parallel to the cross sections. Amongst the profiles of Figs. 3.5 and 3.15 only those with a complete section of the syn-rift sediments (profiles A,B,F and G, for location see Fig. 3.3) have been qualitatively retro-deformed to their original position in Fig. 3.34. Note that profiles A and B are located in the N and F and G in the S of the Combe de la Nova area. The thickness of the Dent d'Arpire Fm. decrease from SE to NW (Fig. 3.34a) and from N to S (compare Fig. 3.34a to Fig. 3.34b). In addition, the Dent d'Arpire Fm. successively overlies younger formations (from Permian to Middle Triassic) towards SE (Fig. 3.34a) and towards N (compare the Dent d'Arpire Fm. in Fig. 3.34a to 3.34b) according to an up-slope direction.

The large-scale features of the developing sedimentary basin corresponding to this geometry are shown in Fig. 3.35. The onset of extensional tectonics affecting the pre-rift series was responsible for the formation of depocenters and topographic highs. The depocenters established in the hangingwall while the topographic highs formed at the footwall and hangingwall crests (Fig.3.35). The basin plain rests in the hangingwall of a normal fault which was active during the deposition of the Dent d'Arpire Fm. However the normal fault was not observed and its location in the E can only be assumed. These observations suggest a NW-SE orientation of the axis of the basin. The occurrence of paleokarst surfaces at the foot of the western slope of Pt. 2620 indicates that the bottom of this basin remained exposed prior to the deposition of the Dent d'Arpire Fm. In addition, the Dent d'Arpire Fm. shows different vertical trends. (1) In the lower part of the formation the matrix is made by an impure schistose limestone, while up section the matrix changes toward a pure carbonate composition. This indicates an up-ward trend from carbonate/crystalline-derived matrix toward a pure carbonate matrix. (2) Metric-sized clasts of coarse calcareous sandstones and large boulders of Middle Liassic limestone are characteristic for the lower part of the formation. Up section the clasts predominantly consist of centimetric sized dolomitic clasts, suggesting an upward-fining of the clast size (compare Fig. 3.7a to Fig. 3.10). The occurrence of clast coarse calcareous sandstone, i.e. intrabasinal rock (Fig. 3.19), in the first meters of the Dent d'Arpire Fm. also indicates that this lithology was not sourced by the fault scarps. However, the occurrence of large boulder of Lower Liassic limestone in the first strata of the Dent d'Arpire Fm., suggests that these boulders were probably sourced from a close topographic high region. From these observations we infer that, during the deposition of the Dent d'Arpire Fm., the provenience of the sediments, partially originated from close topographic highs and partially from far away along the basin axis.

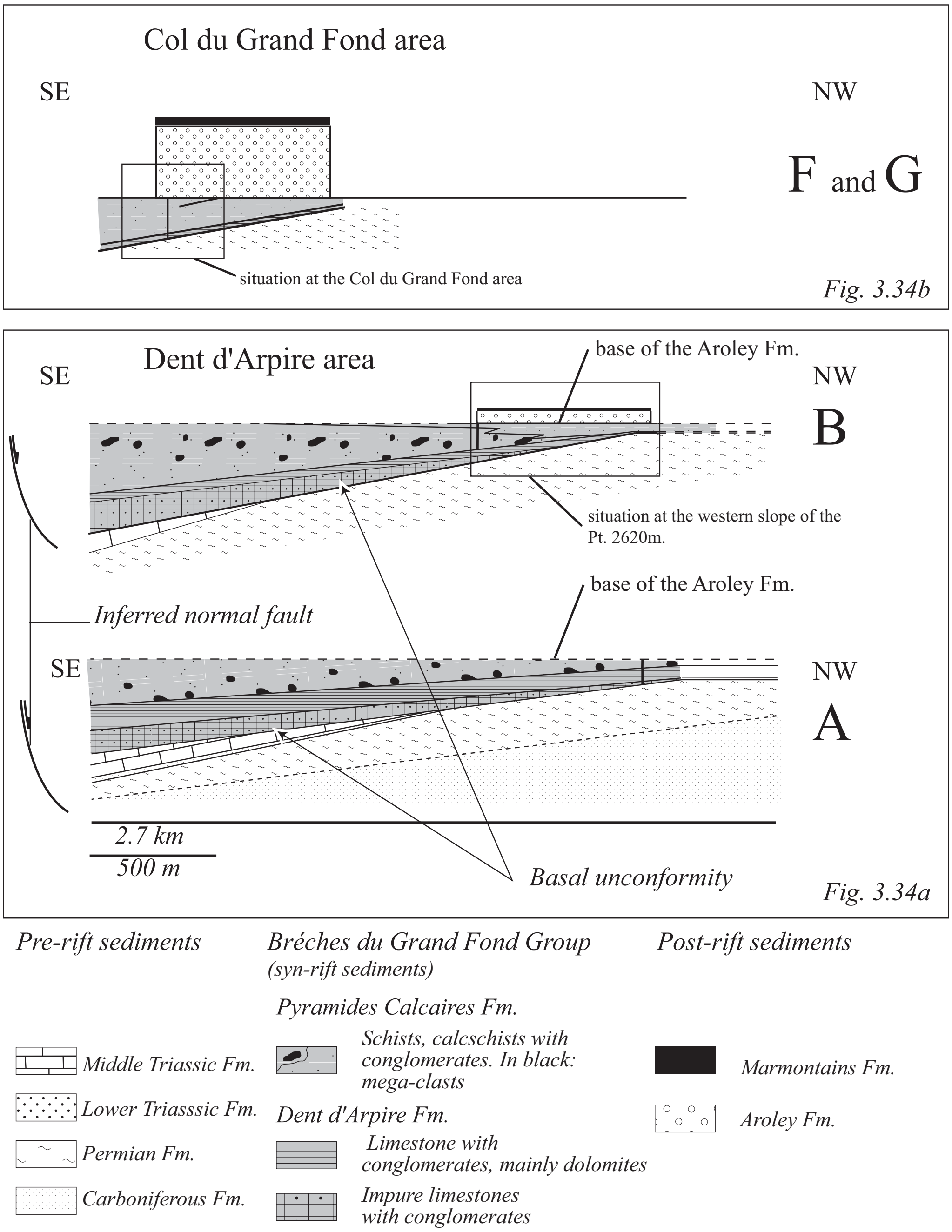


Fig. 3.34 Schematic retro deformation of the Combe de la Nova region illustrating the variation in thickness of the different formations. Fig. 3.34a is based on profiles A and B of Fig. 3.5. Fig. 3.34b is based on profiles F and G of Fig. 3.15. For location, see Fig. 3.3. The width of this portion of the basin (Fig. 3.34a) was deduced from retrodeforming the contact between the Carboniferous and the Permian Fms (profile A of Fig. 3.5).This sedimentary contact, which is not down cut by syn-rift erosion gives a minimal length of 2.7 km. In Fig. 3.34a the Dent d'Arpire Fm. decrease in thickness from SE to NW (Fig. 3.34a) and from N to S (compare Fig. 3.34a with Fig. 3.34b) onlapping the pre-rift sediments according to an up-slope direction. The same trend is observed going from N to S (compare Fig. 3.34a with Fig. 3.34b). Variation in the thickness of the Aroley Fm. (post-rift) is due, mainly, to tectonic reasons. The thicker occurrences of the Aroley Fm. in the S (Fig. 3.34b) corresponds to the hinge region of a large-scale F2 antiformal anticline.

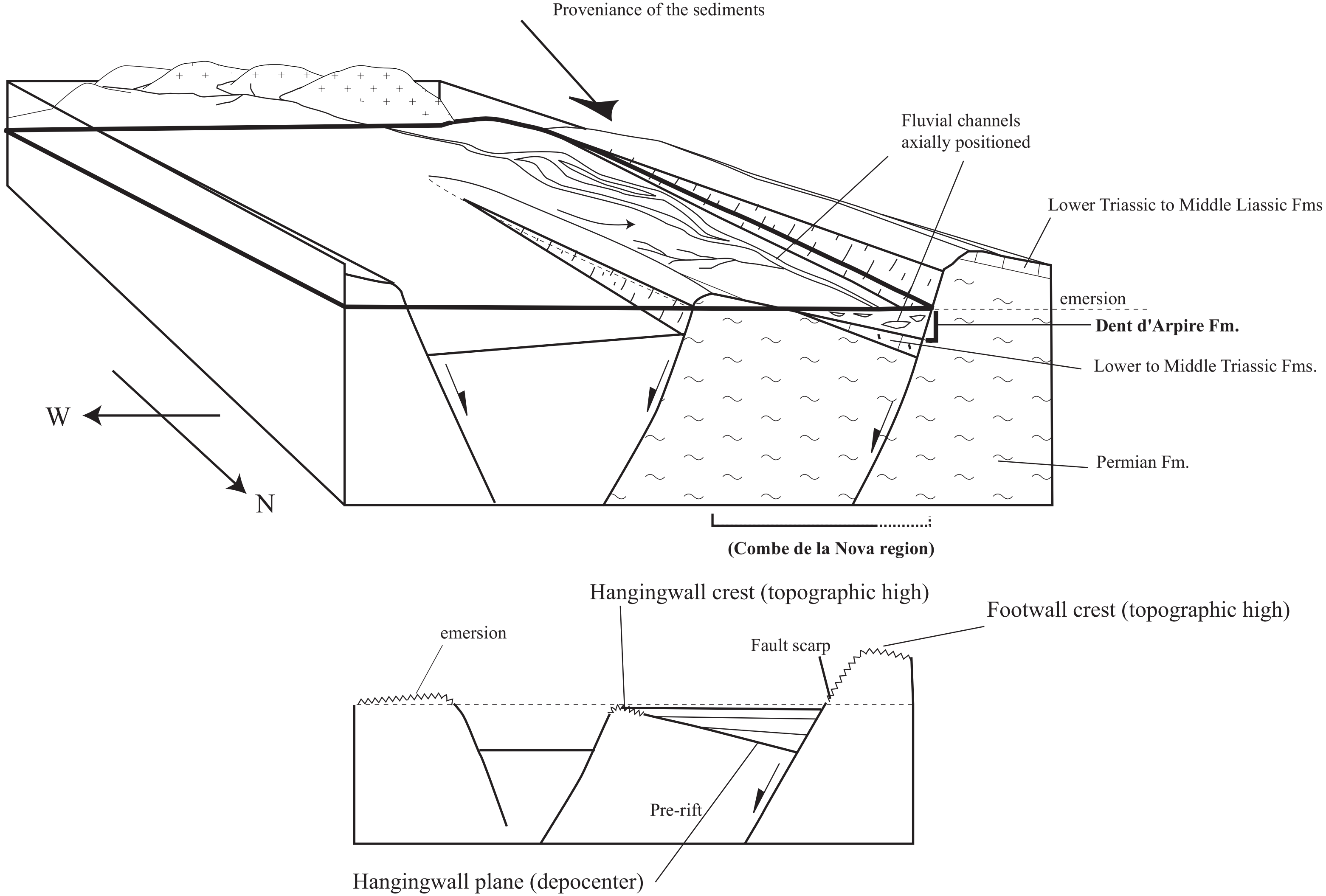


Fig. 3.35 Possible depositional setting during the deposition of the Dent d'Arpire Fm. (based on Prosser 1993). Note the variation in thickness of the Dent d'Arpire Fm. toward W and S.

In all the observed stratigraphic sections, the absence of evidence for emergence in the first meters of the conglomerate and the widespread occurrence of debris-flows, together with fossils of ammonites (Fig. 3.18), indicates that sedimentation occurred under marine conditions below the sea level.

Particularly in the S, in the Col du Grand Fond area, the widespread occurrence of well graded mass flow deposits (Fig. 3.16) is interpreted in terms of a channel fan system. The direction of this channel fan system is undetermined, however it was likely to have been parallel to the basin axis, i.e. roughly N-S. The clinoforms, indicate a provenience of the sediments from SW (i.e. parallel to the basin axis). This suggests that the proportion of the sediments transported along the basin axis was more important than that from the close rift highs (i.e. perpendicular to the basin axis).

The age of the Dent d'Arpire Fm. can only be derived from circumstantial evidence. The age of the pre-rift strata below the unconformity at the base of the Dent d'Arpire Fm. is the result of early rift tilting and erosion of the pre-rift series. In the Combe de la Nova region, the youngest still preserved pre-rift sediments capped by the Dent d'Arpire Fm. are Middle Triassic dolomites. The presence of Middle Liassic limestone as components in conglomerates indicates that erosion and deposition of the Dent d'Arpire Fm. did not take place before Late Liassic times. This observation is in accordance with the finding of ammonites which indicate an Upper Liassic age. The occurrence of reworked clasts of the Dent d'Arpire Fm. in the Pyramides Calcaires Fm. suggests that the Dent d'Arpire Fm. was locally eroded after its deposition. Since no pre-rift sediments younger than Liassic are found amongst the pre-rift sediments, nor as components in this breccia, an Upper Liassic to Dogger age is very likely for the deposition of the Dent d'Arpire Fm.

The Pyramides Calcaires Fm. can be observed over a wider range of outcrops. This formation either overlies the Dent d'Arpire Fm. (Combe de la Nova) or directly the pre-rift sediments (Valleé des Glaciers, Pyramides Calcaires area). In the Combe de la Nova area the Pyramides Calcaires Fm. consists of chaotic breccias and of calcschists associated with fine conglomerates which contain large boulders of pre-rift sediments and occasionally reworked breccias of the Dent d'Arpire Fm. The transition from the chaotic breccia to the calcschists represents an upward and lateral fining trend which, in this area, is characteristic for this formation. Up section, the contact with the overlying post-rift sediments is always formed by calcschists and fine conglomerates. Laterally, roughly along a N-S direction, the chaotic breccia progressively decreases in volume towards S (Fig. 3.34b) and only locally are isolated large and well rounded boulders present (Passage the

Prarozan). The same trend can also be observed along an E-W profile (Fig. 3.34b). In the lowermost outcrop at the base of the W slope of the Pt. 2620 the sediments of the Pyramides Calcaires Fm. consist of calcschists and fine conglomerates (Fig. 3.34a). Therefore, the chaotic facies is characterised by an upward transition to the calcschists as well as toward N and S and toward W.

The occurrence of gravity flow deposits along the formation boundary between the Dent d'Arpire and the Pyramides Calcaires Fms. can be interpreted as an erosional surface. This interpretation is confirmed by the occurrence of the Dent d'Arpire Fm. as reworked clasts within the calcschists of the Pyramides Calcaires Fm. (Figs. 3.20 and 3.21). This reworked conglomerate shows strong facies analogies with conglomerates in the uppermost part of the Dent d'Arpire Fm. The clasts display a well rounded shape which is indicative for lithification of this formation prior to transport and deposition. Hence, some time must have passed between the lithification of the Dent d'Arpire Fm. and its erosion (during sedimentation of the Pyramides Calcaires Fm.).

The increase in clast size toward the E is suggestive for the proximity to a footwall fault scarp located in the E. In general, variations in thickness of the Pyramides Calcaires Fm. accommodate a topography of the basin which was inherited from the time of deposition of the Dent d'Arpire Fm. (Fig. 3.36). The main difference between the Pyramides Calcaires Fm. and the Dent d'Arpire Fm. consists in the source of the clastic material. Within the Pyramides Calcaires Fm. (in the Combe de la Nova area) most of the sediment volume is derived from the topographic highs (at least in the Combe de la Nova region) along the basin forming cliffs according to an input direction perpendicular to the basin axis. This particular feature is interpreted to represent the stratigraphical record of an increase in displacement along the normal faults which controlled the deposition of the Pyramides Calcaires Fm. in this area (Fig. 3.36). It is proposed that this increase corresponds to a reactivation of extensional structures which are related to the onset of the Valaisan rifting (deposition of the Pyramides Calcaires Fm.). In this scheme we propose that the reactivation of faults is the main factor controlling the depositional pattern of the Pyramides Calcaires Fm. Moreover, it is tentatively proposed that the Dent d'Arpire and the Pyramides Calcaires Fm. represent two distinct events of rifting affecting the external Valaisan. The Dent d'Arpire Fm. corresponds to a first syn-rift record related to the Piemont-Ligurian extension. Consequently, the younger Pyramides Calcaires Fm. (at the Pyramides Calcaires area) may have been related to the Valaisan rifting (i.e. may be coeval with the deposition of the Complexe Antéflysch).

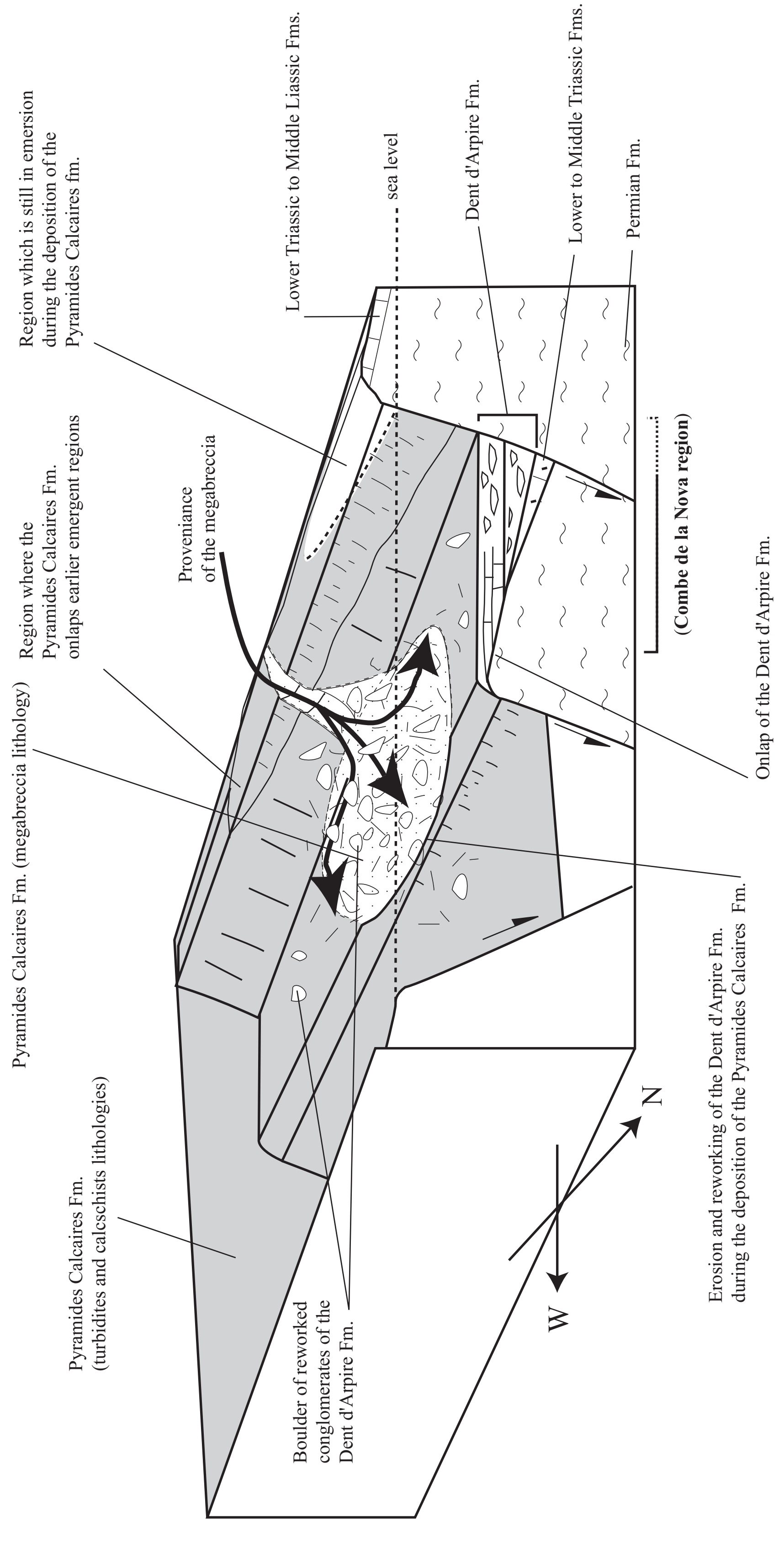


Fig. 3. 36. Possible depositional setting for the Pyramides Calcaires Fm. (based on Prosser 1993)

This scenario leaves many points open for other interpretations. However, this study on the Brèches du Grand Fond Group, based on a limited but consistent set of observations made in the Combe de la Nova area, suggest such a scenario as likely. Master faults related to the deposition of the different formations are not preserved. This limits our interpretation. Following similar lines of reasoning, and in particular the control of fault reactivation on the depositional system, the remaining discussion will focus on the Pyramides Calcaires Fm. in the Valleé des Glaciers. In particular, we will try to apply the concept of a depositional setting controlled by reactivation of faults to some contradicting chronostratigraphic data which come from this area.

In its current position, the Valleé des Glaciers area is tectonically separated from the Combe de la Nova region by a late thrust system related to the Pennine Front (D4). The structural analysis of these contacts (Ch. 6.5) indicates that higher structural levels correspond to more internally paleogeographic regions in respect to lower structural levels. This allows to infer a more internal position for the Valleé des Glaciers in respect to the Combe de la Nova region. Following this line of reasoning a more external position for the Col de Mya region, in respect to the Valleé des Glaciers can be inferred too. Despite the tectonic complications which cannot be quantified, a possible disposition of the different regions from the more external to the more internal includes the Col de Mya, the Combe de la Nova and the Valleé des Glaciers, respectively. The Combe de la Nova represents the only region where both the Dent d'Arpire and Pyramides Calcaires Formations are exposed.

In the Valleé des Glaciers the Pyramides Calcaires Fm. is mainly represented by an alternation of turbiditic fine sandstones and black shales. Minor intercalations of calcschists and the rare occurrence of coarse conglomerates infer a strong facies analogy with the calcschists and fine conglomerates of the uppermost part of the Pyramides Calcaires Fm. as observed in the Combe de la Nova. The observation that the Pyramides Calcaires Fm. stratigraphically overlies Lower to Middle Liassic limestones with a paleokarst surface in the Valleé des Glaciers area and the superposition by the Aroley Fm. leaves a Late Liassic to Early Cretaceous age for this formation possible. When the exposure of the Pyramides Calcaires Fm. is regarded as the homonymous locality (Les Pyramides Calcaires, Val Veni, I) the occurrence of *Orbitolina sp.* causes confusion. In this exposure, the base of the Pyramides Calcaires Fm. yielded fossils of *Orbitolina sp.,* suggesting for a Lower Cretaceous (Barrêmian) age for the unconformity bounding the base of this formation (Elter 1954, and new observations in this study). This observation would indicate that the base of the Pyramides Calcaires Fm. in this area has almost the same age as the base of the post-rift sediments (Aroley Fm., Sodero 1968). Thereby it is assumed that the reworking of these fossils at the base of the Aroley Fm. is unlikely (Sodero 1968).

These chronostratigraphic data need can be discussed in the framework of a depositional setting controlled by diacronous reactivation of faults of the syn-rift stage in the Pyramides Calcaires Fm.. A clear sedimentological feature of the Grand Fond Group consists in the occurrence of paleokarst surfaces related to emersion prior to deposition. the age of these surfaces cannot be determined: in the case of the Pyramides Calcaires exposure, it may have occurred sometimes between Late Liassic and Early Cretaceous. Emersion may represent a long time span. Pre-sedimentary emergence indicates that the basins were characterised by a strong topography, dividing the external Valaisan basin into separate sub-basins. It is possible to imagine that sub-basins were not active at the same time while the deposited sediments were always of the same kind: i.e. black shales, calcareous sandstones, turbidites, calcschists and fine conglomerates. The final result would be to produce sediments which may be indistinguishable in many circumstances but of different ages.

In Fig. 3.37 the Valleé des Glaciers and Combe de la Nova areas are shown along a schematic E-W profile starting from Late Liassic times. At this time corresponds the deposition of the Dent d'Arpire Fm. which in the Combe de la Nova area locally post-dates emergence. Later on, during the deposition of the Pyramides Calcaires Fm., (Late Jurassic (?) to Early Cretaceous) a differential tectonically induced subsidence affects the Valleé des Glaciers-Combe de la Nova-Col de Mya basin. This paleotectonic evolution of the Valais domain will end during the Early Cretaceous, corresponding to the opening of the Valais ocean. Thereafter, the thermal subsidence will affect the entire Valaisan domain corresponding to the deposition of the Aroley Fm. which, as a matter of fact, represent the first sediments found on both continental and oceanic domains.

According to this evolution, the onlap of the Pyramides Calcaires Fm. onto the paleokarst surface in the Valleé des Glaciers area (Pyramides Calcaires area) occurred during a very late stage of the deposition of the syn-rift sediment (Early Cretaceous), immediately before the sedimentation of the post-rift sediments (Aroley Fm.). Such a young age, at least at the Pyramides Calcaires exposure, is also suggested by sedimentological observations. There, the occurrence within the Pyramides Calcaires Fm. of fine conglomerates bearing mainly dolomitic clasts is a feature which is particularly evident also for the Aroley Fm. of the external Valaisan too (Lomas 1992, see Ch. 2.5.1).

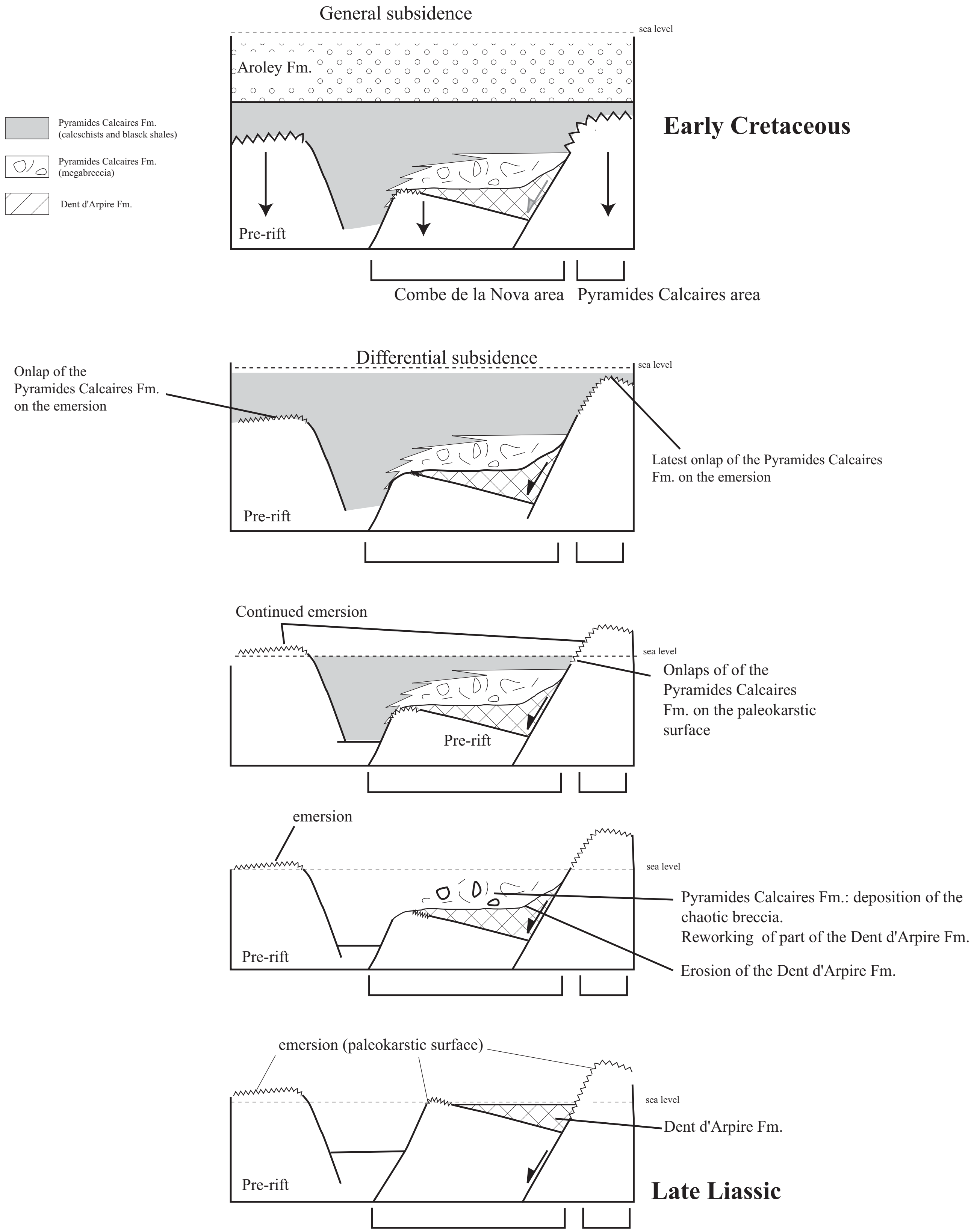


Fig. 3.37 Timing of deposition of the syn-rift sediments of the Brèche du Grand Fond Group (Moûtiers unit, external Valaisan)

Equivalents of the Pyramides Calcaires Fm. at the Pyramides Calcaires area can be found in the internal Valaisan where the corresponding stratigraphic interval is represented by the black schists of the Complexe Antéflysch. For this reason we can assume that part of the Pyramides Calcaires and the Complexe Antéflysch, both grading upward into the Barrêmian Aroley Formation, are part of the Lower Cretaceous succession. Consequently, the oceanic crust (preserved within the Pt. Rossa Complexe) must have formed during the latest Jurassic to Early Cretaceous.

CHAPTER 4: GRADE AND EVOLUTION OF METAMORPHISM

The purpose of this chapter is, in a first part (chapter 4.1), to review the available petrological studies in the Valais domain and surrounding units. Subsequently (chapter 4.2), new observations about the relation between diagnostic mineral parageneses and structural elements of the Pt. St. Bernard and Versoyen units will be described.

Since the late sixties, petrographical studies of the Mesozoic metasediments and the basaltic volcanic rocks of the Valais domain of the western Alps have been carried out in order to define diagnostic mineral parageneses (Loubat, 1968). Since the early attempts up to the most recent studies, the increase of information revealed a polyphase character of Alpine metamorphism (Bocquet and Delaloye, 1974; Bousquet, 1998; Cannic, 1996; Goffé and Bousquet, 1997; Lasserre and Lavergne, 1976; Schürch, 1987). Relics of HP assemblages were first described in the Versoyen unit only, i.e. in the internal Valaisan (Cannic, 1996; Schürch, 1987), but afterwards also in the Pt. St. Bernard unit, i.e. in a part of the external Valaisan (Bousquet, 1998; Goffé and Bousquet, 1997). In the more external Roc de l'Enfer and Moûtiers units, i.e. in most of the external Valaisan, the metamorphic overprint did not exceed greenschist facies conditions (Bousquet, 1998; Goffé and Bousquet, 1997). Recently, several authors (Bousquet, 1998; Goffé and Bousquet, 1997; Oberhänsli et al., 1995; Schürch, 1987) related these observations to a subduction scenario, affecting part of the Valais domain.

However, it is worth noticing that despite the differences of metamorphic grade among the different subunits of the Valaisan, all Valaisan units show evidence for having been affected by three identical deformation phases i.e. D1, D2 and D4 (D3, D5 and D6 being of local extent). In order to discuss the tectonometamorphic evolution of the Valaisan units, particular attention has been given to the relation between mineral growth and deformation phases. Understanding the sequence of mineral growth with respect to the deformation is a prerequisite for paleogeographical reconstructions as well as for possible future interpretations of radiometric ages on syn-kinematically grown minerals.

4.1: Overview of the metamorphism in the Valais domain and surrounding units

In the basaltic volcanic rocks of the Versoyen unit, previous petrographical studies (Bocquet and Delaloye, 1974; Lasserre and Lavergne, 1976; Loubat, 1968) have shown the occurrence of omphacite and glaucophane.

Although these minerals are diagnostic for HP metamorphic conditions, the Versoyen unit was first regarded as a polyphase metamorphic unit which later was overprinted under greenschist facies conditions (Antoine, 1971; Bocquet and Delaloye, 1974; Lasserre and Lavergne, 1976; Loubat, 1968). Also, the part of the Versoyen unit containing eclogites were interpreted by Lasserre and Lavergne (1976) as representing allochthonous slices of Piemont-Ligurian origin emplaced onto the more external Valais domain. According to these various authors, during a post-eclogitic evolution, tectonic slices of Piemont-Ligurian units were first transported onto the Valais, Subbriançonnais and Briançonnais domains. Then, a later out-of-sequence thrust along the Houillèr Front would have brought the Briançonnais and Subbriançonnais units onto the Valais domain (Bocquet and Delaloye, 1974; Fudral and Guillot, 1988; Lasserre and Lavergne, 1976; Schoeller, 1929).

Such models were seriously put into doubt by the findings of (Schürch, 1987). The detailed petrological study by Schürch (1987) demonstrated the widespread occurrence of HP relics in all of the Versoyen units and the subsequent overprint by retrograde assemblages under blueschist to greenschist facies conditions. These observations led Schürch (1987) to propose an alternative model according to which the metamorphism of the Versoyen unit was considered autochthonous in respect to the Valais paleogeographical domain, and compatible with a second more external subduction zone, situated in the Valais paleogeographical domain.

According to Schürch (1987), the metamorphic evolution of the Versoyen volcanic rocks is controlled by two different tectonic settings. The first one was the intrusion and extrusion of the basaltic volcanic to subvolcanics rocks which took place during the rifting of the Valais ocean. The second one was related to a subduction scenario during the Alpine orogenic evolution. The first gave rise to the hydrothermal activity responsible for the concentration of Mn, as the basalts got in contact with the pelitic host rocks. At the same time, contact metamorphism took place. The most interesting consequence of this event is represented by the porcelain-like rocks or the "adinoles" (Schürch 1987, p.36 and references therein) caused by reactions between the sediments and the volcanic rocks. As for the second setting, the Versoyen unit first underwent HP-LT eclogitic facies conditions followed by blueschist and greenschist facies conditions.

The metamorphic evolution recognised in the Versoyen unit by Schürch (1987) was strictly related to observations concerning the basaltic volcanic rocks, apart from contact metamorphism. The geodynamic implication of a subduction zone for the Versoyen unit led following researchers on focus metamorphism in the associated sediments. Recently Goffé and Bousquet (1996) have bridged the gap between eclogitic rocks and associated sediments providing evidence for identical HP-LT metamorphic conditions realised within the Complex Antéflysch of the Versoyen unit and the calcschists of the Pt. St. Bernard unit. It has to be mentioned that the Pt. St. Bernard unit was previously considered as to be characterised by low grade metamorphic conditions. Currently, there is a lack of up-to-date radiometric data which would allow to constrain the age of eclogite facies metamorphism, except that metamorphism post-dates the deposition of the post-rift sediments whose youngest formation are considered Upper Cretaceous-Paleocene (?) (Antoine et al., 1992).

4.1.1: Metamorphic grade and evolution of the Versoyen and Pt. St. Bernard units

Temperature and pressure are taken into consideration for different rock compositions found in the Versoyen and Pt. St. Bernard units, i.e. metabasites, metapelites and acidic rocks. The estimates are summarised in Fig. 4.1. In the mafic rocks, the eclogitic conditions are indicated by the parageneses omphacite + Ca-Fe-Mn-garnet + glaucophane 1 (Cannic, 1996), + zoisite-clinozoisite, + rutile, + quartz (Schürch, 1987) and the PT conditions are estimated to be around 13-16kbar and 425-500°C (Cannic, 1996).

For the metapelites, Bousquet (1998) indicates two different stages of the HP-LT metamorphism. A first stage is indicated by the assemblage carpholite, white mica and chlorite defining peak pressure conditions. This mineral association can be traced only inside quartz segregations, where carpholite can be preserved as a relic during the following decompressional evolution. Estimates of P-T conditions for the peak pressure event range between 14-15kbar and 350-400°C (Bousquet, 1998). The association of chloritoid, chlorite, white mica and pseudomorphs of Fe-carpholite defines a second metamorphic stage. The temperature constrained by the carpholite-chloritoid reaction and by the chloritoid-chlorite geothermometer ranges between 450 and 500°C. The related pressure calculated from the content of Si in white mica indicates 13-15kbar. These conditions are comparable with those of the eclogitic rocks.

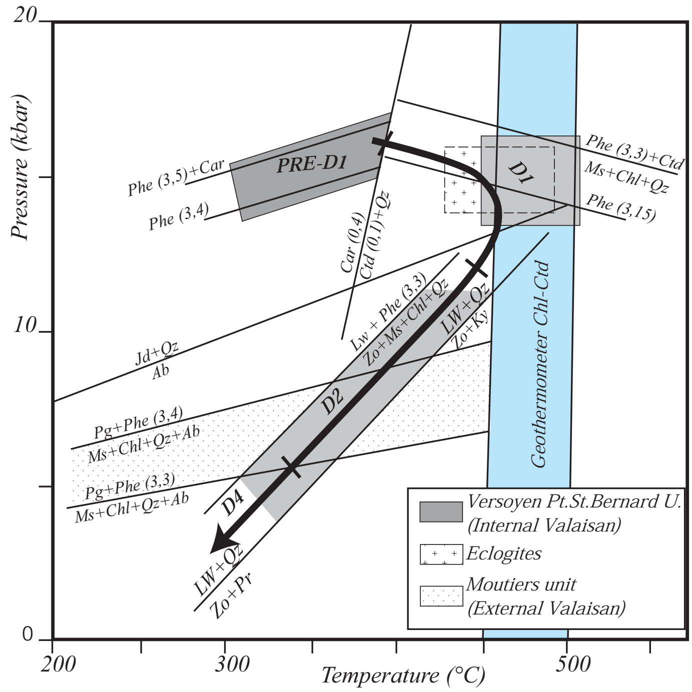


Fig. 4.1 Metamorphic P-T path determined for the Valaisan domains, redrawn after Bousquet and Goffé (1997), Bousquet (1998), Cannic (1996). According to ours findings the P-T condition during pre-D1, D1, D2 and D4 are indicated.

For the sodic metasediments and acidic rocks of the Pt. Rossa basement, similar conditions i.e. at least 12kbar are indicated by the jadeite and quartz association (Cannic, 1996; Saliot, 1979; Schürch, 1987). According to Goffé and Bousquet (1997), these post-HP peak conditions are compatible with the occurrence of eclogitic rocks in the metabasites and with the occurrence of chloritoid in the metapelites. The decrease of pressure between the HP peak associated with decompression heating indicates that the eclogites of the Versoyen unit were formed along a decompression path related to an exhumation process.

The further retrogressive overprint of the HP-LT events in the metabasaltic rocks is indicated by the assemblage glaucophane (2) + zoisite-clinozoisite + phengite + paragonite + albite (Cannic et al., 1996). Lawsonite can be found in the metapelites as pseudomorphs (Cannic et al., 1996; Dalla Torre, 1998; Goffé and Bousquet, 1997) replaced by zoisite, white mica and chlorite. The estimate of pressure and temperature indicates 8kbar and 400°C (Bousquet, 1998). Green amphibole (actinolite-tremolite and Mg-hornblende), epidote (pistacite), biotite, albite, white mica, stilpnomelane, chlorite, tourmaline and prehnite-pumpellite (Cannic, 1996; Schürch, 1987) represent the common assemblage indicating ongoing retrogression under greenschist and subgreenschist facies conditions.

4.1.2: Metamorphic grade of the Moûtiers and Roc de l'Enfer units (external Valaisan)

The PT conditions estimated for the Versoyen and Pt. St. Bernard unit cannot be generalised for the entire area studied (Fig. 4.2). The Moûtiers and Roc de l'Enfer units (external Valaisan) lack evidence for HP overprint. In these units, the common assemblage is quartz, + calcite, + albite, + chlorite and phengite, the latter characterised by a high Si value of substitution (Si 3,3-3,4) (Bousquet et al., 1998).

The phengite associated with albite and chlorite after paragonite suggests a reaction where the PT conditions are estimated to be around 4-6kbar and 350°C (Bousquet, 1998). This reaction, generally associated with a decrease in pressure, suggests that pressure conditions higher than 6kbar could have affected the external units of the external Valaisan. The absence of jadeite, glaucophane or chloritoid as pseudomorphic relics indicates that the pressure conditions during earlier metamorphic stages did not exceed 8-12kbar (Bousquet, 1998).

Further to the S, the Valaisan tectonic units located between the Pennine Front and the Moûtiers units i.e. the Quermoz and Bagnaz units show PT conditions lower than of 530°C/5kbar and 450°C/4kbar, respectively (Gely and Bassias, 1990). As for the external part of the external Valaisan, i.e. the Moûtiers and Roc de l'Enfer units, the pressure conditions decrease from E to W and from higher to lower tectonic units, i.e. towards the Pennine Front.

4.1.3: Metamorphic grade of the adjacent cover nappes (Delphino-Helvetic and Zone Houillère)

The Valais zone is confined between two major tectonic features, namely the Pennine front (PF) to the W and the Houillèr front (HF) to the E (Fig. 4.2). Both tectonic lineaments correspond to changes in the metamorphic grade although their meaning in the tectono-metamorphic evolution of the Valais zone is not the same. To the W the Pennine Front corresponds to a change from a pronounced epizonal metamorphism in the external Valaisan units to an epizonal to anchizonal overprint in the Delphino-Helvetic zone (Gely, 1988; Gely and Bassias, 1990).

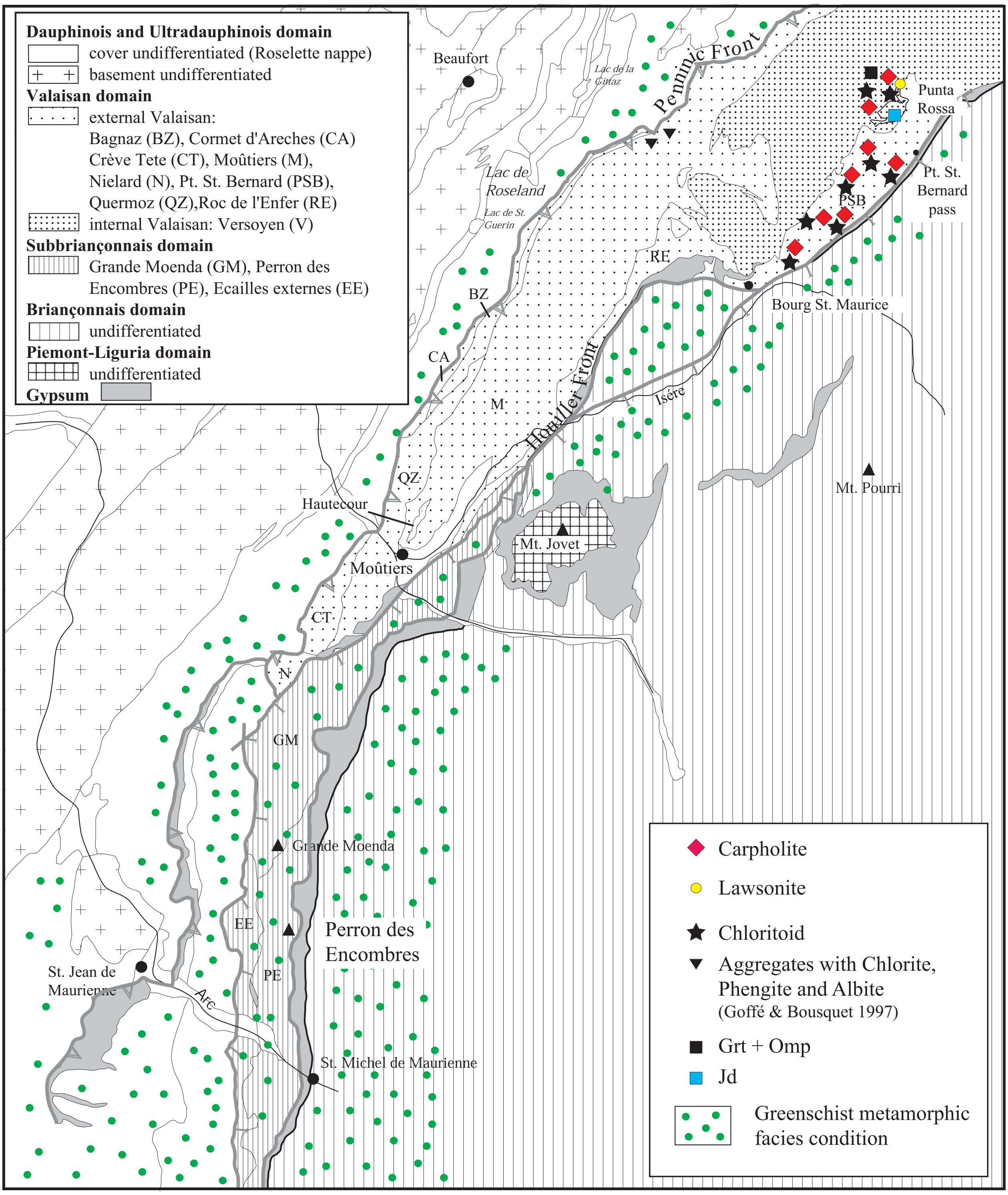


Fig. 4.2 Map of the metamorphic zonation within the Valaisan units and the adjacent cover nappes, i.e. Delphino-Helvetic, Subbriançonnais and Houiller domains. Petrological data from: Delphino-Helvetic domain, Leikine et al. (1983), Gely (1989), Gely and Bassias (1990), Jullien and Goffé (1993), Fry et al. (1999); Valais domain, Saliot (1979), Schürch (1986,1987), Goffé and Bousquet (1997), Bousquet (1998); Subbriançonnais and Houillèr domains Fry et al. (1999).

To the E the Houiller Front corresponds to an abrupt change in the metamorphic conditions between the HP Pt. St. Bernard and Versoyen units to the W and the greenschist metamorphic conditions of the Houillèr zone to the E.

The internal Delphino-Helvetic zone SW of the Mt. Blanc massif, is made up of several allochthonous cover units referred to as the Roselette nappe (Eltchaninoff-Lancelot et al., 1982) (Fig. 4.2) and basement units. The SW-NE striking zone comprises for lower to higher structural levels: the Roselend, the Roselette, the Rocher du Vent and the Crête du Gittes units. All these units show a common metamorphic evolution characterised by two metamorphic events under lower greenschist facies conditions (Gely, 1988; Leikine et al., 1983).

Related to the first event, the Delphino-Helvetic zone preserves a normal metamorphic gradient detected from the basement to the upper most Crête du Gittes unit that is included within the anchizonal to epizonal metamorphic conditions (Gely and Bassias, 1990). During the second event PT conditions were the same in all the units. Leikine et al. (1983) estimate these conditions to be in the range of 400-420 °C and 2,4-3 kb. In the Roselend unit, at the base of the Roselette nappe, the second metamorphic event corresponds to the crystallisation of chloritoid (Leikine et al., 1983). The chemical composition of this mineral indicates low-pressures, typical of greenschist facies condition (Leikine et al., 1983). The age of the second metamorphic event inferred from radiometric data is suggested to be between 13.4 ± 2 Ma and 18.3 ± 2 Ma (Leikine et al.1983 and references therein). Further to the south, the internal Delphino-Helvetic zone exhibits the same metamorphic conditions as in the Roselette nappe. According to Jullien and Goffé (1993) PT conditions range between 4-5 kb and 280-340 °C (Fig. 4.2). The more internal Subbriançonnais zone shows greenschist facies condition (Frey et al., 1999). To the E, the Houiller Front separates the Versoyen and Pt. St. Bernard units with HP relics from the less metamorphosed Zone Houillère. This zone is characterized by greenschist facies metamorphic conditions (Frey et al., 1999). At the Pt. St. Bernard pass well preserved fossils plants at the base of the Zone Houillère (Debelmas et al., 1991; Fabre, 1958) may indicate even lower greenschist facies metamorphic condition. This might suggest the presence of internal tectonic contacts, in the Houillèr zone, yet a detailed study is still missing.

Despite this lack in petrological studies in the Zone Houillère, metamorphic temperatures can roughly be estimated using the available zircon fission-track data (Fügenschuh pers. com.). The results for samples from the Zone Houillère close to the Houiller front are depicted in Fig. 4.3 a and b along a profile from the Col du Pt. St. Bernard in the NNE to the Arc valley in the SSW.

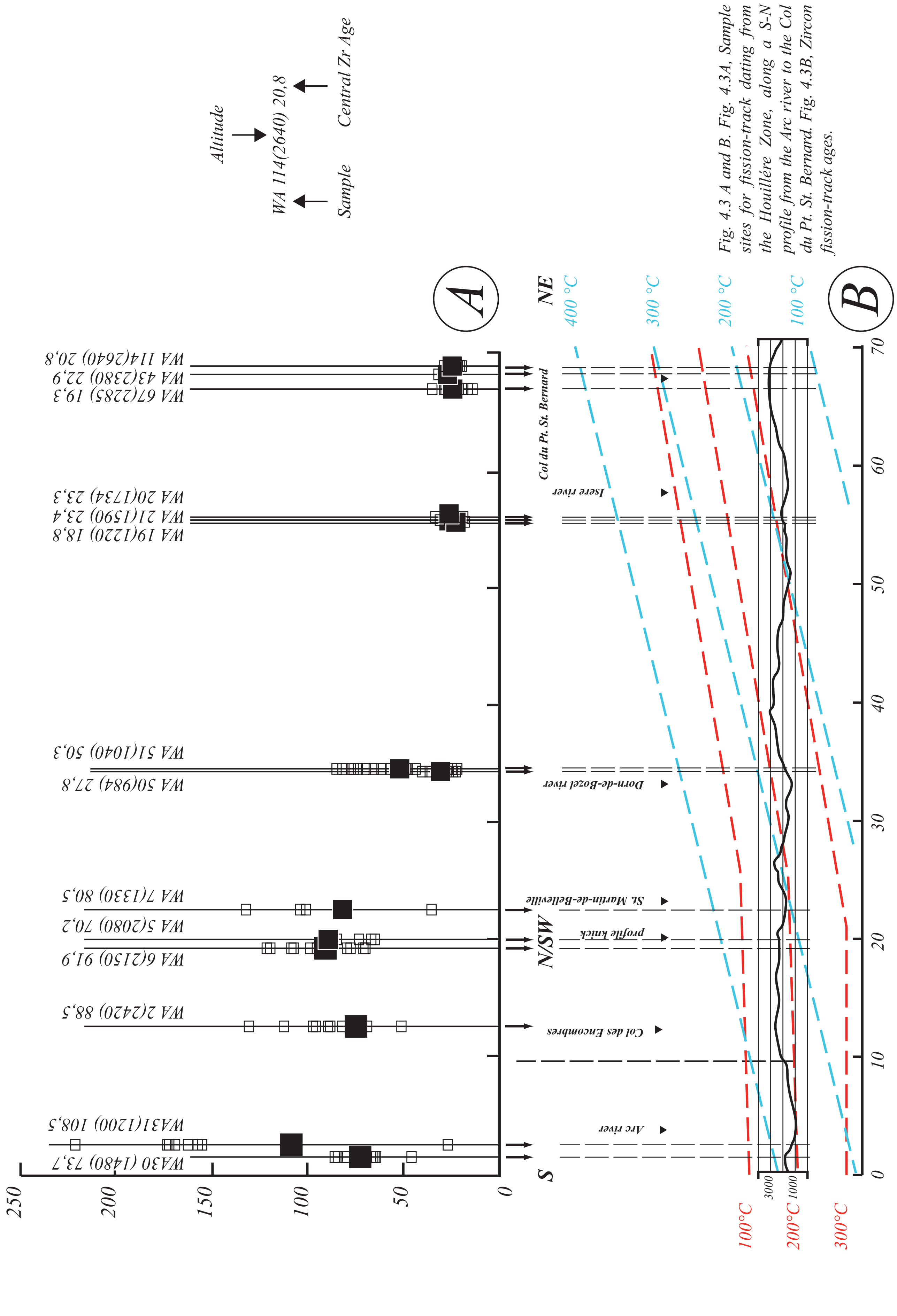


Fig. 4.3 A and B. Fig. 4.3A, Sample sites for fission-track dating from the Houillère Zone, along a S-N profile from the Arc river to the Col du Pt. St. Bernard. Fig. 4.3B, Zircon fission-track ages.

Samples from the south yielded zircon Central ages (Galbraith and Laslett, 1993) >60Ma and most of them fail the Chi-square test, reflecting the great spread in single-grain ages. Central ages for samples WA 50 and 51 from the Bozel valley are 27.8 and 50.3 respectively. Sample WA 51, which fails the Chi-square test contains two populations, the younger one indicating approximately the same age as WA 50. These two samples are interpreted as being partially reset, i.e. they have been subjected to temperatures between 200°C and 300°C well within the partial annealing zone for zircon (Yamada and Tagami, 1995). Central ages for the six northernmost samples gave very consistent ages between 18.8±1.8Ma and 23.4±2.4Ma with small variation in the single grain ages. These samples are therefore interpreted as being fully reset under metamorphic temperatures ≥300°C.

In Fig. 4.3b isotherms for two different geothermal gradients (20°/km blue dotted line, 30°/km red dotted line) have been drawn so as to let the samples lie within the temperature ranges estimated above. For a geothermal gradient of 20°C/km (blue line, Fig. 4.3b) 10°southward dipping isotherms allow for the estimated temperature conditions. On the other hand, for a geothermal gradient of 30°C/km (red line, Fig. 4.3b), 6° southward dipping isotherms fulfil the requirements until St. Martin de Belleville in the south. Yet further to the south ± horizontal isotherms are needed to keep temperatures above 100°C since apatites from this locality are still reset.

4.2: Relation between mineral growth and structural evolution in the Versoyen and Pt. St. Bernard unit

In this section, two different examples will be discussed in order to show the relation between mineral growth and structural evolution. Although in the next chapters (Chapters 5 and 6) a complete description of the structural elements will be given, a basic introduction to the structural features becomes already necessary now. In the internal and the external Valaisan a penetrative S1 cleavage is well known (Andrieux and Lancelot, 1980; Antoine, 1971; Cannic et al., 1996; Lancelot, 1979; Spencer, 1989). This S1 cleavage is generally parallel to bedding and related to isoclinal F1 folds. D2 produced tight to isoclinal folds and an S2 crenulation cleavage. Both D1 and D2, are common to all the Valaisan units.

In the Versoyen unit the most suitable lithologies for observing overprinting relations of different generations of cleavage are the metapelites of the Complexe Antéflysch. Most of the basaltic volcanic sills are thick and competent enough in order to preserve magmatic textures (Cannic et al., 1996). In this case the different foliations are not recorded and the HP minerals crystallised under static conditions. However, suitable rocks for such observations are the glaucophane schists where the mineralogical association is close to that of basaltic rocks.

Sample 1 (Fig 4.4a, b)

This sample was collected close to the hinge of a D2 east-facing synformal anticline within the Versoyen unit (sample coordinates are given in Fig. 4.4a, b). The two different visible cleavages are correlated to the D1 and D2 deformation phases. In thin section, S1 is defined by the crystallisation of chloritoid, chlorite and white mica, while S2, well visible by the alignment of small oxides, is characterised by the occurrence of white mica and chlorite. Chloritoid bent around S2 clearly formed during D1. However, locally chloritoid is re-oriented parallel to the S2 cleavage in which white mica and chlorite grew synkinematically.

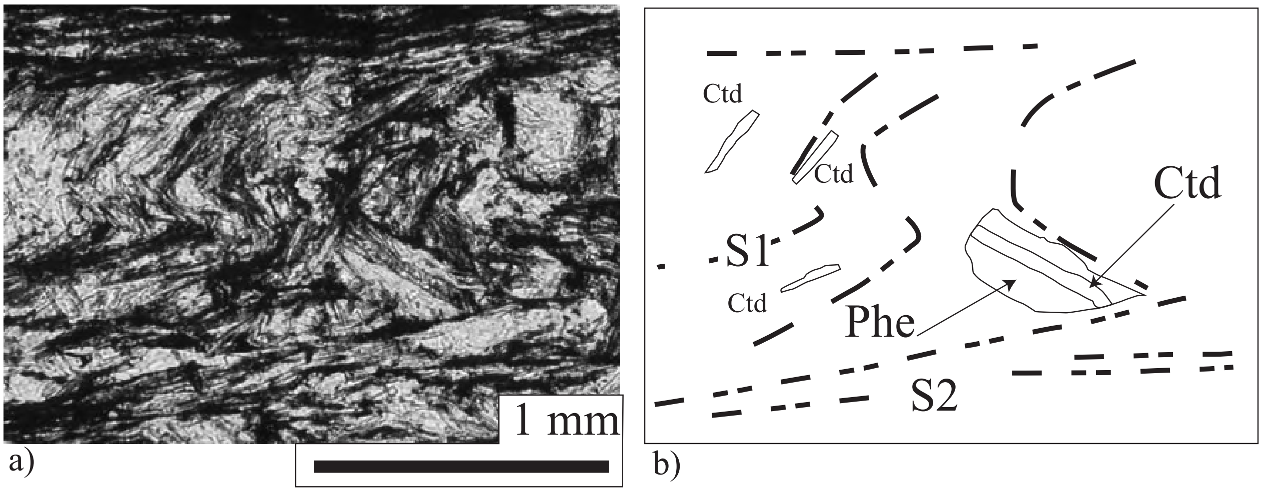
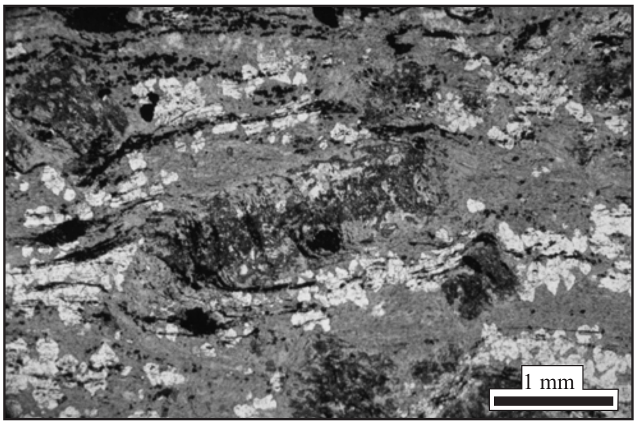


Fig. 4.4 a) and b).Thin section and line drawing, respectively, of chloritoid-bearing black schist (Complexe antéflysch, Versoyen unit, internal Valaisan). Chloritoid has grown along the main S1 cleavage (sample AL9617b, x:948125, y: 2081950, z:1810)

Sample 2 (Fig 4.4c,d and e)

This sample, a lawsonite-glaucophane schists, was originally described by Dalla Torre (1998). Macroscopically a compositional layering parallel to the S1 foliation is visible. In thin section this foliation is marked by small grains of oxides and titanite in a chlorite matrix as well as by helicitic textures produced by inclusion trails within albite porphyroblasts. White mica, clinozoisite/epidote and quartz forms up to millimetre-size pseudorphs after lawsonite. The relics or pseudomorphs of lawsonite and the preserved foliation show a rotated pattern in respect to the matrix and to the albite porphyroblasts. This texture is typical of syn-tectonic mineral growth and suggests that lawsonite grew during the formation of the S1 foliation. Glaucophane is found as relics in the same S1 foliation and as inclusions within albite porphyroblasts.



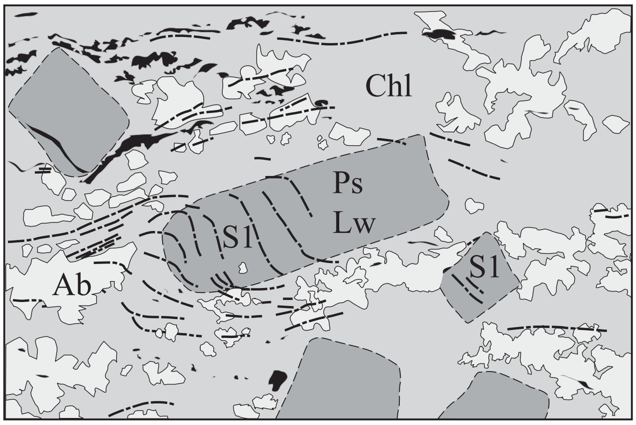


Fig. 4.4 c) and d) Thin section and line drawning of glaucophane-lawsonite bearing schist (Complexe antéflysch, Versoyen units, internal Valaisan). The S1 foliation is materialized by the alignment of oxides and by helicitic textures made up by inclusion trails within albite porphyroblast (Fig. 4.4 e). In Fig. 4.4 c) and d) lawsonite, now preserved as pseudomorphs, and the S1 foliation show rotated pattern in respect to the matrix and to the albite porphyroblasts. This infers that lawsonite grew during the formation of S1 foliation.

These observations indicate that the S1 cleavage characterised by synkinematic growth of glaucophane and lawsonite, formed under blueschist facies conditions whereas albite, chlorite and the pseudomorphic assemblage after lawsonite formed under subsequent greenschist facies conditions, characterising the retrograde part of the PT path.

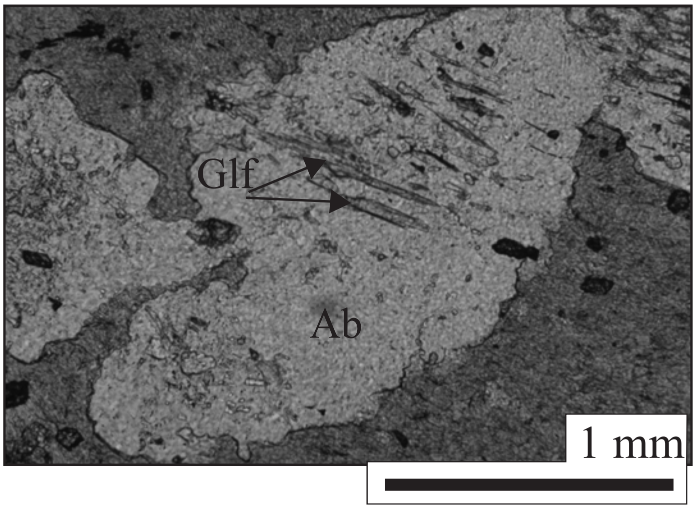


Fig. 4.4 e) Enlargement of the same thin section of Fig. 4.4 c). Along the same foliation (S1) relicts of glaucophane are preserved as are inclusions within the albite porphyroblasts. (specimen of Figs. 4.4 c, d and e was provided by Dalla Torre, 1998).

4.3: Conclusion of the grade and evolution of metamorphism in the Valaisan units

Up to now, no detailed studies correlating mineral growth and structural elements were available for the investigated area. The most recent attempts in this direction indicate that chloritoid is commonly aligned within the main foliation, without specifying this foliation (Goffé and Bousquet, 1997) while Bousquet (1998) correlates it to the D2 deformation phase.

Micro- and macroscopical observations indicate a clear relation between mineral growth and the D1 deformation phase in the Versoyen and Pt. St. Bernard units. Chloritoid in metapelitic rocks (object of this study), and glaucophane and lawsonite in metabasites (Dalla Torre, 1998), lie within the S1 foliation. Integrating these observations into the PT path after Goffé and Bousquet (1997) and Bousquet (1998), the observed syn-D1 chloritoid is proposed to replace carpholite, according to the reaction:

Fe-carpholite = chloritoid + quartz + water,

which post-dates HP peak conditions. The same metamorphic conditions are also indicated by the occurrence of syn-D1 glaucophane and lawsonite (Dalla Torre, 1998).

This study suggests that D1, i.e. the first observable deformation phase, is related to exhumation processes (Fig 4.1). Probably due to the pervasive overprint of D1 and D2 phases, no structural elements related to the growth of carpholite (pre-D1, see also section 6.1) have been preserved within the Versoyen and Pt. St. Bernard units. However, pre-D1 deformation, related to subduction has to be invoked (see Chapter 5, sections 5.4 and 5.5).

In conclusion, during the Alpine orogeny, the Valais domain underwent a multi-stage metamorphic evolution. While Versoyen and Pt. St. Bernard units, internal Valaisan, and part of the external Valaisan respectively, underwent HP metamorphic conditions (Bousquet, 1998; Goffé and Bousquet, 1997; Schürch, 1987) all other Valaisan units the metamorphic conditions did not exceed greenschist facies conditions. The metamorphic gradient presently observed within the Valaisan domain (see Fig. 4.1)

In the Versoyen and Pt. St. Bernard units, however, chloritoid and the association glaucophane-lawsonite, grow during the formation of the S1 foliation. These observations, combined with the metamorphic evolution suggested by Goffé and Bousquet (1997) and Bousquet (1998) led to the conclusion that the first deformation phase D1 already post-dated the HP peak conditions in the Versoyen and Pt. St. Bernard unit. As a consequence, it could be proposed that both units, i.e. Pt. St. Bernard and Versoyen units, have been already in contact before the D1 deformation phase (see Chapter 5).

It has to be noticed that according to our findings (see Chapters 5 and 6), the same S1 foliation, has been observed to be axial plane of F1 folds affecting all the different Valaisan rocks as well as the contacts between the pre-,syn-rift sediments, basaltic volcanic rocks and related sediments of the Complexe Antéflysch, as well as post rift sediments (see Chapter 6.1). This infers that during D1 all the different Valaisan units have been already tectonically juxtaposed and that they have underwent, simultaneously, different metamorphic conditions in different parts of the Valais domain. In conclusion, it is here proposed that the metamorphic grade of the Valaisan units is characterised by a metamorphic gradient originated during the first D1 deformation phase. This gradient is included between the HP condition of the Versoyen and Pt. St. Bernard units (internal Valaisan and internal part of the external Valaisan respectively) and the green schist metamorphic condition of the Moûtiers and Roc de l'Enfer units (more external part of the External Valaisan). Thus, looking in map view (Fig. 4.2), the present day distribution of different metamorphic conditions within the Valais (metamorphic gradient), it must be considered as the result of the superimposed deformation phases (D1 to D6) that lead to the refolding of the metamorphic gradient instead of the tectonic juxtaposition of slices showing different metamorphic conditions.

CHAPTER 5: THE N APPE STACK OF THE VALAISAN UNITS

Overprinting relationships in the field as well as thin sections observations allowed to distinguish several deformation phases, i.e. from pre-D1 to D6. With the exception of the allochthonous Pt. Rossa slice of basement, the age of the rocks of the Valaisan units ranges from Carboniferous to probably Tertiary. Therefore, the deformation history of the Valais domain must pertain to the Alpine paleotectonic and orogenic evolution. The structural analysis of the Valaisan units has been divided into two chapters (Chapters 5 and 6).
The complexity of this multiply deformed area requires the regional geometry of the nappe stack of the Valaisan units to be introduced and discussed in Chapter 5, (section 5.1 to 5.7). The structural elements related to each deformation phase are then discussed in the following chapter (Chapter 6) in detail emphasising features that place constraints on the kinematic evolution of the Valaisan units.

5.1: Main structural features of the study area

This section exclusively deals with the main structural features of the Valaisan units as deduced from the new structural map of the study area (Figs**.** 5.1 and 5.2). Following an inverted chronological order of deformation phases, the geometrical effect of the D4 and D2 deformation phases will be described in more detail along two main tectonic profiles constructed across the Valaisan units (Chapter 5.2). Geometric features older than D2 and formed during the pre-D1 and D1 phases are the most complex and occupy a major part of the descriptions of the nappe stack of the Valaisan units. In a first step these older features are inferred from the analysis of F2 structural facing directions (Chapter 5.3) and in a second step they will be defined (Chapter 5.4) and described by selecting a number of small but crucial areas (Chapters 5.5 and 5.6). The last section (5.7) summarises the data presented in order to achieve a better understanding of the regional scale features within the study area.

Our gradual approach towards the understanding of the nappe stack of the Valaisan units starts by looking at the tectonic map presented in Fig. 5.1. This map shows the present-day disposition of the Valaisan unit. For a complete framework of the structural and geometrical features of the study area (Fig. 5.1) should be combined with the map of the axial planes presented in Fig. 5.2.

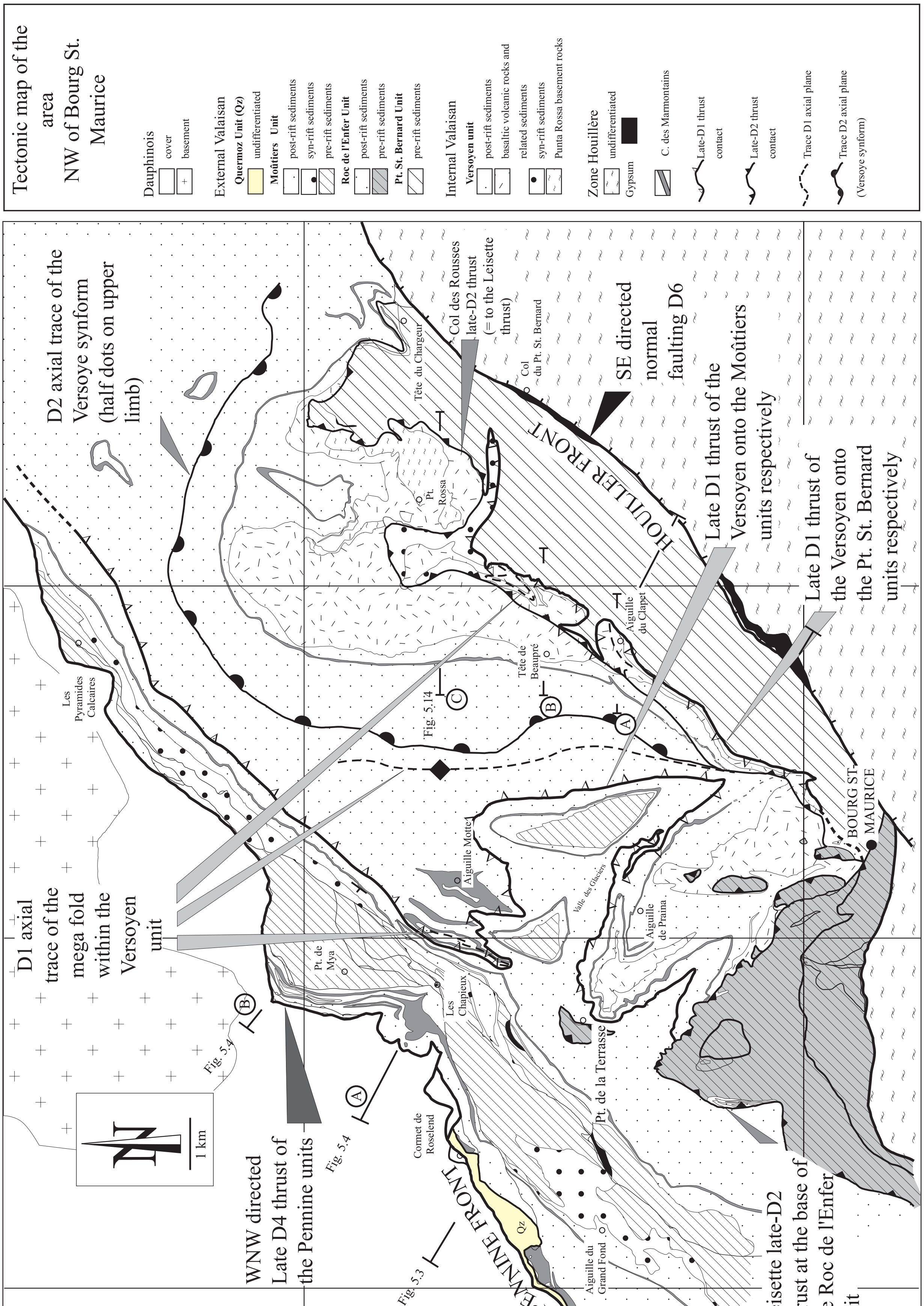


Fig. 5.1 Tectonic map of the study area showing the main structural features of the Valaisan units.

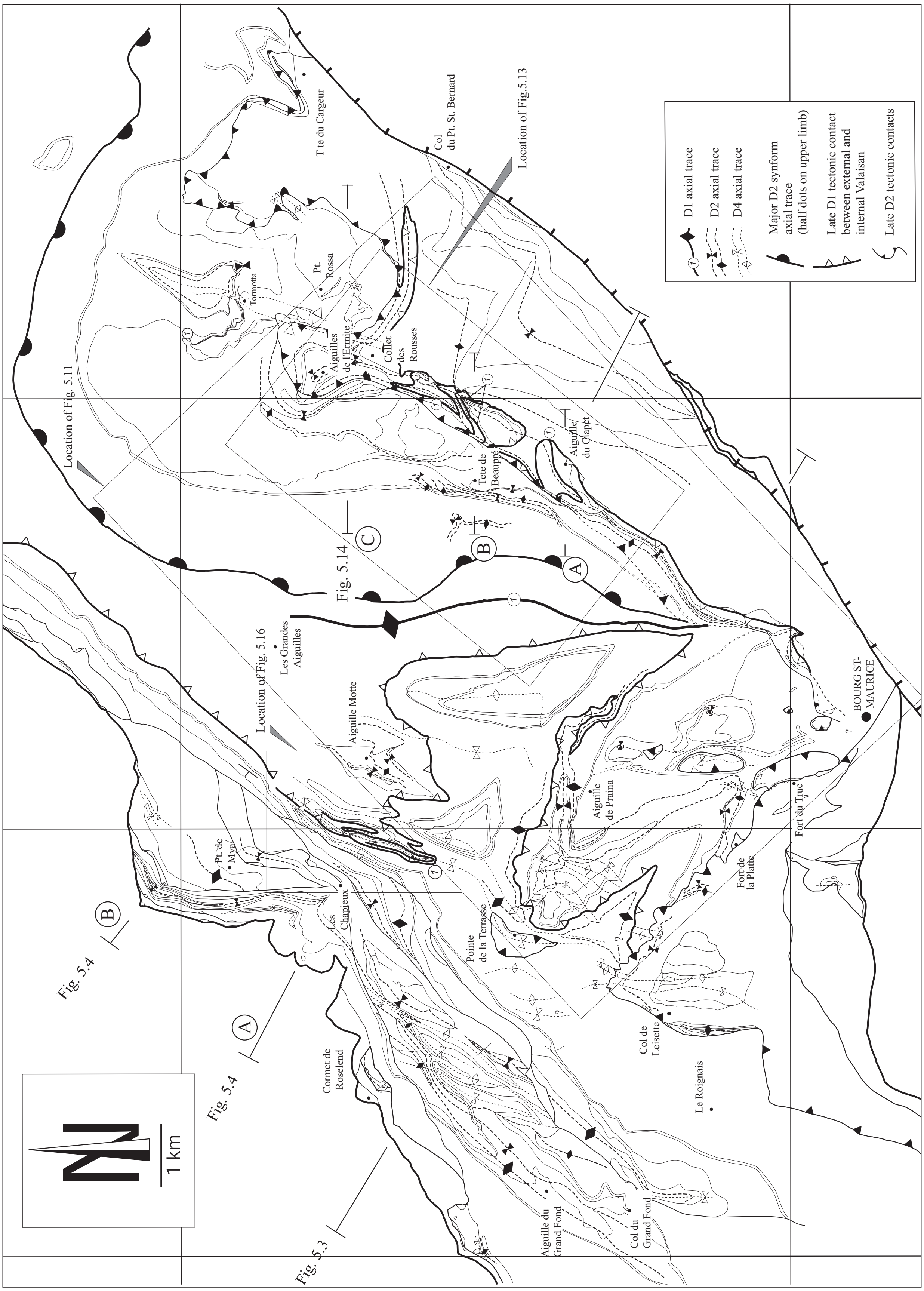


Fig. 5.2 Axial plane traces of the F1, F2 and F4 folds within the Valaisan units

The most prominent large scale structures of the Valaisan units are represented in two WNW-ESE vertical cross sections (Figs 5.3 and 5.4), constructed across the Valais zone from the Pt. St. Bernard area in the E to the Chapieux-Cormet de Roselend area in the W. Amongst the different deformation phases, i.e. pre-D1 to D6, only D1, D2 and D4 have a significant regional impact on the geometry of the nappe stack of the Valaisan units. The major relationships between the different units produced by these deformation phases, i.e. D1, D2 and D4 are summarized in three schematic NW-SE oriented profiles shown in Fig. 5.5. The others phases, i.e. D3, D5 and D6 being of more local extent will not be discussed in this chapter. Nevertheless, the D3, D5 and D6 deformation events have important consequences on the regional geometry and will be presented in Chapter 6 (6.4, 6.6 and 6.7 respectively).

In the SE part of the study area (Fig. 5.1), from Bourg St. Maurice towards the Col du Pt. St. Bernard, the map trace of the Houillèr Front (HF), outlined by the occurrence of Upper Triassic evaporites, can be followed as a SW-NE oriented straight line, cutting all previously formed structures. In the Cormet de Roselend area, in the northwestern part of the study area (Fig.5.1), the map trace of the Pennine Front (PF) indicates a gently SE dipping fault zone, steepening towards NE as it approaches the Mont Blanc massif in the Pyramides Calcaires area. In the study area, between the Cormet de Roselend and the Pyramides Calcaires, the Pennine Front represents the map trace of the late-D4 basal thrust of the Valaisan units over the Dauphinois zone in its footwall (Fig. 5.5, D4 deformation phase).

The most outstanding feature observed between the PF in the W and the HF in the E is a regional scale D2 synform, the Versoye synform (Figs 5.1 to 5.5). In the SE, i.e. the overturned limb of the Versoye synform, the Pt. St. Bernard unit rests above the Versoyen unit. In the W the Moûtiers unit, considered as a more external part of the external Valaisan domain, rests below the Versoyen unit (Fig.5.5, D2 deformation phase).

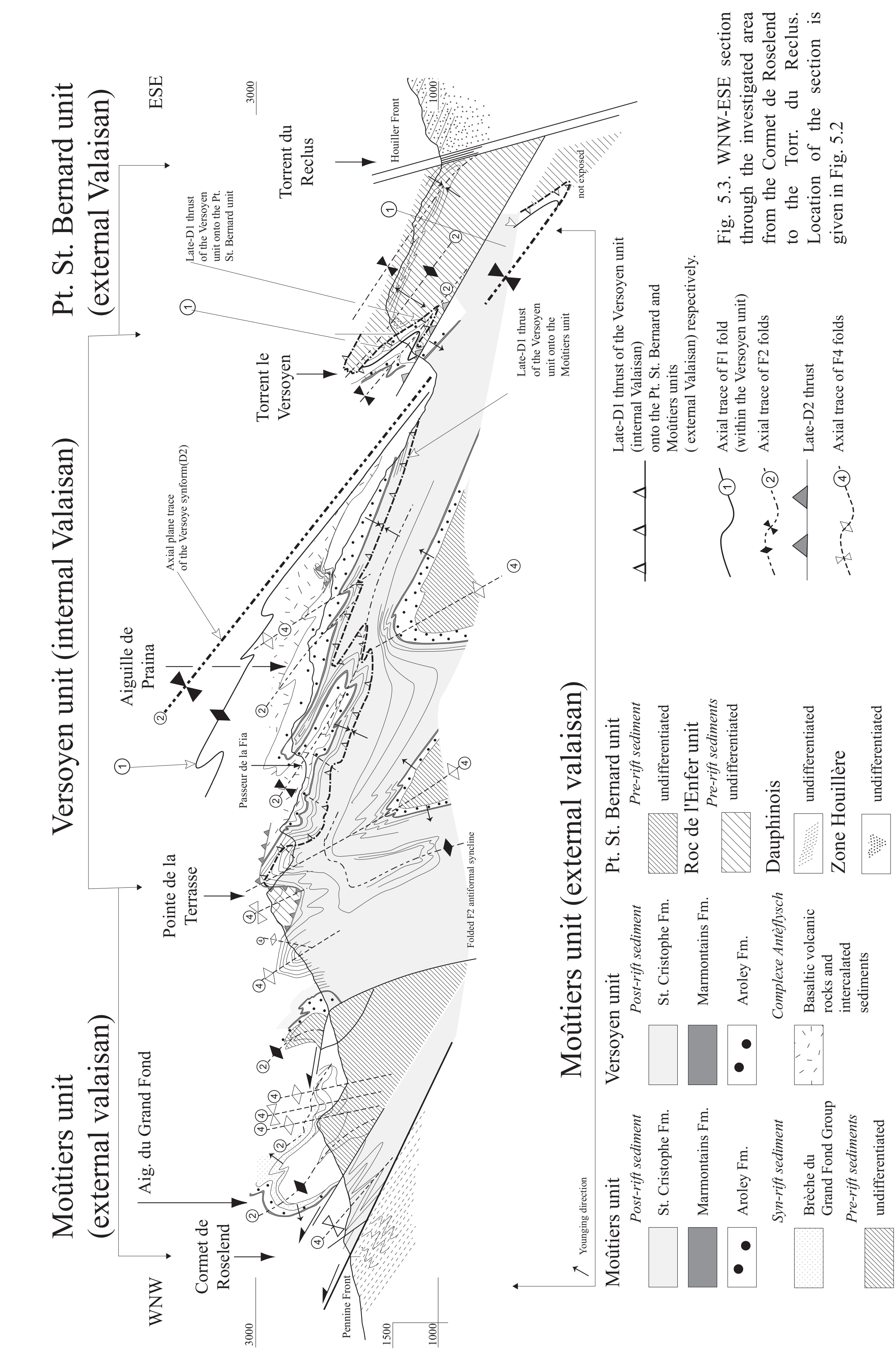


Fig. 5.3. WNW-ESE section through the investigated area from the Cormet de Roselend to the Torr. du Reclus. Location of the section is given in Fig. 5.2

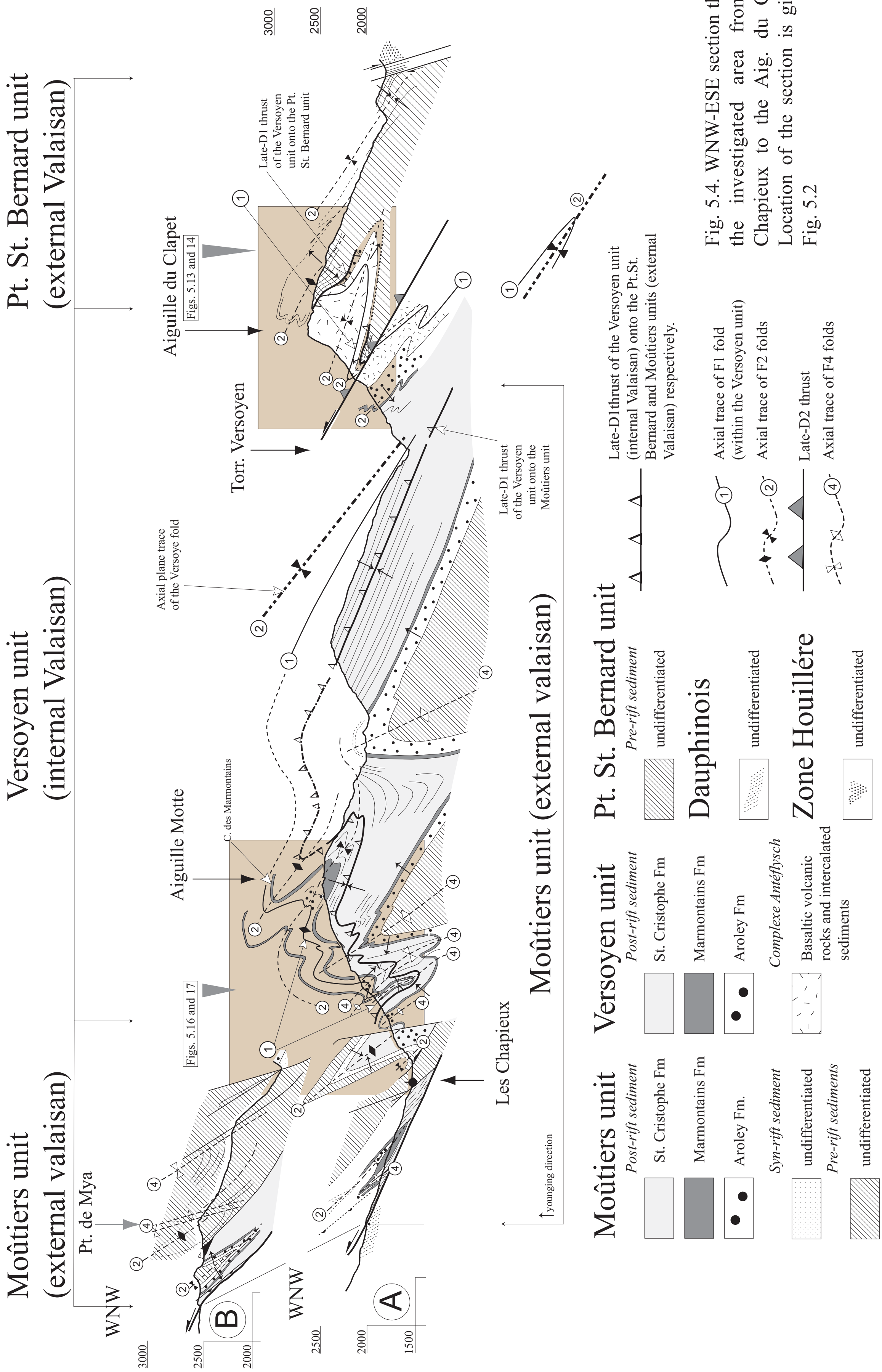


Fig. 5.4. WNW-ESE section through the investigated area from Les Chapieux to the Aig. du Clapet. Location of the section is given in Fig. 5.2

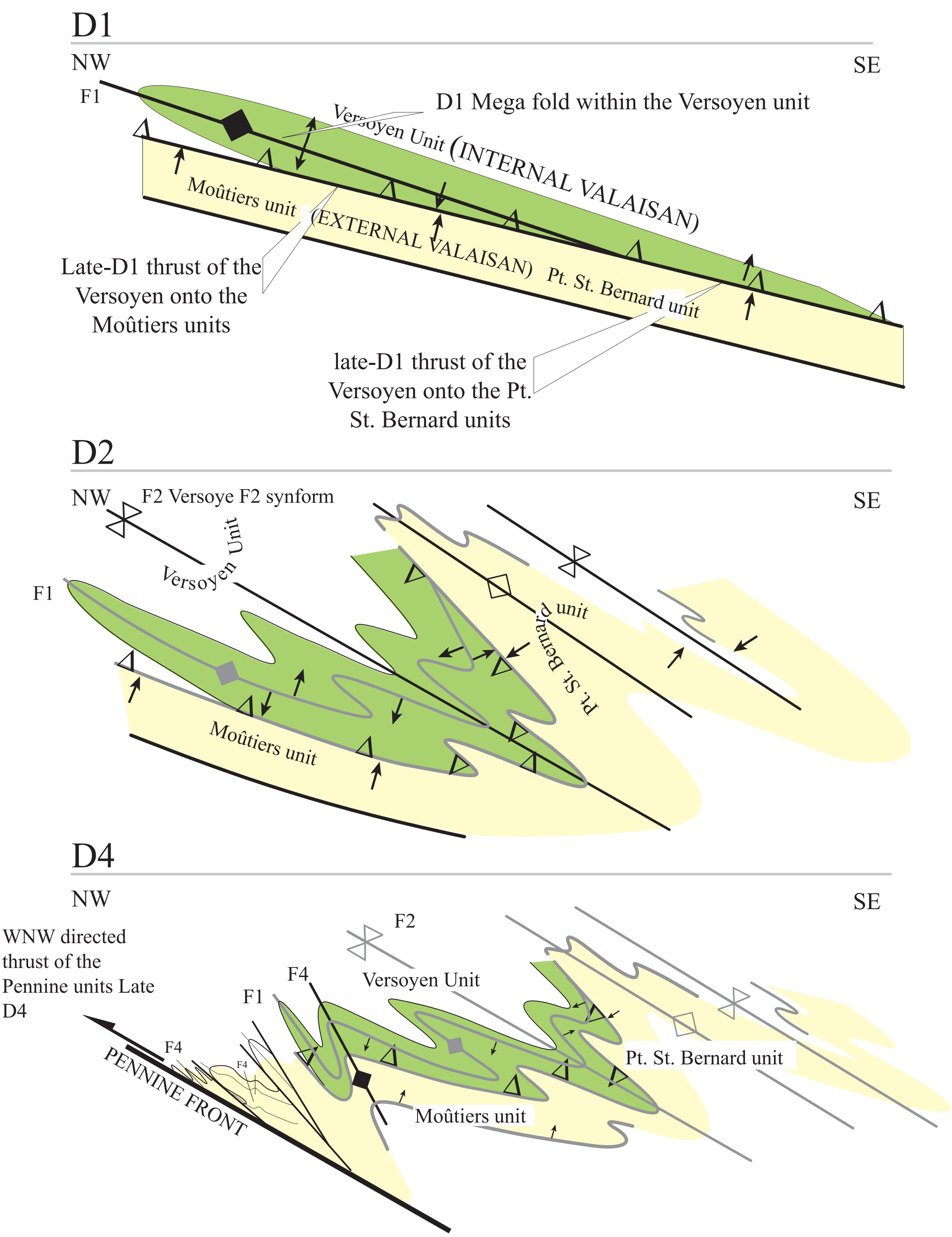


Fig. 5.5 Schematic SE-NW oriented profiles across the Valais Zone showing the regional geometrical relationships during D1, D2 and D4 between the Versoyen, the Moûtiers and the Pt. St. Bernard units respectively. The arrows indicate the younging direction. Notice that the disposition of the F1 inverted and normal limbs within the Versoyen unit is overturned in the upper limb of the D2 Versoye synform.

Consequently the Versoyen unit forms the core of the D2 regional scale Versoye synform. The study of the asymmetry of parasitic F2 folds at the scale of the study area revealed that this disposition of the Pt. St. Bernard and Moûtiers units is compatible with the D2 regional scale E closing Versoye synform (see section 5.2). with the Versoye synform displaying gently E–SE dipping axial planes and gently S to SE plunging fold axes, cause refolding of the previously formed thrust and fold nappe stack of the Valaisan units which originated during D1. In Fig. 5.1 the refolding of the D1 nappe stack of the Valaisan units allows the fact that older D1 features do occur on both limbs of the Versoye synform.

Another important large scale D2 structure within the Valaisan units is visible NW of Bourg St. Maurice, where the Roc de l'Enfer unit individualized from the Moûtiers unit during late-D2 thrusting. Late during D2 the Roc de l'Enfer unit overrides and, hence, crosscuts the early D2 Versoye synform axial trace in map view. This feature is not represented in Fig. 5.5 and will be discussed in section 6.3.2.

Retrodeformation of the D2 Versoye synform allows to focus on the geometrical analysis of the Valaisan units regarding the, so far undetected, prominent regional scale D1 features. These features, revealing an early stage of nappe stacking of the Valaisan units, indicate a late-D1 thrust contact and a D1 recumbent mega-antiform within the Versoyen unit (see Fig. 5.5, D1 deformation stage). In map view (Fig. 5.1) late-D1 thrust contacts are visible between the Versoyen and the Pt. St. Bernard units in the eastern and between the Versoyen and the Moûtiers units in the in the western part of the study area respectively. Mapping of the late-D1 thrust contact (open triangles in Fig.5.1 placed on the hangingwall of the former D1 thrust) indicates that the contact between the Versoyen unit and the Pt. St. Bernard unit (inverted due to D2 folding) may be directly connected with the contact between the Versoyen unit and the Moûtiers unit further in the W folded around the F2 Versoye synform (compare D1 and D2 in Fig. 5.5). The late-D1 thrust therefore represents the limit between the external Valaisan comprising the Pt. St. Bernard and Moûtiers units and the internal Valaisan (the Versoyen unit see D1 in Fig. 5.5).

The axial trace of the D1 mega fold within the internal Valaisan was mapped out in the Aig. du Clapet-Tête de Beaupré area (Figs. 5.1 and 5.2) where it is situated in the upper and overturned limb of the F2 Versoye synform. The other outcrops are located in the lower limb of the Versoye synform and are best exposed in the Aig. Motte area (Figs. 5.1 and 5.2). The F1 axial trace is not continuously outcropping due to erosion. Often the outcrops of the Versoyen unit are only preserved within either the inverted or the normal limb of this F1 megafold. Most of the inverted limb of the D1 megafold within the Versoyen unit outcrops in the lower (right way up in respect to D2) limb of the D2 Versoye synform and situated above the D1 thrust contact to the Moûtiers unit.

On the other hand the former normal limb of the D1 mega anticline is now found in that part of the Versoyen unit which rests on the upper (overturned in respect to D2) limb of the D2 Versoye synform (see Fig 5.5, D2 deformation stage). D1-thrusting postdates D1-folding as evidenced by the fact that, in map view (Figs. 5.1 and 5.2) the F1 axial planes are truncated by the late-D1 thrust contact between the Versoyen and Pt. St. Bernard units. Since this contact postdates F1 folding and was subsequently refolded and overturned during D2 D1 thrusting is referred to as "late-D1" (see Figs. 5.5 and 5.6). All these observations will be extensively supported by field arguments during this chapter (5.1 to 5.7). However, it becomes immediately apparent from the geometries of Figs. 5.1 and 5.5 that the original nappe stack of the Valaisan units developed from an earlier D1 fold nappe stage to thrusting of the internal Valaisan onto the external Valaisan during a later stage of D1. The subsequent superposition of D2 and D4, leading to the present day situation as depicted in Fig. 5.1.

The following section focuses on a particular region within the Valais zone, where the prominent D2 and D4 structural and geometrical features, only briefly mentioned so far, will be described in more detail.

5.2: Major D2 and D4 structures: the refolding geometry of the nappe stack of the Valaisan units

In order to describe the D2 and D4 geometrical features in more detail two WNW-ESE oriented vertical sections have been constructed: the Cormet de Roselend–Torrent de Reclus sections (Fig. 5.3) and the Chapieux – Aiguille du Clapet sections (Fig.5.4). Each of these sections, running from the Pennine Front in the W to the Houillèr Front in the E, cross the three major Valaisan units which are from W to E: the Moûtiers, the Versoyen and the Pt. St. Bernard units, respectively. A fourth Valaisan unit (i.e. the Roc de l'Enfer unit), representing a sub-unit individualized from the Moûtiers unit during a late stage of D2 is shown in the cross section of Fig. 6.19 and its structural features will be discussed in section 6.3.2. Several authors (Antoine, 1971; Antoine et al., 1992; Fudral, 1980; Lancelot, 1979; Spencer, 1989) have already constructed profiles between the Cormet de Roselend and the Torrent De Reclus. Substantial differences in terms of the interpretation of both the stratigraphic record and the structural evolution, as revealed by the present study, asked for redrawing this profile (Fig. 5.3) and construction of an additional one located further to the NE, i.e. the Chapieux-Pt. du Clapet profile (Fig. 5.4, with two branches A and B). Both profiles, document the structural evidence supporting the prominent geometrical features of the nappe stack of the Valaisan unit briefly introduced in the previous section. However, the understanding of the relationships between the different Valais units is also closely connected to the proposed paleogeographical distinction between External and Internal Valaisan (Chapter 2), as well as to the tectono-metamorphic evolution of the study area (Chapter 4).

In all the profiles and detailed geological maps discussed in this chapter, the formations of the Valaisan units will be subdivided into pre-, syn- and post-rift sediments and additionally, into the Complexe Antéflysch (basaltic volcanic rocks and intercalated sediments of the Versoyen unit). In order to facilitate the distinction between different tectonic units, the pre-rift sediments of the Moûtiers unit were given a slightly different signature from those of the Pt. St. Bernard unit in Figs 5.3 and 5.4. The thin post-rift Marmontains Fm. constitutes a key formation for the tectonic interpretation and therefore depicted separately. The Marmontains Fm. are highlighted in order to give evidence for the polarity of the younging direction.

In order to depict the geometrical features that are evident from a comparison between the two profiles, the large scale D4 features schematically shown in Fig. 5.5 and consisting of folding and thrusting (D4 deformation phase) have been enhanced Fig.5.6. The first observation which can be made from Fig. 5.6 is that not all the Valaisan nappe stack was affected by the same intensity of D4 deformation. D4 deformation becomes more intense from E towards W, i.e. towards the Pennine Front, which is therefore interpreted as a late-D4 thrust. This interpretation is supported by the fact that D4 folding becomes pervasive towards the Pennine Front, i.e. towards W.

Conversely F4 folds are almost absent within the Pt. St. Bernard unit further to the E. The only occurrence of F4 folds found in the upper limb of the D2 Versoye synform are located within the Versoyen unit, i.e. in the Tormotta area north of the Pt. Rossa (Fig. 5.2, see Dalla Torre 1998), while further to the W and in the lower limb of the Versoye synform, these folds can be widely observed within both the Versoyen and the Moûtiers units. The Moûtiers unit, exposed in the central part of Fig. 5.6a (at the base of the ridge between the Pt. de La Terrasse and the Aig. de Praina) and of Fig. 5.6b (in the Aig. Motte area), D4 folds overprint the previous D1 and D2 structures. D4 folds essentially consist of two antiform-synform pairs.

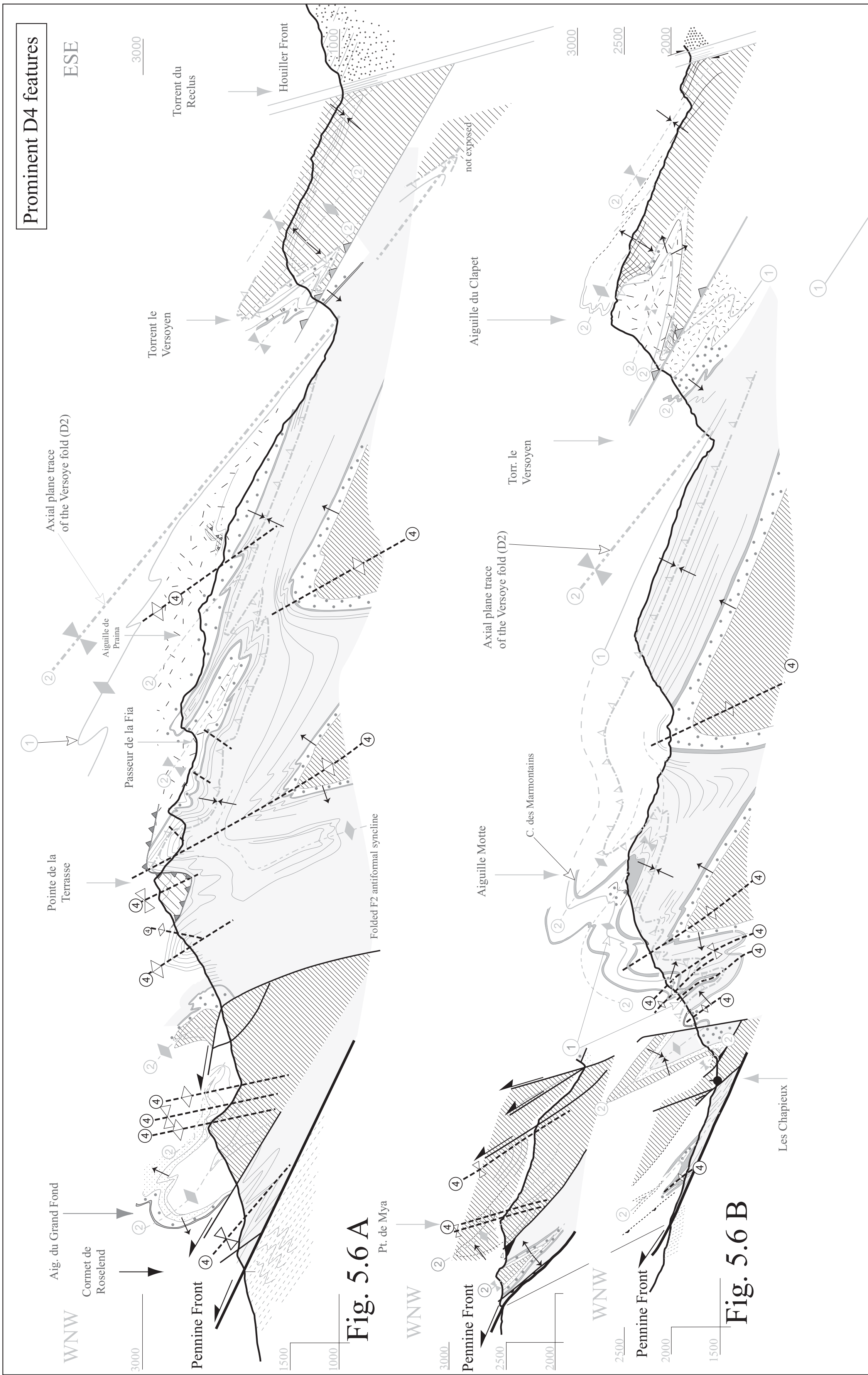


Fig. 5.6 a) and b). Enhancement of the prominent D4 features of the profile of Fig. 5.3 (Fig.5.6a) and of Fig. 5.4 (Fig.5.6b).

In the core of these antiforms the lowest structural and stratigraphic level, i.e. the pre-rift sediments of the Moûtiers unit are exposed, Post-rift sediments form the core of synformal synclines. In map view these two antiforms (Figs. 5.1 and 5.2) form two windows surrounded by post-rift sediments. They represent the most accessible D4 structures and it is possible to identify all their characteristic features. In cross section, these D4 folds display a relatively open geometry and a vergence towards the NW. F4 fold wavelengths range in scale from tens to several hundreds of meters, with steeply SE dipping axial planes and SSW-NNE-striking sub-horizontal fold axes. The limbs of these D4 folds generally exhibit steeply inclined to sub vertical strata. These folds clearly overprint and locally steepen all the previous structures i.e. D1 and D2. F4 folds are also exposed in the higher parts of the cross sections and within the Versoyen unit, i.e. in the central part of Fig. 5.6a around the Pt. de la Terrasse. However, here the D4 folds display steeply dipping axial planes, inclined towards NW or SE. This change in geometry towards higher tectonic levels sometimes leads to box folds as observed near Pt. de la Terrasse. Also very open D4 folds, characterised by open hinges have been observed as well.

Approaching the Pennine Front the F4 folds, progressively become more isoclinal and they are associated with D4 thrusts, including the Pennine Front itself. All of these thrusts share a common transport direction towards the WNW (see section 6.5.1). The strain gradient displayed by D4 structural features towards the Pennine Front is also apparent in map view over the entire study area (Figs.5.1 and 5.2). In the NW part of the study area, i.e. between the Aig. du Grand Fond in the SW and the Pyramides Calcaires in the NE, a 2-3 km wide NE-SW striking corridor of the Moûtiers unit, is totally dominated by D4 structures. This corridor is adjacent and runs parallel to the Pennine Front. The Pennine Front is defined as a discrete mylonitic shear zone. Minor thrusts, together with intense D4 folding are characteristic for the adjacent areas in the hangingwall of the Pennine Front. However, the Dauphinois zone, situated in the footwall of the Pennine Front, lacks discrete thrusts and is characterised by pervasive ductile isoclinal folding. Despite of the fact that the frontal parts of the Valaisan appear to be one of the most complex from a structural point of view, the overprinting relationships observed still allow for a reconstruction of the complete structural chronology. In this part of the Moûtiers unit, D4 folding shows various geometries ranging from isoclinal folds in the less competent post-rift sediments to tight folds in the most competent pre-rift sediments of the Pt. de Mya and the Grand Fond massif (Fig. 5.6). In the latter region the relationships observed indicate the superposition of D4 folds on previous structures i.e. D2 folds. All these structures, D2 and D4 are truncated by thrusts that branch off the SE dipping Pennine Front.

In order to illustrate the main features related to the D2 deformation phase, D4 has been retro-deformed in Fig. 5.5 (D2 deformation phase). Additionally the prominent D2 structures have been enhanced in Fig. 5.7. The gently E-SE dipping axial planes (except in the limbs of D4 folds) and the intense isoclinal folding observed within the entire study area are in clear contrast to the previously described D4 structures. Beautifully exposed examples of interference patterns between F2 and F4 folds are recognizable in the Grand Fond massif Fig. 5.3 and in the Pt. de Mya area Fig.5.4b. Along the ridge between the Pt. de la Terrasse and the Aig. de Praina, i.e. in the central part of Fig. 5.3, and further E towards the Aig. du Clapet and up to the Houillèr Front (Fig. 5.4) D2 structures are perfectly preserved because D4 overprint is weak to absent (Fig. 5.7). In this area, a systematic analysis of small scale D2 structures supports the large scale interpretation in terms of the D2 Versoye synform and highlights the so far undetected D1 mega-fold including and late-D1 thrusting of the internal Valaisan above the external Valaisan. Decametric to kilometric scale amplitude F2 isoclinal folds are clearly recognizable in the Valleé des Glaciers below Les Chapieux (Fig. 5.8) and in the Aig. du Clapet-Tête de Beaupré area (Fig. 5.9). Fig. 5.8 represents a detail of the central part of the profile of Fig. 5.3 (see also Fig. 5.7 a) while Fig. 5.9 represents the eastern part of profile of Fig. 5.4 (see also 5.7 b).

There are two crucial observations that can be made by analysing the D2 structural features. The first concerns the asymmetry of the F2 parasitic folds that, in the regional frame, will help to localise the major D2 Versoye synform. The second concerns the polarity of F2 facing directions which are crucial for the discussion of the D1 features. The first kind of observation will be presented below while the latter will be discussed in the next section (5.3).

The F2 folds shown in Fig. 5.8 exhibit an asymmetry characterised by a Z shape (looking SW). Also along the ridge between the Pt. de la Terrasse and the Aig. de Praina all F2 folds are characterised by a vergence which is compatible with a position of these outcrops in the lower limb of a major SE-closing D2 synform, i.e. the Versoye synform. In order to get an idea of the dimensions of the Versoye synform, one has to keep in mind that the lower limb of the large scale F2 synformal anticline of Fig. 5.8, which merely represents a parasitic fold in the lower limb of the Versoye synform, is about 3 km long. Fig. 5.9, together with the eastern parts of the two profiles, of Fig. 5.7, shows that the asymmetry of the F2 folds switches to an S shape (looking SW) which is compatible with a position in the upper limb of the same Versoye synform. The same change in F2 vergence which occurs at different structural levels within the Valaisan nappe stack has been systematically mapped within the entire study area.

Fig. 5.7 a) and b). Enhancement of the prominent D2 features in the profile of Fig. 5.3 (Fig.5.7a) and of Fig. 5.4 (Fig.5.7b).

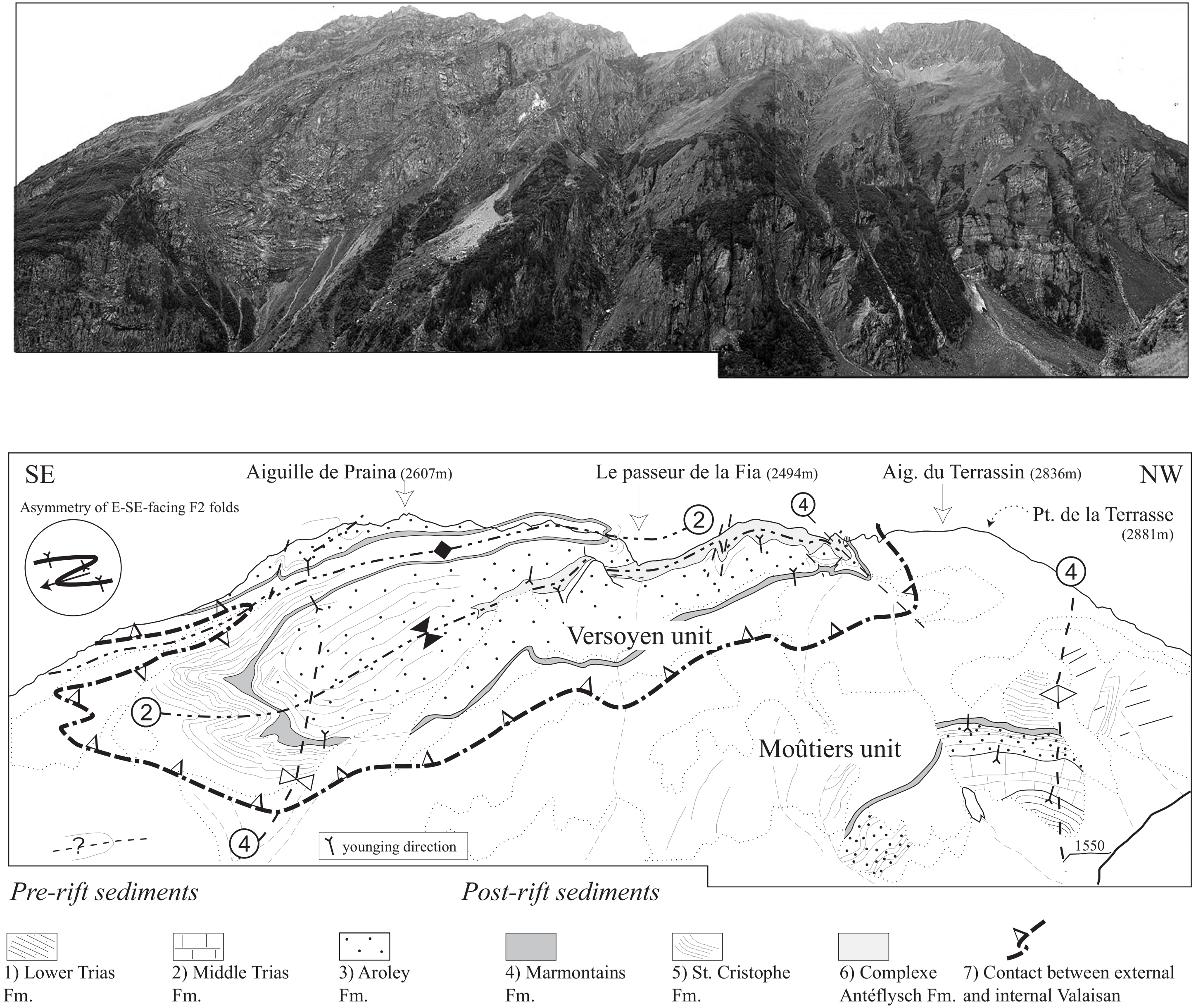


Fig.5.8. Photograph and line drowning of the NE ridge between the Aig. de Praina and the Pt. de la Terrasse (lower part of the Valleé du Glaciers) showing the cliff section where the tectonic contact between the internal and external Valaisan is exposed. From the bottom of the valley up section one can see the Moûtiers unit displaying a normal sequence. This sequence consists of the pre-rift sediments, 1) and 2) , that are followed by post-rift sediments; 3) , 4) and 5) . Further up section the same post-rift sediments display an inverted sequence (5,4,3) resting in stratigraphic contact with the basaltic volcanic rocks and related sediments of the Complexe Antéflysch (6) forming the Versoyen unit. The contact between external and internal Valaisan (7) is considered to be situated within the C. de St Cristophe. Note that both sequences of post-rift sediments are commonly folded during the D2 deformation phase. The asymmetry of the SE-facing F2 folds (looking SW, Z shape, see inset) is compatible with the lower limb of the Versoyen synform (D2). Compare this situation with the inset of Fig. 5.9.

This allowed to map the trace of the axial plane of the regional scale D2 Versoyen synform as presented in Figs. 5.1 and 5.2. In the eastern part of the study area, i.e. within the Pt. St. Bernard unit the vergence of the F2 folds indicates a major antiform situated structurally above the Versoye synform (at Aig. du Clapet in Fig. 5.7) which is than followed by another synform further E (see Fig. 5.7). Hence, the lower limb of the Versoye synform is considerably longer compared to the inverted limb. This overall geometry indicates that F2 folding is W-vergent on the scale of the entire study area.

After having introduced the Versoye synform and associated parasitic folds we now introduce the reader into the earlier deformation phases, i.e. pre-D1 and D1. The Moûtiers unit, forming part of the lower limb of the Versoye synform, consistently shows an upward younging direction (Fig. 5.5, D2 deformation phase and Figs. 5.6 and 5.7).

This indicates that the Moûtiers unit must have occupied a normal position also before F2 folding, i.e. during D1. On the other hand, the Pt. St. Bernard unit, forming part of the upper limb of the Versoye synform (overturned in respect to D2), consistently exhibit a downward younging direction at present. This implies that, after retrodeformation of the Pt. St. Bernard unit into its pre-D2 position, the Pt. St. Bernard unit must have displayed upward younging direction, comparable to that of the Moûtiers unit. Therefore, from a geometrical point of view, the Moûtiers and Pt. St. Bernard units may be connected together around the hinge of the Versoye synform (see Fig. 5.5). This inference indicates that before the onset of the D2 deformation phase both these units occupied the same structural level, both being in a normal position before the onset of D2 folding. This is supported by the fact that both units show similarities in their stratigraphic records. Both are characterised by the presence of pre-rift sediments and by the lack of basaltic volcanic rocks, suggesting that both units belonged to the same paleogeographical domain, i.e. the external Valaisan. The internal Valaisan (Versoyen unit) is preserved in the core of the Versoye synform, and is characterised by the presence of basaltic volcanic rocks and intercalated sediments (Complexe Antéflysch) and by the lack of pre-rift sediments. Retro-deformation of the Versoyen unit to its original position before D2 clearly reveals that the Versoyen was already tectonically juxtaposed onto both the Moûtiers and Pt. St. Bernard units, respectively, during the D1 deformation phase and before the onset of D2 (Fig. 5.5, D1 deformation phase). Since the external and internal Valaisan represent two fundamentally different paleogeographical domains, it seems likely that the boundary between external and internal Valaisan is of tectonic nature that can be related to older (pre-D2) deformation phases, i.e. late-D1 thrusting of the internal Valaisan onto the external Valaisan. These inferences will be supported by further evidences provided in the sections 5.3 and 5.4 . Hence, D2 and D4 refold a nappe stack already formed during the D1 deformation phase.

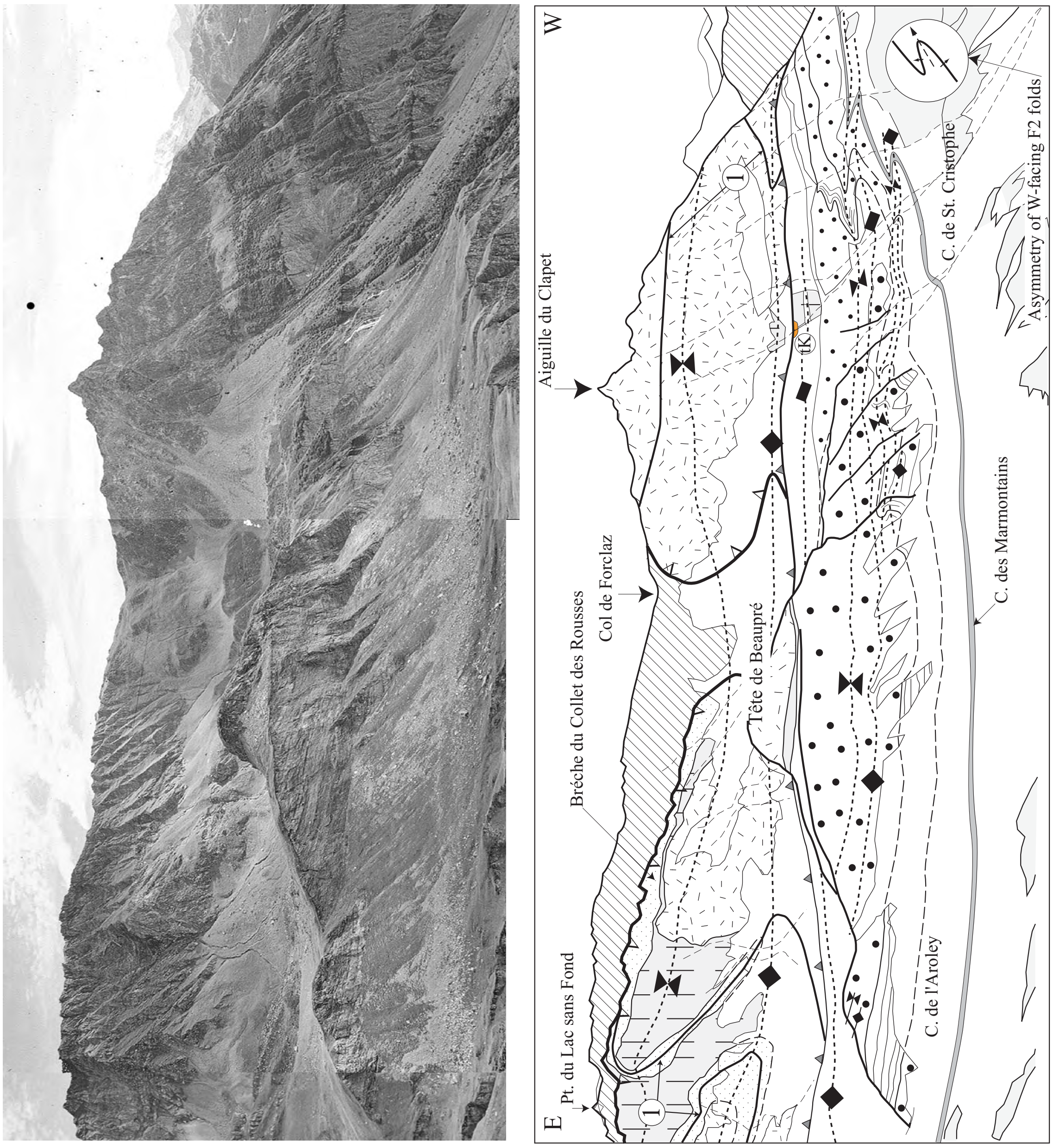


Fig. 5.9 Photograph and line drawing of the Aig. du Clapet-Tête de Beaupr area showing the asymmetry of W-facing F2 folds (looking south, S shape, see inset) that is compatible with the upper limb of the D2 Versoye synform (compare this situation with that of Fig. 5.9) . Note that the contact between the Pt. St. Bernard and the Versoyen units has been folded during D2 and post-date D1 large scale isoclinal folds with the Br che du Collet des Rousses in the core. These observations allow us to define this contact as late-D1. At the base of the Aig. du Clapet the exotic position of Upper Triassic evaporites (tK) marks the late-D2 tectonic contact within the Versoyen unit, responsable for the justapposition of two F2 anticlines. Well bedded horizon within the C. de l'Aroley formation is used in the line drawing as marker for the structures. This area is included in the geological map of Fig. 5.13 (same signatures).

5.3: Major D1 features inferred from their relation to the superimposed folding: the analysis of F2 structural facing directions.

It is often difficult to reconstruct the early stages of the deformation history because frequently they are completely overprinted by later deformation events. In the case of the Valaisan units this is fortunately not always the case. The analysis of the D2 geometrical features provided in the previous sections, points to the presence of pre-D2 deformation phases, characterised by the tectonic juxtaposition of the internal Valaisan, i.e. the Versoyen unit, onto the external Valaisan, i.e. the Pt. St. Bernard and Moûtiers units respectively. The analysis of F2 facing directions offers an additional possibility to receive more information concerning the earlier stages of the deformation history of the Valaisan units.

Before embarking on the discussion of the formation of the D1 nappe stack of the Valaisan units, however, it is necessary to introduce the concept of structural facing. When the stratigraphic record allows to infer younging directions, the application of the concept of facing is a valuable tool in analysing geometry and kinematic evolution of areas which underwent superimposed deformation. This method is particularly suitable in regions, where F1 hinge zones are hard to be detected, as is the case for the study area. Structural facing denotes a vector pointing towards the stratigraphically younger sediments, constructed normal to the fold axis and lying within the axial plane (fold facing). The same kind of information can also be obtained by constructing a vector pointing towards the younger sediments normal to cleavage-bedding intersection and lying in the cleavage plane (cleavage facing, generally subparallel to fold facing). In the present study way up criteria are provided by the well established stratigraphic succession within the metasedimentary sequence. Most of the structural facing directions were constructed within the post-rift sediments of the external and internal Valaisan. Moreover, due to the extremely isoclinal nature of the F1 folds, actual hinges can only rarely be detected, cleavage and bedding are often parallel, and only in a few cases it was possible to directly infer the F1 facing direction. For this reason the analysis of facing directions within the Valaisan units focuses on F2 structures which offer a good chance to investigate the effects of previous deformation phases. Fig. 5.10 illustrates the facing concept in the case of coaxial second phase re-folding of first phase folds and illustrates the relationships between the facing concept and the younging direction. Please note that this generalised illustration is based on the observations made within the studied area and applied to them.

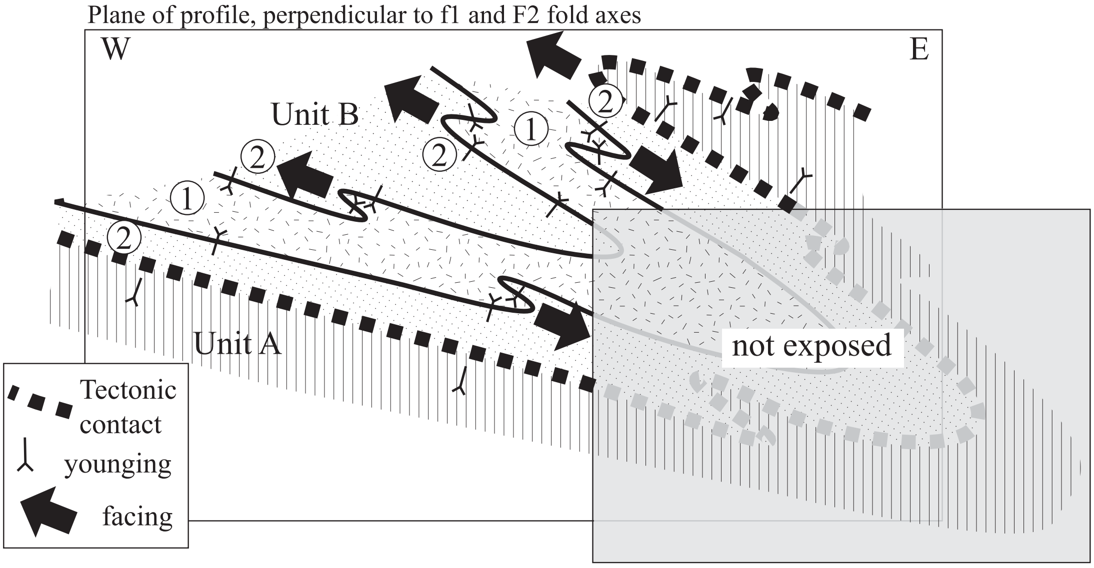
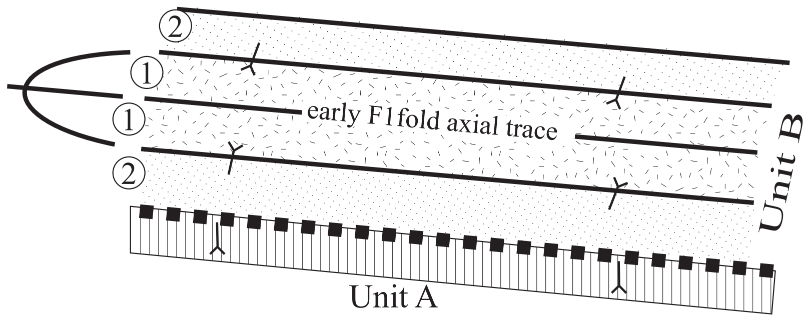


Fig. 5.10 a) In this special case, based on the outcrop pattern, both structural facing and fold asymmetry need to be defined in order to unravel the complex structural history. The order of deposition 1,2 is known. Units A and B which show tectonic relationships since they exhibit opposite younging directions. The parasitic folds show changes in facing and asymmetry (geometry). Facing and asymmetry are indipendent concepts.

Fig. 5.10 b) Changes in facing indicated by the reversal of younging directions, locate early F1 fold axial traces. Younging reversal can be identified by the repetition of stratigraphy. In this special case the anlysis of facing highlights the presence of a D1 fold within the unit A, the inverted limb of unit A being thrusted onto the rigth-way-up unit B.



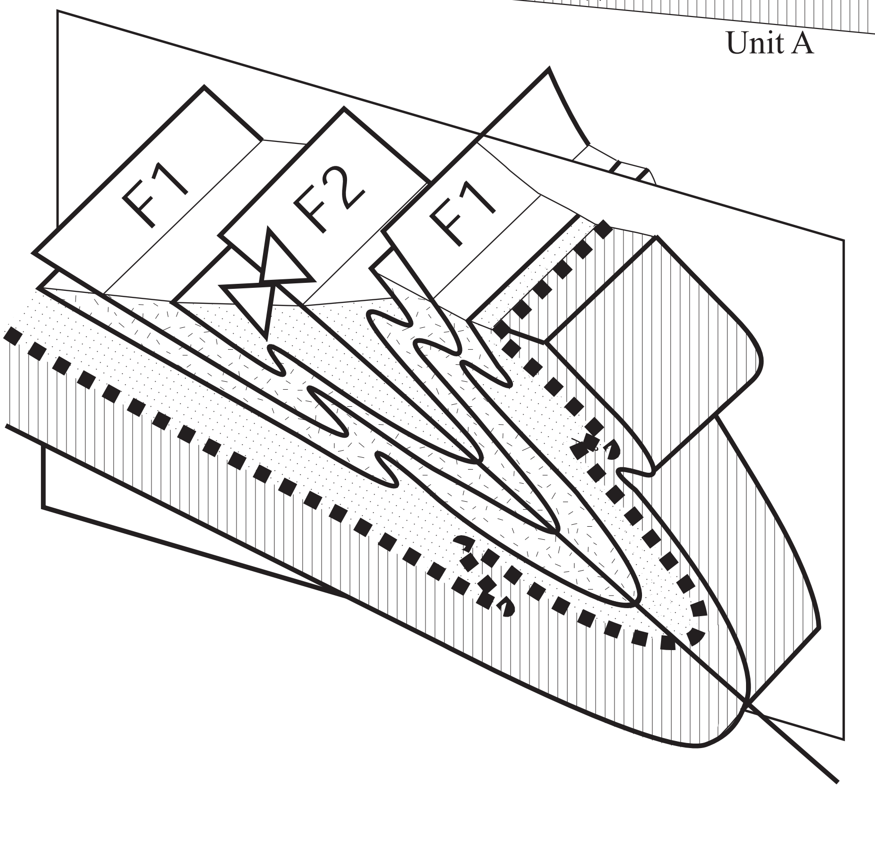


Fig.5.10 c) In this figure the combined information on facing and asymmetry shows the complex structural history. Changes in the asymmetry of minor F2 folds is used to locate F2 major axial trace. Changes in the facing (Fig. 10 b) locate early F1 fold axial traces. The entire deformation history can be described in terms of:
D1 folding and thrusting of Unit A onto unit B.
D2 refolding of the D1 nappe stack forms a major D2 synform.

The arrows displayed in Fig. 5.11 represent the facing azimuths in respect to F2 folds within that part of the Valaisan units which is largely unaffected by D4. The analysis of F2 facing directions reveals that there are areas, where E to SE facing is dominant while in other areas W to SW facing predominates. Since D2 folding alone cannot cause a change in structural facing this change corresponds to regional scale D1 features. As already described in the previous section, the younging direction within the Moûtiers and Pt. St. Bernard units remains constantly upward oriented after retrodeformation of the D2 deformation phase. This is compatible with Fig. 5.11 which show that the F2 facing direction within the Pt. St. Bernard and Moûtiers units consistently show identical westward directed structural facing on both limbs of the D2 Versoye synform. This analysis confirms that the external Valaisan (Pt. St. Bernard and Moûtiers units), was oriented right-way up before the onset of D2. Inspection of F2 facing direction (Fig. 5.11) reveals the disposition of inverted or normal strata related to the D1 deformation phase (see Fig. 5.10). The first crucial information is that F2 w-facing areas correspond to a normal stratigraphic sequence before D2, while areas showing E-facing direction indicate an inverted stratigraphic sequence before D2. This scheme becomes of chief importance when looking at the F2 structural facing in the internal Valaisan, where different facing azimuths are recognizable on both limbs of the Versoye synform (Fig. 5.11). On both limbs of the Versoye synform the Versoyen unit displays, in different proportions, alternatively W or E facing directions.

These areas of opposite facing are separated by D1 axial planes or thrusts with the only exception of a late-D2 contact within the Versoyen unit that locally overprints these boundaries (see Chapter 6). Hence, it can be concluded that, within the Versoyen unit, the F2 folds affect large sequences that were either normal or inverted before the onset of D2 deformation. This shows that it is necessary to also take into account the effect of regional scale D1 folding when reconstructing the original nappe stack within the Valaisan units.

F2 facing direction (Fig. 5.11) reveal the following geometrical disposition of the D1 nappe stack within the Valaisan units. Before D2 the Moûtiers unit and the Pt. St. Bernard unit were in a normal position, leading to W-ward directed F2 facing. The same Moûtiers unit was overthrusted by the inverted limb of a F1 mega fold which formed within the Versoyen unit revealed by E-ward directed F2 facing in the Aig. de Praina area (Fig. 5.11). The Pt. St. Bernard unit also occupied, a normal position and was also overthrusted by the Versoyen unit during D1.

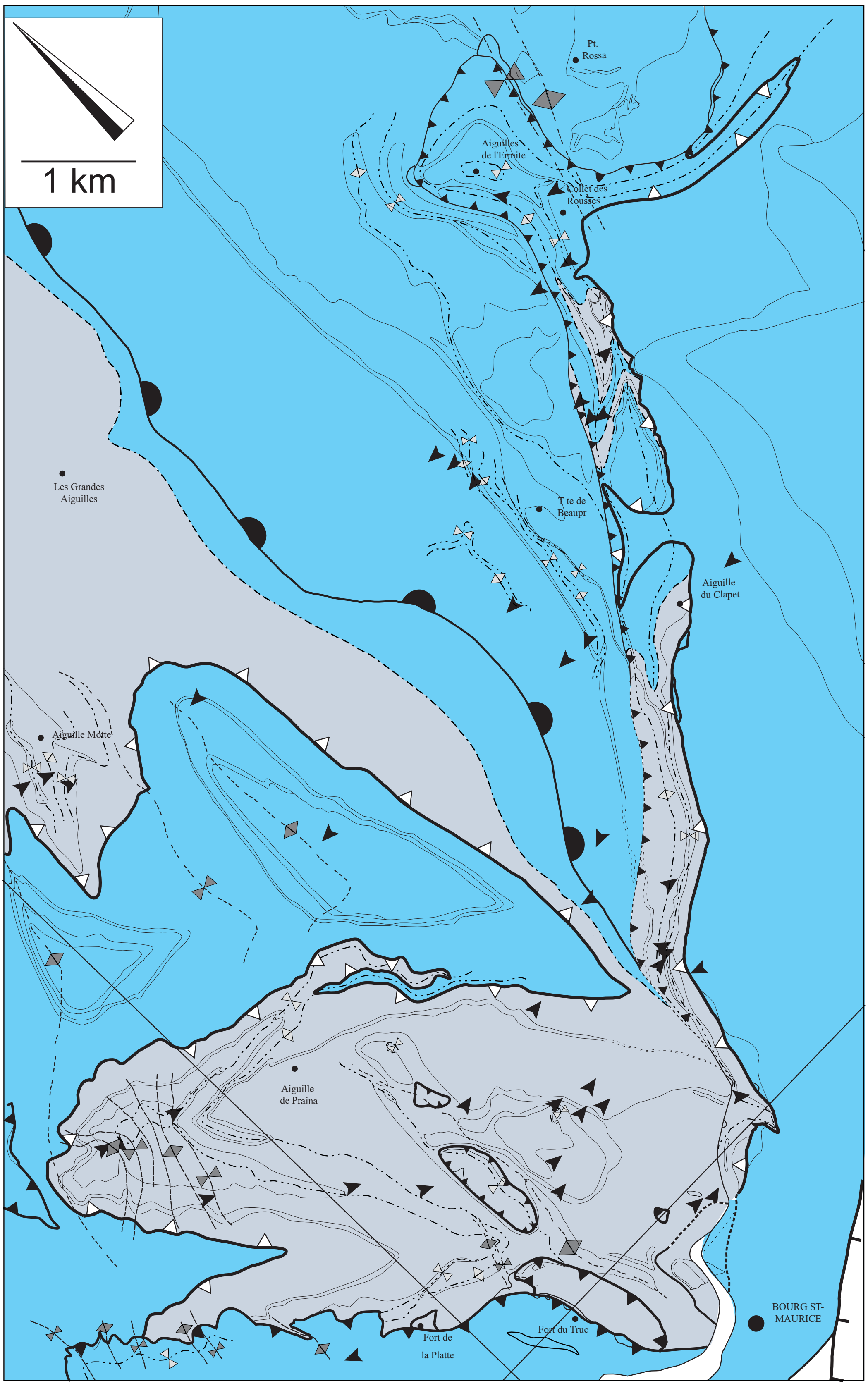


Fig. 5.11 Azimuth of F2 structural facing in the area between B. St. Maurice and the Collet des Rousses pass.
Gray: areas with E-facing. Blue: areas with W-facing. Limit between gray and blue:
F1 axial trace, Late-D1 thrust, Late-D2 thrust.
W-facing areas are part of the normal F1 limb while areas with E-facing are part of the inverted F1 limb

In the Versoyen unit after D1, both F2 W facing and F2 E facing folds have been observed showing the existence of an F1 mega fold within this unit. A pre-existing D1 isoclinal fold was subsequently folded around the D2 Versoye. This led to the present disposition in map view shown in Fig. 5.11. The so far undetected main D1 structural features are the following: a regional scale D1 fold within the internal Valaisan, referred as the D1 mega fold was thrusted onto the external Valaisan (Pt. St. Bernard and Moûtiers units). These D1 features cannot be treated separately from the effect of D2 because field evidence indicates clear overprinting relationships which only emerged after combining observations made in crucial regions together at the scale of the entire study area. Overprinting relationships are the subject of the detailed structural geology presented in the following sections. However, these overprinting relationships can be already inferred from inspection of Fig. 5.11. The D1 tectonic contact between the Versoyen and the Pt. St. Bernard units cuts portions of the inverted and normal sequences of the D1 megafold, including associated D1 parasitic folds in the Versoyen unit (Fig. 5.5, D1 deformation). In Fig. 5.11 this is displayed by the occurrence of areas exhibiting different F2 facing directions indicating the occurrence of a series of F1 folds within the Versoyen unit, close to the tectonic contact with the Pt. St. Bernard unit. This observation clearly suggests that F1 isoclinal folding was followed by thrusting at the tectonic contact between the Versoyen and the Pt. St. Bernard units. This tectonic contact is referred as a late-D1 contact.

5.4: The early stage of the formation of the nappe stack within the Valaisan units: regional scale pre-D1 and D1 folding and thrusting

General remarks

After the discussion of the D4 and D2 geometrical (section 5.2) and structural features (section 5.3) evidence for the pre-D1 and D1 deformation phases will be presented. Selected regions of the study area help to introduce the reader into the details of the geometry of the first deformation phase. For these regions an attempt was made to infer relative age relationships from fold interference patterns. Furthermore indications and evidences pointing to a pre-D1 deformation stage will be presented. Finally an attempt was made to interpret such D1 and pre-D1 features.

The contact between external and internal Valaisan, (late-D1) folded around the D2 Versoye synform is described within two particular regions of the study area. The first and more internal one is represented by the Aig. du Clapet-Tête du Beaupré area (Fig. 5.1) where it is clearly structurally above the Versoye synform. The second and more external one is found along the ridge between the Pt. de la Terrasse and the Aig. du Praina and in the Aig. Motte area (Figs. 5.1, 5.3 and 5.4) where the late-D1 contact is below the axial plane of the Versoye synform. Within both these areas it is possible to reconstruct a chronological order of the structural features in order to take into account the field relationships between late-D1 thrusting and D1 folding, as well as such features which have been active before D1 (pre-D1).

Pre-D1 features are generally related the detachment of the Valaisan units from their former substrate, oceanic for the internal Valaisan and continental for the external Valaisan (see Chapter 2).

5.5: The contact between the Versoyen and Pt. St. Bernard units in the Aig. du Clapet-Tête du Beaupré area.

Structural analysis along the late-D1 contact (Figs 5.9 and 5.12) shows a clear relationship between D1 folding and thrusting. This relationship is documented in the geological map of Fig. 5.13 and outlined by a set of geological profiles (Fig. 5.14) located between the Aig. du Clapet and the Collet des Rousses pass area. The contact between the Versoyen and Pt. St. Bernard units itself appears to be unaffected by D1 folding (Figs. 5.2 and 5.14). On the other hand, between the Aig. du Clapet and the Collet des Rousses pass and within the Versoyen unit local D1 isoclinal folds are truncated by this tectonic contact, a clear evidence for the relative chronology of D1 folding and thrusting. According to the geological map of Fig. 5.13, syn-rift sediments (i.e. the Brèches du Collet des Rousses, marking the detachment horizon of the Versoyen unit) form the core of F1 isoclinal folds (Fig. 5.9 and 5.14). F1 folding is inferred from changes in the younging direction overprinted by F2 isoclinal folds. In Fig. 5.11, between the Aig. du Clapet and the Collet des Rousses pass small areas showing E-facing direction are preserved within a generally W-facing area. This documents the occurrence of F1 large scale isoclinal folds. In some cases the F1 hinge can hardly be detected. F1 axial planes (S1) are sub-parallel to the lithological contacts (S0) and both (S1 and S0) are clearly cut by the tectonic contact with the Pt. St. Bernard unit. The fact that F1 isoclinal folds also affect the detachment horizon implies that the development of these D1 folds postdates the development of the detachment horizon which we therefore consider as a pre-D1 structure.

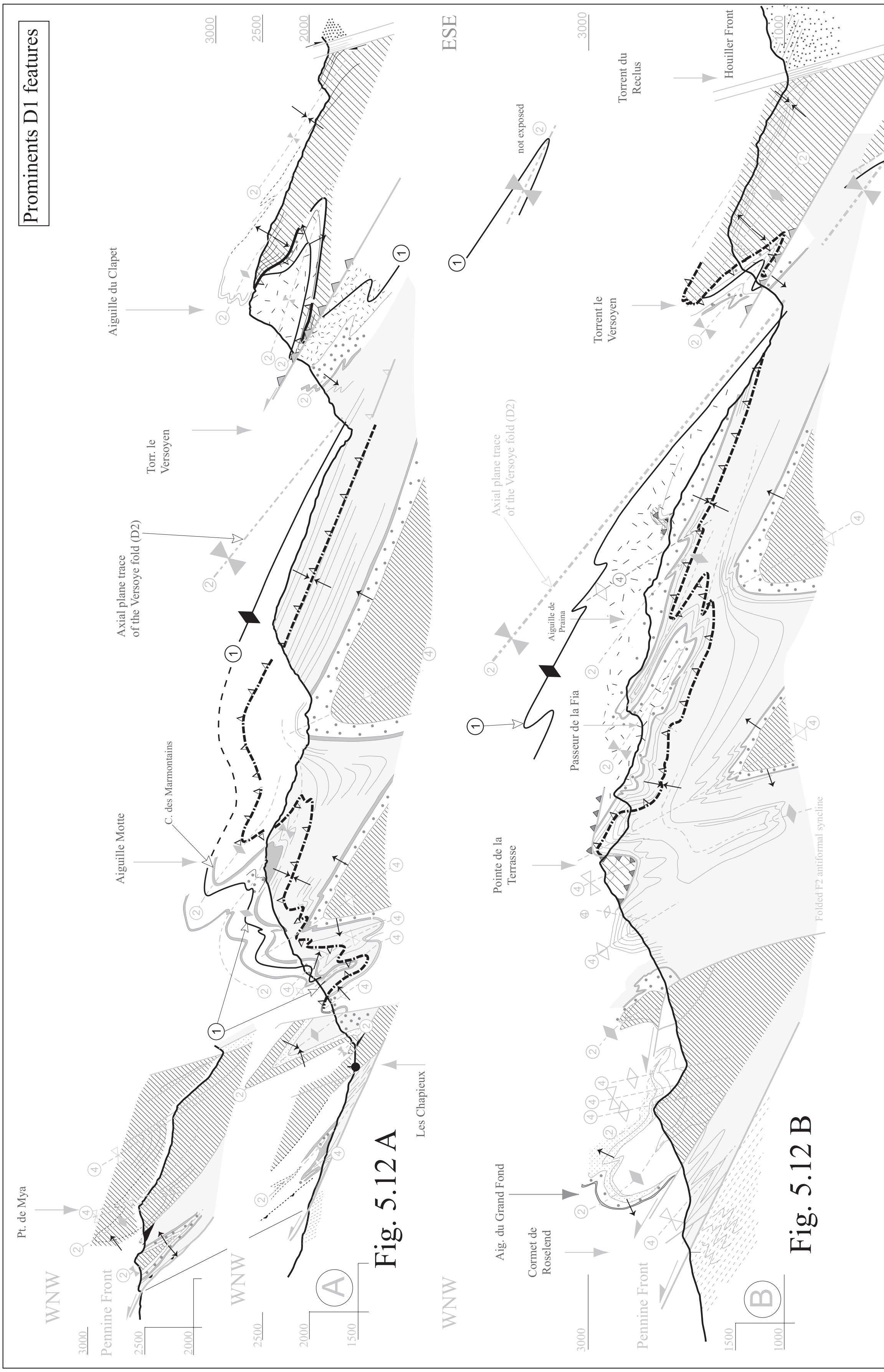


Fig. 5.12 a) and b). The prominent D1 features are enhanced in the profile of Fig. 5.3 (Fig.5.12a) and of Fig. 5.4 (Fig.5.12b).

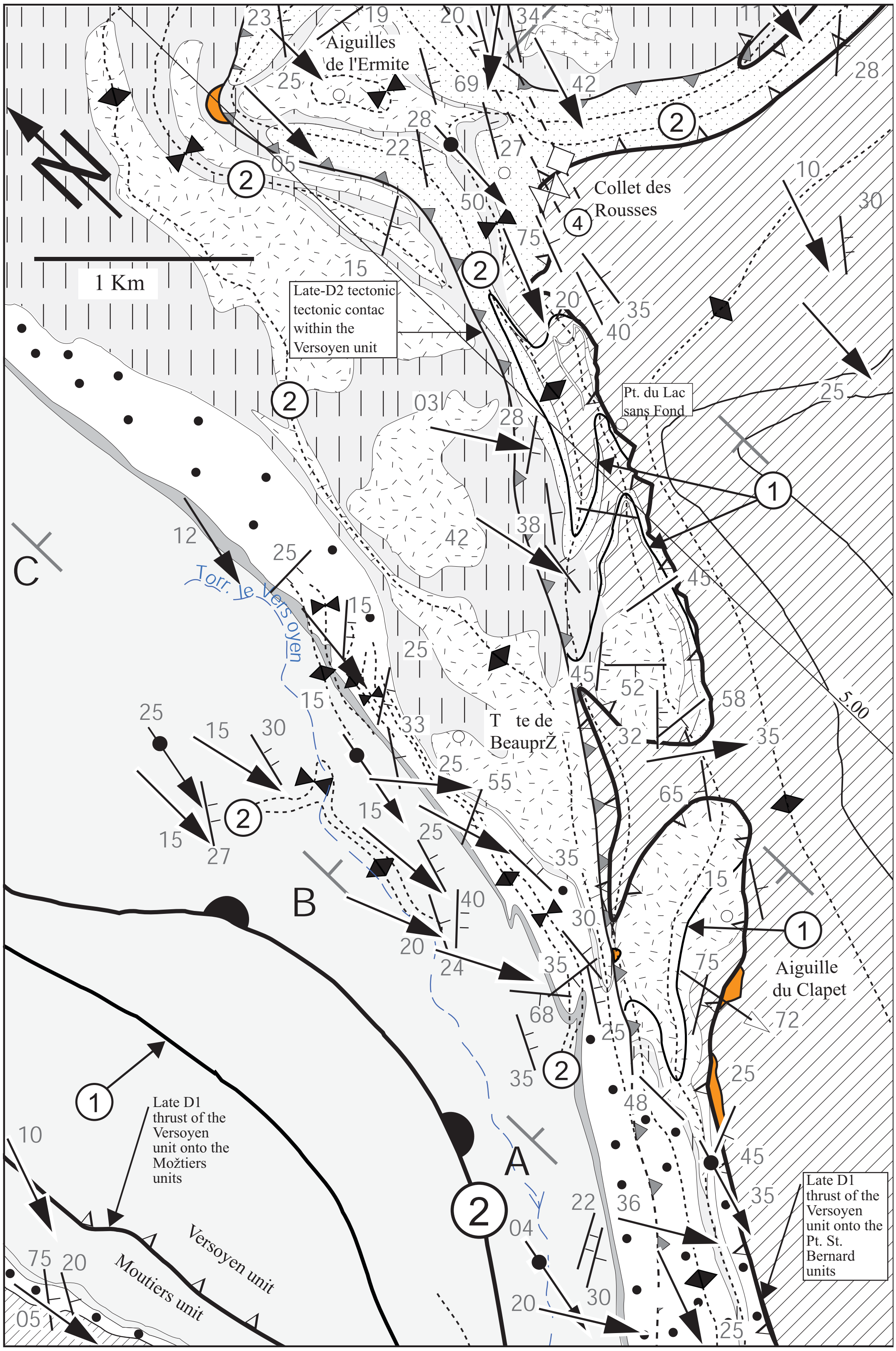
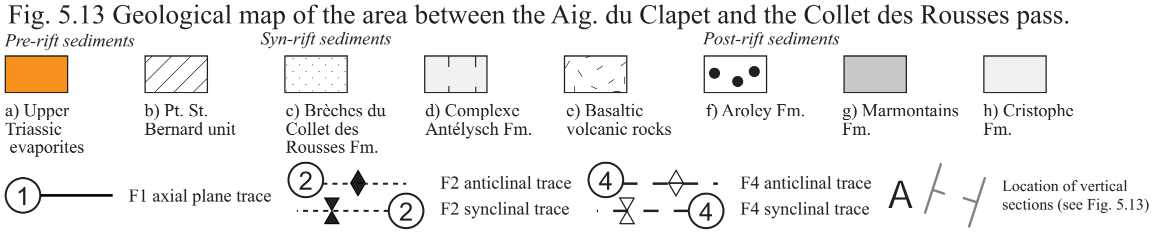


Fig. 5.13 Geological map of the area between the Aig. du Clapet and the Collet des Rousses pass.



Hence, the chronology of early tectonic events is as follows: the detached Versoyen unit is subsequently folded by D1 isoclinal folds (early D1). Immediately after D1 folding, continued late-D1 thrusting of the internal Valaisan above the external Valaisan led to the formation of the present day tectonic contact between the Versoyen and the Pt. St. Bernard units.

Identical relative age relationships between D1 thrusting and folding can be observed (Fig.5.1) along the entire distance from Bourg St Maurice (France) to the Tête du Chargeur (Italy), where the lithological contacts within the Versoyen unit are truncated by the late-D1 tectonic contact. Between Bourg St. Maurice and the Collet des Rousses the oldest preserved lithologies of the Versoyen unit appear to become older towards NE, ranging from post-rift sediments in the SW near Bourg St. Maurice to the Complexe Antéflysch formation near Aig. du Clapet and finally to the Brèches du Collet des Rousses in the NE. Still to the NE, i.e. from the Italian-French border up to the Tête du Chargeur (Italy), the oldest preserved lithologies of the Versoyen unit appear to become younger again, (Complexe Antéflysch to the post-rift sediments). This outcrop pattern within the Versoyen unit, followed from Bourg St. Maurice to the Tête du Chargeur can also be investigated in respect to F2-facing directions (Fig.5.11). In the SW F2 folds in the Versoyen unit, found close to the tectonic contact with the Pt. St. Bernard unit exhibit E to SE facing synformal anticlines and antiformal synclines, respectively, (Fig. 5.11 and 5.14). This implies that these F2 folds affected an already inverted sequence by folding during D1. However, further towards the NE, i.e. towards the Collet des Rousses in the Tête de Beaupré area, F2 folds show a F2-facing direction towards W to SW (Fig. 5.11 and 5.14). Implying a normal position before D2. The disposition of inverted and normal sequences, after the removal of the D2 deformation phase, imaged in their present-day position by the distribution of different F2 facing directions, clearly indicates the presence of two limbs of a D1 mega-fold within the Versoyen unit (Internal Valaisan). The limits between areas of E-facing directions and W-facing directions corresponds to the trace of the axial plane of the F1 mega-fold. Fig. 5.11 also shows that a tectonic complication due to late-D2 thrusting within the Versoyen unit complicates the map pattern of the axial planes related to the D1 mega-fold. On the other hand, within the Pt. St. Bernard unit, the F2-facing directions are consistently W-directed. The sediments of this unit are generally overturned in their present-day position because they rest in the upper limb of the D2 Versoye synform. Near the tectonic contact with the Versoyen, they consistently show a downward younging direction. This suggests that, after the removal of the D2 Versoye synform, the Pt. St. Bernard unit was in an upright position. This clearly stresses the fact that the tectonic contact between the Versoyen and Pt. St. Bernard units (late D1) post-dates large scale F1 folds as well as the D1 mega-fold within the Versoyen unit.

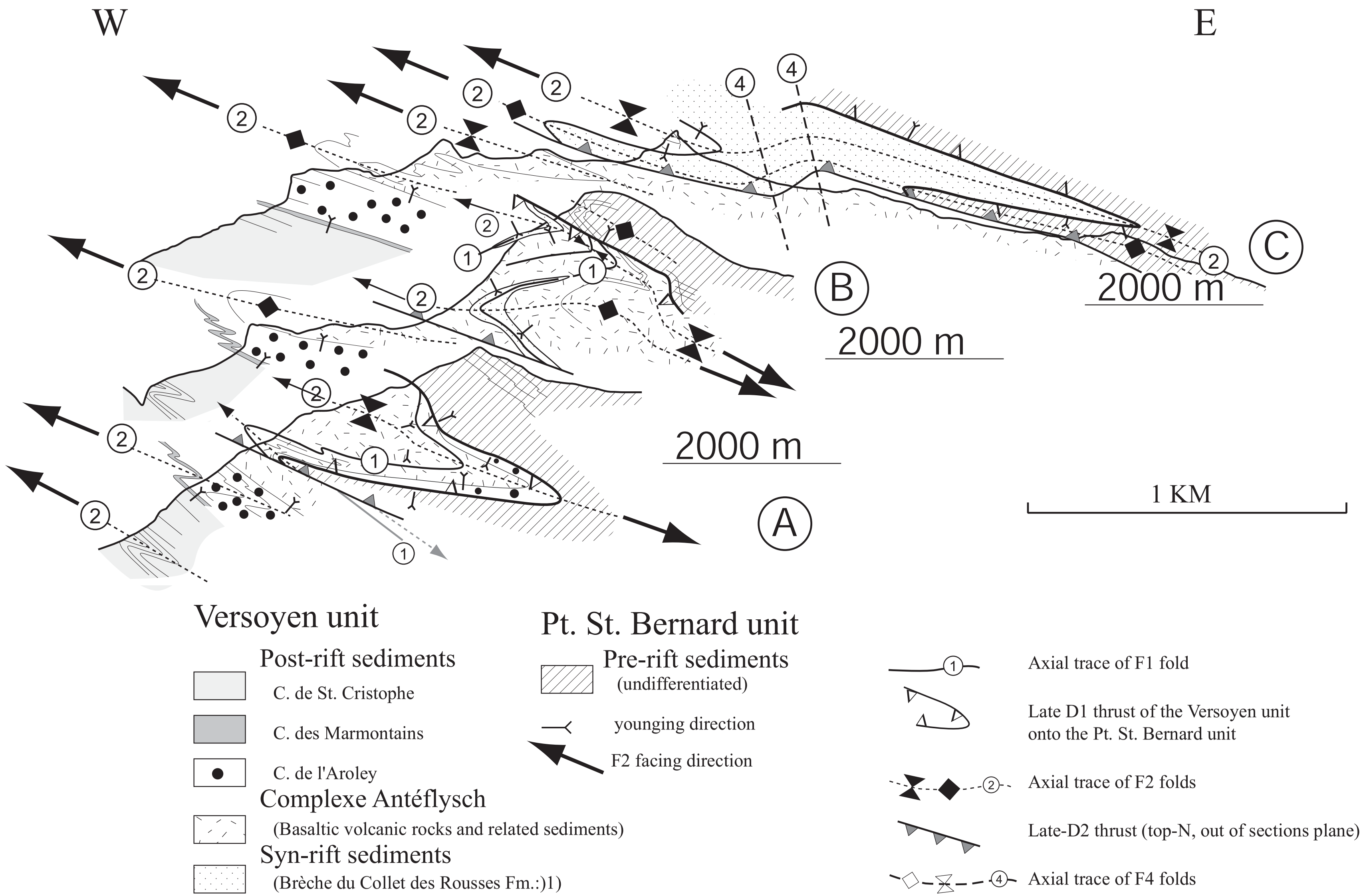


Fig.5.14 E-W cross sections through the Aig. du Clapet area. The traces of these sections are given in Fig. 5.13. These sections inllustrate, within the Versoyen unit, large scale F2-F1 overprinting in the upper limb of the D2 Versoye synform (see Fig. 5.12) . Note the in profiles A and B, lithological contacts, as well as the F1 axial trace, are truncated by the late D1 thrust of the Versoyen unit onto the Pt. St. Bernard unit.

5.6: The contact between the Versoyen and Moûtiers units in the Valleé des Glaciers-Aig. Motte area

Decametric to kilometric scale amplitude F2 isoclinal folds, have been described in chapter 5.2 and depicted in Fig. 5.8 from the lower Valleé des Glaciers area. The asymmetry of these F2 folds is compatible with their position in the lower limb of the D2 Versoye synform. Due to the orientation of the valley, approximately perpendicular to the azimuth of F2 and F4 fold axes, and due to a high relief of more than 1000 m this valley offers a most suitable, natural cross section through a large part of the study area. From the valley floor up to the ridge between the Pt. de la Terrasse and the Aig. de Praina (Fig. 5.8) one crosses the pre-rift and post rift sediments of the Moûtiers unit, both showing an upward younging direction (normal stratigraphic sequence, external Valaisan). The top half of this mountain ridge has the same post-rift sediments in a downward younging direction. (see Chapter 2, Figs. 5.3 and 5.8). At the crest, finally, the stratigraphic contact between the post-rift sediments and the basaltic volcanic rocks and related sediments of the Complexe Antéflysch of the Versoyen unit is exposed in an inverted stratigraphic sequence. The normal and the inverted sequences are intensely folded by parasitic D2 isoclinal folds, the most spectacular of which is found at the Passeur de la Fia (Fig. 5.8). F2 synforms in the inverted sequence have the basaltic volcanic rocks and related sediments of the Complexe Antéflysch in their core, thus forming east-facing synformal anticlines. A good example can be seen in the Passeur de la Fia region (Fig. 5.8). On the other hand in the upright sequence the same F2 display the post rift sediments in the core of west-facing synformal synclines. From this it is concluded that the Versoyen unit between Pt. de la Terrasse and the Aig. de Praina, has been overturned during D1 since this previously inverted sequence is folded by the same F2 folds. This inverted sequence is positioned in the inverted limb of the F1 mega fold found within the Versoyen unit (see Fig. 5.5, D1 deformation phase and Fig. 5.11).

The fact that along the same limb of the D2 Versoye synform two distinct paleogeographical domains can be recognized (i.e. the external Valaisan at the base and the internal Valaisan above) clearly indicates that they have been tectonically juxtaposed, i.e. during D1 (“Late-D1 thrust of the Versoyen unit onto the Moûtiers unit” in Fig. 5.5, D1 deformation phase). There is no evidence that lithological boundaries are truncated by this late-D1 tectonic contact. Thus, the tectonic boundary between the inverted part of the Versoyen unit and the right-way up Moûtiers unit formed parallel to the S1 foliation.

Due to the monotonous and homogeneous character of the C. de St. Cristophe formation and due to the intense post-D1 deformation, this tectonic contact cannot be documented by direct field observations. It is proposed that this contact represent a diffuse shear zone located within the youngest sediments of the C. de St. Cristophe formation. The detailed structural analysis of the subsequent D2 and D4 structures allows for a reconstruction of the D1 geometry and points out that the frontal part of the Versoyen unit has been affected by a large scale F1 isoclinal fold still preserved in the Aig. Motte area. As schematically depicted in Fig. 5.5 (D1) this axial trace of the F1 isoclinal fold is situated structurally above the inverted part of the Versoyen unit. Fig. 5.15 shows a photograph and line drawing of the detailed structures in this area while a geological map and a detailed geological cross section of the same area are found in Figs. 5.16 and 5.17 (for location see Fig. 5.2). Due to the extreme isoclinal nature of the F1 folds, fold hinges are not clearly visible, especially in this area where post-D1 deformation, i.e. D2 and particularly D4, was very intense. However, the presence of D1 structures is deduced from the fold interference patterns between F1 and F2. The asymmetry of F2 folds in this area (central part of profiles in Fig. 5.3 and Fig. 5.4) indicates a position in the lower limb of the Versoye synform, where the Versoyen unit always forms the core of F2 E–SE facing synformal anticlines. After retro-deformation of D4 (Figs. 5.18 and 19) it becomes evident that this F1 isoclinal fold is a W-facing antiform with the oldest preserved lithologies, i.e. the C. de l'Aroley, in its core (Fig. 5.17).

The observations in the Aig. Motte area allow for the deduction of details about the D1 configuration of the this part of the Versoyen unit. Most of the Versoyen unit exposed in the Valleé des Glaciers area is located in the lower limb of the F2 Versoye synform. With respect to D1 it mostly represents the inverted limb of the F1 mega-fold. The normal limb of this F1 mega-fold is no more preserved due to erosion. Only in the Aig. Motte area, D4 folding allowed for the preservation of this D1 structure.

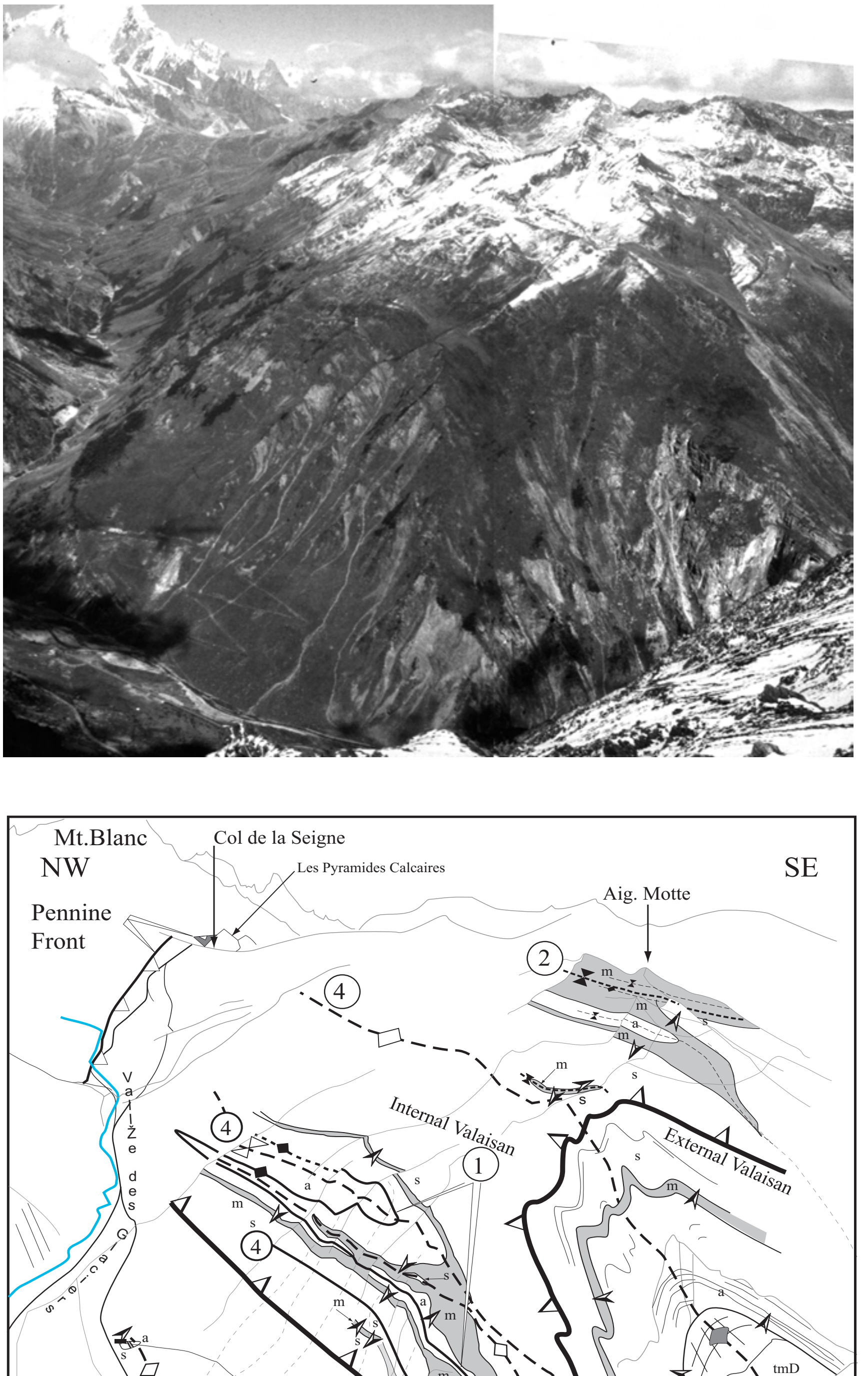


Fig. 5.15 Photograph and line drawing of the SW side of the Aig. Motta showing the large scale F1 fold and the contact between the external Valaisan (Moûtiers unit) and internal Valaisan (Versoyen unit). Towards SE the post rift sediments forming the Aig. Motte peak are in contact with the Complexe Antéflysch (see fig. 3.9). 1 , 2 , 4 are axial plane traces of D1, D2 and D4 deformation phases, respectively. Photograph taken from Pt. 2679 (941'325/2'084'150)

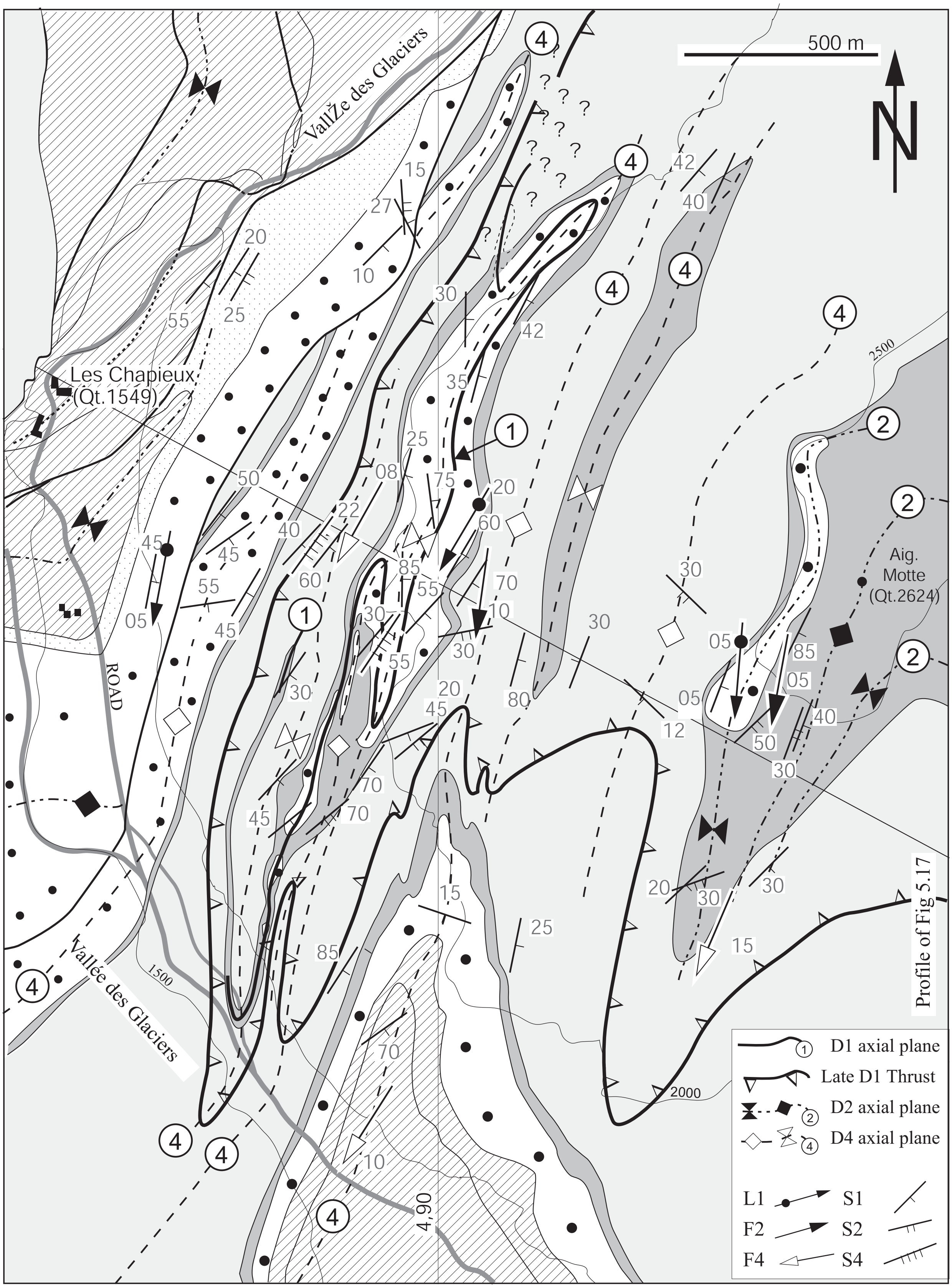


Fig. 5.16 Geological map of the area between Les Chapieux and Aig. Motte, for location see Fig. 5.2.

Post-rift sediments (of the Moûtiers (external Valaisan) and Versoyen units (Internal Valaisan)) — *Syn-rift-sediments* — *Pre-rift sediments* (of the Moûtiers unit)

a) St. Cristophe Fm.
b) Marmontains Fm.
c) Aroley Fm.
d) undifferentiated
e) undifferentiated

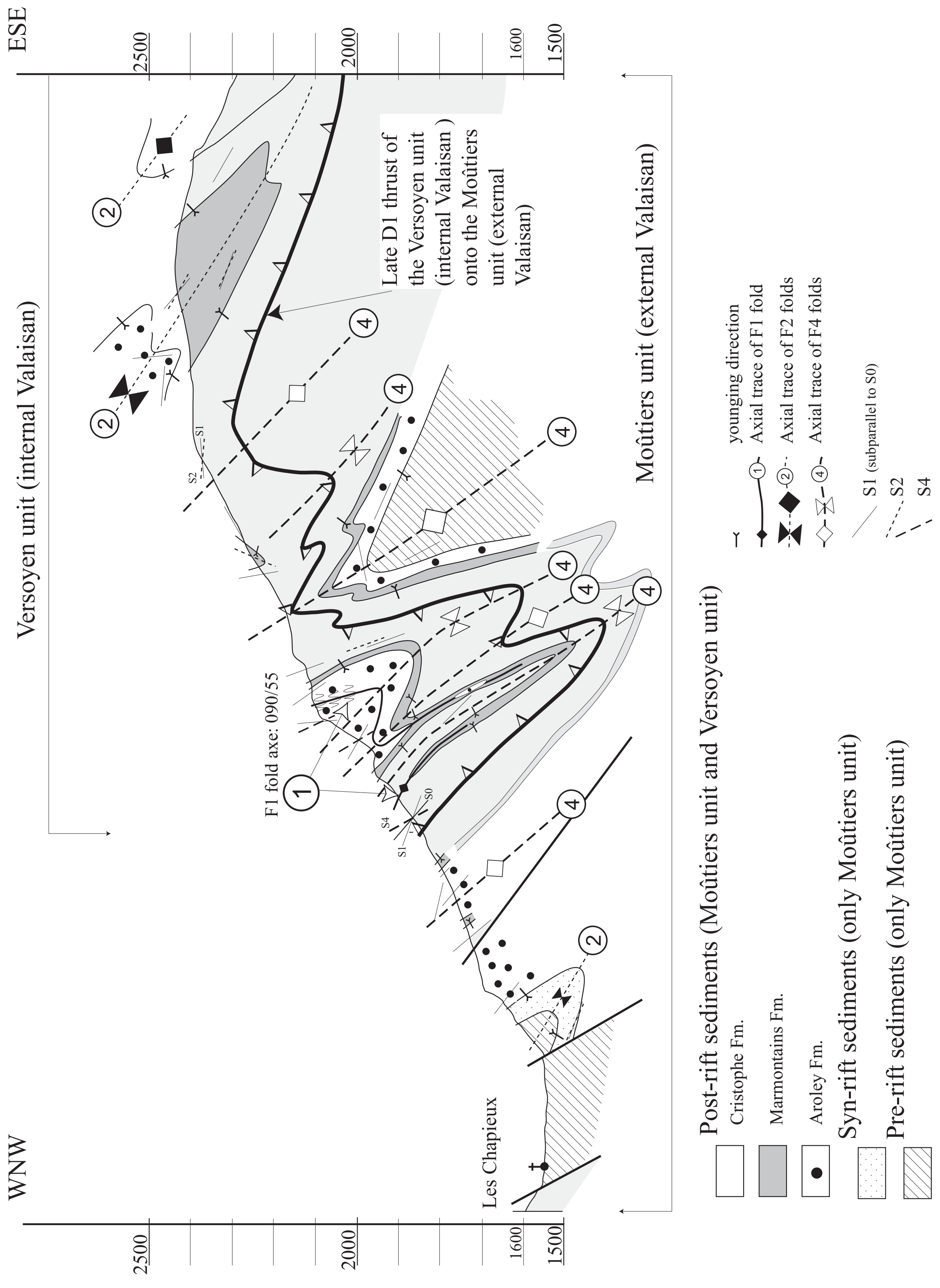


Fig. 5.17. Enlargement of the WNW-ESE section through the area between Les Chapieux and the Aig. Motte given in Fig. 5.4 The location of this section is given in Fig. 5.16)

5.7: Conclusions on the regional geometry of the Valaisan units

Until now the nappe stack of the Valaisan units, even if regarded as the result of the superposition of several deformation phases (D1 to D4 of Lancelot 1979) was poorly understood in terms of the early emplacement history of the different tectonic units. However, structural evidences, reported in this study, point out a complex situation so far undetected. It includes, during the early stages of the formation of the nappe pile of the Valaisan units, a subdivision into geometrical and structural features that belong to the pre-D1 and D1 deformation phases. Ongoing deformation, on a regional scale, is related to the D2 and D4 deformation phases. The pre-D1 phase, related to initial detachment of the Versoyen unit from its oceanic substratum, is preserved as large scale F1 isoclinal folds with the Brèches du Collet des Rousses in the cores of the. The D1 deformation phase, related to fold nappe formation and thrusting of the internal Valaisan, onto the external Valaisan is preserved as a F1 W-facing recumbent mega anticline within the Versoyen unit with a core formed by the basaltic volcanic rocks and related sediments (also including the syn-rift Brèches du Collet des Rousses). The thrust of the internal Valaisan post dates the F1 mega fold and is hence referred to as late-D1. The analysis of the D2 geometries and distribution of the F2 facing directions is of chief importance for the understanding of the nappe stack of the Valaisan units.

The geometry of F2 folds indicates the presence of the regional scale Versoye synform, responsible for refolding of the D1 fold and thrust nappe pile. The D1 fold and thrust nappe pile is preserved in its original position in the lower limb of the D2 Versoye synform, while it is overturned in the upper limb of the Versoye synform. The analysis of the F2 facing direction reveals the D1 disposition in terms of inverted and normal sequences.. The D4 deformation phase did not significantly alter the already formed nappe stack of the Valaisan units built during D1 and D2. However, where the D4 is more intense, i.e. towards the Pennine Front the D1 and D2 nappe stack was severely overprinted by F4 folds which display an isoclinal geometry, together with late D4 thrusts, interpreted to be related to the Pennine Front, in terms of a late-D4 feature.

The prominent D2 structural features are summarized in Fig. 5.18 (see also Fig. 5.5, D2) and essentially consist in the formation of the Versoye fold, a regional scale E closing D2 synform. According to the analysis of the F2 asymmetries the trace of the axial plane of the Versoye fold has been mapped out (Fig. 5.1) in order to make it possible to correlate D1 structures in the upper limb with D1 structures in the lower limb.

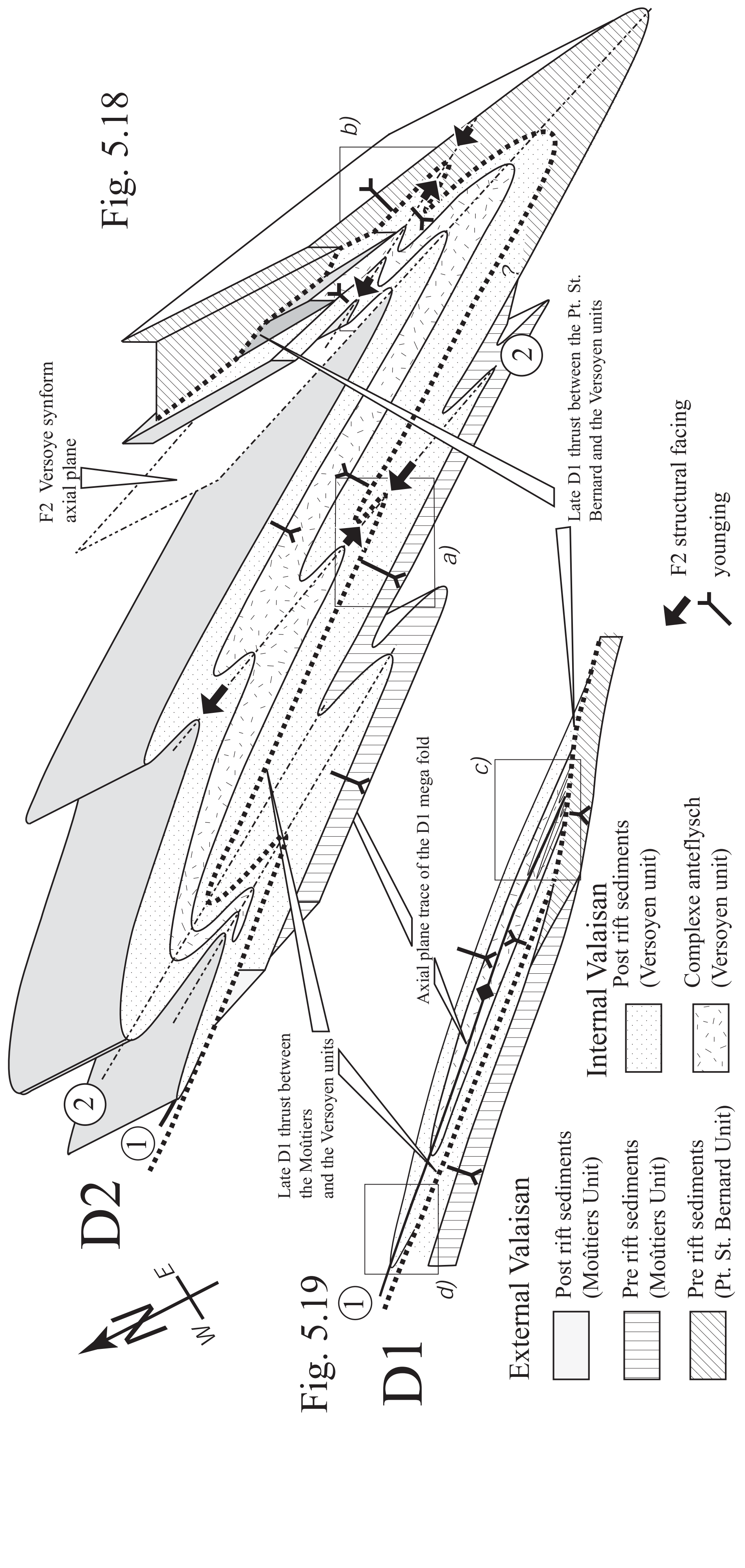


Fig. 5.18
b)
F2 Versoye synform
axial plane
Late D1 thrust between the Pt. St.
Bernard and the Versoyen units
?
2
a)
F2 structural facing
younging
c)
Axial plane trace of the D1 mega fold
D2
2
1
Late D1 thrust between
the Moûtiers
and the Versoyen units
Internal Valaisan
Post rift sediments
(Versoyen unit)
Complexe anteflysch
(Versoyen unit)
N
E
W
Fig. 5.19
1
d)
D1
External Valaisan
Post rift sediments
(Moûtiers Unit)
Pre rift sediments
(Moûtiers Unit)
Pre rift sediments
(Pt. St. Bernard Unit)

Fig. 5.18 and 5.19. Schematic block diagram showing the main structural features of the D1 (Fig. 5.19) and of the D2 (Fig. 5.18) deformation phases in the Valaisan units. Insets refer to crucial geological situations observed in the field and described in the text, that allow us to define the nappe stack of the Valaisan units during D1 and D2 deformation phases. The inset a of Fig. 5.18 shows an identical situation to that depicted in Fig. 5.9 (see also Fig. 5.14) as usually observed in the Valleé des Glaciers (section 5.1.6). The inverted stratigraphic sequence in the Versoyen unit rests in tectonic contact above the normal sequence of the Moûtiers unit, as indicated by the opposite F2- facing direction. Inset b) shows a situation which is identical to that observed in the Tête de Beaupré-Aig. du Clapet area (section 5.1.5 and Figs. 5.11 to 5.14). In this area the occurrence along the same F2 axial plane of two opposite F2-facing directions (Fig. 5.13 and 14) locate earlier (D1) axial planes within the Versoyen unit. Compare the asymmetry of minor F2 folds shown in inset a) and b) which allow to locate the axial trace of the major D2 Versoye synform. Fig. 5.19 (D1) inset c) shows the geometry of the D1 nappe stack as deduced after removal of the D2 deformation phase. The younging direction within the Pt. St. Bernard unit is now restored to its original orientation before D2 and is identical with that of the Moûtiers unit. The D1 axial plane of the mega fold within the Versoyen unit is cut by the tectonic contact with the Pt. St. Bernard unit that is therefore considered as late-D1 (Fig. 5.9 to 11). Inset d) shows the observations made in the Aig. Motte area (Fig. 5.15 to 17, see section 5.1.6). In this region the occurrence of large scale D1 folds is interpreted to indicate the location of the hinge zone of the D1 mega fold within the frontal part of the Versoyen unit.

The most outstanding D2 feature, the Versoye synform, is characterised by a long lower limb and a short upper limb. The lower limb is made up of those parts of the Versoyen unit (internal Valaisan) which tectonically rest above the Moûtiers unit (external Valaisan). The upper limb, however, is made up by the Pt. St. Bernard unit (external Valaisan) that presently rests above the Versoyen unit (internal Valaisan). In this study, according to important differences in the stratigraphic record between external and internal Valaisan, i.e. continental and oceanic domains respectively, and as result of the D2 geometrical observations, we propose to connect the Moûtiers unit in the lower limb of the Versoye synform with the Pt. St. Bernard unit in the upper limb. The Versoyen unit presently forms the core of the Versoye synform.

This correlation is also supported by the analysis of the F2-facing directions. In the Moûtiers unit, as well as in the Pt. St. Bernard unit, F2-facing directions are consistently oriented towards W-SW and the F2 folds form antiformal anticlines and synformal synclines. This implies that after the removal of the D2 deformation phase the Moûtiers and the Pt. St. Bernard units have occupied the same position both exhibiting an upward younging direction, (i.e. normal sequence before D2). The same analysis of F2 facing directions, carried out in the Versoyen unit reveals that part of this unit, had already been overturned by D1 folding and thrusting prior to D2. The inverted F1 limb is mainly preserved in the lower limb of the Versoye synform where it is represented by the inverted part of the Versoyen unit which rests above the Moûtiers unit. This is supported by the constant orientation towards E-SE of the F2-facing direction within this part of the Versoyen unit, indicating refolding during D2 of an already inverted F1 limb. In the same lower limb of the Versoye synform, the normal F1 limb of the mega-fold is only locally preserved due to erosion. However, a large scale F1 isoclinal fold detected in the frontal part of the Versoyen unit, in the Aig. Motte area, can be interpreted as the hinge zone of the D1 mega-fold itself. In the upper limb of the Versoye synform, on the other hand, F2-facing directions within the Versoyen unit are not constantly oriented and they point out the presence of normal and inverted F1 limbs. The normal F1 limbs are presently preserved mainly as W-facing F2 antiformal anticlines, while the inverted F1 limbs produce E to SE-facing F2 antiformal synclines. The map trace of the limits between areas with different F2-facing direction represents the F1 axial plane (Fig. 5.11). However, a late-D2 thrust complicated the D1-D2 fold interference pattern because it brought different parts of the F1 limbs in direct contact without the occurrence of an F1 axial plane nowadays resting in the footwall of these thrusts.

Figs. 5.9, 5.13 and 5.14 clearly show that a late-D1 contact between the Pt. St. Bernard and the Versoyen units truncates lithological contacts and early D1 folds within the Versoyen unit in the Aig. du Clapet area. The structural analysis of these D1 folds revealed the occurrence of the detachment horizon at the base of the Versoyen unit, i.e. the Brèches du Collet des Rousses in the core of these F1 folds. This observation reveals details on the chronological order during the first stages of the deformation history of the Versoyen unit. The F1 folds affect the detachment horizon and therefore post-date the detachment of the Versoyen unit from its substratum during a pre-D1 deformation phase. After F1-folding the tectonic contact with the Pt. St. Bernard unit did cut these F1 folds during a late-D1-stage. These isoclinal F1 folds show the detachment horizon, i.e. the lower most preserved stratigraphic levels within the Versoyen unit in their cores, and can be interpreted as the remnants of the hinge zone of the F1 mega-fold. In this study the large scale F1 folds of the Aig. du Clapet area are proposed to be correlated with the large scale isoclinal folds found in the frontal part of the Versoyen unit in the Aig. Motte area. Finally the W-facing mega-fold within the Versoyen unit is truncated at its base by the late-D1-thrust which brought the Versoyen onto the Pt. St. Bernard unit. On-going thrusting of the internal Valaisan above the external Valaisan finally brought the Versoyen unit also above the Moûtiers unit. This D1 configuration has been successively refolded by the D2 Versoye synform, thus giving the present day configuration of the Valaisan units, as they can be seen in the tectonic map of Fig. 5.1.

CHAPTER 6: DETAILED STRUCTURAL ANALYSIS OF THE VALAISAN UNITS

The final geometry of the nappe stack of the Valaisan units presented in Chapter 5 represents the summary of several deformation phases. In this Chapter (Ch.6), based on a combination of both regional mapping and more detailed investigations in selected parts of the study area, with good outcrop continuity, each of these deformation phases will be described (Ch. 6.1 to 6.8). The structural analysis of the Valaisan units deals with the discussion of the structural elements related to the deformation phases encountered (D1 to D6) with particular attention to those features which place constraints on the kinematic evolution. (Ch. 6.2.2, 6.3.7, 6.4, 6.5.2, 6.6 and 6.7).

6.1: The pre-D1 deformation phase

As already discussed in Chapter 4, parts of the Valaisan domain, i.e. the Versoyen and the Pt. St. Bernard units, underwent HP-LT eclogitic metamorphism. Furthermore, the D1 deformation phase was shown to have post-dated peak pressure conditions. The pre-D1 deformation phase, as discussed in Chapters 5.4 and 5.5, is therefore very probably related to peak HP metamorphic conditions.

Microstructural observations suggest that there is indeed a pre-D1 microfabric with a mineral association which is clearly distinct from that stable during the D1 phase. A planar fabric associated with this pre-D1 phase is defined by the mineral association carpholite, chlorite and white mica, occasionally preserved within quartz segregates (Bousquet, 1998). In Fig. 6.1 relicts of carpholite within a quartz grain show a preferred orientation that could be interpreted as related to a pre-S1 fabric, i.e. a foliation or fibres. Although details on the geometry of this pre-D1 deformation phase are impossible to be deciphered due to the subsequent overprinting, some conclusions can be drawn nevertheless.

The large-scale considerations discussed in Chapter 5 suggest that the pre-D1 deformation phase represents the oldest preserved Alpine deformation phase and possibly the most complex one that has affected the Valaisan units. However, there are no direct observations of reliable shear-sense criteria associated with the pre-D1 phase and the kinematics of this phase remains unresolved. This phase was primarily responsible for the detachment and the individualization of the Versoyen unit from its oceanic basement within the internal Valaisan paleogeographical domain.

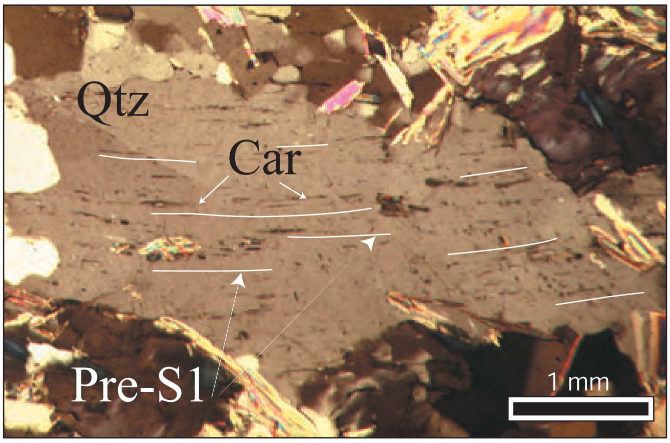


Fig. 6.1 Thin section showing relicts of carpholite preserved within quartz aggregates (calcschists of the Pt. St. Bernard unit). Carpholite generally shows a preferred orientation that can be interpreted as a pre-S1 foliation (sample provided by R. Bousquet)

6.2: The D1 deformation phase

Apart from a pervasive regional S1 foliation, clearly recognizable within all the Valaisan units (including the Pt. St. Bernard unit), other structural elements related to this phase are strongly overprinted by later deformation phases and are sometimes hard to be detected. This is probably the reason why in all the previous studies little attention was paid to this D1 phase. All Authors agreed that a first deformation phase was related to the formation of a S1 foliation (Lancelot, 1979; Spencer, 1989), but the regional impact of this phase remained unclear.

In Chapters 5.1 to 5.7 this phase was demonstrated to be related to the formation of the regional scale late-D1 thrust which emplaced the Versoyen unit (internal Valaisan) onto the Pt. St. Bernard and Moûtiers units, respectively, post-dating an F1 mega-fold within the Versoyen unit itself. Interestingly, the structural elements related to this first deformation phase (D1) must have developed under different metamorphic conditions within different parts of the Valaisan units. In the internal Valaisan (i.e. the Versoyen unit) and inside a part of the external Valaisan (i.e. the Pt. St. Bernard unit) the S1 foliation formed under blueschist facies metamorphic conditions (see

Chapter 4.3). In a more external part of the Valaisan (i.e. the Moûtiers and Roc de l'Enfer units), however, the S1 foliation formed under greenschist facies metamorphic conditions (Gely and Bassias, 1990). The simultaneous formation of the S1 foliation, under different metamorphic conditions, is supported by the structural evidence that the S1 foliation within the Versoyen unit is coeval with the formation of the F1 mega-fold.

Together with the late-D1 basal thrust of the Versoyen unit, the F1-mega structure is responsible for an earlier juxtaposition of the internal Valaisan onto the Pt. St. Bernard unit (occurring under blueschist facies metamorphic conditions) and later thrusting onto the less metamorphic part of the external Valaisan, i.e. the Moûtiers unit. At the end of the D1 deformation phase the HP-LT unit of the internal Valaisan, the Versoyen unit, was in tectonic contact with the more external and less metamorphic part of the Valaisan domain, i.e. the Moûtiers unit. Therefore it seems logical to conclude that the formation of the S1 foliation is related to the exhumation of the high-pressure Valais units due to progressive top-N thrusting. (see Fig. 7 in Fügenschuh et al. 1999).

Additional structural evidence confirming the more or less contemporaneous formation of the S1 foliation within the entire Valaisan domain is supplied by the presence of large-scale F1 folds inside the Moûtiers unit. Although these large-scale F1 folds are not directly observable in the field, their presence can be inferred using the F2 facing structural directions. Chapter 5.2 mentioned that the Moûtiers unit rests in the lower limb of the F2 Versoye synform, and that before the onset of D2 deformation it displayed an upward younging direction. This pre-D2 disposition of the Moûtiers unit is responsible for the W-facing F2 folds already described in Chapter 5.3. In the westernmost part of the Moûtiers unit, where the superposition of the D4 phase is more intense, D2 (and older) structures are transposed and dislocated by a system of late-D4 thrusts that obliterates any structural correlations between different thrust sheets. Despite these tectonic complications, in the Les Chapieux and the Pt. de Mya areas (Fig. 5.7 b) F2 folds displaying an ESE directed F2-facing direction, forming antiformal synclines and synformal anticlines can be observed. The presence of ESE facing F2 folds suggests that these F2 folds have affected a part of the Moûtiers unit which had already been overturned during D1. This overturning of parts of the Moûtiers unit during D1 can be explained by the formation of large-scale F1 folds which are contemporaneous to the D1 phase in the Versoyen unit. Unfortunately, these observations are limited to these two small areas and cannot be directly correlated with larger scale D1 structures.

6.2.1: Small-scale D1 structures: F1 folds

Evidence for D1 deformation in the external and internal Valaisan also includes small-scale D1 structures. Small-scale F1 isoclinal folds occur within all the units of the Valais domain and the F1 fold axes are depicted in the pole Figure of Fig. 6.2.

However, especially in the post-rift sediments of the Moûtiers unit and in the Liassic sequences of the Pt. St. Bernard unit, F1 hinges are sometimes difficult to be detected. This is probably due to the homogeneity of these lithologies as well as to the very intense D2 deformation. F1 folds are best preserved near the hinges of F2 folds. At every location F1 fold axes are oriented subparallel to the L1 stretching lineation (also compare statistically similar orientation in the pole Figures of Figs. 6.2 and 6.3).

However, the orientation of F1 fold axes, L1 and the S1 foliation (sub-parallel to S0) can vary considerably between different localities within the Valaisan units (see maps of Figs. 6.2 and 6.3). This is partly due to non-cylindrical folding and, more importantly, to later overprinting by D2 and D4 folding. Nonetheless, these small-scale observations can be correlated with larger scale structures. Figures 6.4 to 6.6 show small-scale examples of D1 structures from external and internal Valaisan units.

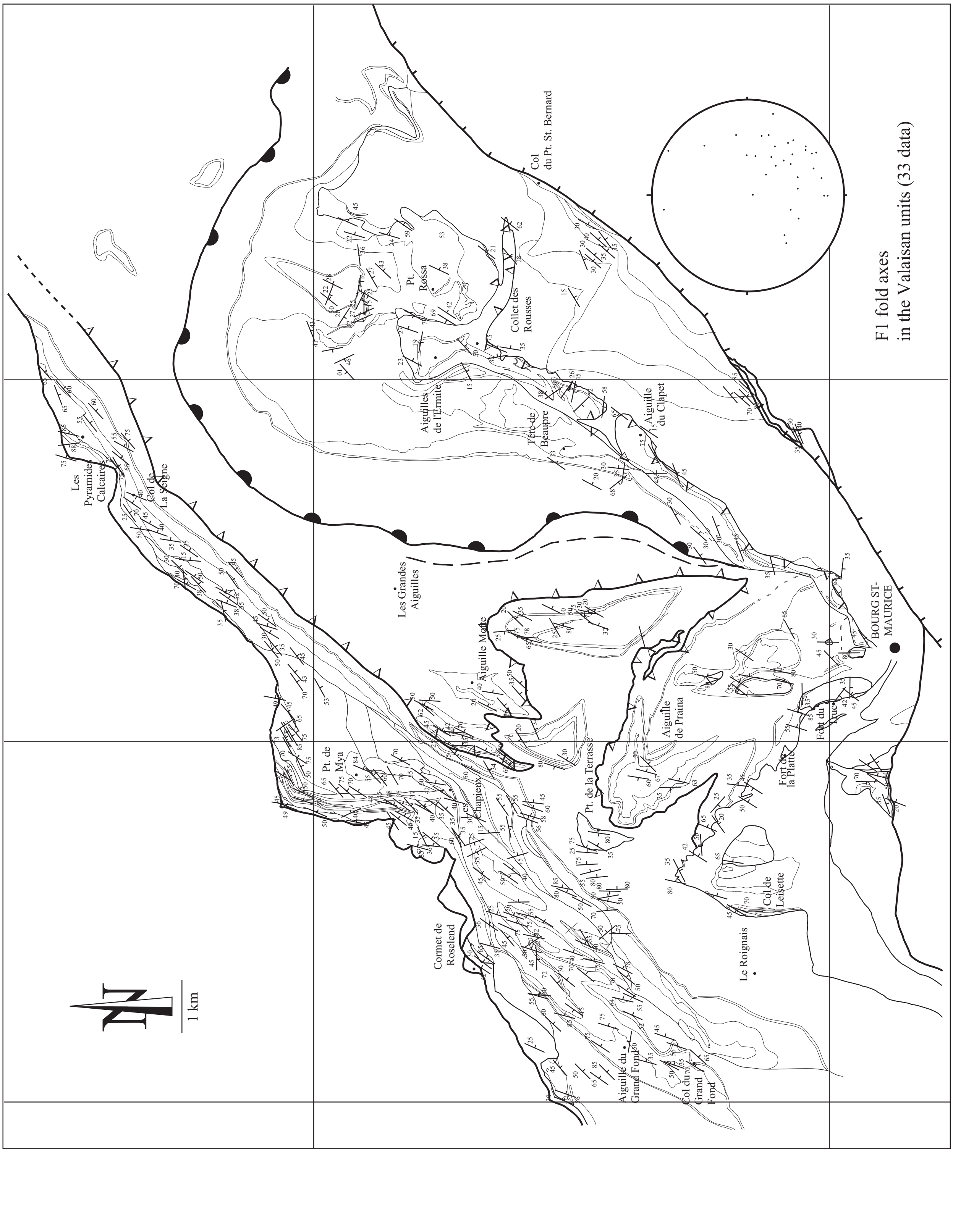


Fig. 6.2 Representative measurements of the S1 foliation (map) and F1 fold axes (stereoplot) within the Valaisan units

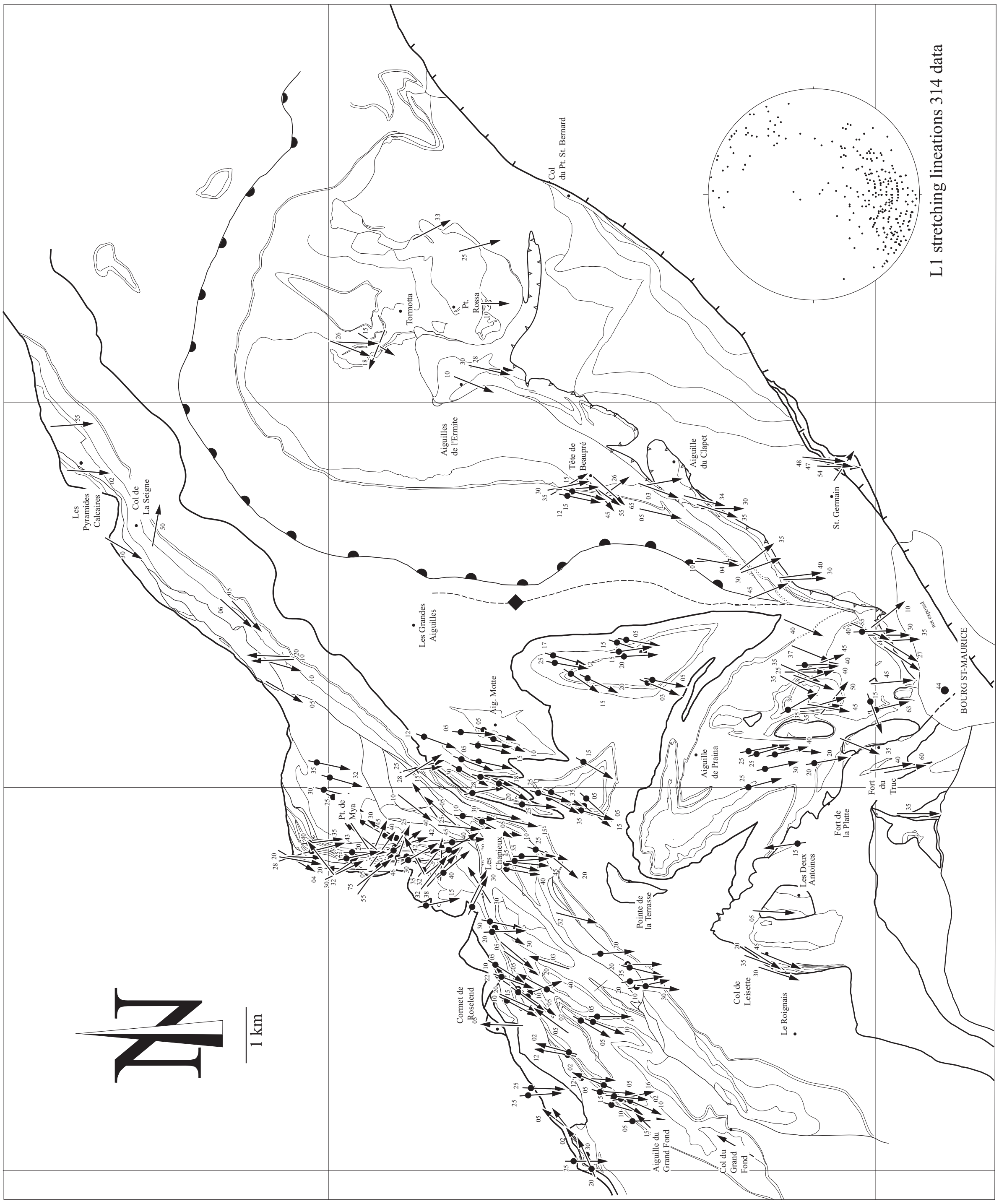


Fig. 6.3 Azimuth and dip of L1 stretching lineations: mineral lineations interpreted as stretching lineations (arrow) and long axes of breccia components (arrow with a solid dot)

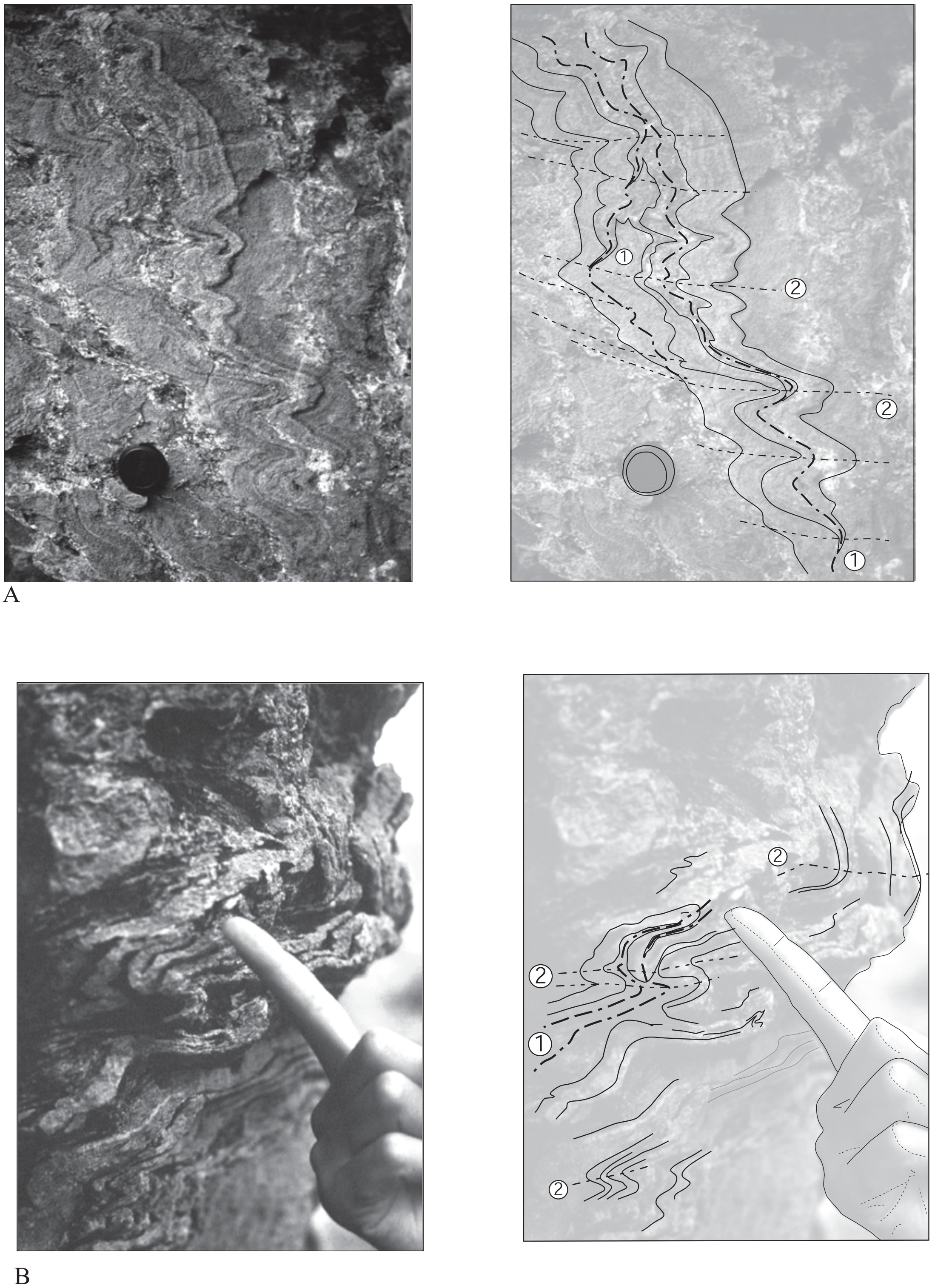
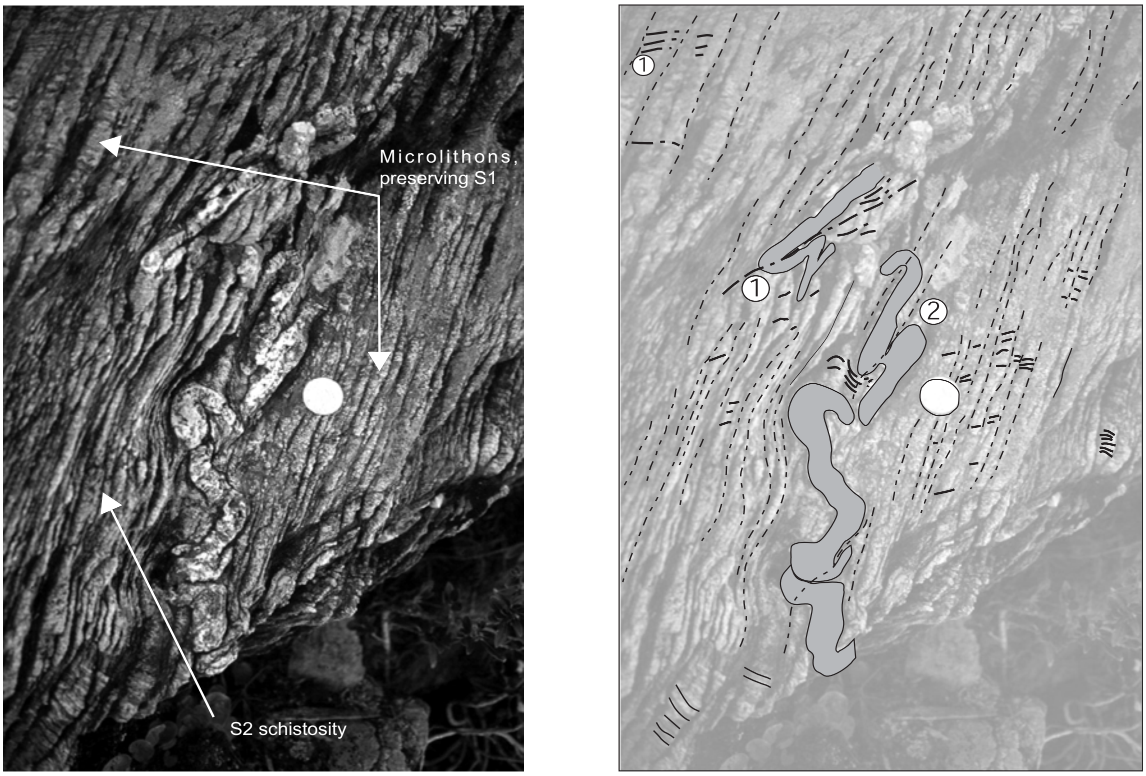


Fig. 6.4 Photographs and line drawings of F1 small scale folds in the C. de St. Cristophe formation (post-rift sediments) of the Versoyen unit (internal Valaisan), overprinted by D2 folds. Tête de Beaupré area (Fig. 6.4a: 949'150/2'085'750, 6.4b: 949'120/2'085'650)



A

B

Fig. 6.5 Photographs and line drawings of F1 small scale folds of the Moûtiers unit (external Valaisan).
(A) Overprinting relationships in Permian conglomerates of the pre-rift sediments. A sub-horizontal S1 foliation (1) is preserved within microlithons related to the formation of a spaced S2 schistosity (2). Roselend pass area (940'000/2'085'600).
(B) F2 fold (2) overprints F1 isoclinal fold (1) in the C. de St. Cristophe formation of the post-rift sediments. Les Chapieux area (944'075/2'084'025).

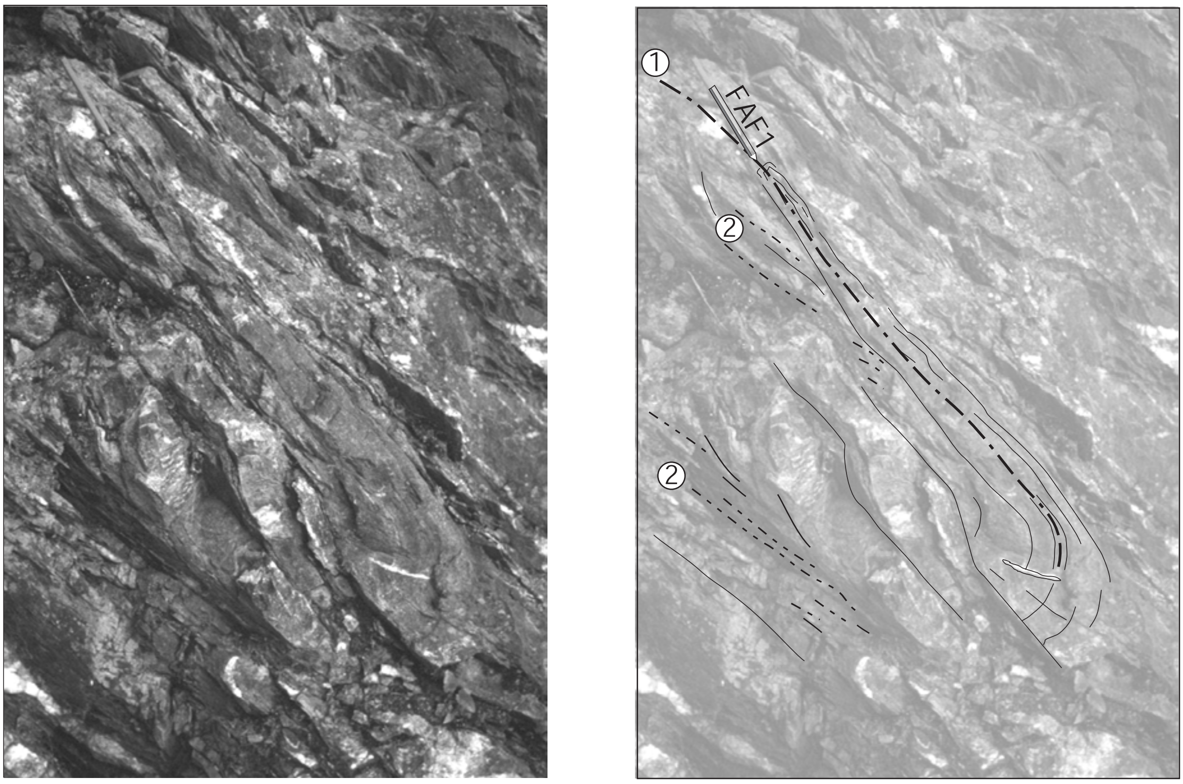


Fig. 6.6 Small scale F1 folds in the Middle Liassic sediments of the Pt. St. Bernard unit (external Valaisan) overprinted by F2 crenulation. Pte. du Clapey area (947'350/2'081'250)

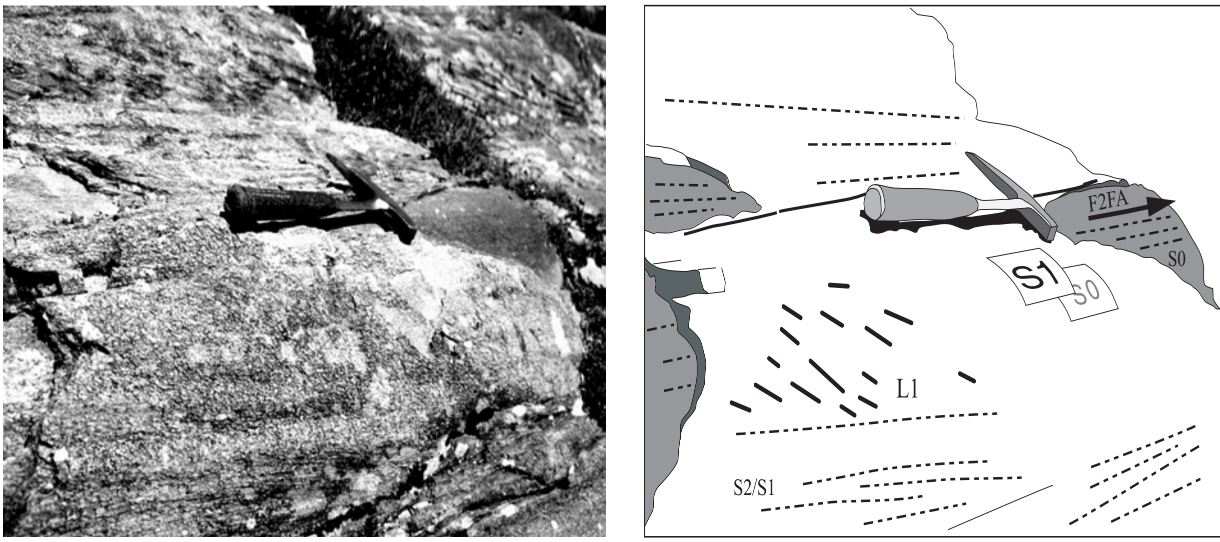


Fig. 6.7 Photograph and line drawing of first phase mineral and stretching lineation (L1) folded by second phase folds in the C. de St. Cristophe of the Moûtiers unit (external Valaisan) of the Chapieux area. (2'085'250/942'450).

6.2.2: Small scale D1 structures: L1 lineations

Two types of lineations are related to the first deformation phase (Fig. 6.3): a) mineral lineations interpreted to be parallel to the principal stretching and b) lineations produced by the intersection between S1 cleavage and bedding.

In the Valaisan units a mineral lineation is often present on S1 cleavage planes. However, it is important to note that, where overprinting criteria are not directly observable, it is not evident whether the lineation is solely due to the D1 deformation phase or if it is also related to strain during the D2 deformation event. This is additionally complicated by the fact that the L1 lineations generally tend to be subparallel to both D1 and D2 phase fold axes (see Chapter 6.3.1 for comparison). The mineral lineation consists of aligned calcite aggregates or elongated quartz grains. The most suitable lithologies in which such mineral lineations can be observed, are: the Permian and the calcareous schists of the C. de St. Cristophe in the Moûtiers unit, the calcschists of the Pt. St. Bernard unit (external Valaisan) and the metasediments of the Complexe Antéflysch of the Versoyen unit (internal Valaisan).

In the Brèches du Grand Fond formation of the Moûtiers unit, as well as in the C. de l'Aroley of both external and internal Valaisan units, the long axes of detrital components lie within the S1 cleavage and are subparallel to the mineral lineation. This proves that this mineral lineation represents a stretching lineation. The second type of L1 lineation is produced by the intersection between S1 and bedding (S0), which however, is parallel to L1 stretching lineation.

The azimuth of the long axes of pebbles measured on bedding planes are often seen to be oriented parallel to the S1/S0 intersection (Fig. 6.3). The S1 foliation, containing these components and a subparallel mineral lineation, is passively folded by D2 and D4. A rare example of refolded L1 lineations is given in Figure 6.7. The effect of the reorientation of the L1 lineation by F4 folding and thrusting becomes particularly evident when looking at the L1 lineation map (Fig. 6.3). Some of these L1 lineations situated close to the Pennine Front, i.e. near the Les Chapieux village, show a spread in their azimuth as the result of D4 deformation.

In this area, this re-orienting effect is particularly evident because the D4 deformation phase becomes more intense. However, Figure 6.7 demonstrates that the L1 lineation may also be passively folded by D2 structures, in regions not affected by D4 folds.

Such a region, not affected by D4 folding, is found in the eastern part of the study area, i.e. far away from the Pennine Front (Versoyen unit situated between Bourg St. Maurice and the Italian-French border) The data depicted in Fig. 6.8 show that the poles to S1 (subparallel to S0) containing the L1 lineation, vary considerably. The distribution of S1 is consistent with folding of the S1 foliation, and consequently also of L1, around a S-SSW oriented (187/33) fold axis, typical for the F2 fold axes in general (see Chapter 6.3.1). This clearly suggests that the finite strain represented by the L1 lineation has been passively folded also by D2 deformation and that hence the L1 lineations can be interpreted to have formed during D1.

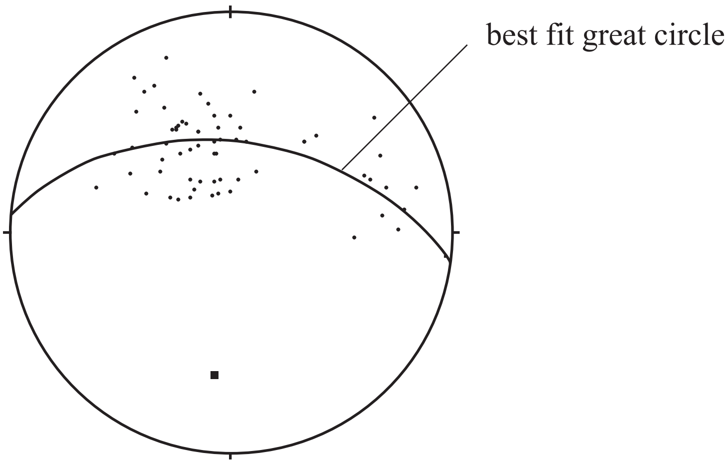


Fig. 6.8 Stereoplot of the poles to the S1 foliation (64 data, equal area, lower emisphere), containing the L1 lineation for an area situated between Bourg St. Maurice and the Italian-French border (for location see Fig. 5.12). This area within the Versoyen unit has been chosen because there is no effect of the D4 deformation phase. The S1 foliation is folded around a mean F2 fold axis with a N-S (187/33) orientation. This shows that S1, and consequently also L1, are folded during the second deformation phase (D2). Therefore the stretching lineations have to be assigned to the first phase of deformation (D1).

In conclusion, the stretching lineations in this area are not due to a combination of D1 and D2 but they are formed during D1. Because the D2 fold axes are subparallel to L1 this D2 folding only had a minor effect in reorienting L1.

This interpretation is also supported by the following remarks regarding the metamorphic grade. In most lithologies, D1 is associated with a pervasive S1 foliation, whereas D2 lacks such a pervasive foliation and S4 is even less penetrative. Because subsequent folding phases are accompanied by a decrease in metamorphic grade and supposedly strain intensity, one would expect that the finite strain state was largely produced during the early D1 deformation event. The L1 stretching lineations are N-S to NNE-SSW oriented in regions away from the Pennine Front (Fig. 6.3). Because D2 has a minor reorienting effect this orientation is likely to be close to the original orientation of L1 before overprinting due to D2. D4 deformation, however, reorients L1 into a NW-SE direction, particularly near the Pennine Front. Since the orientation of the long axes of components is generally subparallel to that indicated by aligned mineral aggregates, and assuming simple shear, we interpreted L1 to represent the direction of tectonic transport during the D1 deformation phase. No shear sense indicators associated to the D1 phase were observed and hence it is not possible to infer an unambiguous determination of the sense of movement during D1. However, large scale considerations suggest top-N shearing during D1 (see Fig. 5.5). This intense shearing was associated with the rotation of D1 fold axes into parallelism with the transport direction.

6.3: The D2 deformation phase

The D2 deformation phase within the Valaisan units is superimposed onto the structural elements discussed in Chapters 6.1 and 6.2. The prominent D2 features formed during two stages. Structures related to the early-D2 stage consist in F2 folds, among which the most prominent feature is the regional scale F2 Versoye synform. Chapter 6.3.1 will discuss small and large-scale early-D2 structures, respectively, and completes, together with Chapters 5.2 and 5.3 the analysis of the early stage of the D2 deformation phase. During a later D2 stage, thrusts did develop, post-dating earlier F2 folds. These tectonic contacts will be referred to as late-D2 thrusts. Large part of the discussion of late-D2 thrusts is dedicated to tracing a thrust contact at the base of the Roc de l'Enfer unit which corresponds, in this study, to the late-D2 Leisette thrust (Chapter 6.3.2). The structural analysis of this thrust contact implies the redefinition of its hangingwall block (i.e. the Roc de l'Enfer unit) which is made up by a complex thrust sheet system (Chapters 6.3.3 to 6.3.5). Chapter 6.3.6 discusses the impact of the late-D2 thrusts (mainly the Leisette thrust) on the geometry of the Valaisan nappe stack and concludes the discussion on late-D2 features. Information regarding the transport direction during the D2 deformation phase (including early- and late-D2 stages) will be provided in Chapter 6.3.7, completing the analysis of the D2 deformation phase. The subject of Chapter 6.3.7, already approached during the discussion of the Leisette thrust (Chapter 6.3.2), will be completed with additional evidence coming from the study area.

6.3.1: D2 folding and cleavage formation

D2 structures, overprinting D1 structural elements, are also observed on the meso- and microscale. F2 folds crenulate the pre-existing S1 cleavage and are generally associated with a millimetric spaced (1-5 mm) crenulation cleavage, i.e. S2. This cleavage is best developed in the hinge region of F2 folds affecting Mesozoic pelitic and calc-pelitic rocks of the Valaisan units (Figures 6.9 and 6.10). In rocks of different mineralogical composition, i.e. massive calcareous limestone, the S2 cleavage is generally difficult to be recognised. Where phyllosilicates are missing, the S2 cleavage is generally not visible and S1 remains the dominant foliation.

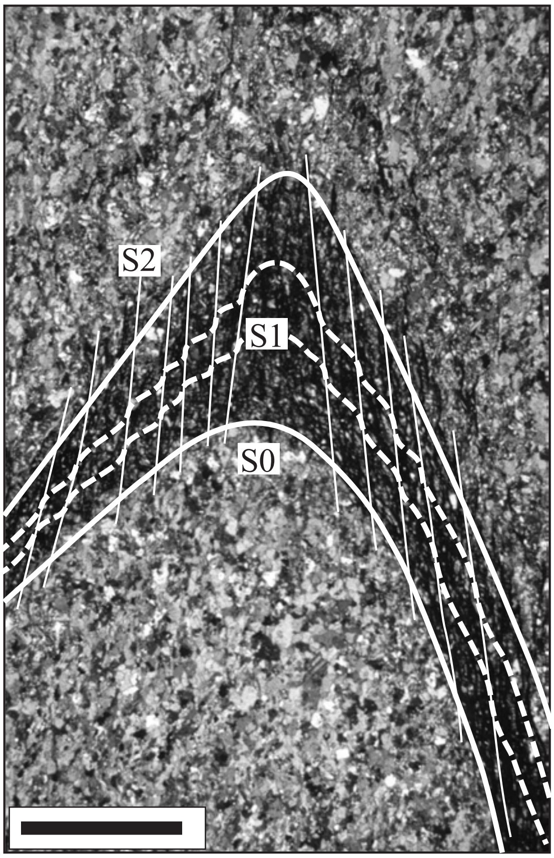
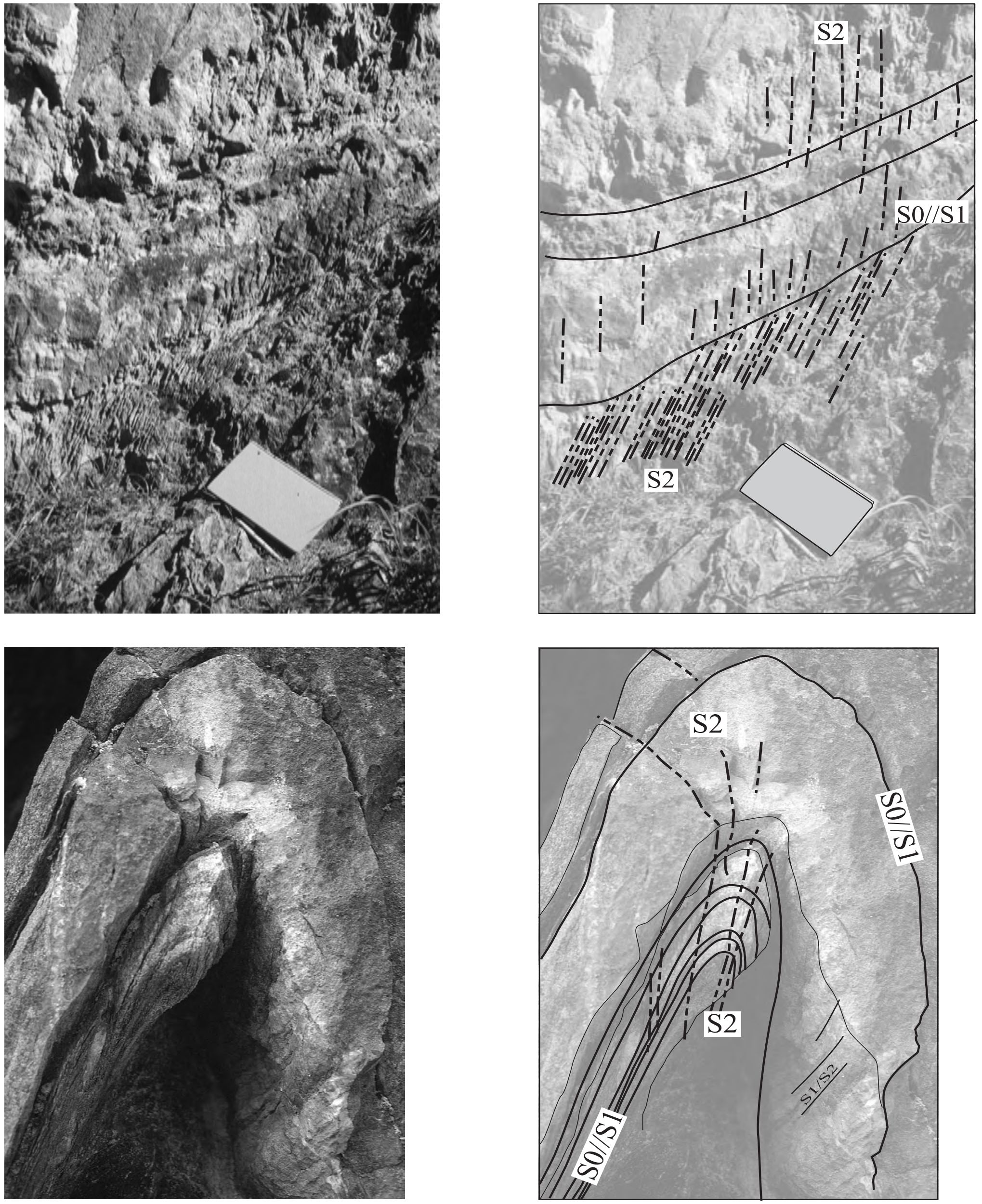


Fig. 6.9 F2 micro fold with S2 crenulation cleavage in mica-rich levels in the St. Christophe Fm. of the Moûtiers unit. Scale bar is 0.8 mm.

Hence, the S2 crenulation cleavage is generally less pervasively developed than the S1 foliation (see Chapter 6.2.2). The S2 cleavage is subparallel to the axial plane of the second phase folds and, except in the hinge zone of the F2 folds, it is usually subparallel to both the S1 cleavage and the sedimentary bedding (S0). Representative orientations of the gently inclined second phase axial plane (S2) are given in Fig. 6.11. S2 generally dips toward E and ESE, where not reoriented by D4 folds. While the dip azimuth of the S2 foliations is predominantly to the SE, the dip angle displays a large variation (Fig. 6.11). Apart from local tectonic complications, the angle of dip generally increases towards the NW: the S2 foliation changes its dip angle from about 30° to 40° degrees in the E and SE, to values between 50° and 80° in the W and NW part of the study area. This general trend is an effect of the superposition by the D4 deformation phase (F4 folds are characterised by steep axial planes) onto the D2 structural elements (see Chapter 6.5). An additional, but local effect, on the S2 dip angle is provided by the late uplift of the Mt. Blanc massif (D5 deformation phase, see Chapter 6.6).

For these reasons, information concerning the original dip of the S2 foliations and D2 axial planes must be searched far S to SE of the Mt. Blanc massif and in an area where the post-D2 overprint (mainly due to D4) is not so pronounced, i.e. away from the Pennine Front. As in the case of the L1 lineations (Chapter 6.2.2) this area corresponds to the Tête de Beaupré region where all the D1 and D2 structures are not affected by D4 structures. In this area, the S2 axial plane cleavage dips about 35° toward E and SE.

Mapping of the study area reveals that amongst the different structural elements, the D2 folds produced the dominant geometrical features (see Fig. 5.7 illustrating the regional impact of D2 folding on a regional scale). Therefore, F2 folds play a key role in deciphering the deformation history of the Valaisan units. Analysis of the geometry of F2 folds, combined with that of the F2 facing direction, represented a powerful tool (Chapter 5.3) to discern pre-D2 features (pre-D1 and D1) from structures assigned to subsequent phases (D3 to D6).



A

B

Fig. 6.10 Examples of S2 cleavage from F2 folds of C. de St. Cristophe formation of the Moûtiers unit. Note that both examples show an F2 axial plane cleavage (S2) that is now sub-vertical due to later D4 overprint. a) An S2 cleavage at high angle with bedding (S0//S1) in the hinge zone of decametric F2 fold. S2 cleavage is well developed in the lower and less competent strata. In this case millimetric spacing is diagnostic for S2. After the change in the orientation of cleavage towards more competent layers, the S2 cleavage becomes well spaced and less evident. b) Core of an F2 isoclinal fold. S2 cleavage is difficult to be recognised in the less competent strata (core of the fold) and this is even weaker in the more competent strata (outer hinge of the fold). Note that in the less competent strata the S2 cleavage can be recognised only in a narrow area close to the hinge line. Toward the fold limbs, S2 becomes subparallel to the S0,S1 cleavage. In the more competent strata, however, the S2 cleavage strongly changes its orientation and is often no more observable.

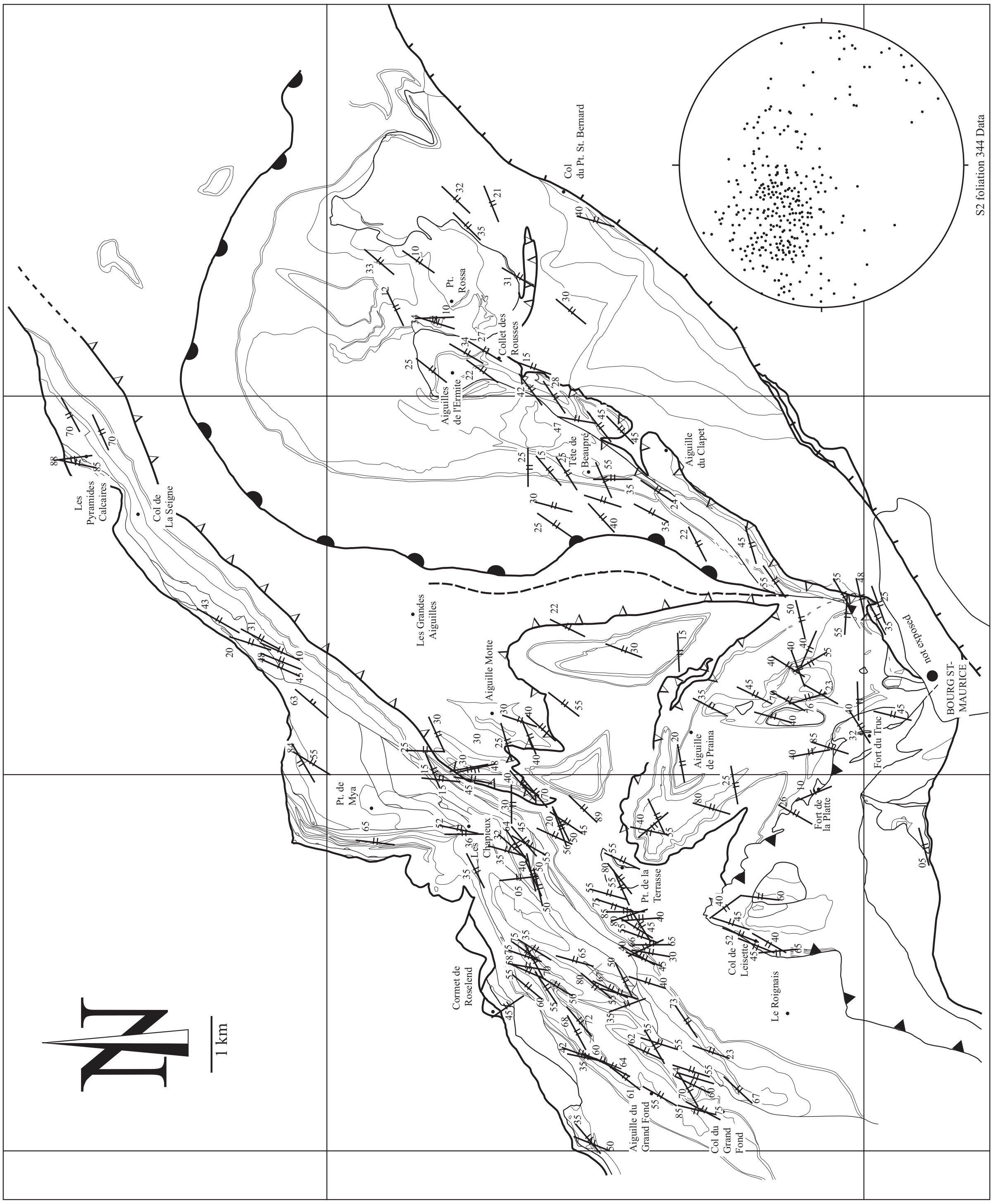


Fig. 6.11 Orientation of S2 foliations and F2 axial planes in the Valaisan units

In the Valaisan units, near-isoclinal F2 similar folds can be clearly recognized on metric (Fig. 6.12) to kilometric scale amplitude. The most outstanding F2 fold shown in Fig. 5.8, can be observed from the road between Bourg St. Maurice and the Cormet de Roselend pass (road D902 in correspondence of the Crêt Bettex village). Other large scale D2 folds can be observed in the Tête de Beaupré area (Fig. 5.11) and within the Complexe Antéflysch in the Aig. de l'Ermite area (Fig. 5.14, profile C). In the western part of the study area, a valid example of D2 large scale folding is represented in the Grand Fond area (Fig. 5.7a). Most of the large-scale isoclinal F2 folds affecting the Valaisan units (Fig. 5.7) were interpreted, at a regional scale, as parasitic F2 folds of the Versoye D2 synform (Chapter 5.1.2).

The orientations of F2 fold axes are shown in Fig. 6.13. They plunge gently toward 190°-220° in the western to north-western areas and toward 180°-150° in the eastern and south-eastern part of the study area. They are oriented subparallel to the L1 principal stretching direction (compare the pole Figure of F2 fold axes in Fig. 6.13 to that of the L1 stretching lineations of Fig. 6.3). The plunge of D2 fold axes is gentle (typically 5-20°; locally, N of Bourg St. Maurice, it increases to values between 30-50°). This local increase of the plunge of fold axes corresponds to a comparable increase in the S2 foliation (Fig. 6.11). This steepening is an effect of late- or post D2 structural overprint.

Fig. 6.12. Example of F2 fold style. F2 folds are generally tigth with narrow, rounded hinges, although the curvature of the hinges dependes strongly on lithology. In the above figure, similar and near-isoclinal F2- folds formed within the C. de St. Cristophe formation of the Versoyen unit.

Another geometrical feature of the F2 fold axes is represented by their change in orientation F2 fold axes between the north-western (190°-220°) and the south-eastern part (180°-150°) of the study area. This regional trend was first noticed by Lancelot (1979). Andrieux and Lancelot (1980) related this change in orientation to the occurrence of late-D2 tectonic contacts, i.e. their Phase 2'. These Authors interpreted this tectonic contact ("*contact cissaillant 2'* ", Figs. 2 and 3 in Andrieux and Lancelot 1980), associated with a late-D2 folding, as responsible for the re-orientation and refolding of the F2 folds axes from 220° in the north-western part to 160° in the south-eastern part of the study area. However, own field data (Fig. 6.13) indicate that the orientation of F2 folds, from NW to SE within the study area, change progressively and smoothly. Therefore a different explanation is needed (see below).

In Chapter 5.1.2, the major D4 features were described in terms of a strain gradient, increasing from SE to NW and associated with the Pennine Front. Towards NW, this results in a more pronounced overprinting of the previous structural features, i.e. pre-D1 to D2. Implying that in the SE, the strike of F2 folds, i.e. 180°-150°, was not reoriented during D4. Hence it is proposed that the original azimuth of the F2 fold axes was N-S to NNW-SSE prior to the onset of D4 deformation phase.

Fourth phase folds, striking NE-SW, developed oblique to the trend of the earlier F2 folds (180°-150°) and led to a type-3 interference pattern (Ramsay, 1967). Therefore, the progressive change of the strike of F2 folds, from 160° in the SE to 220° in the NW of the study area, could simply be an effect of the refolding by F4, whereby the F2 axes rotate from 160° towards NE-SW, i.e. 220°. According to our structural data, there is definitely no necessity for the late-D2 (D2') tectonic contact and folding proposed by Andrieux and Lancelot (1980). It should be stressed that, both in cross section and in map view, the trace of the late-D2 (D2') tectonic contact of these authors (Andrieux and Lancelot, 1980) strongly reminds that of the late-D1 thrust of the internal Valaisan onto the external Valaisan as proposed in this study (Figures. 5.1 and 5.3). Despite this similarity the two contacts should not be confused. Nevertheless, F2 folds are found in association with late-D2 thrusts which share a similar transport direction (Chapter 6.3.7). However, the F2 folds are not footwall synclines and hangingwall anticlines. Instead, the thrusts cut through the previously folded F1 and F2 strata.

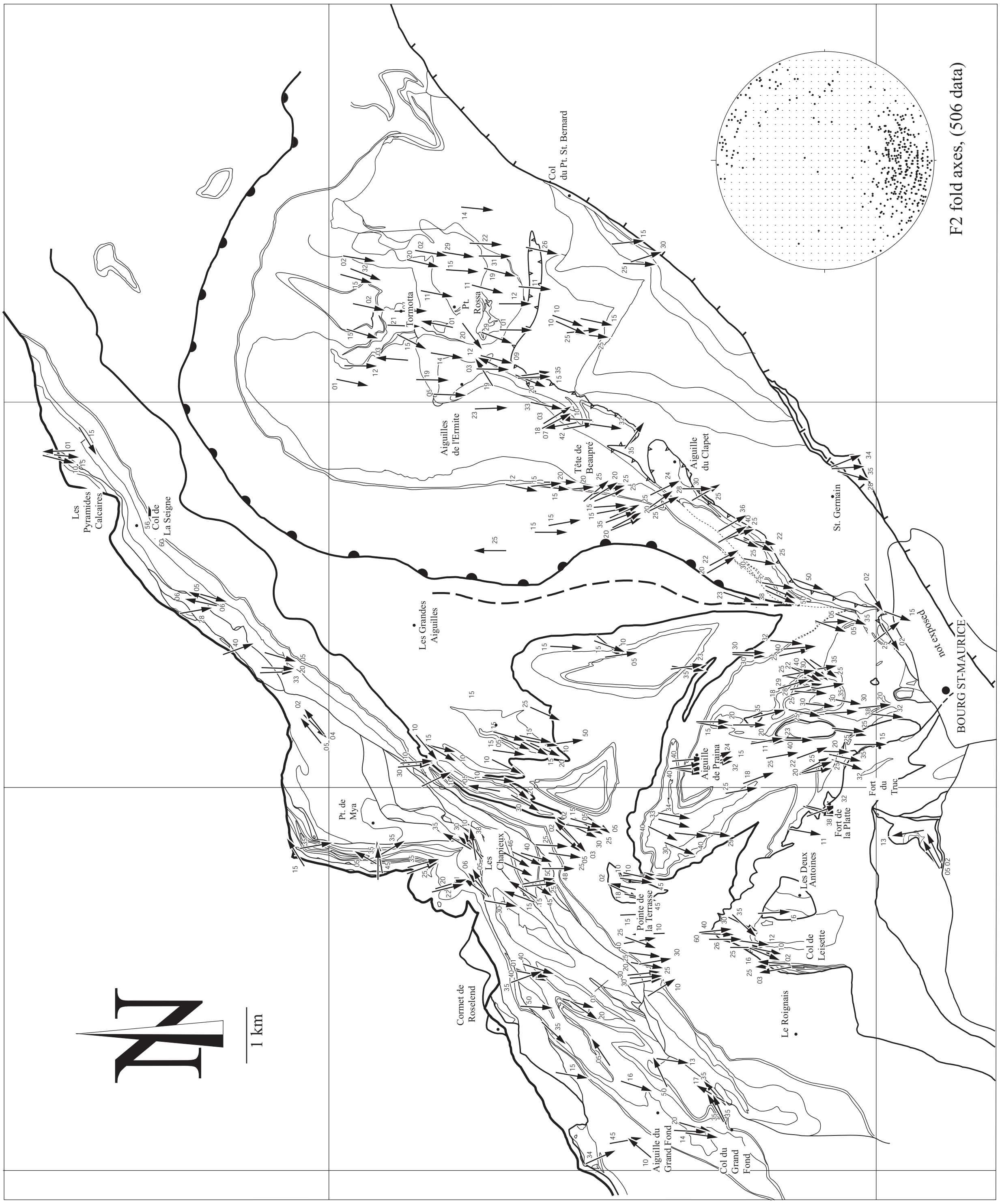


Fig. 6.13. Azimuth and dip of second phase fold axes in the Valaisan units

At this stage some field observations concerning the D2 fold style will be discussed in order to document the structural connection between early- and late- D2 stages. In Fig. 6.14 three occurrences of the same F2 large scale fold, from within a distance of 6-7 km, have been schematically drawn together. The occurrence shown in Fig. 6.14b represents the example depicted in Fig. 5.9 and shows the F2 antiformal syncline observed at the NE side of the ridge between the Pt. de la Terrasse and the Aig. De Praina (Chapter 5.2). The hinge zone of this structure can be followed, parallel to the fold axis orientation both toward N into the Aig. Motte area (Fig. 6.14a), and towards S (Fig. 6.14c).

In Fig. 6.14b, the wavelength of this large-scale D2 structure roughly corresponds to the drop in topography over a distance of about 1000 m (Fig. 5.9). Further north, occurrences of the same D2 fold can be observed around the Aig. Motte peak; the same structure is only partially preserved due to erosion and its wavelength is not quantifiable (see Fig. 5.16 and 5.17). However, the observable preserved hinge zone, formed by the Marmontains Fm., displays a much greater thickness compared to that displayed in the central occurrence (Fig. 6.14b). In the southernmost occurrence, however, (Fig. 6.14c) close to the late-D2 Leisette thrust (see Chapter 6.3.2), the same D2 fold structure (here preserved) is considerably tighter.

Despite the qualitative nature of these observations, it seems that the wavelength of F2 folds decreases from N to S (Fig. 6.14). This can be associated with the appearance of a late-D2 thrust towards the S, among which the Leisette thrust is the most evident feature (Figures 5.1 and 5.2). Moreover, and at a regional scale, also the axial plane of the Versoye synform is truncated by a late-D2 thrust towards SE (see section 6.3.2).

In conclusion, these observations possibly indicate a complex geometrical shift in the D2 deformational style from an early D2 stage characterised by folding to a late-D2 stage associated with thrusting. Late-D2 thrusts, seen in association with F2 folds, could be interpreted as a structural feature which developed when the folds locked and thrusting started to play a predominant role in accommodating further shortening. The fact that this superimposition took place along a roughly N-S-directed trend should give a first indication on the kinematics during the D2 deformation phase, discussed later in this Chapter (Chapter 6.3.7).

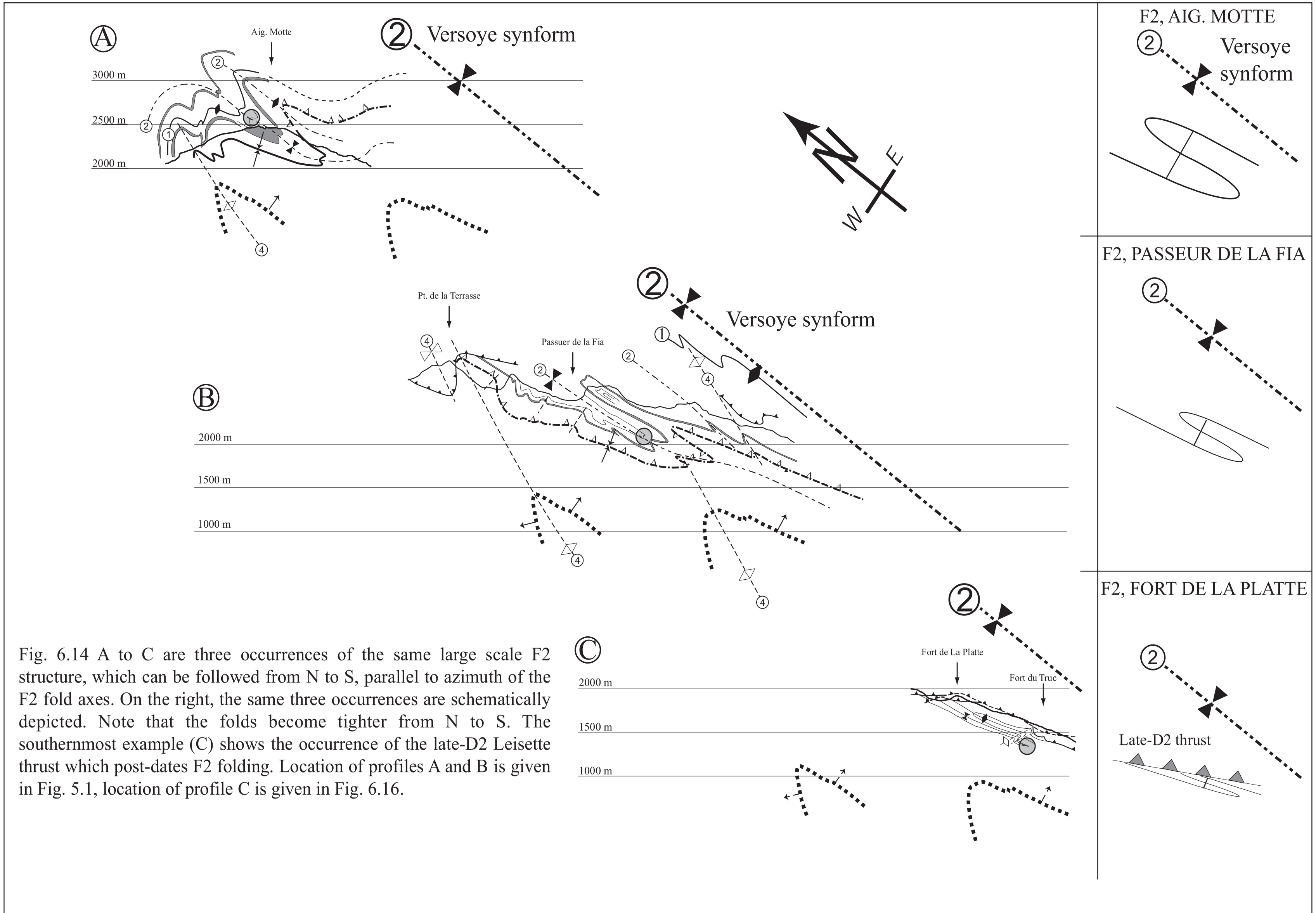


Fig. 6.14 A to C are three occurrences of the same large scale F2 structure, which can be followed from N to S, parallel to azimuth of the F2 fold axes. On the right, the same three occurrences are schematically depicted. Note that the folds become tighter from N to S. The southernmost example (C) shows the occurrence of the late-D2 Leisette thrust which post-dates F2 folding. Location of profiles A and B is given in Fig. 5.1, location of profile C is given in Fig. 6.16.

6.3.2: Late D2 thrusting: the Leisette thrust and the Roc de l'Enfer unit

Late-D2 structures predominantly occur W and NE of Bourg St. Maurice, in the southern part of the study area. This region is characterised by bad outcrop conditions due to gravitational sliding (regarding this gravitational sliding see the geological map of Schöller, 1930). Several, so far unresolved, geological complications (Antoine et al., 1992; Antoine et al., 1993; Fudral, 1980) can be explained in terms of late-D2 features in this area. In the following the Roc de l'Enfer structural unit, currently individualized from the rest of the Valaisan by the Leisette late-D2 thrust contact, will be redefined. This Roc de l'Enfer unit extends from Bourg St. Maurice all the way to Moûtiers, forms a thin sliver of mainly Carboniferous schists (and occasionally also Mesozoic pre- and post-rift sediments), and is structurally located between the Houiller zone in the hangingwall and the rest of the Valaisan units in the footwall (external and internal Valaisan units). According to our findings three different thrust-sheets can be recognized within the Roc de l'Enfer unit (Figures 6.15 and 6.16). The stacking order of the different thrust sheets indicates that the lowermost thrust sheet, referred to as the Plan Andre thrust sheet, was overthrusted by the Forclaz thrust sheet towards the NNE. This Forclaz thrust sheet mainly corresponds to the Roc de l'Enfer unit, as classically defined by Barbier (1948) and Antoine et al. (1992) (see Chapter 6.3.4).

On top of the Forclaz thrust sheet the uppermost Deux Antoine thrust sheet can be mapped out (Fig. 6.15). All these sub-units, i.e. Plan Andre, Forclaz and Deux Antoine thrust sheets, rest in the hanging wall of the late-D2 Leisette thrust. The Forclaz, Plan Andre and Deux Antoine thrust sheets will be discussed separately in terms of their structural evolution, distribution and stratigraphic record (Chapters 6.3.3 to 6.3.5 respectively). The late-D2 Leisette thrust occurs in the region included between the Forêt de l'Arbonne and the Charbonnet Valley NW of Bourg St. Maurice (Figures 6.15 and 6.16). Between the Col de Leisette and N of the Fort de la Platte, the late-D2 Leisette thrust defines the base of the Forclaz thrust sheet which rests directly on top of the Moûtiers and Versoyen units. Note that the Leisette thrust cuts discordantly through the D1 thrust between the Moûtiers and Versoyen units.

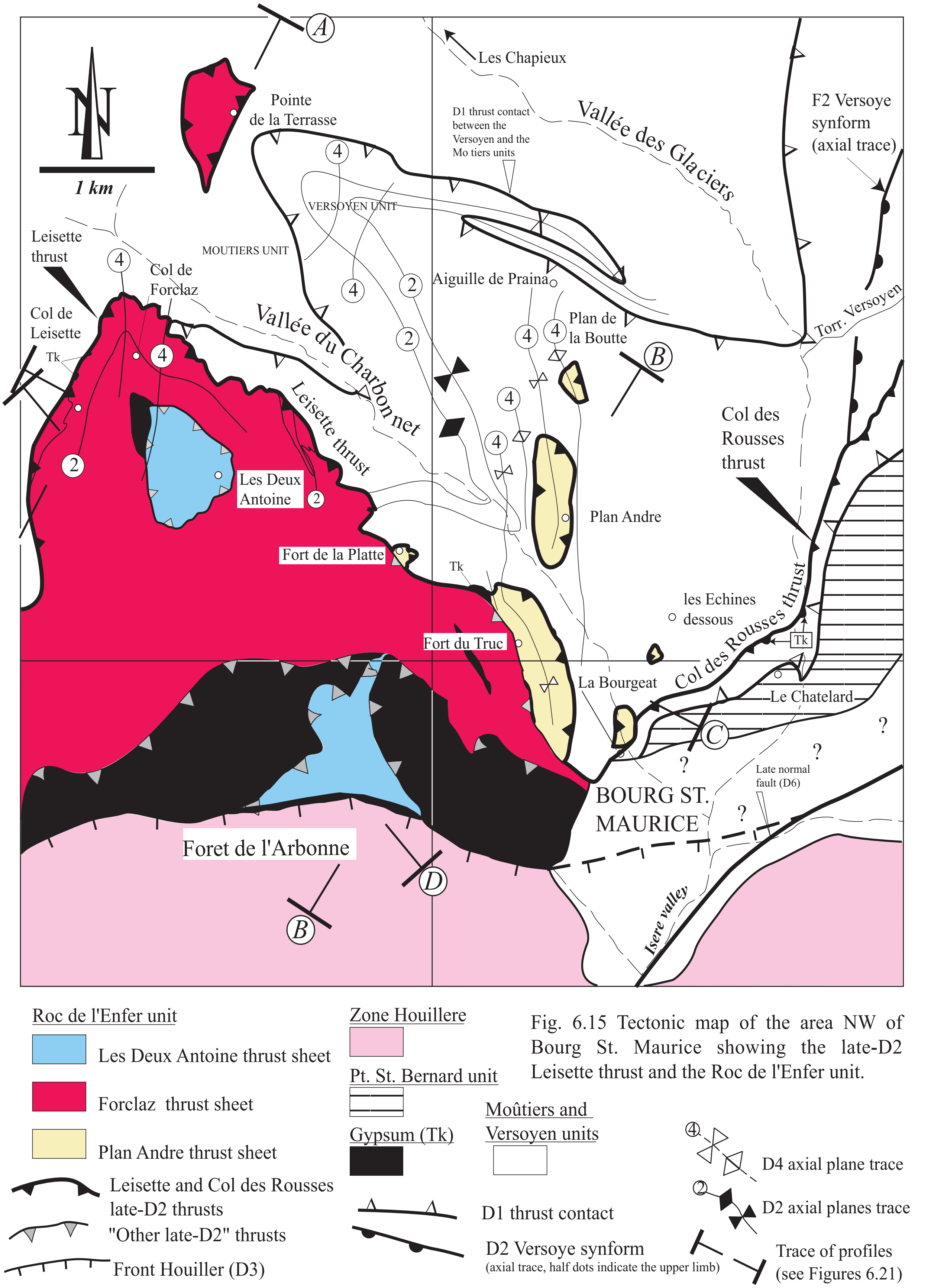


Fig. 6.15 Tectonic map of the area NW of Bourg St. Maurice showing the late-D2 Leisette thrust and the Roc de l'Enfer unit.

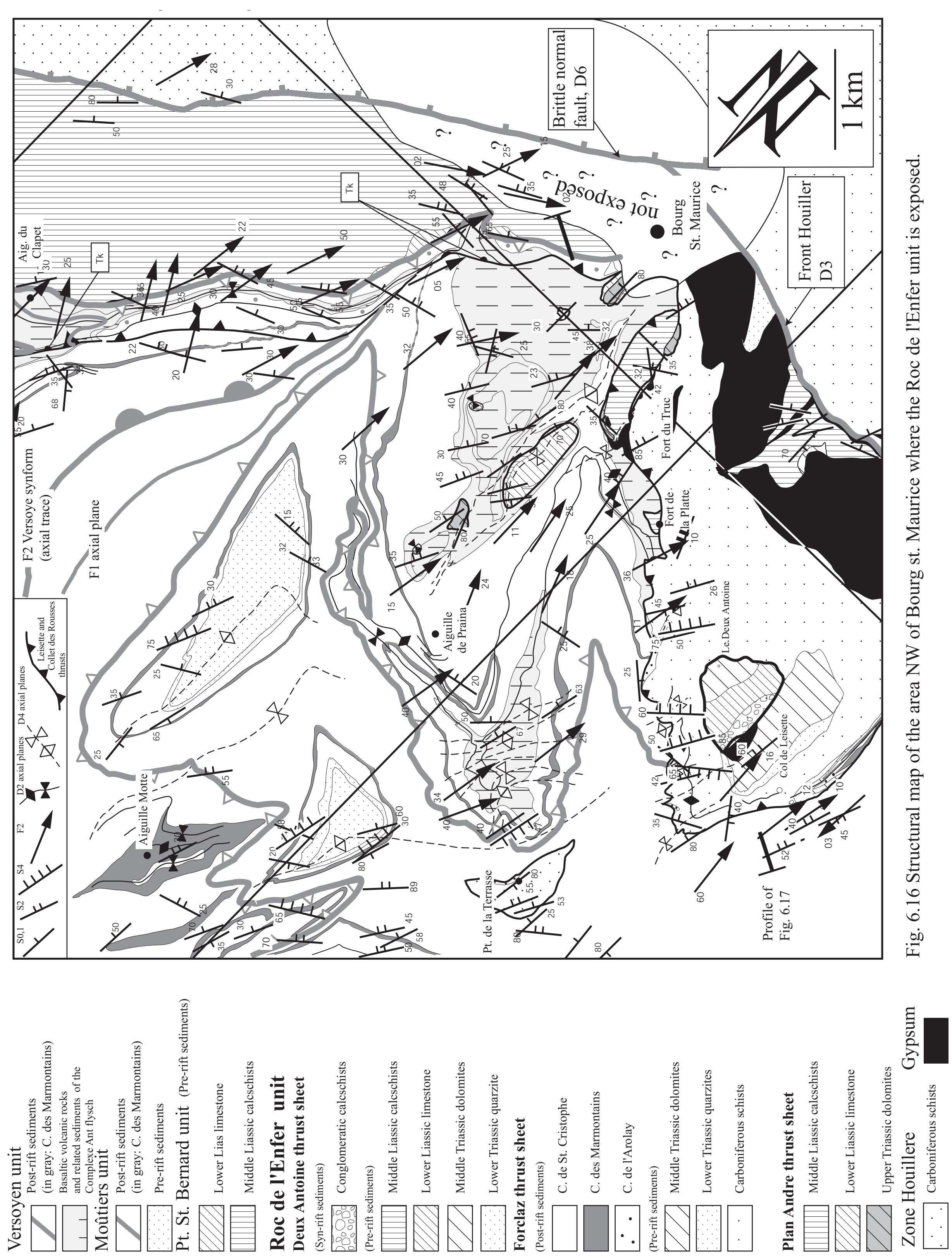
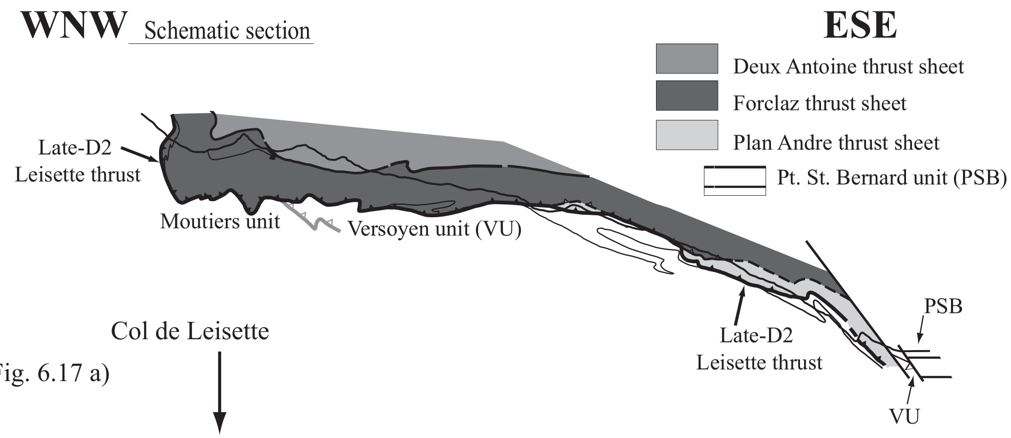


Fig. 6.16 Structural map of the area NW of Bourg st. Maurice where the Roc de l'Enfer unit is exposed.

At the Fort de la Platte and further SE at the Fort du Truc, the Plan Andre thrust sheet is inserted between the Forclaz thrust sheet and the underlying strata of the Versoyen unit (Figures 6.15 and 6.16). Another set of small but crucial outcrops was studied along the NE side of the Charbonnet Valley, from the Aig. de Praina peak to Bourg St. Maurice (Fig. 6.15). These outcrops, also belonging to the Plan Andre thrust sheet, are tectonically emplaced onto the Versoyen unit (Fig. 6.15).

D1 and D2 structures in the footwall of the Leisette thrust, are part of the south-westward continuation of the structures described in Chapter 5 (Chapter 5.2, Fig. 5.2) and they are portrayed in the profile of Fig. 6.17. The axial planes of the large-scale F2 folds shown in Fig. 5.9 and described in section 5.2 can be traced across the Valleé de Charbonnet up to a point N of the Fort de la Platte, where they are truncated by the Leisette thrust (Fig. 6.15). In the Col de Leisette area, the generally gently SE dipping Leisette thrust underlining the Forclaz thrust sheet is affected by a D4 large scale synform (Fig. 6.17) responsible for the steepening of the thrust contact in this area. The fact that the Leisette thrust contact post-dates the F2 folds and is overprinted by F4 folds, unambiguously ascribes this structural feature to a late-D2 stage of deformation. The Plan Andre thrust sheet shows the same relative age relationships with the underlying folded strata of the Versoyen unit. These relationships can be observed at the Fort du Truc and at Plan Andre further N. Fig. 6.17 shows that the basal thrust of the Plan Andre thrust sheet, post-dates F2 folded lithologies of the Versoyen unit and is overprinted by D4 folding at the Fort de la Platte.

For all those localities where the structural relationships with the underlying strata are not exposed, it is proposed that they form isolated fragments of the same hanging wall block. This is supported by two observations. Firstly, all these outcrops rest on top of the previously folded (F1 and F2) strata of the Versoyen unit, i.e. they occupy the same structural level and secondly all the different klippen show the same stratigraphic record (Fig. 6.16). For these reasons it is proposed that N and W of Bourg St. Maurice, the late-D2 Leisette thrust is located at the base of the Plan Andre thrust sheet while further NW, where the Plan Andre thrust sheet is missing, the same late-D2 Leisette thrust contact is located at the base of the Forclaz thrust sheet (Fig. 6.15).



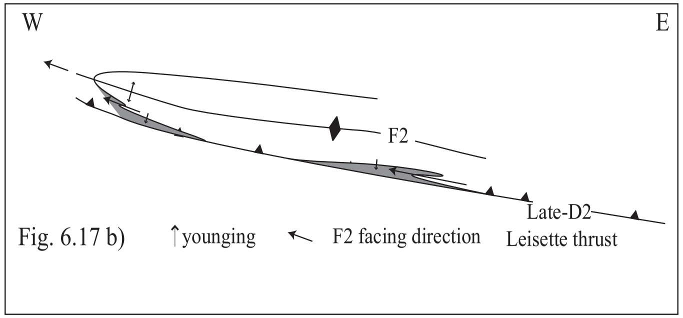
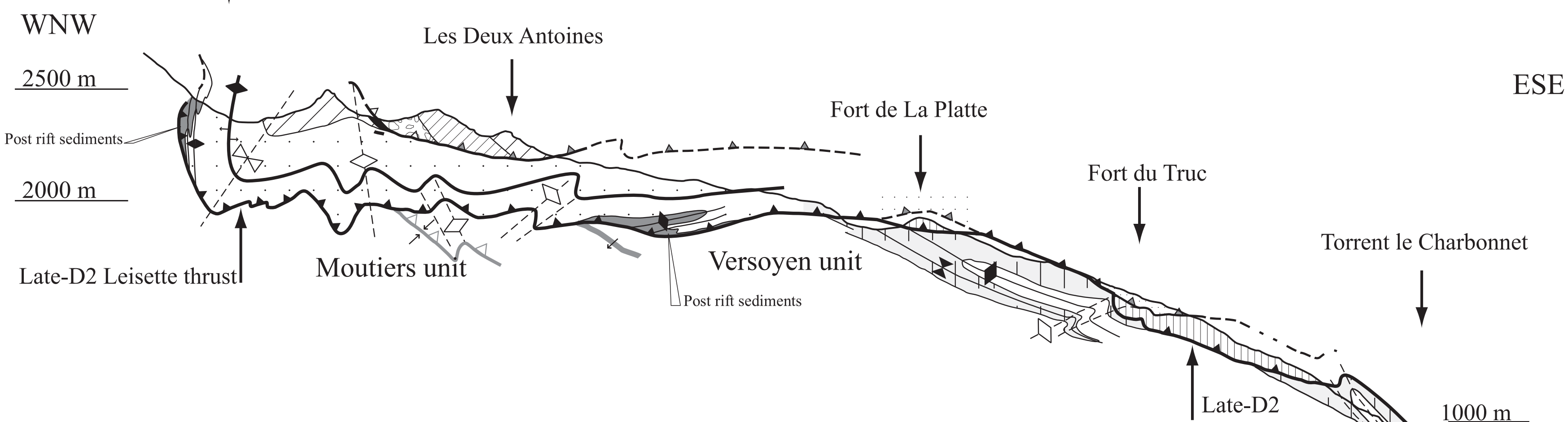


Fig. 6.17 a)



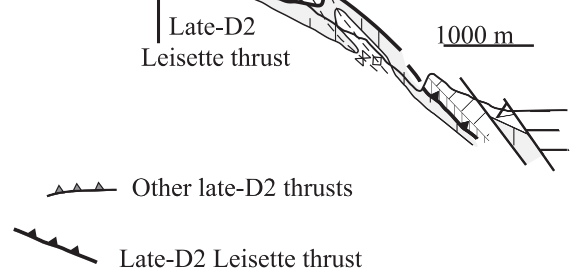


Fig. 6.17a) WNW-ESE oriented geologic section across the Roc de l'Enfer unit together with a schematic section (top left) showing the geometry of the different thrust sheets forming the Roc de l'Enfer unit. Signatures and location of the profile are given in Fig. 6.16. The uppermost thrust sheet (the Deux Antoine thrust sheet) is preserved, at the Deux Antoine massif, on top of the Forclaz thrust sheet. Below the Forclaz thrust sheet, the lowermost Plan Andre thrust sheet is preserved at the Fort de La Platte and Fort du Truc areas. In the WNW the late-D2 Leisette thrust is found at the base of the Forclaz thrust sheet while in the ESE it underlies the Plan Andre thrust sheet.

Fig. 6.17b) Qualitative reconstruction, only for the Forclaz thrust sheet, before the onset of the D4 deformation phase. Note that the F2 facing direction is toward W, similar to that of the Moûtiers unit shown in Fig. 5.11.

Information concerning the transport direction along the Leisette late-D2 thrust is found in the Col de Leisette area. The sigma clast shown in Fig. 6.18 was observed in the C. de St. Cristophe, i.e. post rift sediments belonging to the Forclaz thrust sheet (i.e. part of the Roc de l'Enfer unit). Fig. 6.18 shows a subhorizontal S2 cleavage, overprinting an older S1 cleavage which is only preserved as microlithon structures. Within the S2 cleavage a more competent layer is observed to form σ-clasts.

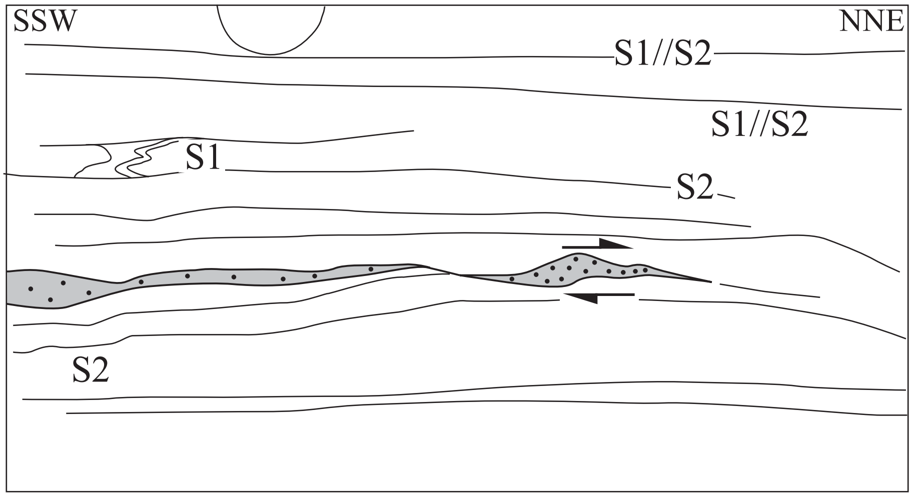


Fig. 6.18 Photograph and line drawing showing shear sense indicators observed along the Leisette late-D2 thrust within the C. de St. Cristophe formation. The sigma-clast like aggregate of quartz and calcite is subparallel to the S2 cleavage. The inferred sense of shear is top to the NNE. This indicates a top-to-the N transport direction during the second (D2) phase of deformation.

This observation, allows to unambiguously deduce a top-to-the NNE (lineation 210/20) movement direction during the late-D2 phase of deformation.

This shear direction, however, does not represent the true original late-D2 movement direction, because it occurs near the hinge zone of large scale F4 folds and was slightly reoriented. At the Fort du Truc, the tectonic contact between the Forclaz and the Plan Andre thrust sheets shows a tectonic transport direction which is SSW-NNE oriented (lineation 202\35) while the uppermost Deux Antoine thrust sheet is thrusted onto the Forclaz thrust sheet towards N (lineation 170/60, top-to-the N).

An extended discussion concerning the tectonic transport direction during the early and late stages of the D2 deformation phase is presented in Chapter 6.3.7. In conclusion, the stacking of the thrust sheets forming the Roc de l'Enfer unit is interpreted to have occurred during a late-stage of the D2 deformation phase (late-D2) because the internal tectonic contacts between the different thrust sheets display a roughly N-directed transport direction subparallel with the transport direction of the late-D2 Leisette thrust.

6.3.3: The Plan Andre thrust sheet at the base of the Roc de l'Enfer unit

The Plan Andre thrust sheet consists of a series of outcrops N, NW and W of Bourg St. Maurice. All these outcrops exhibit a characteristic stratigraphic record and are in an identical structural position. From Bourg St. Maurice towards the Aig. du Praina all the occurrences of the Plan Andre thrust sheet are aligned along a N-S direction. All the localities (from S to N: La Bourgeat, Les Echines dessous, Plan Andre and Plan de la Boutte) rest structurally above the basaltic volcanic rocks intercalated within the sediments of the Complexe Antéflysch of the Versoyen unit (internal Valaisan). The basal thrust of the Plan Andre thrust sheet coincides with the Leisette late-D2 thrust.

In all the situations observed, the lithologies included exclusively Upper Triassic to Liassic strata (Fig. 6.16). Dolomites are only locally preserved (La Bourgeat, Les Echines Dessous and Plan de la Boutte). Due to the lack of lithologies older than the dolomites, it is not possible to decide whether these dolomites belong to a Middle Triassic stage (Ladinian) or to the Upper Triassic. The sporadic presence of dolomitic breccia (Les Echines dessous) points to the second option. The Liassic strata are represented by a massive white limestone (Lower Lias, Antoine 1971) followed by a sequence of belemnite-bearing Middle Liassic calcschists (Antoine et al., 1992) and Upper Liassic black schists (new findings of belemnite-bearing black schists at La Bourgeat, this study). The tectonic contact of the Upper Liassic calcschists with the Versoyen unit, is only exposed at La Bourgeat.

Close to this contact, within the Lower Liassic limestone, a N-S oriented stretching lineation was observed (lineation 175/44). However, neither the transport direction nor the relative age of this thrust could be interpreted (late-D2?). The Plan Andre thrust sheet is made up by pre-rift sediments which belong to the external Valaisan. However, there is no possibility to discern if they originated from the paleogeographic domain similar to that of the Moûtiers unit or that of the Pt. St. Bernard unit.

6.3.4: The Forclaz thrust sheet forming the main body of the Roc de l'Enfer unit (="faisceau de Salins")

The Forclaz thrust sheet mainly corresponds to a unit which was referred to as "fascieau" de Salins (Barbier 1948) or the Roc de l'Enfer unit (Antoine et al., 1992; Antoine et al., 1993) in previous studies. Within the study area, the Forclaz thrust sheet has a large extension along the ridge between the Col de Leisette and the Fort du Truc, W of Bourg St. Maurice (Fig. 6.15). Northward, an occurrence of the Forclaz thrust sheet is preserved in form of an isolated klippe at the Pt. de la Terrasse peak. In the south the Forclaz thrust sheet often rests directly below gypsum of the Deux Antoine thrust sheet (Torrent de l'Arbonne, Fig. 6.15 and 6.16).

In this southern area slices of the uppermost Deux Antoine thrust sheet are associated with several meters thick layer of upper Triassic, sandwiched between the Forclaz thrust sheet (below) and the Carboniferous of the Zone Houillère (above). At a regional scale, the Forclaz thrust sheet can be followed southward up to Moûtiers where it wedges out, together with the other Valaisan units.

The structures within the Forclaz thrust sheet indicate the occurrence of three deformation phases corresponding to D1, D2 and D4 as observed in the rest of the Valaisan (Fig. 6.19).

An initial phase (D1), evidenced by a foliation (S1) oriented subparallel to bedding (S0), is overprinted by a subsequent set of isoclinal folds (F2) which are N-S oriented. Parasitic F2 folds can be observed at the westernmost occurrence of the Roc de l'Enfer unit, i.e. at the Col de Leisette area, and WNW of the Fort the la Platte (Fig. 6.17a).

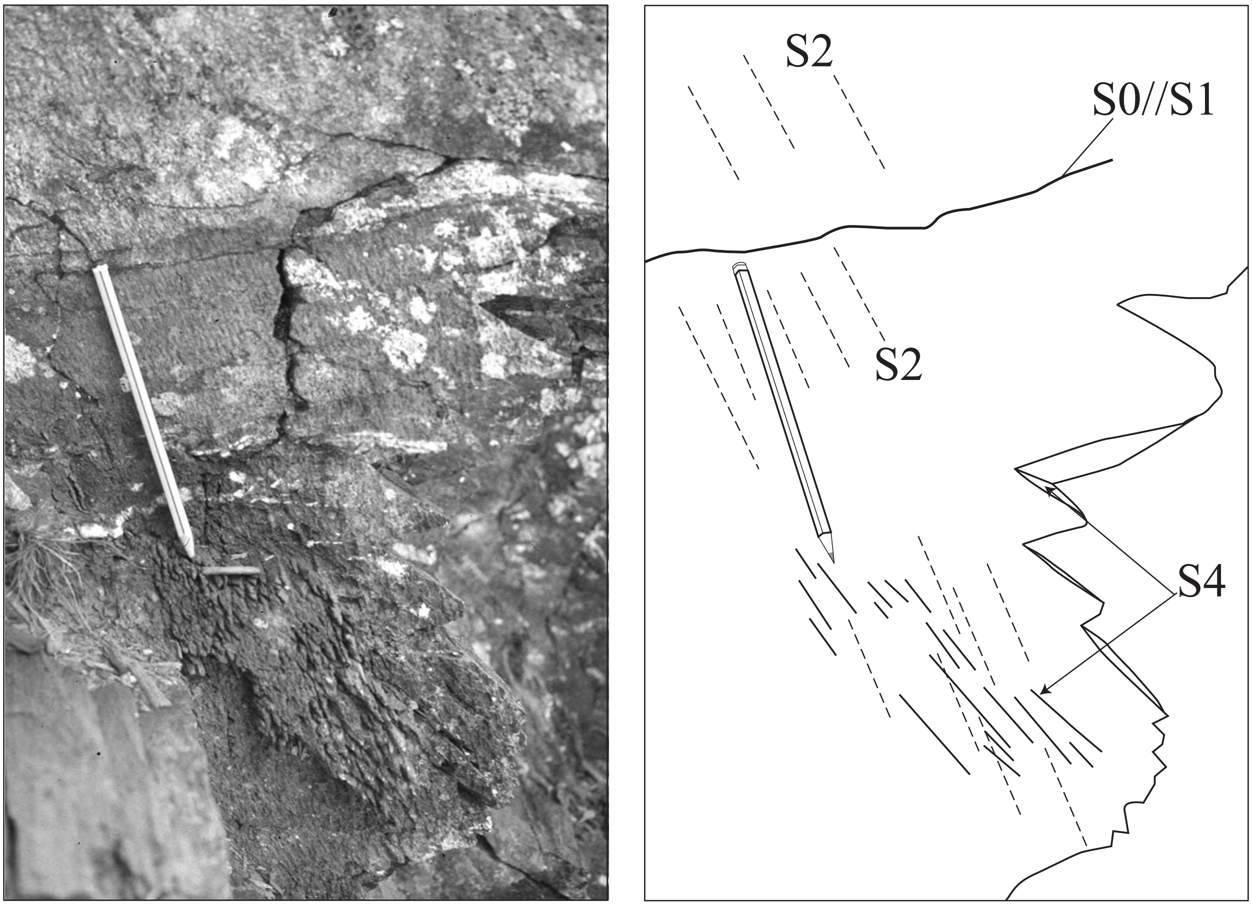


Fig. 6.19.Three deformation phases (D1, D2 and D4) recorded in the Carboniferous schists of the Roc de l'Enfer unit at the Pt. de la Terrasse peack. The S1 cleavage is subparallel to the compositional bedding (S0//S1) and is overprinted by a pervasive S2 cleavage. Actually, the subvertical orientation of S2 is due to a succesive overprinting by F4 folds (S4 cleavage).

After retro-deforming the effect of D4 they show a W-directed facing direction for F2 and their symmetry is compatible with the occurrence of a large scale W-closing F2 antiformal anticline with the Carboniferous schists, i.e. the oldest preserved lithologies, in the core of this thrust sheet (Fig. 6.17 b). The hinge zone of this structure is not preserved due to erosion but it can be deduced to be situated close to the Col de Leisette area. In the upper limb of this large scale isoclinal F2 fold the Carboniferous schists are stratigraphically followed by Lower Triassic to Upper Triassic lithologies showing an upward younging direction. On the contrary, the lower and inverted F2 limb shows a direct stratigraphic contact of the Carboniferous schists with remnants of the post-rift sediments (Col de Leisette area, Fig. 6.17a) thus indicating, after retro-deformation to their original position before the D4 phase, a downward younging direction (Fig.6.17b). The final structures affecting this thrust sheet are sets of open D4 folds which post-date all the previous structural features.

The stratigraphic record of the Forclaz thrust sheet depends on the exact location of the thrust at its base. According to Antoine (1971) and Antoine et al. (1992, their Fig. 10) this tectonic contact is systematically located between pre-rift sediments and post-rift sediments. Therefore according to these authors the stratigraphic record of the Forclaz thrust sheet consists of pre-rift sediments only. N of the Fort du Truc the Leisette thrust emplaces Carboniferous schists over the Versoyen unit while further NW, at the Col de Leisette, it emplaces Lower Triassic quartzite onto the Moûtiers unit (Antoine et al 1992). However, Fudral 1980 interpreted part of the pre-rift sediments, i.e. the Triassic rocks at the Col de Leisette, as the stratigraphic base of the underlying and overturned post-rift sediments of the Moûtiers unit. Therefore he rejected the existence of a Forclaz thrust sheet ("fascieau de Salins" of Fudral, 1980) as an independent tectonic element.

Our new structural data presented here propose a different location for the basal thrust of the Forclaz thrust sheet in this area. In the Fort du Truc area the occurrence of Upper Triassic evaporites along the thrust plane clearly locates a tectonic contact at the base of the Carboniferous schists belonging to the Forclaz thrust sheet. In the Col de Leisette area, the basal thrust of the Forclaz thrust sheet coincides with the Leisette thrust and is no longer marked by the presence of Upper Triassic evaporites. However, detailed field observations indicate that between the Forclaz thrust sheet and the underlying strata, slices of a calcareous conglomerate are rudimentally preserved. Most of these slices have extremely small cartographic dimensions, forming lens-shaped outcrops ranging between 1-5 meters. These slices are generally made up of a calcareous matrix embedding mainly Middle Triassic components. In some cases they are directly in contact with the underlying the Cristophe Fm. belonging to the Moûtiers or the Versoyen units. The same conglomerate slices can be found at the Col de Leisette where their significance becomes clear. There, the conglomerates assume cartographic dimensions (Figures 6.16 and 6.17) and are associated with the typical Marmontains Fm. followed by the St. Cristophe Fm.. This crucial observation reveals that these conglomerates in fact represent the

Aroley Fm., i.e. the base of the post-rift formation. Hence, it is proposed that a small portion of post-rift sediments, formed by the Aroley Fm. and locally (at the Col de Leisette) by a complete sequence, represents the autochthonous post rift cover of the pre-rift sediments (Carboniferous and Lower to Middle Triassic rocks) forming the Forclaz thrust sheet.

These new observations are in contrast to those made by Antoine et al. (1992) while they partially agree with those of Fudral (1980). The fundamental conclusion from our structural interpretation is that the preserved sequence of post-rift sediments in the Col de Leisette area, so far considered to belong to the underlying Moûtiers unit, rests in stratigraphic contact with the pre-rift sediments of the Forclaz thrust sheet. As a result, the tectonic contact should be searched between the St. Cristophe Fm. belonging to the Forclaz thrust sheet and the St. Cristophe Fm. of the Moûtiers unit. This situation is shown in Fig. 6.17.

Another situation similar to that of the Col de Leisette was observed at the Pt. de la Terrasse peak. In this area, lithologies that where previously interpreted as Marmontains Fm. (Antoine et al. 1992; 1993) are now considered in this study as representing Carboniferous schists belonging to the Forclaz thrust sheet. This is strongly suggested by the fact that the schists, at first sight similar to those of the Marmontains Fm., carry detrital white mica typical for the Carboniferous. Additional evidence is given by the fact that, along the contact with the underlying St. Cristophe Fm., lens shaped slices of conglomerates can again be found. These conglomerates are in the same tectonic position as those observed along the Leisette thrust of the Forclaz thrust sheet, i.e. they do not show any stratigraphic relationship with the underlying St. Cristophe Fm. Therefore, they are interpreted as Aroley Fm. representing remnants of the autochthonous post-rift cover of the Forclaz thrust sheet.

According to this study, the Forclaz thrust sheet is made up by pre-rift sediments formation (Carboniferous and Triassic) directly overlain by post-rift sediments above an angular unconformity. Both formations, (pre- and post-rift sediments) are identical to those observed in the Moûtiers unit, i.e. the external Valaisan. According to these strong stratigraphic affinities, the Forclaz thrust sheet probably as to be correlated with the Moûtiers unit, i.e. the external Valaisan.

6.3.5: The Deux Antoine thrust sheet at the top of the Roc de l'Enfer unit

At the Deux Antoine massif, above the Forclaz thrust sheet, the uppermost Deux Antoine thrust sheet comprises steeply dipping Triassic and Liassic strata, together with younger strata of unknown age (Antoine et al. 1992) (Figures 6.15, 6.16 and 6.17a). W of Bourg St. Maurice, at the "Forêt de l'Arbonne", mainly middle Liassic calcschists and some Triassic dolomites outcrop again and are unambiguously on top of the Carboniferous schists of the Forclaz thrust sheet. Although the middle Liassic strata represent tectonic slices within the gypsum, their structural level is comparable to that of the Deux Antoines massif.

At the Deux Antoine massif some structural observations can be made which could possibly give a new insight to explain the structural relationships between the Forclaz and the Deux Antoine thrust sheets. In this area the tectonic contact, which is generally not well exposed, is characterised by the occurrence of Upper Triassic evaporites (NW of the Les Deux Antoine peak, Figs. 6.15, 6.16 and 6.17a). Moreover, at the Forêt de l'Arbonne the first gypsum above the Carboniferous schists of the Forclaz thrust sheet centimetre-size σ-clasts of dolomite allow to establish top-to-the-N-directed transport of the Deux Antoine thrust sheet above the Forclaz thrust sheet (lineation 170/60, top-to-the N). Unfortunately for the others regions where this tectonic contact is exposed, neither the transport direction nor the cinematic indicators were observed. At the Deux Antoine massif, it is clear, however, that this tectonic contact is folded by F4 folds (Fig. 6.17) thus indicating that the upper Deux Antoine thrust sheet was emplaced before D4. Since the underlying Forclaz thrust sheet was thrusted onto the Moûtiers and Versoyen units during a late-D2 stage it may be inferred that the upper Deux Antoine thrust sheet was carried (top-to-the N) on top of the Forclaz thrust sheet during or before the late-D2 stage.

The situation at the Forêt de l'Arbonne is characterised by the occurrence of large amounts of upper Triassic gypsum. This gypsum clearly marks a tectonic contact between the uppermost Carboniferous of the Zone Houillère (hangingwall) and the Roc de l'Enfer unit (footwall) i.e. with the uppermost Deux Antoine thrust sheet. This major tectonic contact between the Carboniferous schists of the Zone Houillère and the Valaisan (Roc de l'Enfer unit) represents a complex feature that has recorded several deformation events. This results in the presence of a folded foliation in the gypsum and in the occurrence of Liassic tectonic slices belonging to the Deux Antoine thrust sheet. The nature of the tectonic contact with the overlying Zone Houillère is not exposed. However, the geometric relationships with the Roc de l'Enfer unit (see the next Chapter, 6.3.6) and fission track dating (Fügenschuh et al., 1999) suggests that D3 normal faulting took place at the contact between the Zone Houillère and the Roc de l'Enfer unit.

6.3.6: The Col des Rousses thrust and its correlation with the Leisette thrust

At the Col des Rousses and at the base of the western side of the Aig. Du Clapet (Fig 5.13) upper Triassic cargneules, observed within the basaltic volcanic rocks and associated sediments of the Complexe Antéflysch of the Versoyen unit, outline a tectonic contact. This contact overprints and post-dates F2 folded strata of the Versoyen unit but is overprinted by open D4 folds (Fig. 5.14, profile C). Hence the Col des Rousses thrust clearly formed during a late stage of the D2 deformation phase (late-D2), analogous to the Leisette thrust. In the same region, other occurrences of evaporites were observed at the NE side of the Aig. de l'Ermite (Dalla Torre 1998) and further N in the Tête du Chargeur area (Italy) (Bucher 1999).

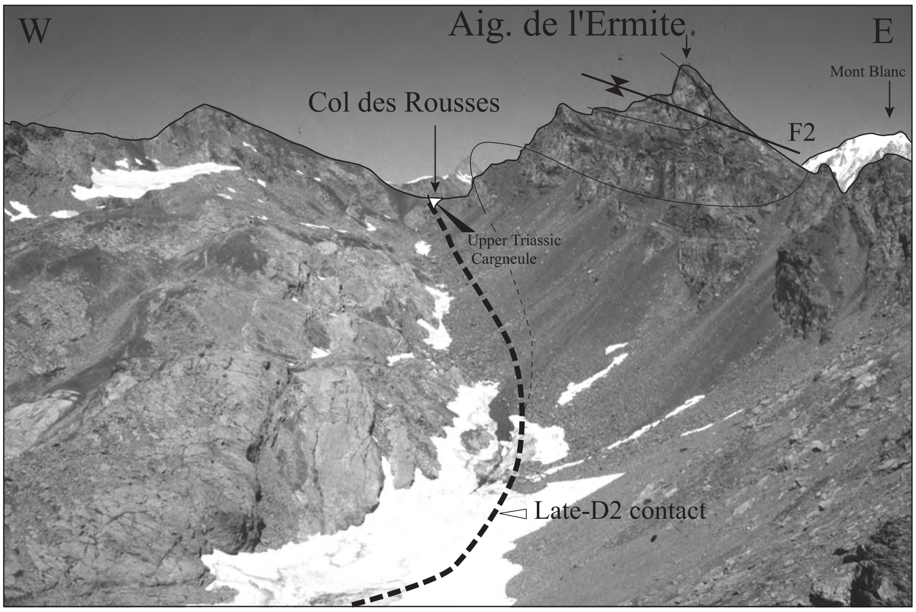


Fig. 6.20a) Late-D2 Col des Rousses thrust within the Versoyen unit, S of the Col des Rousses.

From the Col des Rousses (Fig. 6.20a) and S of the Aig. De l'Eremite area, towards Bourg St. Maurice, the cargneules provide small but crucial outcrops along the left side of the Torr. le

Versoyen (at the 870 m level point) and further S, across the river, N of the village of le Châtelard (x: 946.575, y: 2079.900, z: 970) (Figs. 6.15 and 6.16). Finally, further W the last occurrences of Triassic evaporites are found close to the Leisette thrust (Fig. 6.16).

Earlier workers have already described some of these occurrences of cargneule (Antoine, 1971; Antoine et al., 1992; Antoine et al., 1993; Fudral, 1998; Lasserre and Lavergne, 1976), but their significance at a regional scale was generally neglected. Our observations shed light on this subject because they indicate that all the afore mentioned outcrops of Triassic cargneules form a single late-D2 thrust (the Col des Rousses thrust) that can be traced in continuity from the Tête du Chargeur (I) all the way to Bourg St. Maurice (F) where it can be linked with the Leisette thrust.

From S to N, the Collet des Rousses thrust crosscuts the D2 nappe stack of the Valaisan units from higher to lower structural levels (Fig. 5.1 and 5.2). At the Le Châtelard village in the S, this late D2 contact overprints lithological contacts folded by F2 and belonging to the lower limb of the F2 Versoye synform. Towards NE, first the axial plane of the F1 mega fold within the Versoyen unit and successively the Versoye F2 axial plane (Figs 6.15 and 6.16) are truncated by this contact which, even further NE occurs above the upper limb of the F2 Versoye synform (Figs 5.14 and 15). Across the Italian-French border the Col des Rousses thrust overprints the contact of the Versoyen unit with the Pt. St. Bernard unit (Fig. 5.1) and finally, at the Tête du Chargeur it is associated with a complex late-D2 thrust system which is partially responsible for the northern termination of the Pt. St. Bernard unit (Bucher, 1999). These large scale observations suggest that this thrust is of regional importance. The deformed clasts of the Brèches du Collet des Rousses formation, observed close to the contact, indicate that the hanging wall was thrusted with a roughly top-to-the N transport direction (Fig. 6.20b).

Analysis of the profiles shown in Fig. 5.13 can help to constrain the transport direction along the Col des Rousses thrust. These E-W oriented profiles (Fig. 5.13), actually are not correctly oriented for observing top-to-the N oriented displacement. They can only take into account the E-W component of movement. Actually, such a component of top W movement can be observed in all the profiles although this component cannot be quantified. Therefore, a transport direction towards west or north (i.e. top-to-the NNW or NW?) has to be inferred for this late-D2 thrust contact.

Fig. 6.20b) Close to the late-D2 tectonic contact shown in Fig. 6.20a), shear sense indicators provided by deformed components of the Bréche du Collet des Rousses formation indicate a top-to-the N sense of movement (late-D2).

The Col des Rousses thrust within the Versoyen unit displays structural features which are identical to those seen further SW along the Leisette thrust. Both thrusts formed at a late stage of D2 and display roughly top-to-the N directed transport. A slight difference in the transport direction between the Leisette thrust (top-to-the NE) and the late-D2 contact described before (top-to-the N to NW) reflects differences in the amount of re-orientation due to the D4 deformation phase. In Chapter 6.3.1 it has been demonstrated that moving from NW to the SE of the study area, the trend of the F2 fold axes changes from NE-SW to N-S and finally to NW-SE. Since the transport direction of the late-D2 stage of thrusting is subparallel to the azimuth of the F2 fold axes (see the next section) a similar change from top-to-the NE towards a top to the N to NW is expected going from NW to SE within the study area. For these reasons, and observing that at Bourg St. Maurice (at the locality of La Bourgeat, Figs. 6.15 and 6.16) the western termination of the Col de Rousses thrust comes very close to the eastern termination of the Leisette thrust, it is proposed that both thrusts define a single late-D2 thrust plane.

At this point it should be emphasized that the geometry in this southern part of the study area is not only affected by the late-D2 Leisette thrust but also controlled by a subsequent set of normal fault systems. Two events of normal faulting have fundamental differences in age and in structural style. D2 structures are first overprinted by D3 low angle ductile normal faulting which has been first recognised at the Tête du Chargeur (Bucher, 1999). Our data suggest that the tectonic contact below the Houillère Zone at the Forêt de l'Arbonne can also be interpreted as a D3 normal fault (see discussion below). At a very late stage of the evolution of the Valaisan units another regional scale normal fault crosscuts all the previous structures, overprinting the Houiller Front from La Thuille to Bourg St. Maurice. The Forêt de l'Arbonne area rests in the footwall of this late normal fault (D6, see Fig. 6.15). This late brittle normal fault, steeply dipping toward SE, plays an important role for the eastern and southern termination of the Valaisan units below the Houillère Zone. Therefore, a complete description of the area around Bourg St. Maurice requires the geometric and kinematic relationships between late-D2 thrusting, D3 ductile normal faulting and D6 late normal faulting to be elucidated, a discussion which is found in Chapters 6.4, 6.7 and 6.8.

We now will further discuss late-D2 thrusting along the Leisette thrust referring to Chapter 6.8 for a complete discussion also involving D3 and D6. The geometry of late-D2 thrusting between Bourg St. Maurice and the Col de Leisette, in particular the Leisette thrust, offers the opportunity to analyse the three dimensional geometry of late-D2 thrusting. The geometry of the Leisette late-D2 thrust plane is schematically shown in Figures 6.21 and 6.22. Figure 6.21 shows a set of geological profiles constructed across the Roc de l'Enfer unit. It should also be noted that profiles A and B illustrate the geometry of the Leisette late-D2 thrust in a plane close to the tectonic transport direction. On the contrary, profiles C and D are oriented roughly perpendicular to the F2 and F4 fold axes. They primarily illustrate overprinting of the Leisette thrust by D4 folding. For simplicity only the Plan Andre, Forclaz, Deux Antoine and the Houiller Zone are schematically represented in Fig. 6.21. The four profiles (A to D) are combined into the three-dimensional diagram shown in Fig. 6.22 which shows the regional relations of Leisette thrust plane to the major structural features in its footwall. Note the truncation of the late-D1 thrust contact of the Versoyen unit onto the Moûtiers unit and of two large-scale parasitic F2 folds by the Leisette thrust.

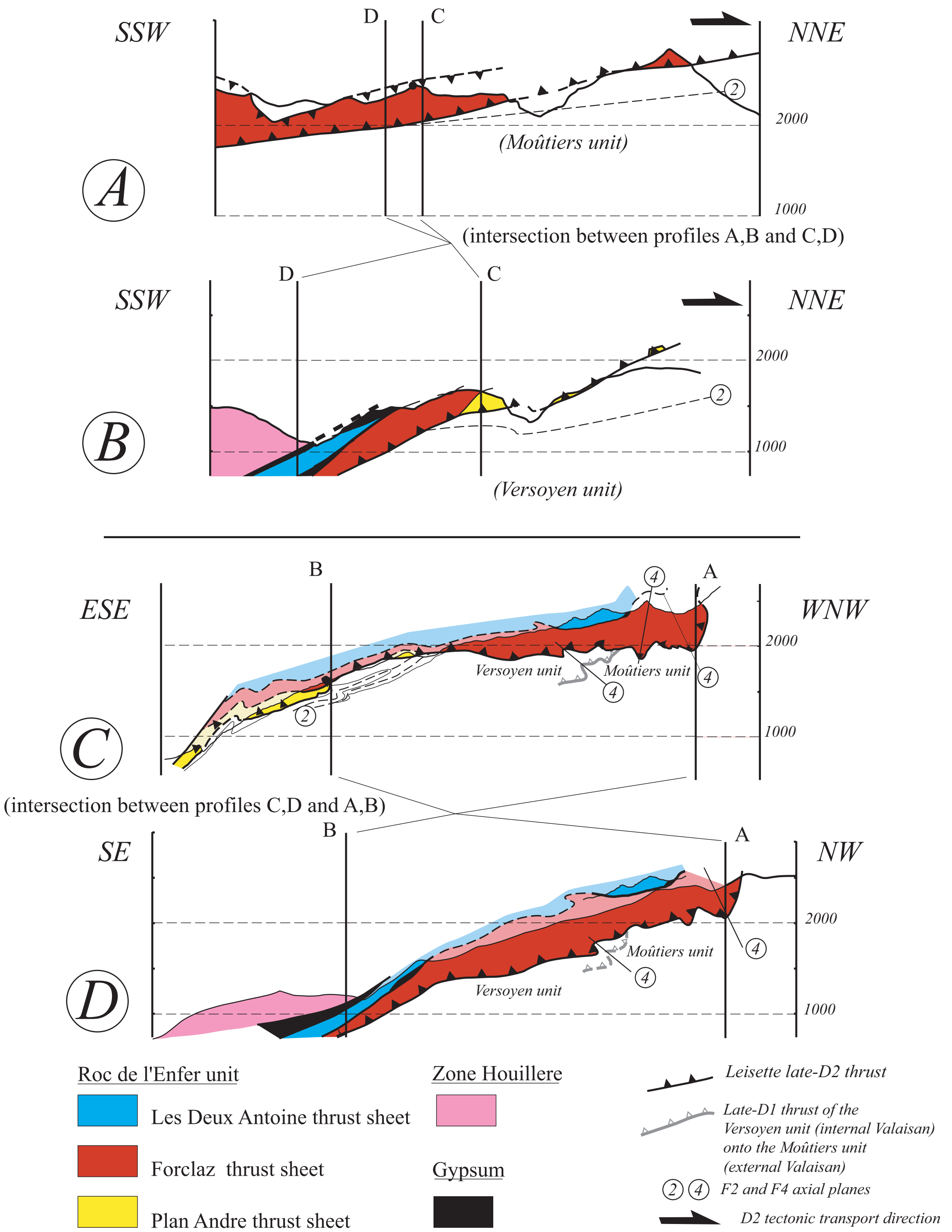


Fig. 6.21 Set of geological profiles showing the Roc de l'Enfer units in the hangingwall of the Leisette late-D2 thrust. Note that profiles A and B are oriented close to the tectonic transport direction during D2 (including early-D2 folding and late-D2 thrusting). They illustrate the stacking of the thrusts sheets forming the Roc de l'Enfer units. Profiles C and D are oriented roughly perpendicular to the F2 and F4 fold axes. They illustrate refolding of the Leisette late-D2 thrust by D4. The same profiles are portrayed in a three dimensional diagram shown in Fig.6.22. The location of the profiles is given in Fig. 6.15.

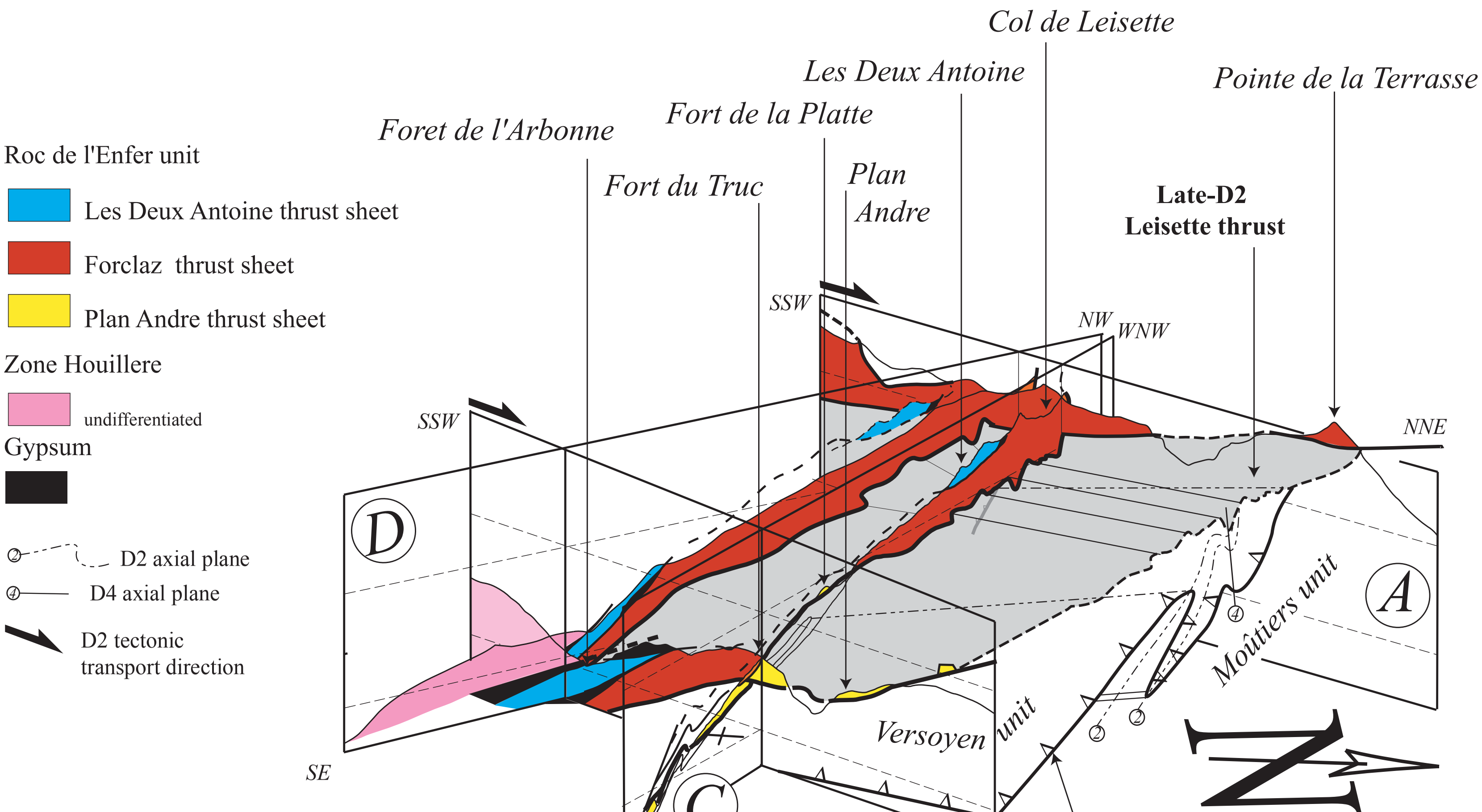


Fig. 6.22. This three-dimensional diagram shows the geometry of the late-D2 Leisette thrust plane (in gray) and the distribution of the thrust sheets forming the Roc de l'Enfer unit. Profiles A, B, C and D are shown in Fig. 6.21. Note truncation the late-D1 thrust contact of the Versoyen unit onto the Moûtiers unit. Two parasitic F2 folds are schematically shown.

The truncation of the D2 axial planes occurs both from SE to NW (profile C, Fig. 6.21) and from N to S, i.e. parallel to the F2 fold axes (profiles A and B, Fig. 6.21, see Fig. 6.14 where the same two F2 parasitic folds are depicted). At a regional scale, these observations indicate that the Leisette thrust is largely responsible for the disappearance of the Versoyen unit towards S, where the Versoyen unit is buried underneath the Roc de l'Enfer unit (Fig. 6.15).

In conclusion the three dimensional analysis, indicates that a late-D2 thrust plane (Leisette and Col des Rousses thrust) is geometrically discordant to all the older planar structures of the D2 nappe stack. Along the profiles parallel to the transport direction, schematically shown in Fig. 6.23a, the late-D2 contact, which is steeper than the F2 axial planes and the older late-D1 contact (see also Fig. 6.22) overprints the D2 nappe stack from S to N and from lower to higher structural levels.

When the same late-D2 thrust plane is observed along the profiles which are perpendicular to the transport direction (Fig. 6.23b), the thrust plane is dipping less steeply in respect to the previous (D1 and D2) planar structures. In this situation, illustrated in Fig. 6.23b, the late-D2 thrust plane takes place at different structural positions within the D2 nappe stack, depending if the profile is drawn S or N of the study area. These observations suggest that the hangingwall of the Leisette thrust is made by formerly lower structural elements which are emplaced onto upper structural elements during late D2 thrusting, the overall shape of these slices being controlled by the intersection between the late-D2 out-of-sequence thrust and the older late-D1 tectonic contact.

The role of the present-day topography is also important. From Bourg St. Maurice up to the Tête de Chargeur area, the map trace of the Col des Rousses thrust is NE-SW oriented. In addition, this particular exposure of the late-D2 thrust, which is oriented roughly subparallel to the transport direction, prevents field observations of three dimensional features. In the area, between Bourg St. Maurice and the Col de Leisette, however, the Leisette late-D2 thrust is three-dimensionally exposed.

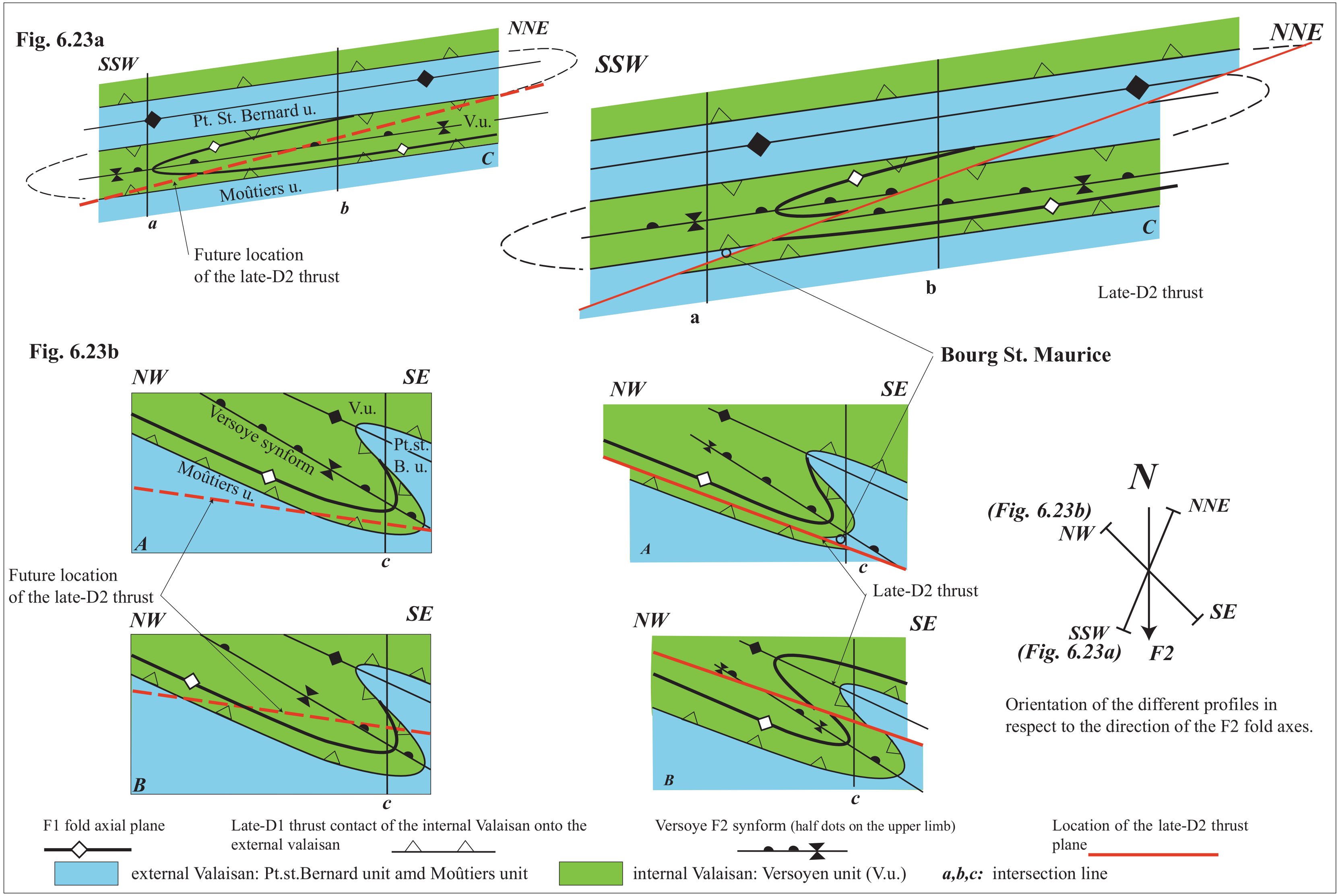


Fig. 6.23 The effect of late-D2 thrusting at the scale of the study area. Fig.6.23a shows schematic profiles parallel to the late-D2 transport direction while Fig. 6.23b illustrates two schematic profiles constructed perpendicular to the late-D2 transport direction. Note that in Fig. 6.23b, profile A is drawn in the S of the study area while profile B is drawn in the N of the study area. Lines indicated by small caps represents intersection between profiles with different orientations

6.3.7: Kinematic analysis of the D2 deformation phase

The kinematic analysis of the second (D2) deformation phase provides evidence for the movement direction inferred for both, the early-D2 stage (folding) and the late-D2 stage (thrusting). The results of this analysis in the Valaisan units are shown in Fig. 6.24. The black arrows indicate the relative movement of the upper block (where the shear sense is known) while the open arrows indicate the dip direction and point towards the outcrop locality. Data come from the lower and the upper limb of the Versoye F2 fold. Shear sense criteria were most frequently observed in the lower limb. The relative transport direction is inferred from kinematic indicators represented by σ-clasts and only rarely by shear bands.

Apart from the σ-clast shown in Fig. 6.18, there are other localities, found in the lower limb of the Versoye F2 synform (Fig. 6.25a and b), where D2 related σ-clasts predominantly indicate NNE-directed movement directions. However, they were found in a region where, again, the effect of the D4 deformation phase is more intense and was probably responsible, as in the case discussed before, for a reorientation of the original D2 transport direction (Fig. 6.24). Not enough data regarding the F4 fold mechanism formation are available to exactly reconstruct the original D2 movement direction. Assuming that D4 deformation phase slightly reoriented the original D2 transport direction towards the strike of D4 folds (i.e. towards NE), the D2 shear sense indicators are inferred to have been originally NNW-SSE oriented. This direction is also more compatible with the large-scale scenario (i.e. sinistral transpression, see discussion in Fügenschuh et al. 1999).

In the upper limb of the Versoye synform σ-clasts were observed mainly within the Brèches du Collet des Rousses, i.e. the syn-rift sediments of the Versoyen unit. In the case illustrated in Fig. 6.25c S1 and S2 are subparallel and together form an average main cleavage in levels of pelitic composition. A calcareous component is internally detached by slip along surfaces close to the shear plane. The intersection line between the early main foliation and the shear plane is generally sub perpendicular to the stretching lineation of the rock. The shear sense is then deduced from the relative movement of the single parts of the clast (top-to-the NNE).

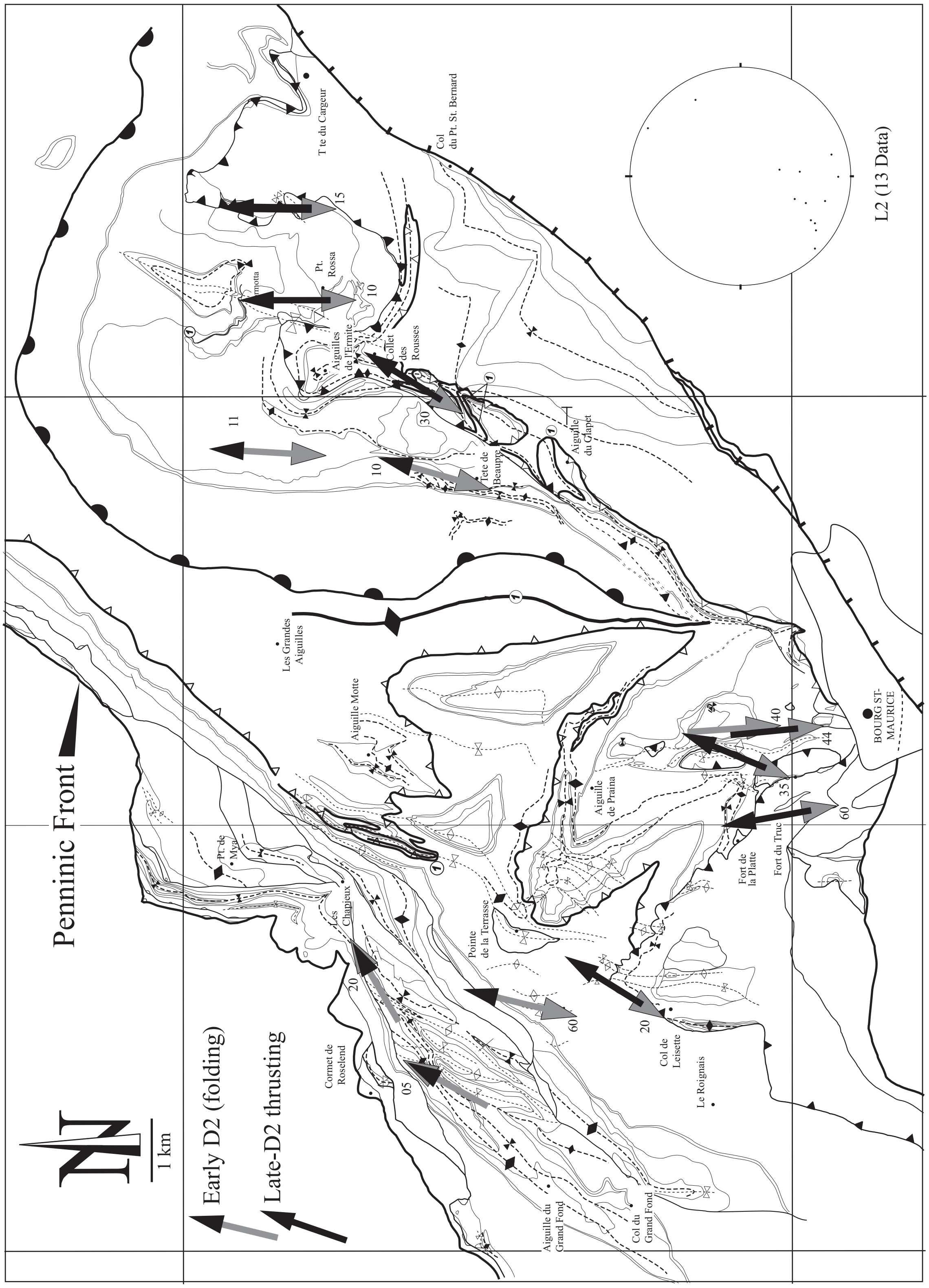


Fig. 6.24. Map of the transport direction during D2. Grey arrows point to the sample locality and indicate the dip direction, while black arrows indicate, where available, the sense of movement

Within the structural framework of the sample this shear plane is subparallel to the locally observed S2 foliation and therefore the inferred transport direction can be ascribed to the early-D2 deformation phase.

Shear bands were observed only in a few cases (e.g. Fig. 6.25d). The shear band foliation deflects the main foliation that is formed by S1 and S2. suggesting that the shear band developed during the final stage of the D2 deformation phase and can therefore be ascribed to this phase.

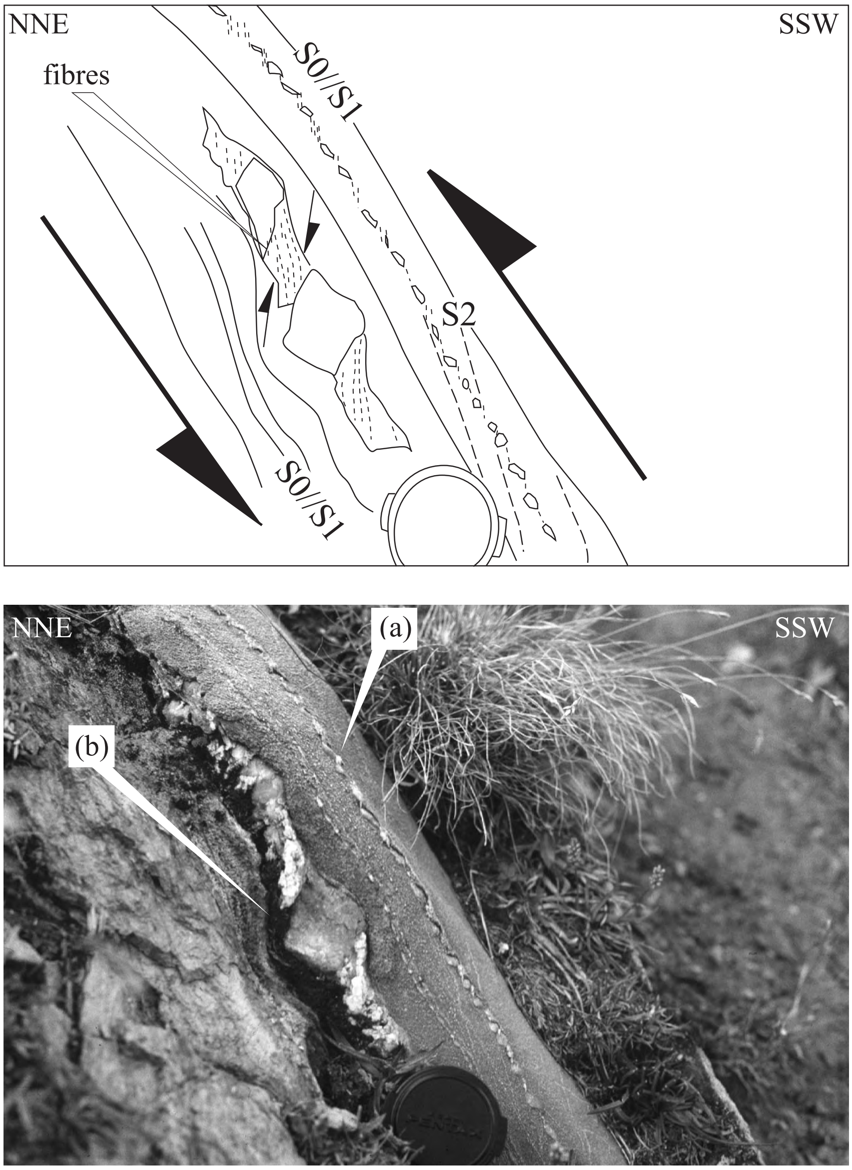


Fig. 6.25a Shear sense indicators in the C. de St. Cristophe formation of the Moûtiers unit. Two more competent levels (a) and (b) are broken and subjected to shear. An S2 cleavage has been observed sub-parallel to the shear plane. The individual parts display a shear motion which is opposite of the bulk shear sense according to a bookshelf sliding structure (Ramsay and Huber 1987). The inferred sense of shear, therefore, related to the D2 deformation phase, is top to the NNE. This D2 situation has been overprinted during D4, leading to the following present-day orientation: S0,1 56/105, S2 80/094, fibres 194/60.

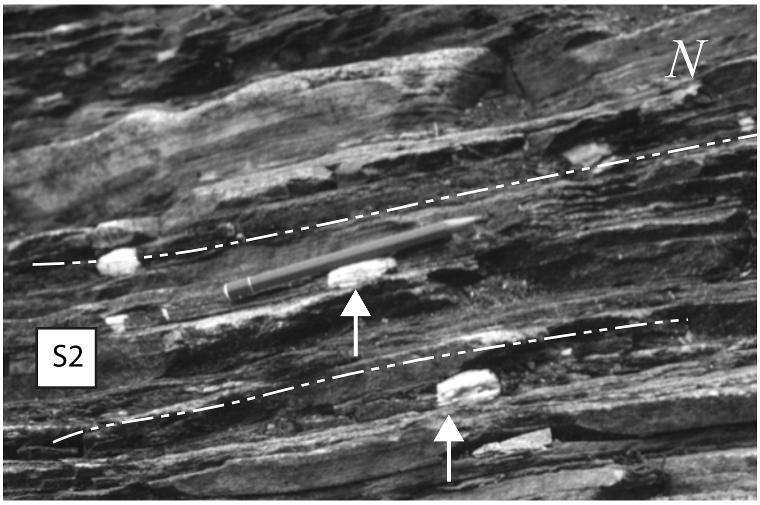
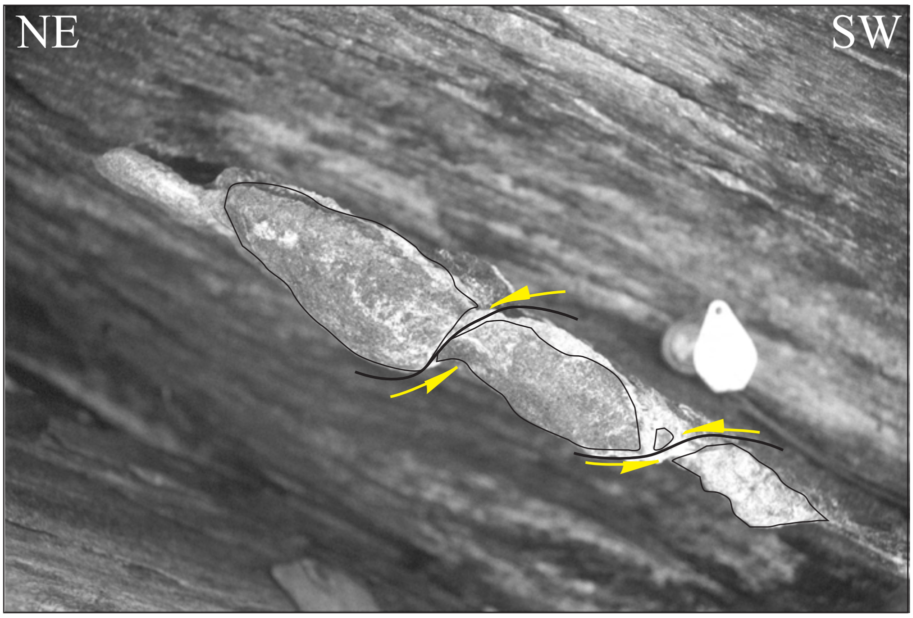
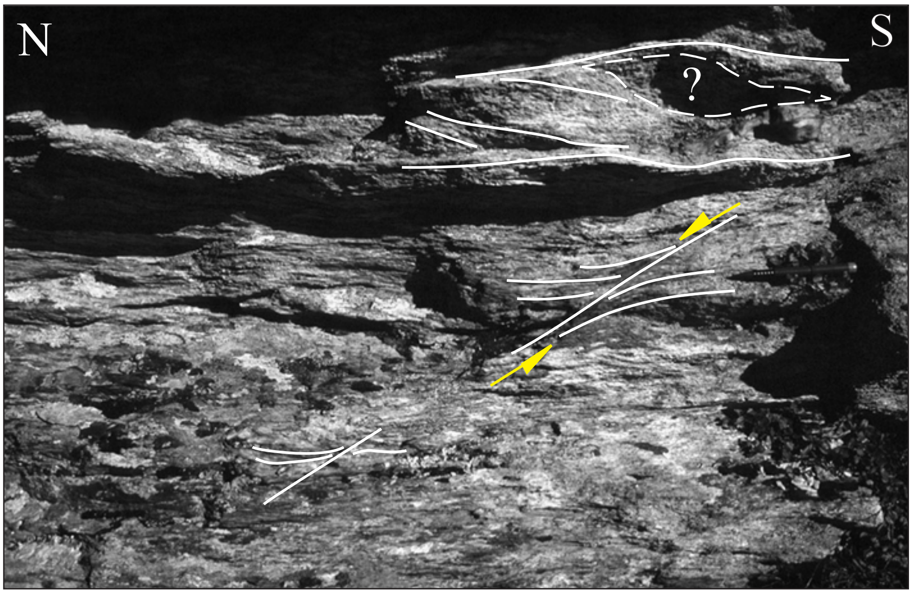


Fig. 6.25 b. The plane of photograph is oriented roughly perpendicular to the second phase (D2) axial planes cleavage (S2) but sub-parallel to the F2 fold axes. The rock, consisting in the C. de St. Christophe lithology, shows alternating calcareous and pelitic layers resulting from isclinal D2 folding. The pelitic strata (Fig. 6.25b, above) contain block shaped extensional fibers (open arrows) oriented sub-parallel to the F2 fold axes and genetically related to the late stage of the D2 deformation phase. During ongoing of the deformation these extensional fibres are occasionally overprinted by shear deformation, revealing information about the transport direction during the D2 deformation phase (Fig. 6.25b, below). The inferred transport direction (late-D2) is top-to-the ENE.(coordinates: 941450 x, 2086675 y, 1700 z)



6.25c



6.25d

Fig. 6.25c Bréche du Collet des Rousses Formation (Versoyen unit), containing stretched component. Inferred sense of movement of the individual parts is top to the NE. Away from the broken clast (looking at the less competent matrix) the shear planes become strongly subparallel to the main foliation, which is the result of the superposition of S0, S1 and S2 cleavages.
Fig. 6.25d Shear bands in the Pt. Rossa crystalline massif (Versoyen unit). The inferred sense of shear is top to the N. In both examples (c and d) it is not possible to determine whether the main foliation, which is deflected by shear planes (Fig. 6.25c) and shear bands (Fig. 6.24d), is S1 or S2. Therefore it is not possible to unambiguously assign the observed shear criteria to either D1 or D2. However shear bands are generally considered to develop during the final stages of high strain deformation (Simpson and Schmid 1983). It is therefore proposed that top to the N-directed shearing is related to the final stage of D2 deformation phase rather than D1.

The map of the D2 transport directions in Fig. 6.24 shows that the shear sense criteria exhibit a similar transport direction (roughly top-to-the N) in both limbs of the F2 Versoye synform. In the Versoye upper limb, they indicate predominantly N to NNE transport directions, while in the lower limb near the Pennine Front some of them are more strongly deflected towards NE, due to F4-overprint. The original transport direction (top NNW to NW, see earlier discussion) is only slightly reoriented N of Bourg St. Maurice (locality "Fort du Truc"). In Fig. 6.24 the overall consistency of the shear directions in both limbs of the Versoye fold supports the interpretation that these shear sense are associated with D2 deformation and not with D1. Reorientation of the D1 movement direction by F2 would result in a reversal of the shear sense because the F2 fold axes are parallel to earlier L1-lineations.

Roughly top to the N directed transport during d2 has been observed in all Valaisan units. Top-N directed transport movements may be interpreted in the light of sinistral transpression. L1 stretching lineations also strike roughly N-S (compare the pole Figures of Fig. 6.3, 6.13 and 6.24), Top to the N directed transport is proposed for D1 as well, based on the similarity between D1 and D2.

6.4: The D3 deformation phase

The D3 deformation phase was first analysed in detail by Bucher (1999) who also provided the structural data discussed here. Evidence for the D3 deformation phase is observed only in the NE part of the study area situated in Italy, i.e. between the Tête du Chargeur up to the Colle St. Carlo, N of the La Thuille village (Fig. 6.26). D3 deformation is characterised by F3 folds (Bucher, 1999) genetically related to a greenschist-facies mylonitic shear zone indicating top-to-ESE directed normal faulting (Bucher, 1999).

This normal fault, referred to as the Pont Serrand normal fault zone, overprints D2 structures and a former thrust at the base of the Zone Houillère. Kinematic data are provided by lineations observed within the mylonitic foliation and by shear sense criteria formed by σ-clasts and shear bands, mainly observed within post rift sediments of the Versoyen unit (Fig. 6.27).

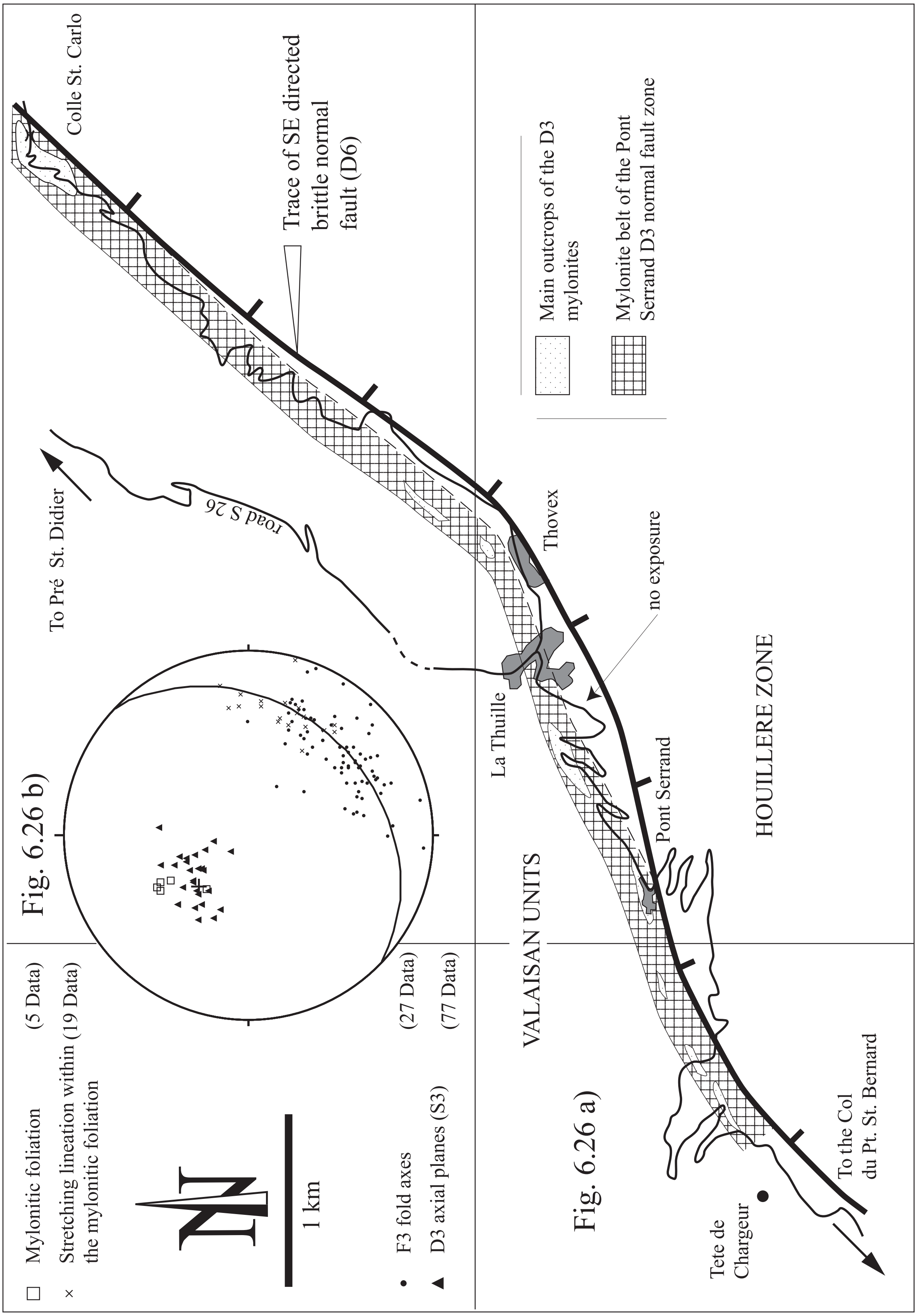
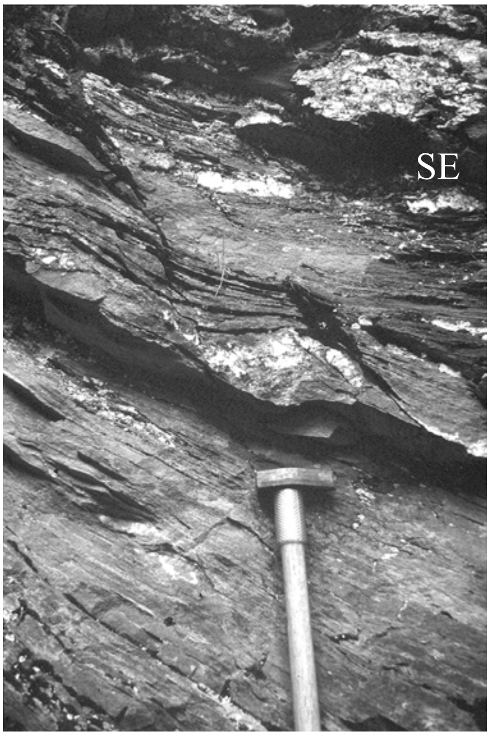


Fig. 6.26 a) and b). Fig. 6.26 a) Map of the Pont Serrand D3 normal fault zone and 6.26 b) Pole figure (for discussion see text,section 6. 4)



A)

B)

Fig. 6.27. Shear bands (a) and shear sense indicators (a) and (b) in the C. de l'Aroley at the Colle St. Carlo (I). Inferred sense of shear is top to SE for a) and top to the E for b).

D3 folds are characterised by F3 fold axes which show a maximum concentration around N140 and gently SE-dipping axial planes which are subparallel to the S3 mylonitic foliation (20°-40° toward SE). D3 and D2 folds are roughly coaxial and the interference pattern is of "Type 3" (Ramsay 1967).

D3 folds phase are commonly related to open folds while D2 folds are tighter and the associated S2 axial-plane cleavage is more intense compared to S3. Overprinting relationships as discovered by Bucher (1999) are shown in (Fig. 6.28).

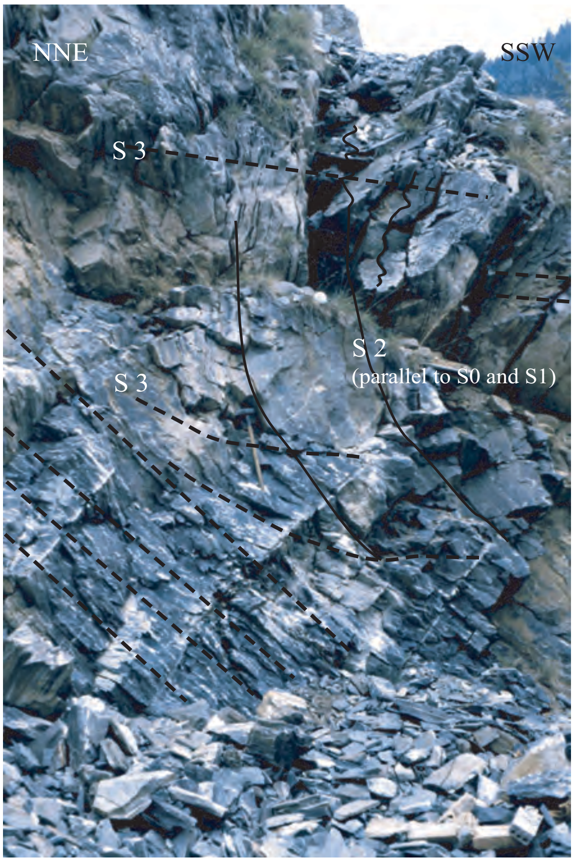
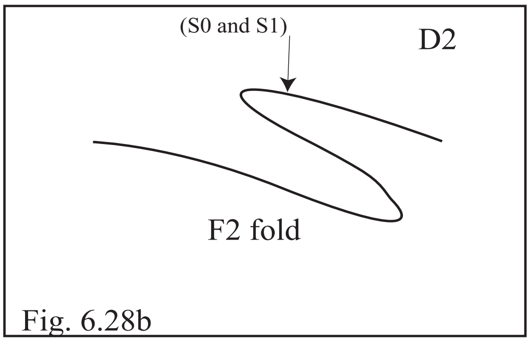


Fig. 6.28a

D2



D3

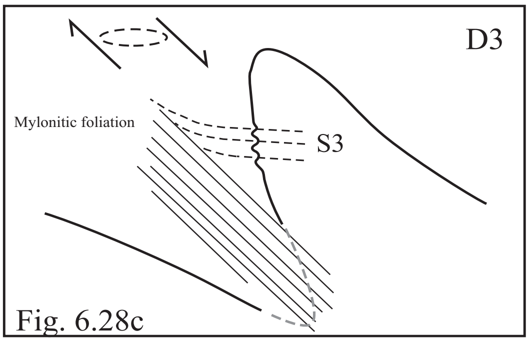


Fig. 6.28 a) Relationships between the D3 Pont Serrand normal fault zone (lower left) and the D3 folds (upper rigth). The older, now subvertical, main foliation (S0 and S1) represents the short limb of a previously formed F2 structure (Fig. 6.28b). During D3 (Fig. 6.28c) the mylonitic foliation is formed and the F2 folds are overprinted and partially re-opened. F3 folds are associated with a subhorizontal axial plane cleavage (S3). Toward the shear zone the S3 cleavage merges into parallelism with the mylonitic foliation, inferring a dextral sense of movement (top-to-the SE).(modified after Bucher 1999)

In Fig. 6.28 the pre-existing steep limb of a parasitic F2 fold (roughly NW-SE oriented) is cut by the Pont Serrand shear zone with a top-to-SE sense of movement. The subvertical F2 limb, oriented perpendicular to the S3 schistosity, is vertically shortened, leading to the formation of open D3 folds: The orientation of F3 fold axes is thus strongly controlled by the initial strike of the layers. The newly developed foliation (S3) becomes subparallel to the mylonitic foliation towards the shear zone. Deflection of the S3 foliation provides a shear sense criterium in agreement with the shear sense deduced within the mylonites.

On a larger scale, similar relationships between D3 folds and the Pont Serrand normal fault zone as observed in Fig. 6.28, can be deduced from the analysis of the pole Figure of Fig. 6.26b. This pole Figure combines F3 axial planes and F3 fold axes together with the pole Figure of the mylonitic foliation and the stretching lineation observed within the mylonites of the Pont Serrand normal fault zone. Fig. 6.26b clearly shows that the mylonitic foliation coincides with the F3 axial planes. The trend of the F3 fold axes shows a maximum value towards N140 and their dispersion within a great circle can be interpreted as an effect inherited from the original strike of the layers.

On the basis of these observations, the D3 folding is interpreted to have been caused by vertical shortening associated with SE directed extension, according to the scheme discussed in (Froitzheim, 1992). The significance of this phase will be discussed on regional scale in section 6.8. The age of the third deformation phase can be estimated from the available radiometric age data. Ar/Ar age data on phengite are available for the Valaisan (Versoyen and Pt. St. Bernard units) (Cannic et al., 1996) forming the footwall to this ductile normal fault in the Pt. St. Bernard pass region.

The samples yielding “good” and “reasonable” plateaus (Cannic, 1996, p.129) indicate consistent ages of about 34 Ma. These ages were interpreted by Cannic (1996) as cooling ages. Freeman (Freeman et al., 1998) performed Rb-Sr dating of white micas from sheared rocks of the Pt. St. Bernard unit immediately below the Zone Houillère, forming the hangingwall of this normal fault. These authors interpreted Rb-Sr white mica ages of about 33 Ma (between 27.1 and 32.9) in terms of formation ages related to NW-directed thrusting of the Zone Houillère over the Valaisan.

However, according to Bucher (Bucher, 1999) the white mica ages of Freeman (Freeman et al., 1998) probably formed during top-to-the SE directed normal faulting (D3). Since the formation ages of Freeman (Freeman et al., 1998) are compatible with coeval cooling in the footwall, as indicated by the data of Cannic (1996), we suggest that this normal fault was active during a time span from about 34 to 27 Ma (early Oligocene).

6.5: The D4 deformation phase

6.5.1: General characteristics regarding D4

D4 structures developed obliquely to the trend of the earlier structures (pre-D1 to late-D2) and to the different tectonic units of the Valaisan zone. Prominent large scale D4 structures were already introduced in section 5.2 (Figs. 5.5 and 5.7). Fig. 6.29 gives an example (from the Moûtiers unit) of the superposition of the D4 deformation phase onto older structures.

Fourth phase folds (F4) consistently strike NE-SW (Fig. 6.30) leading to Type-3 interference patterns (Ramsay, 1967) with earlier structures (pre-D1 to late-D2).Around the Pt. Rossa crystalline massif (Fig. 6.30), in the NE of the study area the strike of fourth phase folds changes into a ENE-WSW to E-W orientation (Dalla Torre, 1998) with variable plunge. Since the F2 fold axes are N-S oriented (Fig. 6.13) a dome and basin like interference pattern can be observed (Type-1 according to Ramsay 1967). At a regional scale, clear examples of Type 1 fold interference patterns are confined to this region. The different orientation of F4 in this area is probably due to local effects controlled by the competence contrast between the Pt. Rossa crystalline massif and the surrounding basaltic volcanic rocks and related sediments of the Complexe Antéflysch.

A crenulation cleavage and a new axial planar cleavage (S4) generally oriented at a high angle to the previous planar structures (Fig.6.31), is associated with fourth phase folds (F4).

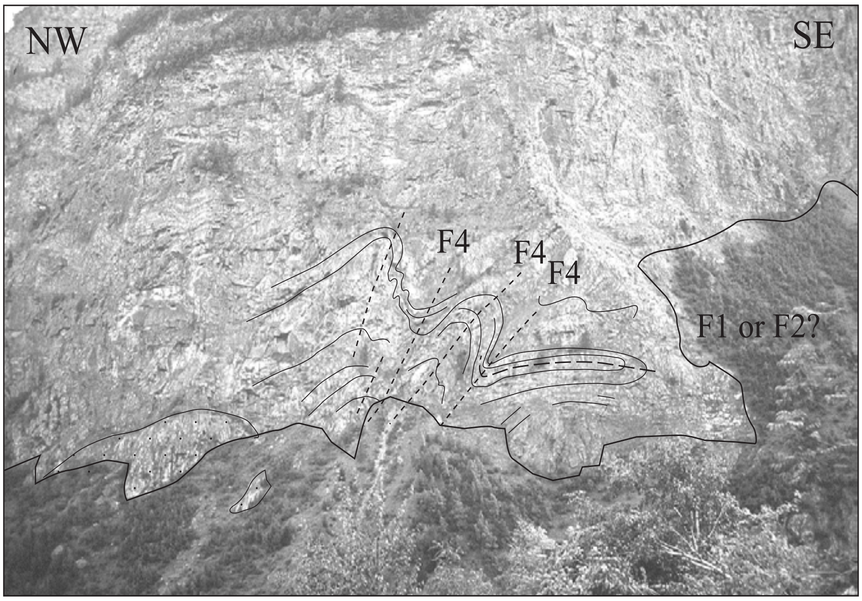


Fig. 6.29 Large-scale F4 overprinting of a previous isoclinal fold (F1 or F2?) within the post rift sediments of the Moutiers unit. The well bedded lithology are C. de l'Aroley while in the lower left Lower Triassic quarzite is indicated with dots. The photo is taken from the road between Bourg St. Maurice and Les Chapieux (x: 945375, y: 2084150, z: 1380), a few kilometers after Les Glinettes.

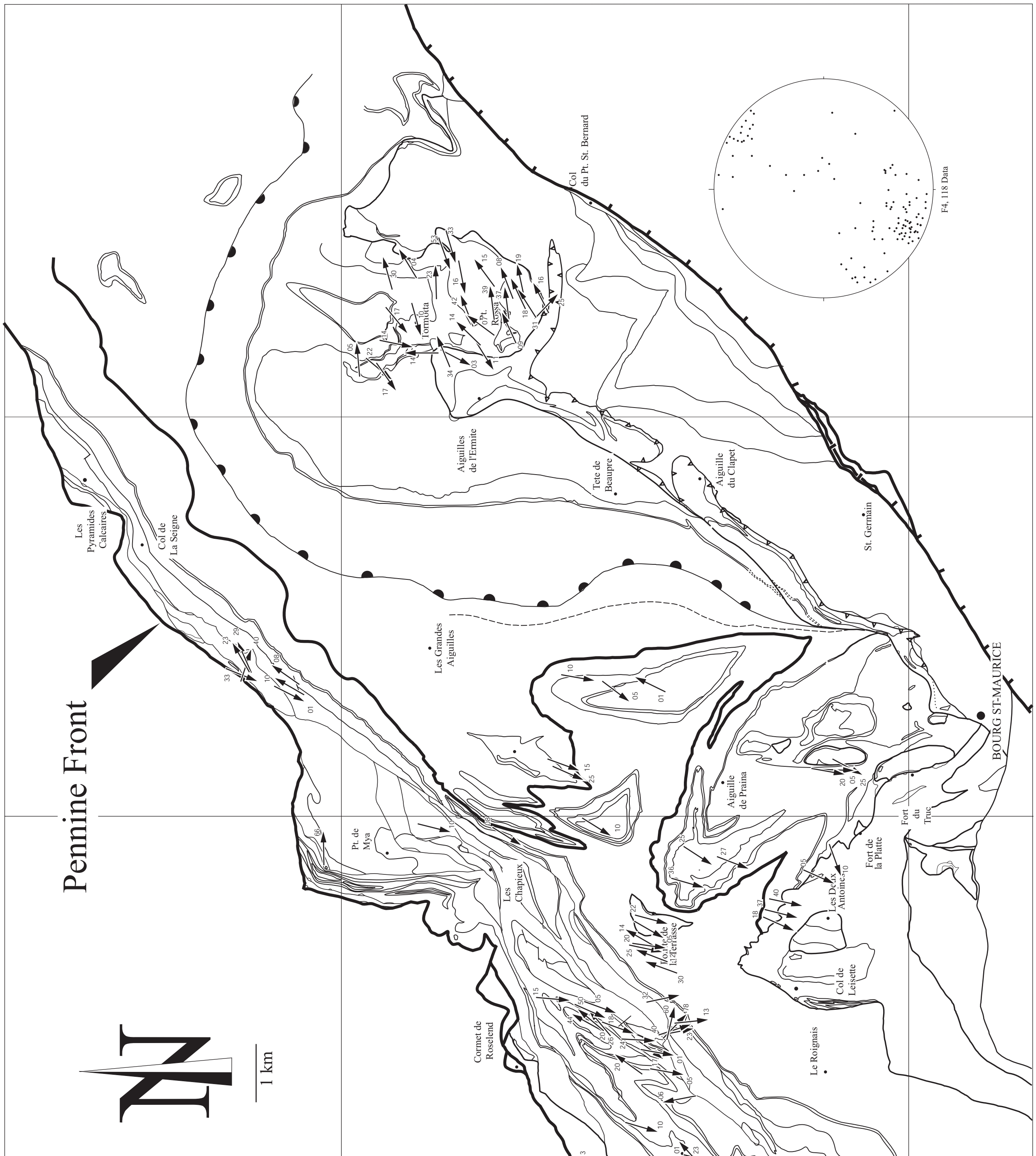


Fig. 6.30 Representative mesurements of the D4 fold axes within the Valaisan unit and related pole figure.

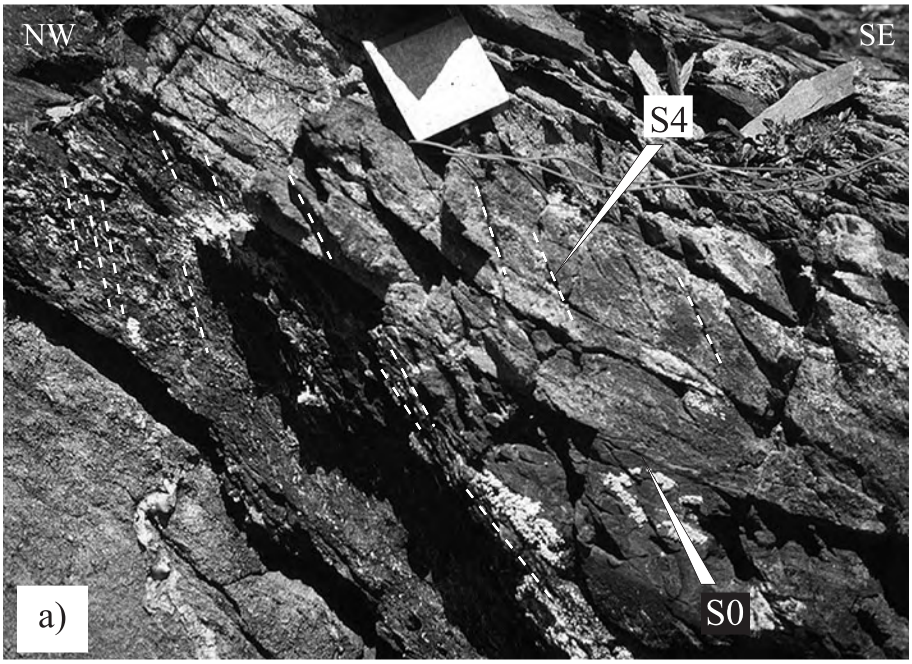


Fig. 6.31 Example of S4 cleavage in the C. des Marmontains of the Moûtiers unit (external Valaisan) and (b) in the sediments of the Complexe Antéflysch of the Versoyen unit (internal Valaisan). In competent layers (upper level in Fig. 6.31a) the S4 cleavage is well spaced (2-4 cm) while, in less competent levels (b) it is more penetrative (< 1 cm). The "spacing" and, obviously, the steep orientation are diagnostic for S4.

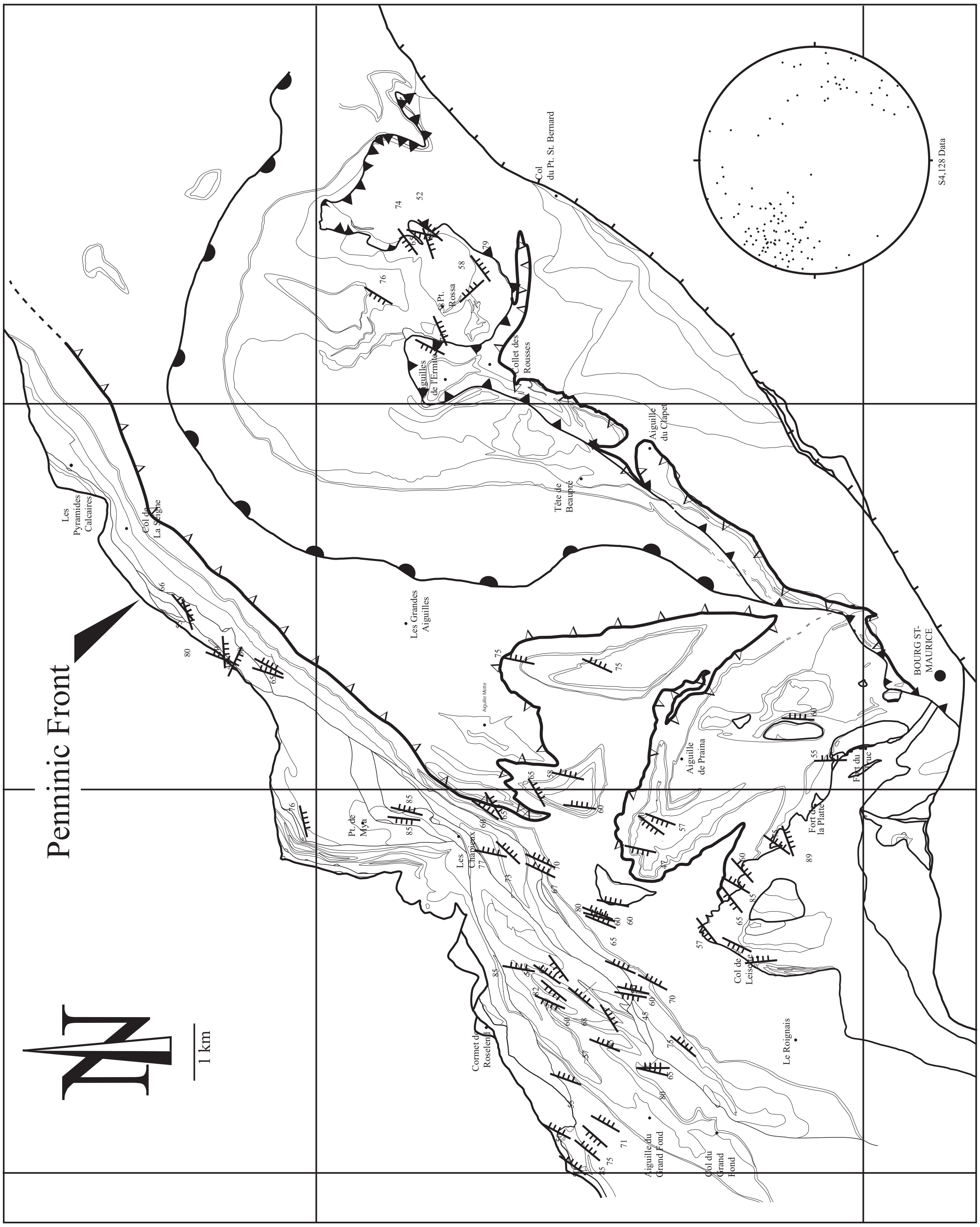


Fig. 6.32. Representative mesurements of the S4 foliation within the Valaisan units and related pole figure

The S4 cleavage generally strikes NE-SW (Fig. 6.32) and generally dips steeply (about 60°) towards SE, or, in some cases, toward NW.

Close to the Pennine Front, the S4 foliation, merges into parallelism with the previous planar structures (S0,S1 and S2), uniformly dipping toward SE. The fourth phase structures can be traced over a large part of the study area (Figs. 6.30 and 6.32), but they do not affect all the Valaisan units homogeneously. The fourth deformation phase led to the formation of folds, which display a gradient of increasing strain toward NW (Chapter 6.5.2), i.e. toward the Pennine Front which, in the study area, corresponds to a top-to-the NW directed late-D4 thrust (Chapter 6.5.3).

6.5.2: D4 Strain gradient

Changes in the character of D4 structures will now be discussed by using different thin sections and outcrop scale examples situated along a NW-SE oriented profile (Fig. 6.33). These examples will be discussed with respect to their distance from the Pennine Front and in terms of their vertical position within the nappe stack of the Valaisan units (vertical segments). The structural setting of the observations presented in Fig. 6.33 coincides with the profile of Fig. 5.6b. Each inset is separately presented in Figs. 6.33a to e.

In the SE part of the study area and within the Pt. St. Bernard unit D4 is only weakly expressed. D4 structural elements at the meso- and large scale were not observed with one exception: the Pt. Rossa area. Moving north-westward, from Bourg St. Maurice towards the Cormet de Roselend, the first large scale D4 structure consists in a NW verging D4 antiform. The axial plane of this structure can be traced across the entire nappe pile of Valaisan units from the bottom of the Valleé des Glaciers up to the Aig. De Praina. The lowest structural levels of this D4 structure (Fig. 6.33a) displaying a weak crenulation cleavage (S4), steeply dipping towards SE, displayed in micaceous layers of Lower Trias rocks only.

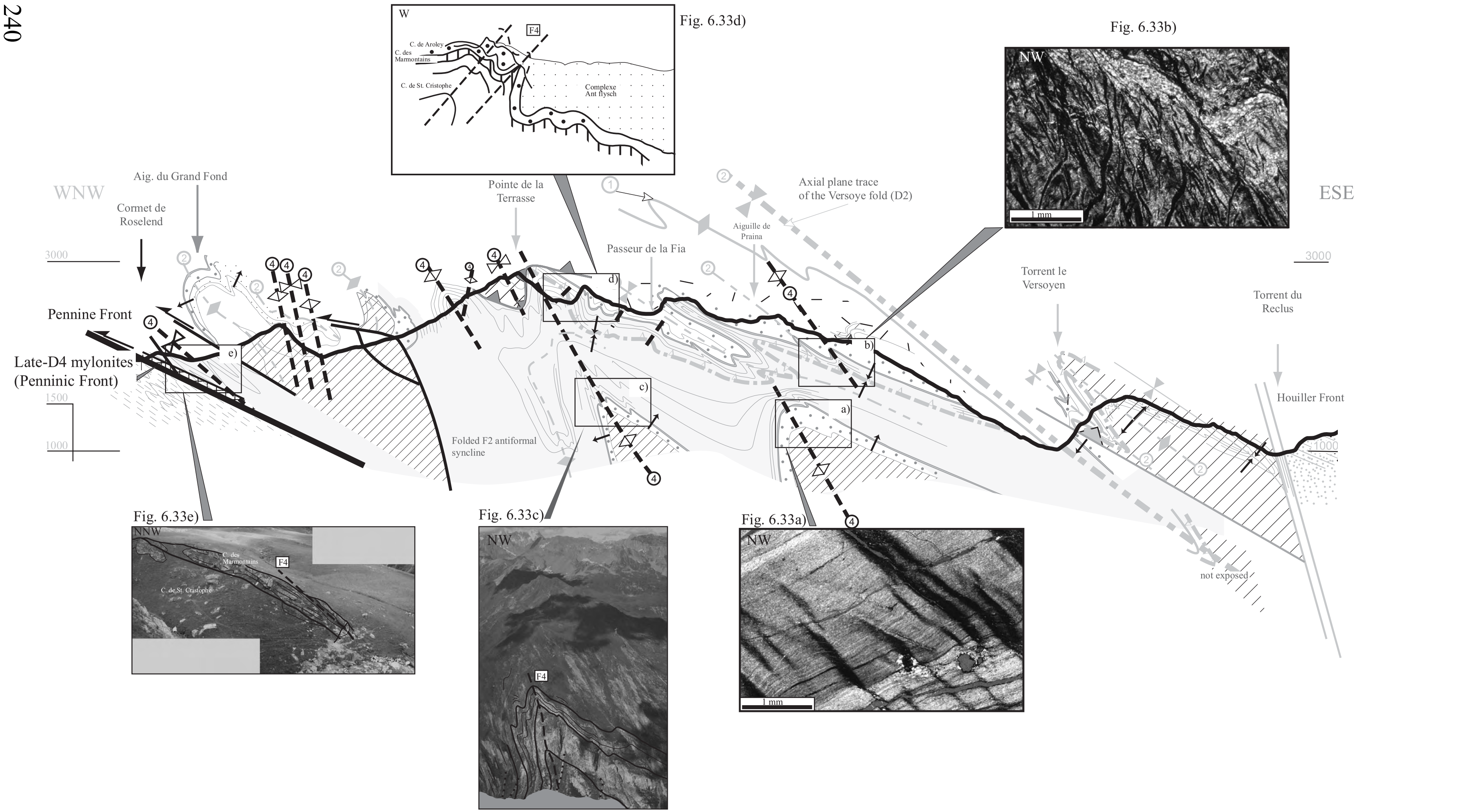


Fig. 6.33 From ESE to WNW and toward the Penninic Front, the D4 structural elements display a strain gradient which is illustrated by thin sections and field examples show in insets a) to e). The outcrops have been observed close to the profile of Fig. 5.6b. In the ESE, inset a), a weak S4 crenulation cleavage, dipping steeply toward SE, is observed. At the same distance from the Penninic Front but in higher structural levels (inset b) the S4 cleavage becomes more evident, dipping predominantly toward SE or occasionally toward NW. In this case, SE and NW dipping S4 cleavages are considered to be coeval. In this particular area large-scale F4 folds exhibit only SE-dipping axial planes. Toward the Penninic Front and in the central region of the profile, F4 folds in the lower structural levels becomes tighter (compare the hinge regions of the two D4 antiforms displayed near insets a) and c). In higher structural levels (inset d), box-like F4 folds with conjugate SE- and NW-dipping axial planes become observable as large-scale structures, responsible for the steepening of large parts of the D1 and D2 nappe stack. Finally, close to the Penninic Front (inset e), F4 folds show an isoclinal style and the S4 cleavage, dipping uniquely toward SE, trends into sub parallelism with this large fault zone. For the localization of inset e) see the text (section 6.5.1). The Penninic Front is interpreted as a late-D4 thrust (see section 6.5.2).

Up-section, the SE dipping S4 crenulation cleavage becomes more intense and a discrete coeval NW-dipping cleavage is also observed (Fig. 6.33, inset b).

Further to the NW, the next large-scale F4 antiform near inset c (Fig. 6.33) is tighter already indicating possible increase in deformation intensity. In the lower structural levels, F4 folds show tight and well defined hinge zones related to a SE dipping S4 cleavage (Fig. 6.33, inset c). Higher in the section within the same D4 structure, the F4 folds start to display a box-like folding geometry with both SE and NW dipping axial planes (Fig. 6.33 inset d).

This gives rise to broader and more rounded hinge zones. Approaching the Pennine Front, the style of the D4 structural features changes dramatically and F4 folds are hardly discernible in style from older folds. This is due to the fact that, as for example in the Chapieux area (Fig. 6.33 inset e), F4 folds become nearly isoclinal.

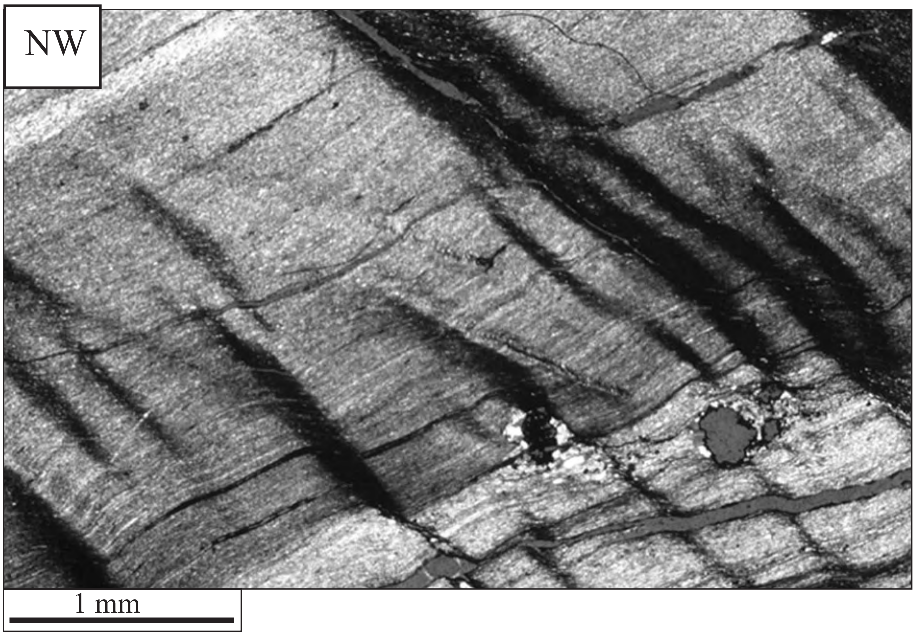


Fig. 6.33a Thin section of spaced and weak crenulation cleavage (S4) in Lower Triassic quartzite, Moûtiers unit, External Valaisan.

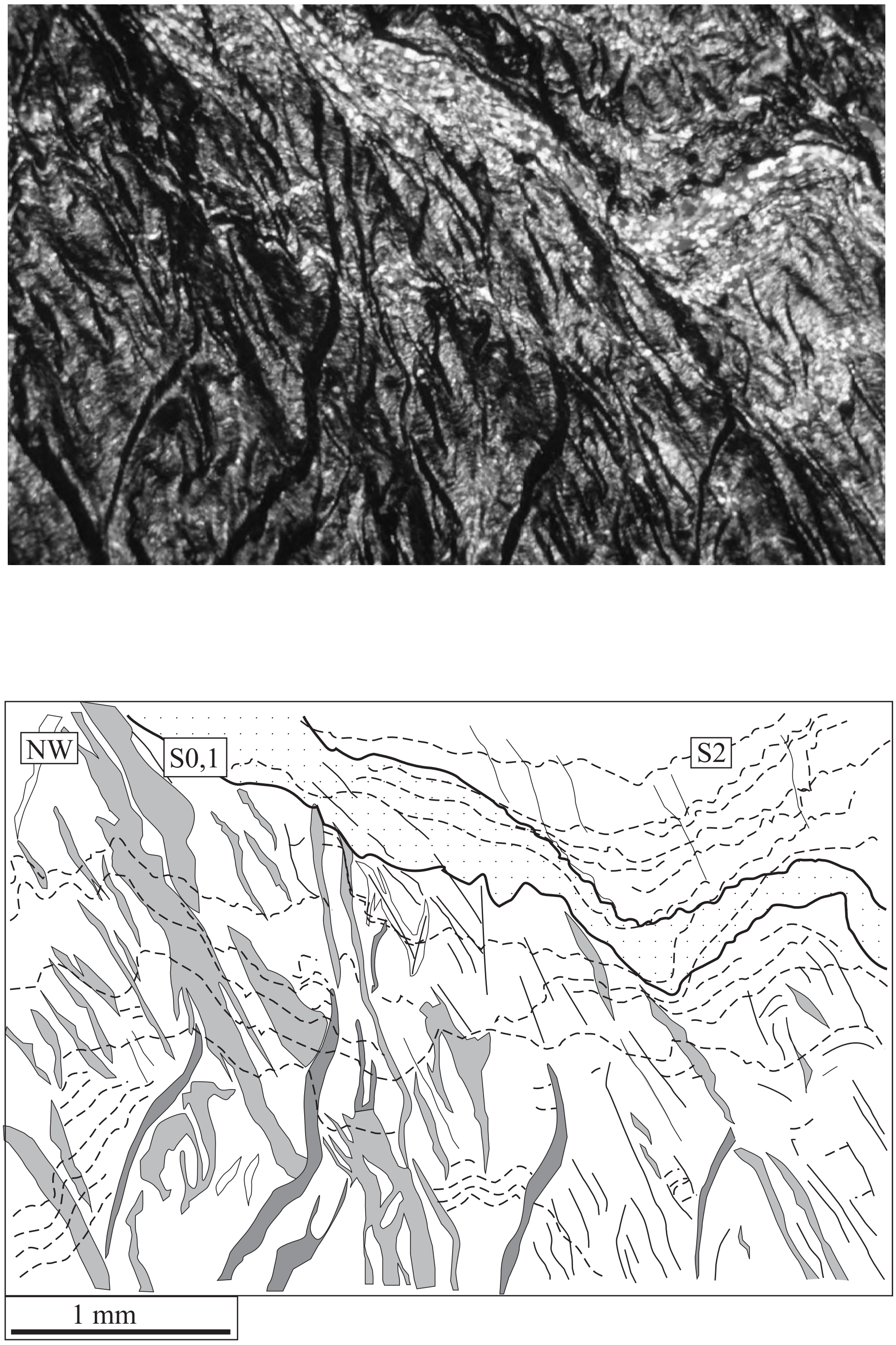


Fig. 6.33b) Thin section and line drawing of black schists from the Complexe antéflysch, Versoyen unit (Internal Valaisan), showing the relationships between different sets of cleavages. Dark gray indicates NW dipping S4 crenulation cleavage domains; ligth gray corresponds to SE dipping S4 crenulation cleavage domains. Both, NW and SE dipping S4 cleavages are interpreted to be coeval.

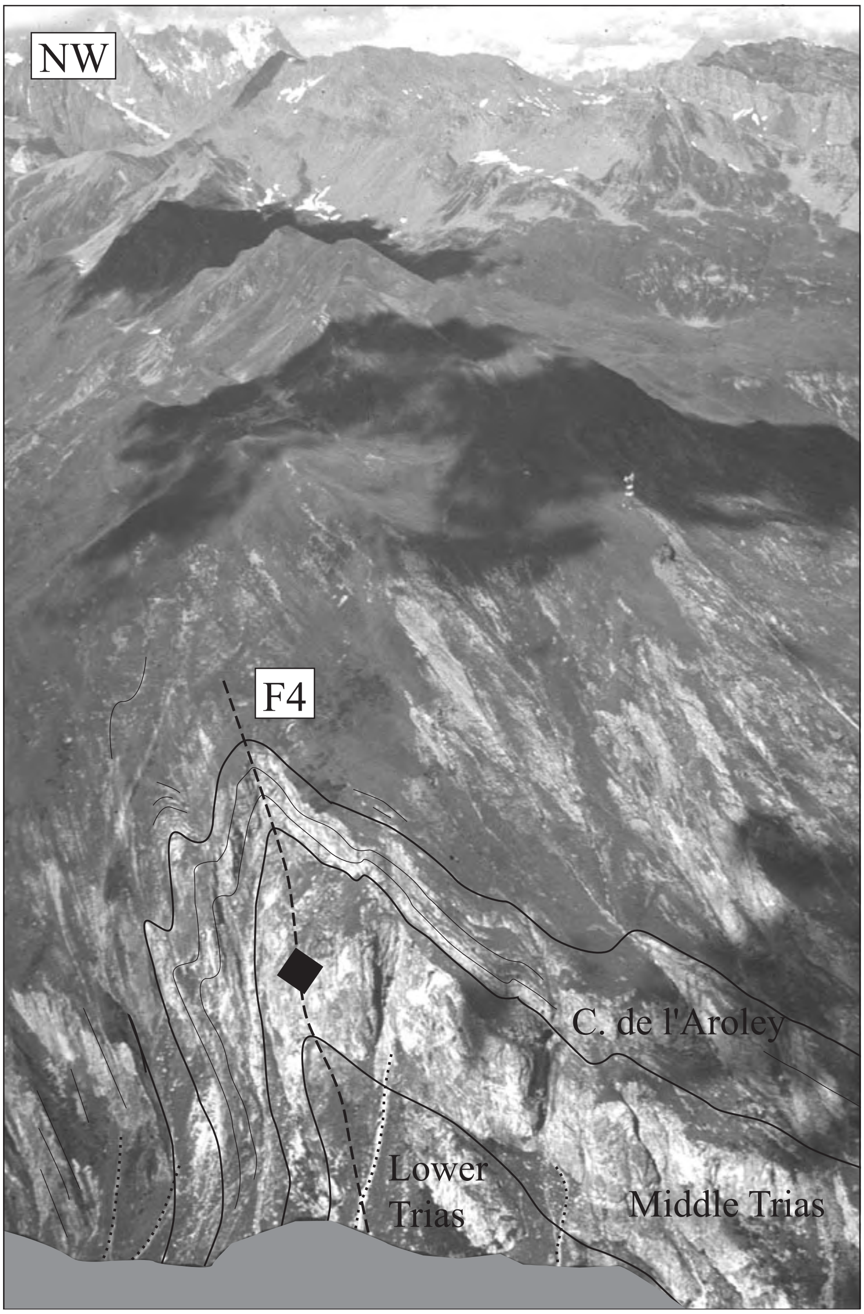


Fig. 6.33c Large scale example of F4 similar fold in the Moûtiers unit (external Valaisan). This structure, characterised by an unique SE dipping axial plane (S4) shows a sharp hinge region. Indicated, are well bedded layers of the C. de l'Aroley formation. Note that the core of this fold is formed by Lower Triassic quartzite followed by Middle Triassic dolomite.

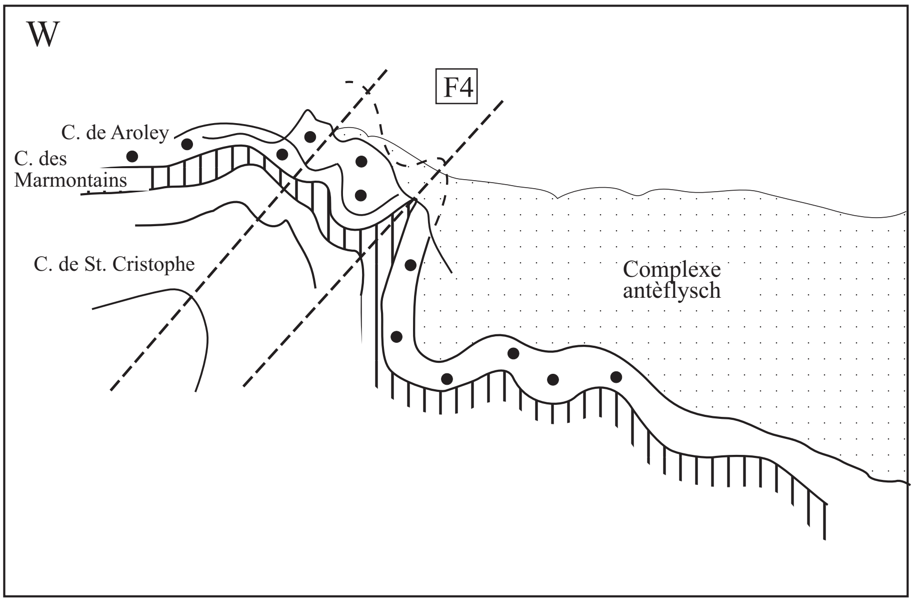


Fig. 6.33d). Large scale F4 folds within the Versoyen unit showing NW dipping axial planes (S4, Passeur de la Fia). At larger scale (see Fig. 6.32) also SE dipping axial planes are found in this area. Both D4 structures, SE and NW dipping axial plane folds, result in a box-fold like geometry.

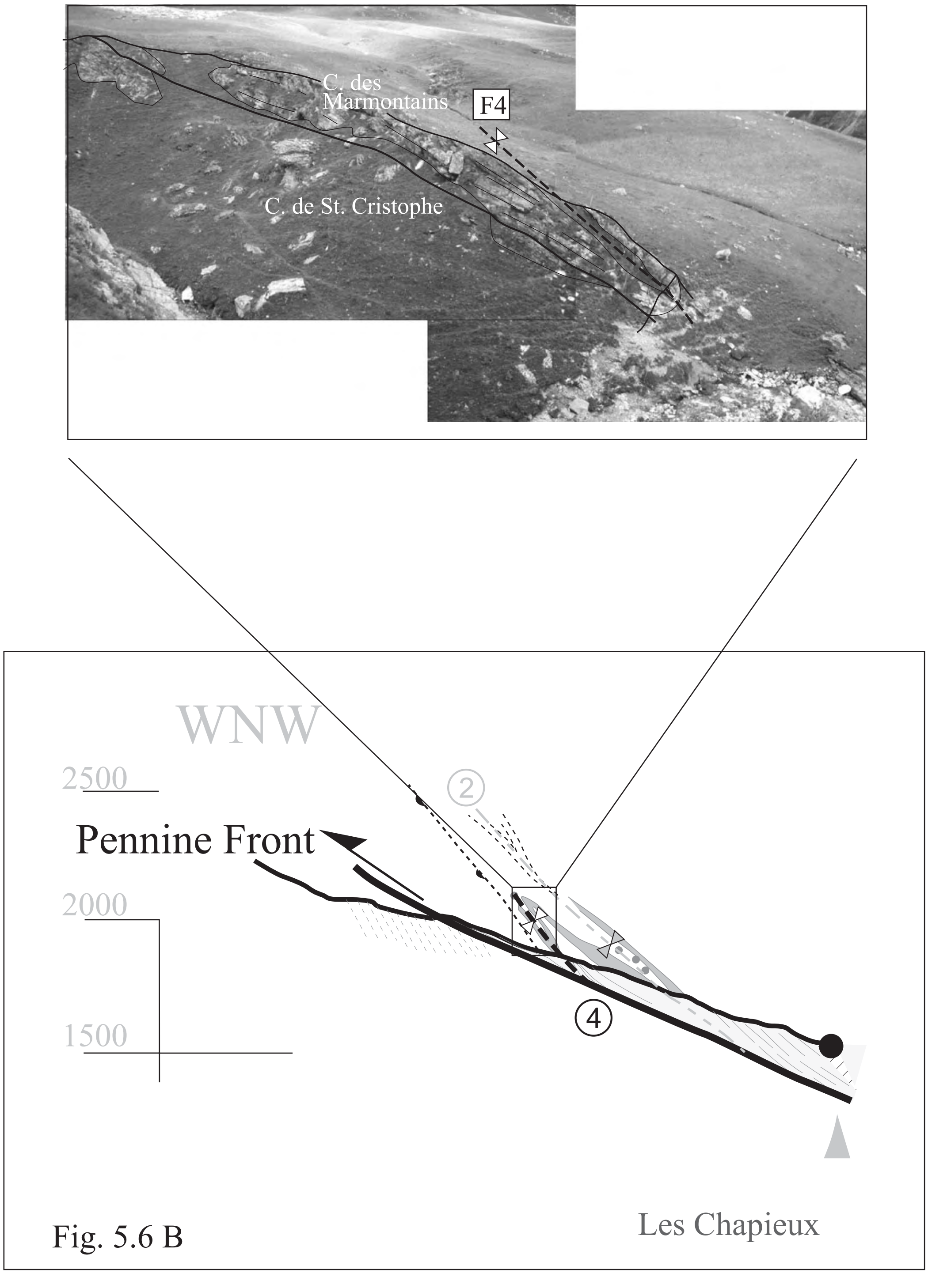


Fig. 6.33e). Above, photograph and line drawing of a tight to isoclinal F4 folds close to the Penninic Front (Moûtiers unit). The inset below shows the local structural context of this example (see Fig. 5.6B). This D4 structure is not located directly on the profile of Fig. 6.31. However, the D4 fold style is the same. In this area, F4 folds are characterised by a consistently SE-dipping axial plane (S4) which flattens and asymptotically approaches the orientation of the Penninic Front.

6.5.3: Late-D4 thrusting along the Pennine Front: The Roselend thrust

The present study covers only a small segment of the Pennine Front which can be traced all along the western Alps (Ceriani et al., 2000 for a discussion). Geometry, kinematics framework and the relative timing of the different deformation phases along the Pennine Front has been studied in the area between the Pyramides Calcaires in the NE and the Cormet de Roselend in the SW (Fig. 5.1).

Since the Pennine Front represents a complex fault zone, the results from this particular segment cannot be simply extrapolated and applied to others parts of the “Pennine Front” (Ceriani et al., 2000). For these reasons we use the term “Roselend thrust” in order to denote the Pennine Front as exposed in our working area.
South of Moûtiers, the continuation of the Roselend thrust does no more coincide with the limit between Pennine and Dauphinois units (Ceriani et al., 2000). However, in the area between the Pyramides Calcaires and the Cormet de Roselend pass area, the Roselend thrust is in fact identical with the Pennine Front in its classical sense, i.e. the Roselend thrust defines a WNW-directed thrust (Fig.6.34) which separates the Dauphinois from the Valais paleogeographic domains (Fig. 6.35). Hence, the terms *Pennine Front* and *Roselend thrust* will be used synonymously.

The Pennine Front, defining the tectonic limit between two different paleogeographic domains, dips down to a great depth as imaged by the ECORS-CROP seismic profile (Nicolas et al., 1990; Schmid and Kissling, 2000). This seismic profile, intersecting the Pennine Front at the surface in the region of the Cormet de Roselend pass, displays a constant dip of the reflectors related to this fault plane by about 30° toward SE. The Pennine Front as observed at the surface does not correspond to a discrete thrust plane but corresponds to a less than one kilometre wide NE-SW trending corridor of intensely deformed rocks of the Moûtiers unit, together with shearing in the underlying Dauphinois units. These mylonites dip with an average angle of 35° to the SE in the Cormet de Roselend area.

At the surface, the structural features related to the contact between Valaisan and Dauphinois are represented by mylonitic rocks which concentrated in a zone several tens to 100 m wide. Further to the SE and away from the contact with the Dauphinois a set of minor tectonic thrusts parallel to the Pennine Front (Fig. 6.34) is observed.

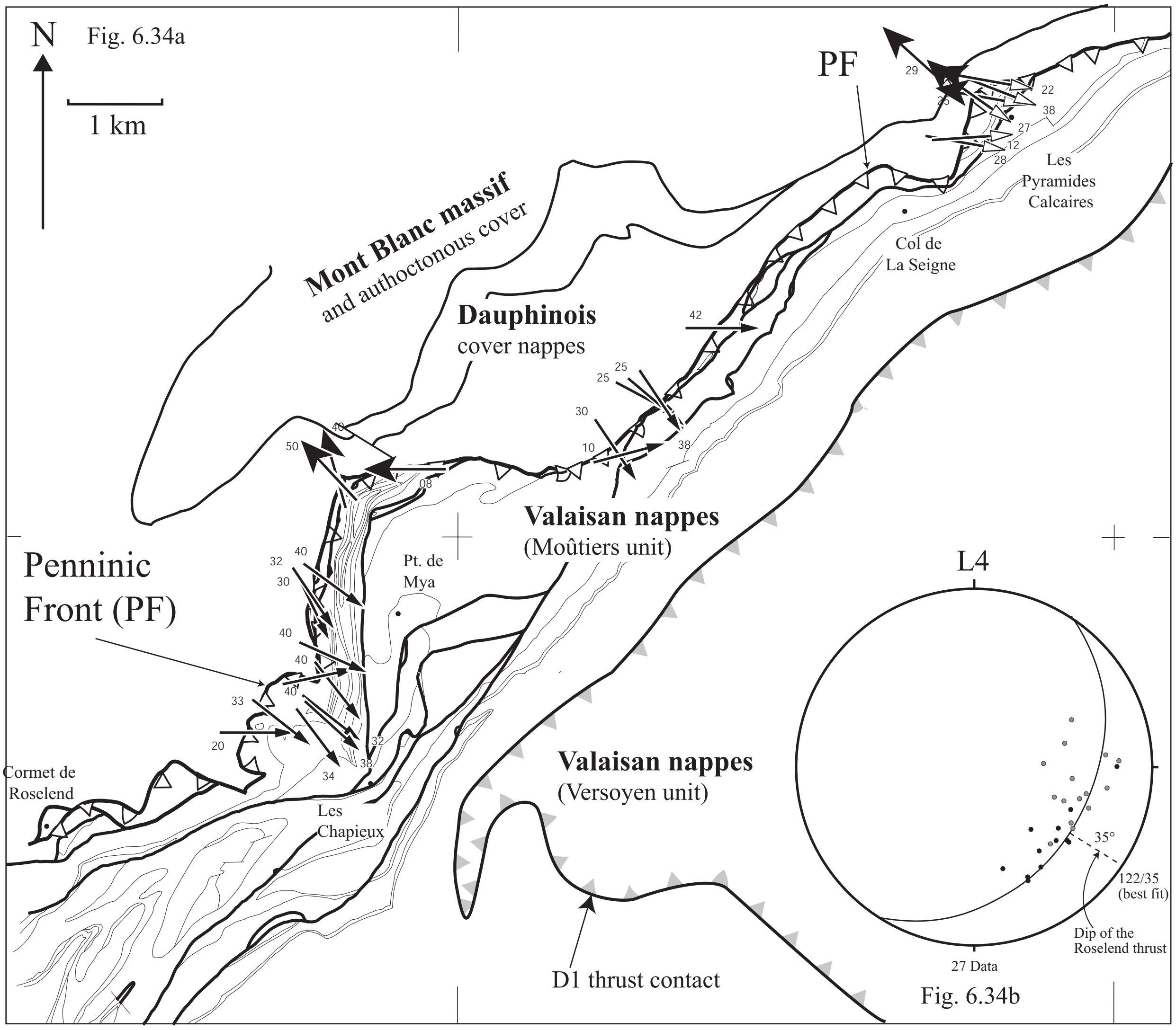
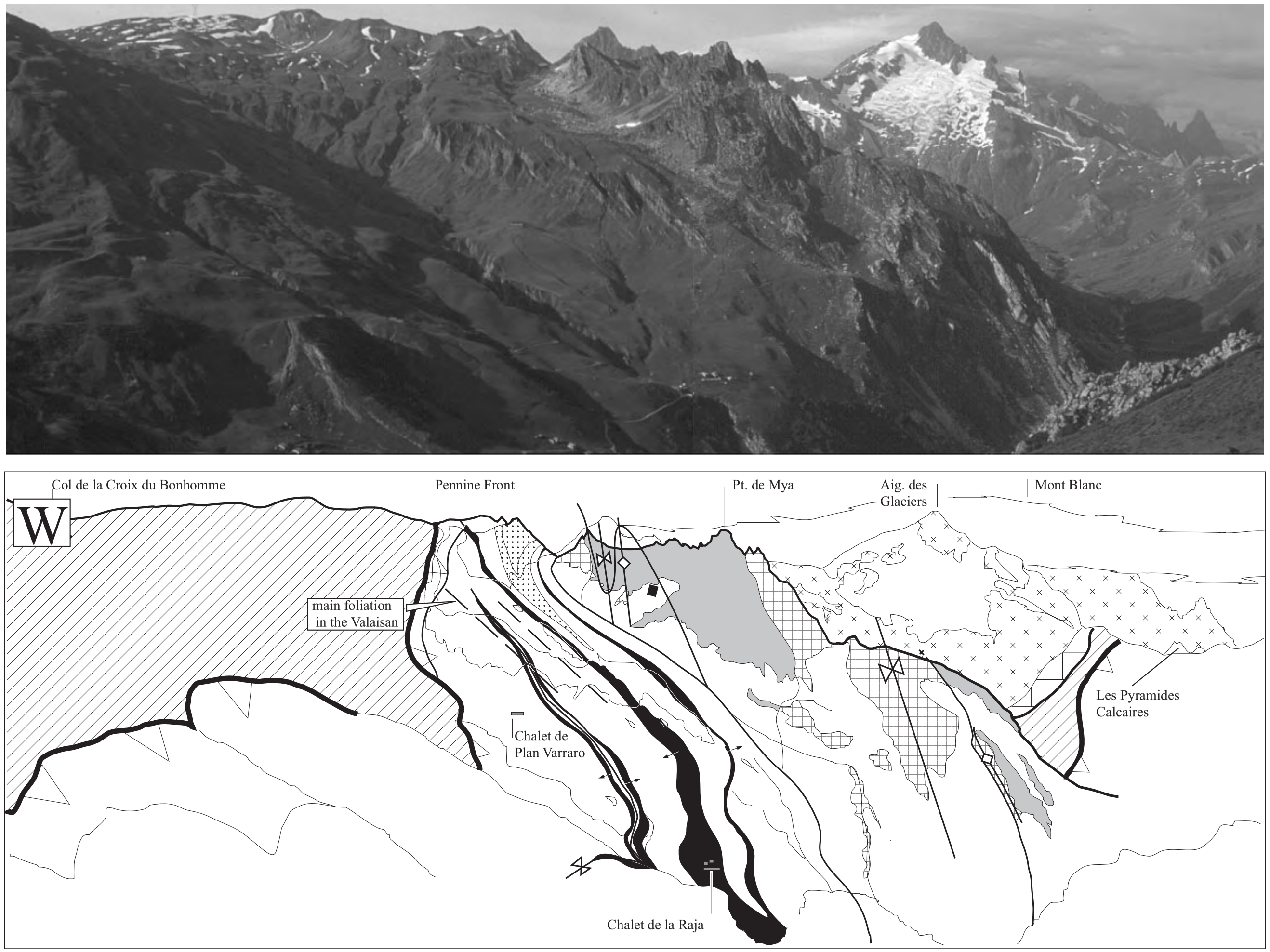


Fig. 6.34a. Azimuth and dip of L4 stretching lineations related to the top-to-the W to NW directed thrust of the Penninic Front in the area between Les Chapieux and the Col de la Seigne. Fig. 6.34b Relative pole figure.

- Transport direction during D4
- L4 stretching lineations
- L4 stretching lineations retrodeformed into their original orientations before the uplift of the Mt. Blanc Massif.
- Minor D4 thrust

Shear sense indicators observed along the Pennine Front include shear bands and asymmetric clasts (Fig. 6.36). Shear bands, indicating top-to-the WNW thrusting are widespread between the Pyramides Calcaires area and further SW in the Cormet de Roselend region.



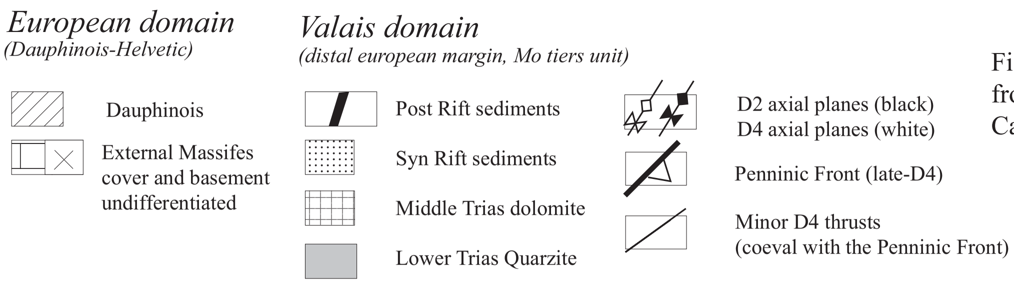
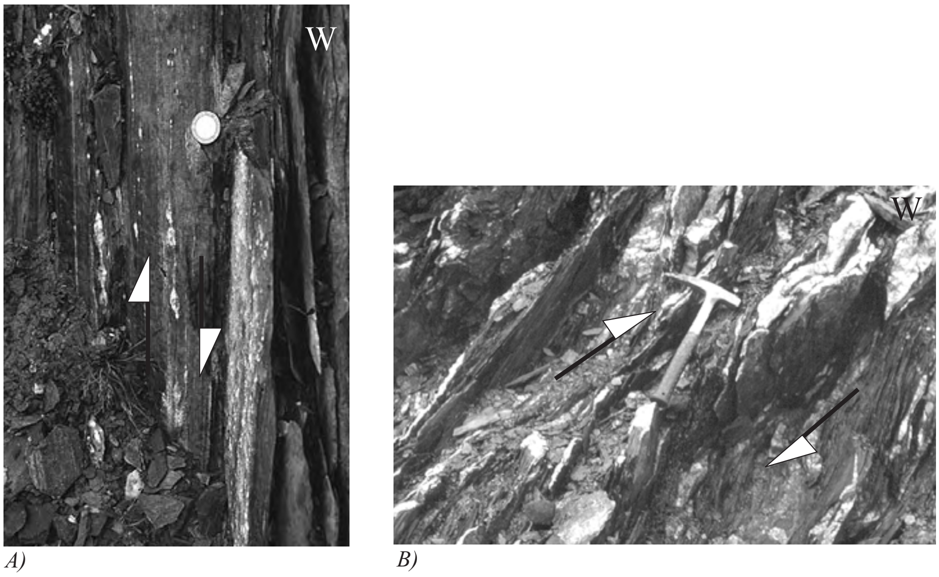


Fig. 6.35. Photograph and drawning of the Penninic Front from the Cormet de Roselend area toward the Pyramides Calcaires (from SW toward NE).

The field examples of Fig. 6.36 come from the Pyramides Calcaires (Fig. 6.36a and b), the Col de Mya (Fig. 6.36c) and the Cormet de Roselend region (Fig. 6.36d) respectively and cover the whole studied segment of the Pennine front from NE to SW.



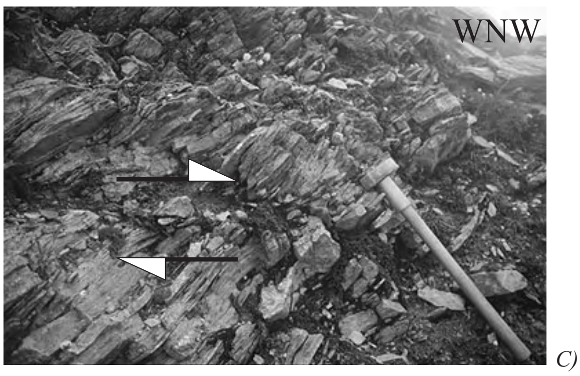
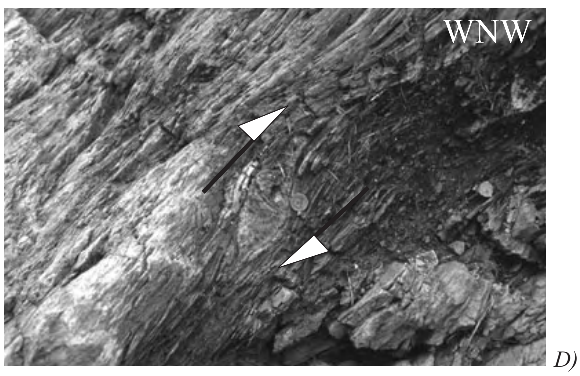


Fig. 6.36. Sigma-clasts (A) and shear bands (B) observed from the Dauphinois at Pyramides Calcaires. Inferred sense of shear is top-to-the-W. Shear bands (C, Col de Mya) and sigma clast (D, Cormet de Roselend) observed from the Valaisan. Sense of movement is top-to-the WNW. Note the difference in orientation of the mylonitic foliation in A) compared to B,C and D).



Shear bands indicating top-to-the WNW thrusting were also found close to the Pennine Front along more internal minor late-D4 thrusts (Fig. 6.36c). Fig. 6.36d shows a σ-clast from near the Cormet de Roselend.

Often σ-clast are much less asymmetric (e.g. Fig. 6.36a, Pyramides Calcaires). Yet they still indicate consistent top-to-the WNW directed thrusting. In the footwall of the Pennine Front isoclinal folds are widespread at the Cormet de Roselend while, close to the Mont Blanc massif mylonites were also observed in the Mt. Blanc basement as well as in the sediments of the Dauphinois below the Roselend thrust.

Along strike and from SW towards NE mylonitic foliation defining the Pennine Front continuously steepens. Values of about 35° towards 130° were found in the Cormet de Roselend area while at the Pyramides Calcaires a subvertical orientation (88° towards 104°) was measured. In the Pyramides Calcaires area, the lineation plunges down-dip with 74° to 085 (Fig. 6.37a) preventing a direct comparison of these measurements with those made in the Roselend area where they plunge at about 35° towards 147° (Fig. 6.34a).

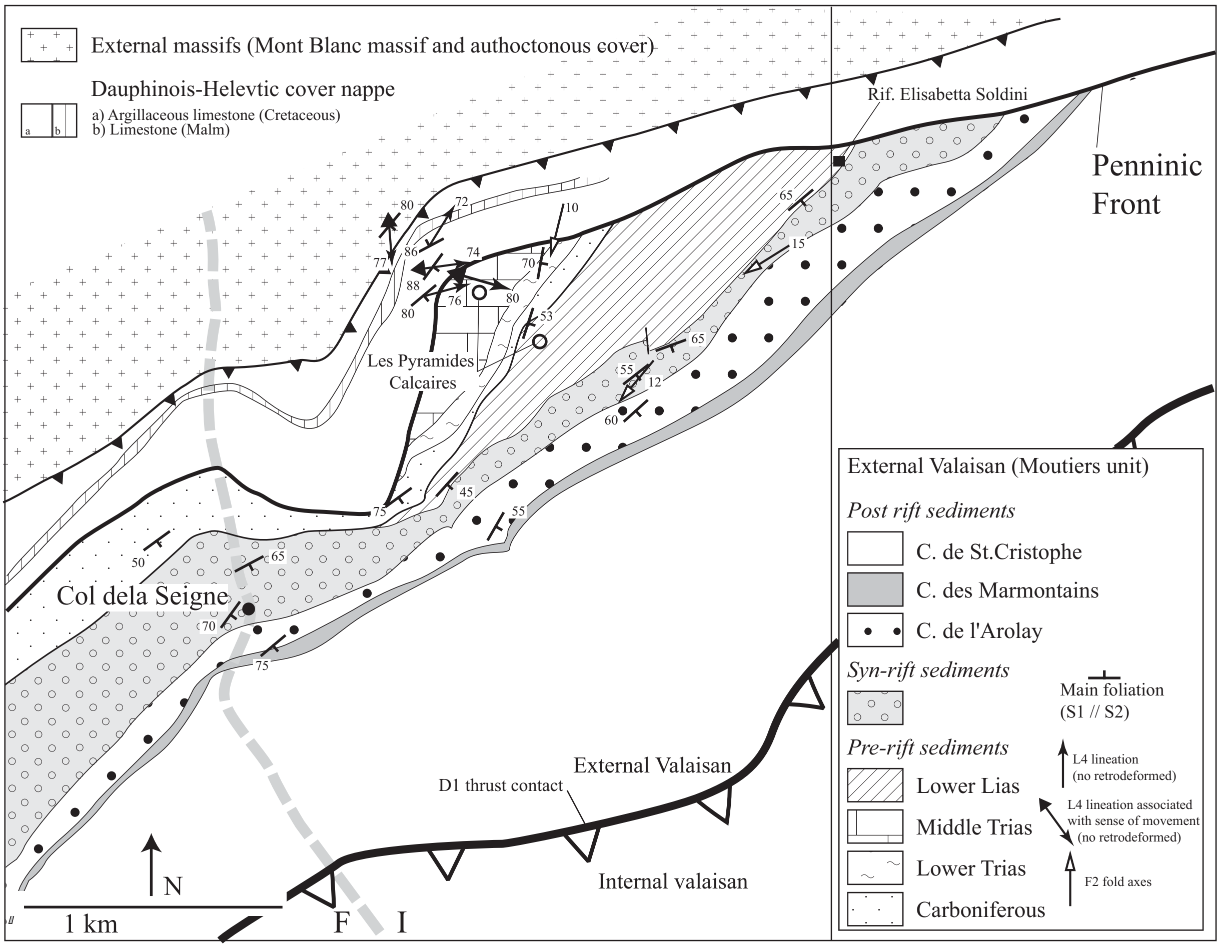


Fig. 6.37a. Geologic map of the Pyramides Calcaires area showing the today's position of L4 stretching lineations associated with Penninic Front.

The steepening of the Pennine Front found close to the Mt. Blanc massif is due to D5 (Fig. 6.37b).

This steepening can be regarded to as a late feature developed in response to the exhumation of the external massifs (D5 deformation phase). Thus at Pyramides Calcaires a deeper part of the Pennine Front is exposed compared to the shallower level of exposure at the Cormet de Roselend area which corresponds to the axial depression between the Mont Blanc in the NE and the Belledonne massif in the SW (Flaine depression, Ramsay, 1989).

In the region of the Pyramides Calcaires it is necessary to correct the measurements of linear structures (L4) obtained from folded strata. The technique employed depends on the style of folding and the mechanism of formation (Ramsay, 1967) (Fig.6.37b).

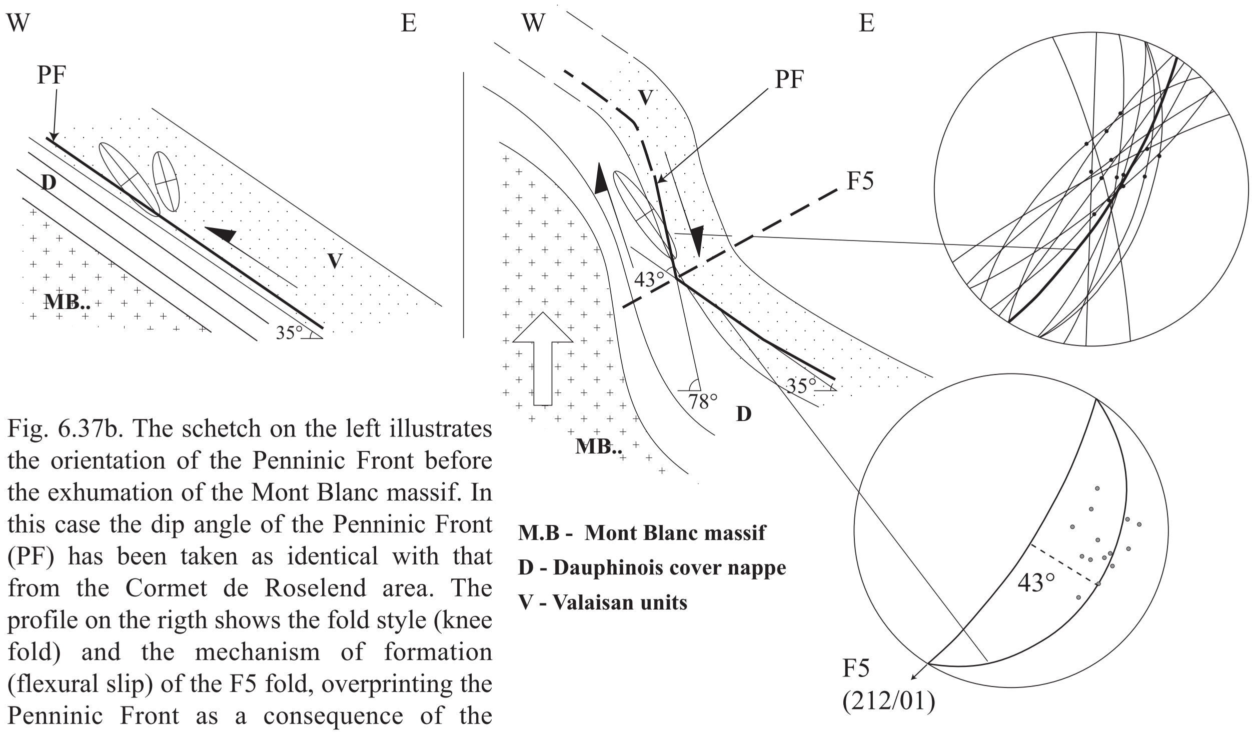


Fig. 6.37b. The schetch on the left illustrates the orientation of the Penninic Front before the exhumation of the Mont Blanc massif. In this case the dip angle of the Penninic Front (PF) has been taken as identical with that from the Cormet de Roselend area. The profile on the rigth shows the fold style (knee fold) and the mechanism of formation (flexural slip) of the F5 fold, overprinting the Penninic Front as a consequence of the exhumation of the Mont Blanc massif.

The pole figure (rigth) shows the L4 lineations and the mean value of the strata forming the steep limb of the F5 fold. The same mean value is plotted together (pole figure below) with the main value of the strata forming the unfolded limb of the F5 fold. Consequently, the intersection between these two main values, gives the orientation of the F5 fold axis (01°towards 212°). Moreover, the angle between the same two mean values (43°) represents the the ammount by which L4 lineations have been rotated around the F5 fold axis to the unfolded position. For further explanation see Chapter 6.6.

The fold is characterised by a single hinge region and by a sharp, angular deflection forming the steep limb with the other limb in its original position (knee fold, Ramsay 1967). Moving towards SW along strike the fold dies out and near the Les Chapieux the Pennine Front remained unfolded. The orientation of the fold D5 axis was determined by plotting a mean value for the steep limb (78° towards 128°) of the folded strata, against a mean value for the unfolded situation (35° towards 122°). From this a folded axis plunging 01° towards 212° can be constructed (Fig. 6.37b). The mechanism that best describes the formation of this fold is flexural-slip. In a flexural-slip fold internal deformation is very weak since slip takes place at the interface parallel to the layer boundaries. The fold forming mechanism allows for a passive rotation of L4 around a sub-horizontal fold axis (see Fig. 6.37b).

The relative chronology between WNW directed thrusting (Roselend Thrust) and D4 within the Valaisan units can be inferred from Fig. 6.35. This view from Cormet de Roselend towards the Pyramides Calcaires, illustrates that the gently SE dipping Pennine Front crosscuts the steeper main foliation (mainly D4) within the Valaisan rocks. As mentioned in the previous Chapter, the S4 axial planar cleavage, merges into parallelism with the Pennine Front in response to WNW-directed thrusting. This means that the main foliation visible in Fig. 6.35 approximates the orientation of the D4 axial planes, which band into parallelism with the Pennine Front towards lower structural levels.

Similar structural observations can be made in the hangingwall immediately next to the late-D4 thrust at the base of the Pt. de Mya massif (Fig. 6.35). This minor D4 thrust, also related to a top-to-the WNW directed thrusting (Fig. 6.36d), clearly post dates F2 folds and is also discordant to earlier formed F4 folds (Fig. 6.35), affecting Permian and Triassic lithologies. The more steeply inclined SE-dipping lithological contacts subparallel to the F2 and F4 axial planar cleavages, are truncated by the underlying, more gently dipping, Roselend thrust. This supports that thrusting post-dates F4 (i.e. it is late-D4).

Further information concerning the relative timing of thrusting at the Pennine Front can be obtained from observations made within the folded rocks of the Dauphinois units resting in its footwall and, on a regional scale, from the kinematic analysis available from previous work (Fig. 6.38). At the junction between the Mont Blanc, Aiguilles Rouges and Belledonne massifs, NW of the Cormet de Roselend region, the present structural configuration of the Dauphinois nappe pile, results from two deformation phases. These Dauphinois nappes (the Mont Joly-A. Croche and the Roselette nappes), represent the uppermost cover nappes overriding the external crystalline massifs and their autochthonous cover (Mont Blanc, Aiguilles Rouge and Belledonne). However, the lowermost crystalline units (external massifs) also moved relative to the main European continent (parautochthon units, Ramsay 1989). Towards NNE, the Helvetic nappes (equivalent to the Dauphinois nappes) are themselves overlain by the Ultrahelvetic nappes, derived from a part of the European margin located internally in respect to the delphino-helvetic depositional region. In the region addressed in this study only the Dauphinois and the external crystalline massifs are exposed, the Ultrahelvetic being buried under the Valaisan units (see Fig. 6.38 and Ramsay 1989).

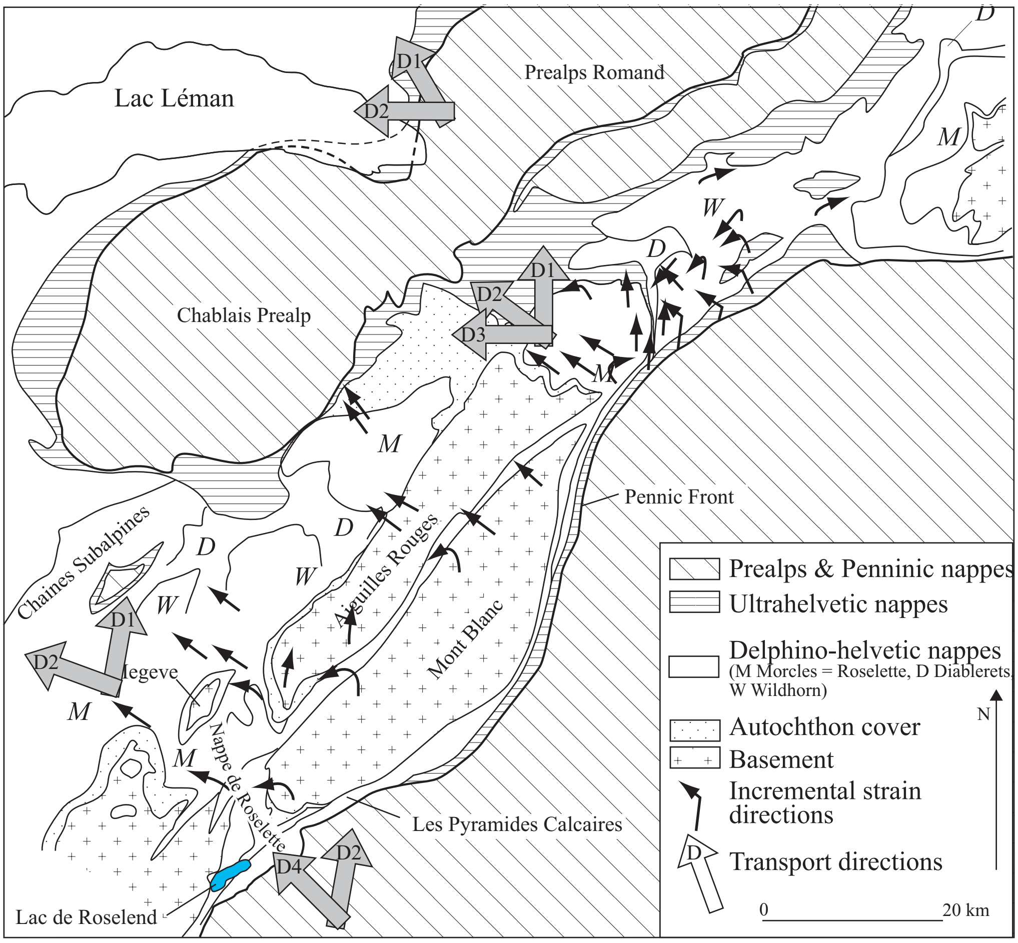


Fig. 6.38 Tectonic map of the external part of the Western Alps (modified after Ramsay 1984). Incremental strain directions from: Ramsay 1984, Dietrich and Casey 1984, Gourlay 1986. Transport directions from: external penninic nappes, this study; external cristalline massifs and Daufinois nappes, Gourlay 1986; Helvetic nappes, Ramsay 1984; Prealps Romande, Mosar 1994). The chronology of each transport direction (D1, D2....) referres to the internal deformation history of each tectonic zone. (M: Morcles, D: Diableret, W: Widhorn nappes)

D1 in the Dauphinois an initial phase (D1) is characterised by a penetrative subhorizontal foliation, associated with isoclinal folding. D1 was observed within the lowermost crystalline rocks and its cover, as well as in the uppermost Dauphinois nappes (Eltchaninoff-Lancelot et al., 1982; Gourlay, 1986). Stretching lineations belonging to this phase, were interpreted in terms of a tectonic transport direction of the detached cover units towards NNW according to a scenario of sinistral transpression active during D1 (Gourlay, 1986). Superimposed on this phase is a subsequent deformation (D2), characterised by the formation of a steep to sub-vertical foliation dipping towards E. D2 becomes more intense towards higher structural levels of the Dauphinois units, i.e., towards the Pennine Front (Gourlay, 1986). A tectonic transport direction toward NW is associated to this phase (D2), which characterises the southern termination of the Mont Blanc as well as its western portion where the Mont Blanc massif is thrusted onto the Aiguilles Rouges massif (top-to-the NW, Gourlay 1986).

Concerning the Helvetic nappes further to the NNE, Ramsay (1989) describes a strain pattern similar to that from further to the south. During the first stages of deformation, after the emplacement of the Ultrahelvetic nappes onto the Helvetic domain, E-W striking folds developed with a N to NNW-directed transport direction (stage 1 and 2 of Ramsay 1989). Later on, during the migration of deformation towards more external areas (i.e. Helvetic nappes and external crystalline massifs) the transport direction of the Helvetic nappes changed from N-S to NW-SE, together with top-to-the NW directed thrusting of the Mont Blanc massif (stage 3a, Ramsay 1989)and the Aiguilles Rouges massif (stage 3b, Ramsay 1989).

Finally E-W shortening, responsible for the N-S oriented depressions and culminations occurred (stage 4, Ramsay, 1989). From N to S they included the Wildstrubel depression, the Rhône culmination and the Flaine depression. In conclusion, the coherent change in transport direction (first northward and than westward) recognised in the basement and cover nappes in the footwall of the Pennine Front shows strong similarities to the change in the transport direction between the D2 (and D1) and D4 as observed in the Valaisan units, found in the hangingwall of the Pennine Front.

These similarities are important for discussing the relative age relationships of the Pennine Front with the Dauphinois rocks in its footwall. In particular, the activity along the Pennine Front offers the possibility to correlate the deformation within the Pennine nappes with that of the more external Dauphinois nappes. The D2 deformation phase within the Dauphinois shows strong affinities with the D4 deformation phase within the Valaisan. These analogies consist in a common transport direction, i.e. towards NW, and in a similar orientation of the NE-SW trending fold axes (F2 concerning the Dauphinois, F4 concerning the Valaisan) including the associated steep axial planes. The Pennine Front is a late feature in respect to the Valaisan units (late-D4), but it was active contemporaneously with the D2 deformation phase of the Dauphinois units.

Detailed observations within the upper Triassic cargneule accompanying the Pennine Front, reveal information concerning the last increments of movement along this thrust contact. The outcrop in Fig. 6.39 shows gently dipping upper Triassic cargneule truncating steeply inclined S2 cleavage of the Dauphinois rocks.

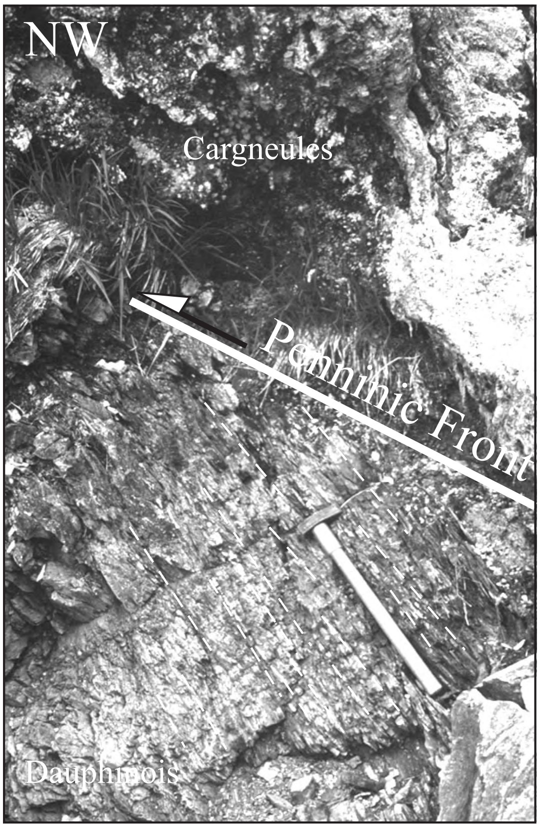


Fig. 6.39. The Cargneule (upper part of this outcrop) outlines the W-NW-directed thrust along the Penninic Front which is observed to post-date the dominant foliation in the underlying Dauphinopis units.

Therefore we can infer that, later thrusting along, the Pennine Front partly post-dates D2 in the Dauphinois. Hence, the Pennine Front is syn- to late-D2 in respect to the internal deformation history of the Dauphinois units and syn-to late D4 in respect to deformation in the Valaisan.

However, the Pennine Front clearly post-dates D1 nappe stacking within the Dauphinois units and hence represents an out-of-sequence thrust.

Several authors, working on the interpretation of the ECORS-CROP seismic profile (Mugnier, 1996; Mugnier et al., 1993; Roure et al., 1990; Schmid and Kissling, 2000) also regarded the Pennine Front in terms of an out-of-sequence structure. In order to better understand the local structural observations, large scale considerations have to be taken into account. NW-directed shortening in the European foreland (Jura, Châines Subalpines and the more internal parts of the Dauphinois nappe system) is taken up along detachment horizons located within the cover sequences as well as within the crystalline massifs which appear to be kinematically linked with the Pennine Front.

It is interesting to note that the amount of shortening estimated from balanced cross sections for the cover sequences (Guellec et al., 1990) exceeds the amount of shortening estimated for the basement. This suggests that the Pennine Front was initiated as an out-of-sequence structure in a more internal zone in respect to the external crystalline massifs (Mugnier et al., 1990). An out-of-sequence structure has also been proposed at the scale of the present geometry of the Alps (Roure et al., 1990; Schmid and Kissling, 2000). Early in-sequence thrusting within the Dauphinois is considered to have taken place during the Aquitanian. Therefore out-of-sequence thrusting along the Pennine Front cannot have been active before the Early Miocene (Tardy et al., 1990). This age estimate is compatible with an Early Miocene age for the D4 deformation phase in the Valaisan units which post-dates D3 normal faulting (proposed to have taken place at 34-27Ma, see Chapter 6.4).

6.6: The D5 deformation phase

Deformation structures post-dating D4 are represented by the already discussed large scale fold, deforming the Pennine Front in the Pyramides Calcaires area and by locally observed small scale young folds. Both features overprint the D4 structures, but their formation my not be contemporaneous.

Small scale young D5 folds can be found, in the NW part of the study area. There F5 folds are generally angular or kinky and coaxial with F4. F5 fold axes plunge gently towards SSW. F5 axial planes are subhorizontal to gently SE dipping. A new axial plane cleavage is usually not developed. A good example of F5 folds can be observed immediately NW of the Col de Leisette (Fig. 6.40) and in the region of the Combe de la Nova.

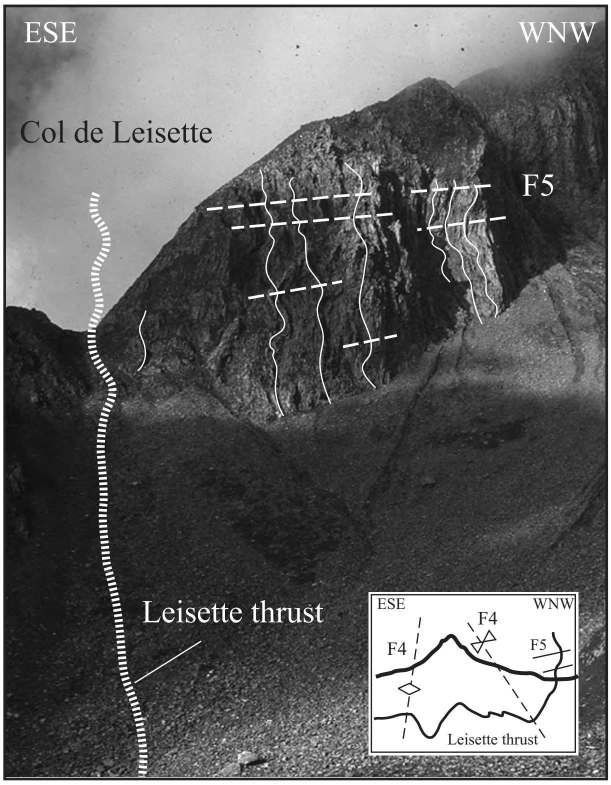


Fig. 6.40. Photograph of fifth phase folds at the Col de Leisette. Axial planes with moderate dip to the ESE. Inset represents a detail of Fig. 6.17 showing, at large scale, the overprinting relationships between F4 and F5 folds in the same area.

Generally, examples of D5 structures were observed where the regional main foliation (S1 and S2) was already in a steep orientation, such as in subvertical to steep limbs of F4 folds.

This suggests that this deformation phase is related to vertical shortening (Froitzheim, 1992). From this point of view it is not impossible that they are contemporaneous with brittle D6 normal faulting (see next Chapter). Since this phase (D5) does not significantly alter the large scale geometry, which is mainly due to the D1, D2 and D4 deformation phases, it is of minor tectonic significance.

However, the tectonic significance of the large scale F5 fold in the Mont Blanc area is more important because it overprints the Pennine Front.

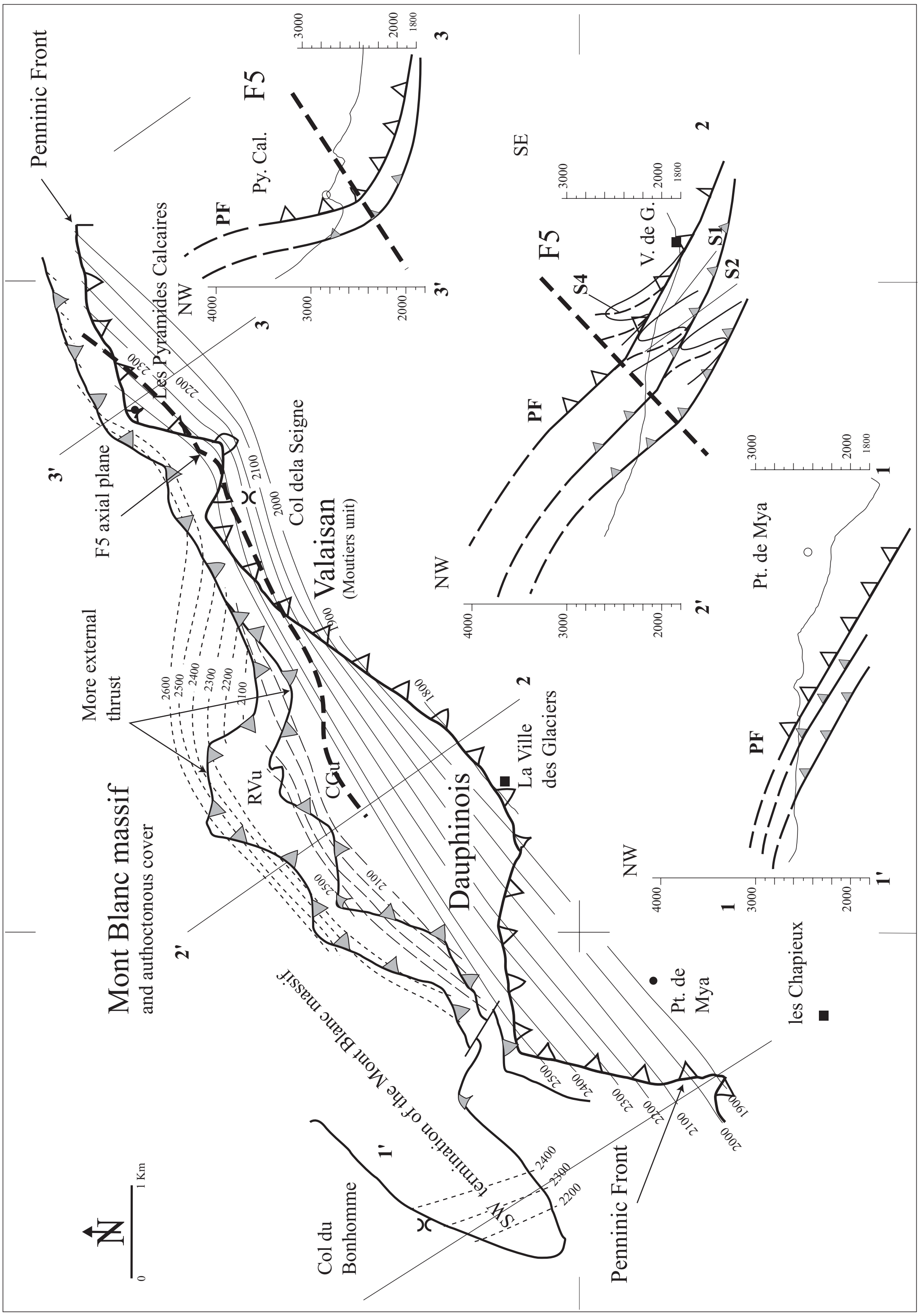


Fig. 6.41 Geologic contour map for the Penninic Front and more external tectonic contacts within the Dauphinois and the Mont Blanc massif. Profiles 1-1' to 3-3' illustrate, from NE to SW, the geometry of the F5 fold

This fold can be visualized by looking at the structural contour map of Fig. 6.41, constructed for the Pennine Front, the thrust of the Dauphinois units onto the Mont Blanc crystalline massif (including its autochthonous cover) and for another thrust contact within the Dauphinois nappe system, i.e. within the Roselette nappe (this thrust contact is responsible for the emplacement of the Rocher du Vent unit onto the Crête des Gittes unit, Eltchaninoff et al. 1982). The Pennine Front, from the Chapieux village to the Pyramides Calcaires area, constantly dips towards SE, exhibiting a steepening in the angle of dip as it approaches the Mont Blanc crystalline massif in the NE.

The thrusts within the Dauphinois, which also roughly dip towards SE, show a constant dip angle. They are closer to the Mont Blanc massif than the map trace of the Pennine Front and hence they are steeper.

Micro-structural observations from strongly deformed rocks coming from the steep limb of the large scale F5 fold give further information on the D5 deformation phase. The overall sense of movement along the late-D4 Pennine Front is top-to-the W to NW as has been described in the previous section. The following micro structural observations refer to a local and later overprinting of the Pennine Front (D5 deformation phase) according to a different kinematic framework, observed in the Pyramides Calcaires area only.

Fig. 6.42 shows a thin section cut parallel to the XZ plane of a calc-mylonite (sample AL 96-33) from the footwall of the Pennine Front. In this sample a top-to-the SE sense of shear is inferred from the δ-clast-like aggregate of large isometric calcite grains (500µ). This aggregate can be interpreted as inherited from a compositional layering that can be traced to be subparallel to the mylonitic foliation. Note that in Fig. 6.42 the orientation of the thin section is the same as that of the outcrop. The larger calcite grains display twin structures while the surrounding matrix, also made up by calcite but with a considerably smaller grain size (20-30 µm by 50-70 µm), is characterised by the elongated shape of the grains. This shape preferred orientation within the matrix defines a SE-dipping fabric which forms a small angle (18°-20°) with the compositional layering.

Since calcite fabrics are generally assumed to record the very late stages of the deformation (Simpson and Schmid, 1983), the fabric defined by the shape preferred orientation in section AL 96-33 cannot be related to planar elements older than the mylonitic foliation. Consequently, it represents a late feature which probably overprints the mylonites of the Pennine Front during the formation of the large-scale F5 fold.

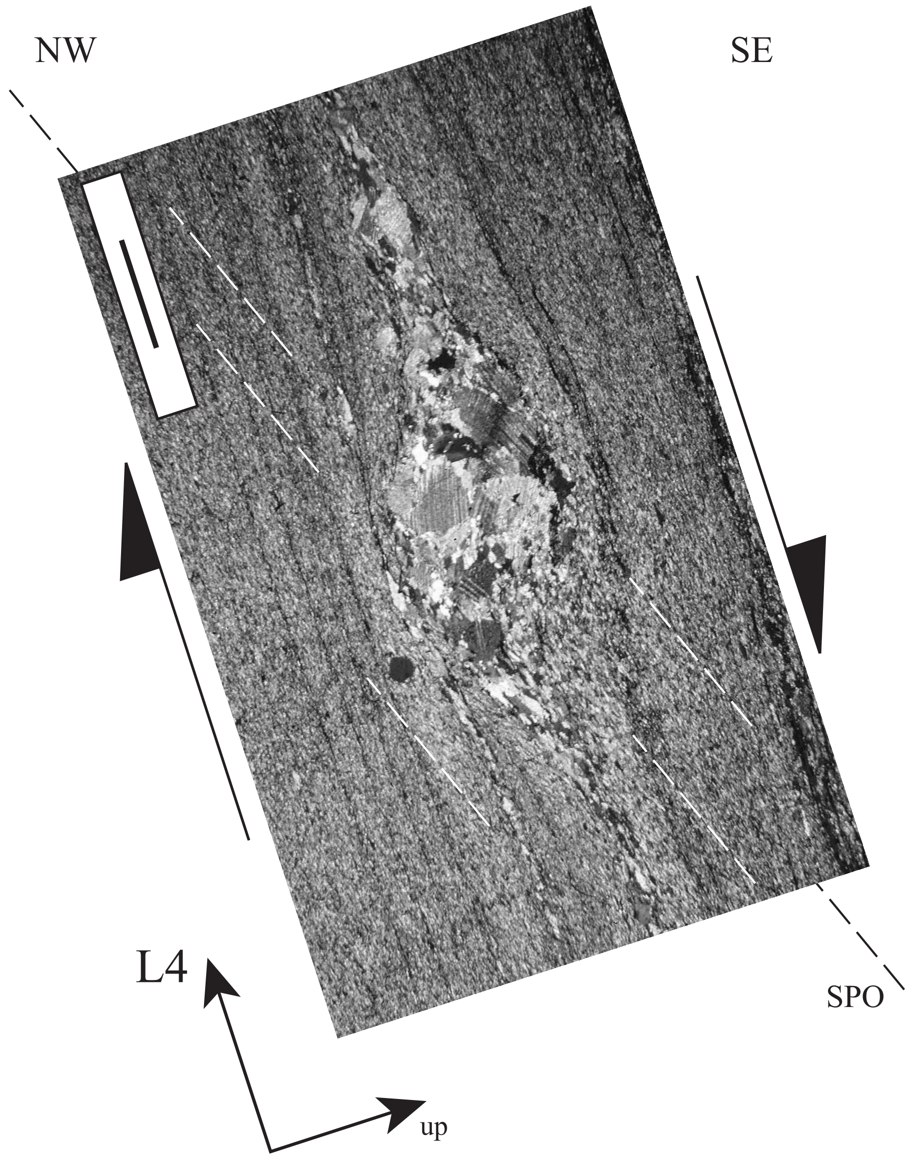


Fig. 6.42 Thin section photograph of the XZ plane from a sample (AL 96-33) collected at the Pyramides Calcaires in the footwall of the Penninic Front. The mylonitic foliation, parallel to the long side of the thin section, dips 72° towards 134 (SE) and the photograph is accordingly oriented. In this sample shear deformation indicating a top-to-the SE sense of movement is inferred from the delta-clast like aggregate of large isometric calcite grains. The fine grained matrix is characterised by the elongated shape of the grains (SPO). This shape preferred orientation within the matrix defines a SE-dipping fabric which, in agreements with the delta-clast, indicates a top-to-the SE shearing, post dating early top-to-the W to NW thrusting along the Penninic Front (late-D4).
The top-to-the SE sense of shear represents a normal fault component which is interpreted to be related to the formation of the F5 fold (see also Fig. 6.37b) according to a flexural slip mechanism and the uplift of the Mt. Blanc massif..

The geometrical relationships between the mylonitic foliation and the fabric defined by the shape preferred orientation of calcite grains within the matrix are compatible with a top-to-the SE directed shearing in agreement with the sense of shear inferred from the δ-clast. Therefore, during the formation of the F5 fold a normal-fault component took place within the steep limb of this fold. This normal fault component is interpreted to be related with the exhumation of the Mont Blanc massif.

In the region of the Mont Blanc and Belledonne crystalline massifs, the inspection of fission track data allows for a discussion of the very youngest deformation in this part of the Alpine chain (Fügenschuh et al., 1999; Seward and Mancktelow, 1994). Seward and Macktelow (1994) recognized, in the NE of the Mont Blanc massif a change in fission track ages (apatite) which has been interpreted in terms of late normal faulting which overprints the Pennine Front (Fig. 3 in Seward and Mancktelow 1994).

Recently, Fügenschuh et al (1999), extended the work of Seward and Mancktelow (1994) further S along strike and from the Pennine Front to the Zone Houillère. Samples for fission track dating were also taken along a transect parallel to the ECORS-CROP profile. Fügenschuh et al (1999) showed that the change in fission track ages described by Seward and Mancktelow (1994) could equally well be interpreted as an overall increase in ages from NW to SE, without the necessity to invoke the reactivation of the Pennine Front in terms of a normal fault (Fig. 11 in Fügenschuh et al. 1999). The overall increase in ages is interpreted to indicate tilting of the whole nappe system toward SE, related to the exhumation of the external crystalline massifs.

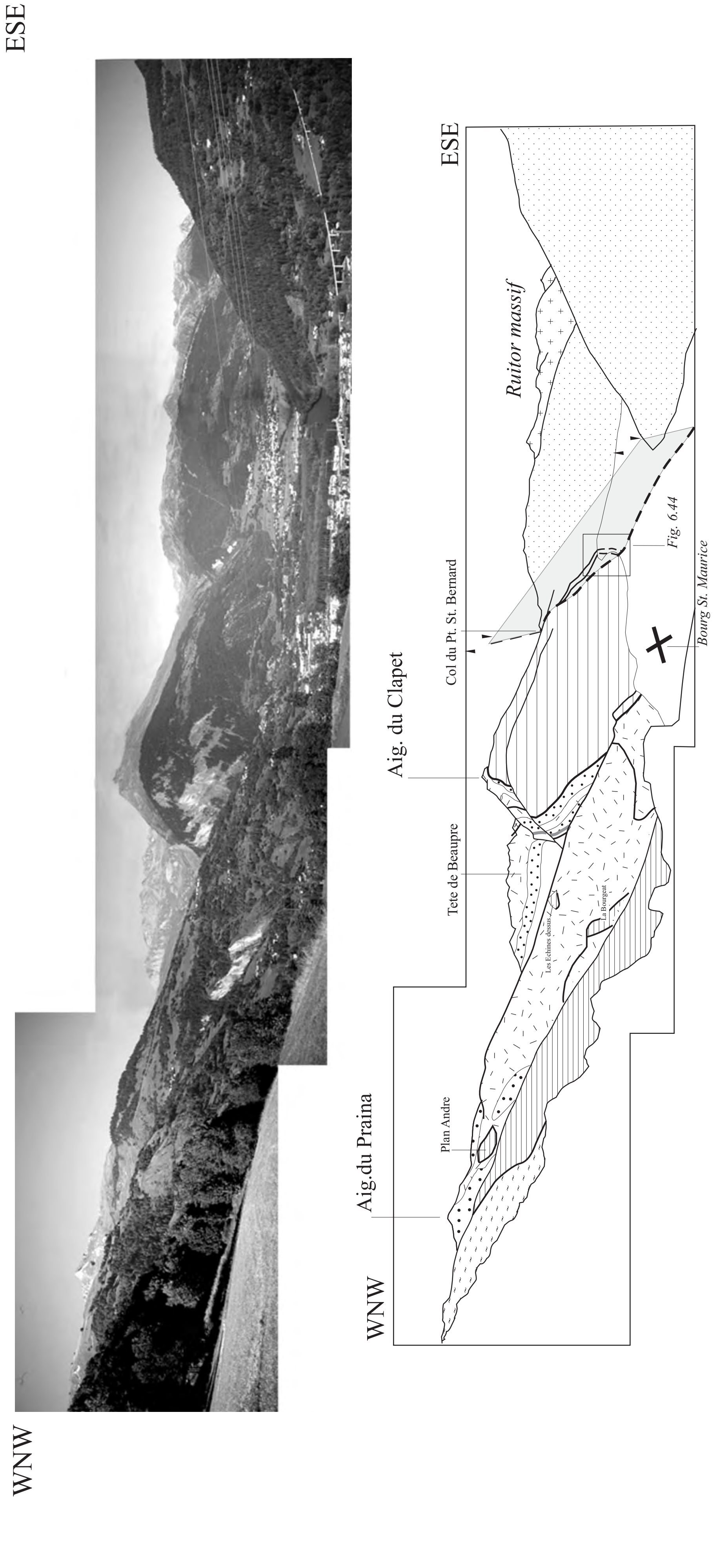
Since the sample localities are located at the NE (Seward and Mancktelow, 1994) and at the SW termination of the Mont Blanc massif (Fügenschuh et al., 1999) these samples were collected from regions with different amounts of exhumation regarding the Mont Blanc massif. Hence, the two transects (in the NE the transect shown in Seward and Mancktelow 1984 and in the SW the transect of (Fügenschuh et al., 1999) cannot be directly compared.

In the area of the Cormet de Roselend, our structural evidence suggests that the high reflectivity zone in the ECORS-CROP profile corresponds to a NW-directed thrust. In this area the Pennine Front is not folded by post D4 deformation and there is no evidence for a significant reactivation of this contact in terms of a normal fault. Toward NE, at the Pyramides Calcaires, microstructural evidence for a normal faulting component is associated with the formation of the large scale F5 fold. From these it is proposed that the normal fault component along the Pennine Front suggested by Seward and Mancktelow (1994) could be interpreted, firstly as a local structure which cannot be related to the deep reflectors of the Pennine Front, and secondly to be genetically related to the formation of steep limb of the F5 large scale fold.

E and SE of the Mont Blanc massif, a general steepening of the entire nappe stack in correspondence with the steepening of the Mont Blanc massif, is observable from the Pyramides Calcaires region towards NE in the Mont Chetif area (Perello et al., 1999) and further NE up to the study area of Seward and Mancktelow (1994).

6.7: D6 deformation phase

The D6 deformation phase consists in a SE-directed very late normal fault, which crosscuts all the previous structures. It can be followed from Moûtiers all the way to Bourg St. Maurice and across the French-Italian border up to La Thuille and is located at the base of the Zone Houillère, overprinting the Houiller Front (Fig. 6.43). Within the study area (and further S until Moûtiers) the Houiller Front represents the SE tectonic boundary of the Valaisan units with the Zone Houillère, which forms the westernmost unit of the Briançonnais Zone. This major fault zone represents also the tectonic limit between the low grade metamorphosed rocks of the Zone Houillère and the Alpine HP-LT rocks of the Valaisan units, i.e. the Pt. St. Bernard and Versoyen units. Since the vertical offset along this D6 normal fault, according to fission track analysis (Fügenschuh et al., 1999) can be roughly estimated to be on the order of 3-4 km, it is not sufficient to produce such a large jump in grade of metamorphism. Therefore the jump in metamorphic grade has to be primarily taken up by the D1 and D2 deformation and only partially from D3-normal faulting (Chapter 6.4). Since we cannot quantify the vertical offset along the D3 ductile normal fault, we cannot exclude that other normal fault structures active during D2 or even D1 worked together in achieving the exhumation of the Valaisan units.



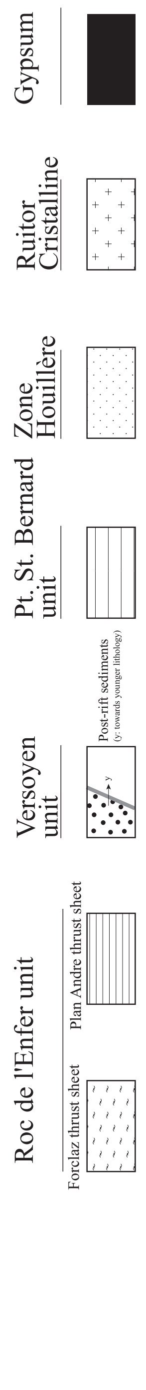


Fig. 6.43 Photograph and line-drawing of the eastern part of the study area, showing a late normal fault (D6) overprinting the Houiller Front. Note that the same D6 late normal fault is also, partially, responsable for the E to SE termination of the Valaisan units.

The existence of D6 normal faulting taking place under brittle conditions and evidenced by fission track data (Fügenschuh et al., 1999) can also be inferred from field observations. In map view, at the scale of the entire study area, the very late character of this feature has already been emphasised in Chapter 5.1.1 and shown in Fig. 5.1. At outcrop scale, the best example of such a D6 normal fault is found at the Torrent de Reclus E of the St. Germain village (Fig. 6.44).

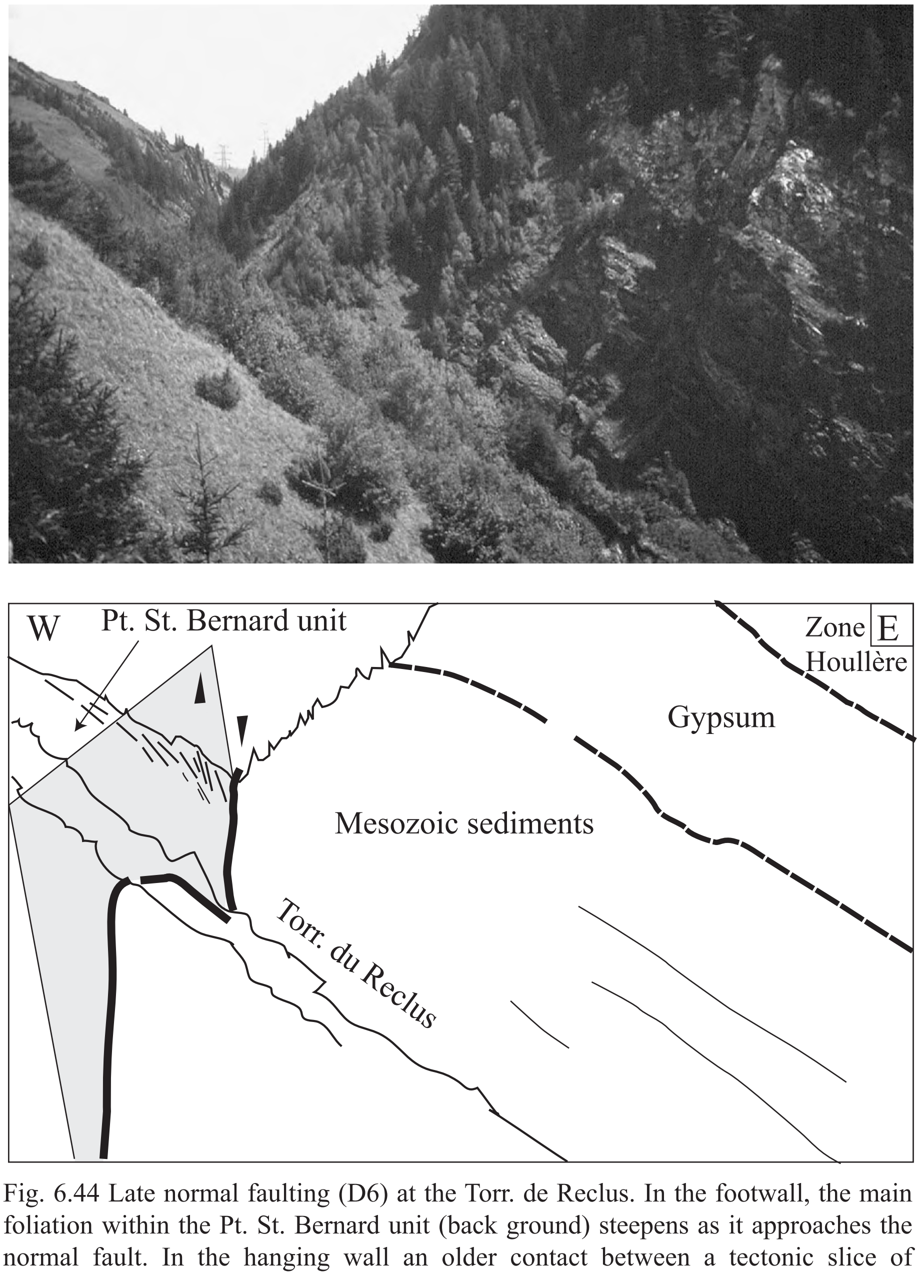
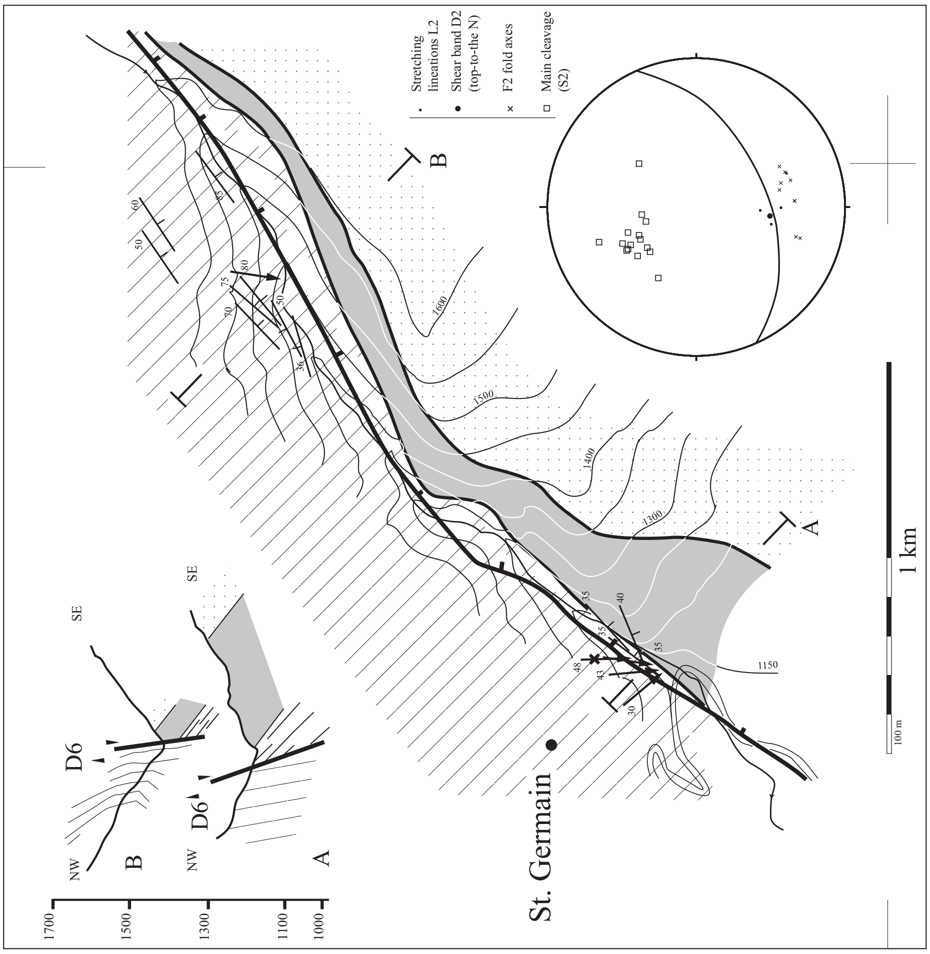


Fig. 6.44 Late normal faulting (D6) at the Torr. de Reclus. In the footwall, the main foliation within the Pt. St. Bernard unit (back ground) steepens as it approaches the normal fault. In the hanging wall an older contact between a tectonic slice of Mesozoic sediments and the Zone Houillère is preserved (see section 6.7).

In this area the late normal fault represents a sub vertical to steeply SE dipping (70°-80°) fault zone, up to 50 m wide, where all the previous structures are mechanically fractured. Within the footwall, the steepening of all previous structures towards the fault plane (Fig. 6.45) does indicate normal faulting. Unfortunately, no direct evidence for the direction of displacement could be found.



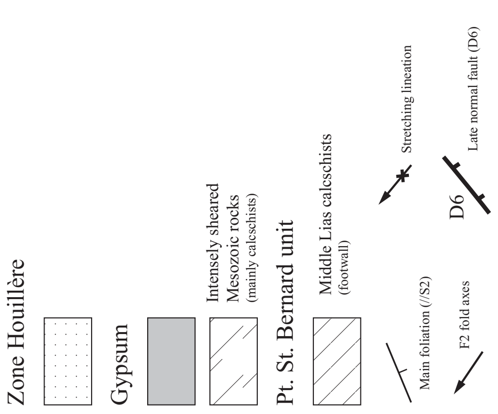


Fig. 6.45. Geological map and NW-SE oriented cross sections of the area along the Torr. du Reclus, near the St. Germain village. The location of this region corresponds to Fig 6.44 (see also Fig. 6.43). The stereoplot refers only to the intensely sheared Mesozoic rocks. In these rocks N-S oriented F2 fold axes and L2 stretching lineation have been found.

6.8: Discussion on the eastern and southeastern contact of the Valaisan units with other tectonic units

According to the geometry of the early-D2 deformation phase, it is to be expected that the high pressure rocks of the internal Valaisan, forming the core of the Versoye synform, have to continue underneath the Houillère Zone (Fig. 6.46a). However, the present-day situation shows that the first Valaisan units to be found below the Zone Houillère are those of the Pt. St. Bernard unit (Figs. 5.1 and 6.46b) in the NE, and those of the Roc de l'Enfer unit in the SE near Bourg St. Maurice (Fig. 6.16). Therefore, the abrupt eastern and southeastern termination of the Valaisan units, and particularly the SE termination of the Versoyen unit in map view, is related to structural features which post-date early D2 folding. They involve N-directed thrusting (late-D2) and two episodes of SE-directed normal faulting (D3 and D6). This section aims to address this overprint which is stronger than suspected until now. This overprinting affects the regional scale D2 folds in the E and SE part of the study area, i.e. the Versoye synform and the more internal F2 antiform (Fig. 6.46a).

In the area between Bourg St. Maurice and La Thuille (Fig. 6.46b) two geological cross section have been drawn (Figs. 6.46c and d). The geometry of this area implies an extremely complex three dimensional configuration due to the discordance between the late-D2 thrust planes and the D3 and D6 normal faults. Therefore, these profiles are only schematic and vertical offsets along the major thrusts and normal faults are not in scale due to the lack of reliable markers. However, where possible, some crucial localities were marked (insets 1 to 5) in order to get a clearer picture of this area.

It must also be noticed that several pieces of information which have been shown in the hangingwall of the D3 normal fault and in the hangingwall of the D6 normal fault, correspond to geometries which are not exposed but have to be deduced from the available outcrop pattern. Especially D6 late normal faulting plays a mayor role when compared to D2 thrusting and D3 detachment, presently hidden below the Houillère Zone and in the hangingwall of this brittle fault. They are partially exposed in its hanging wall.

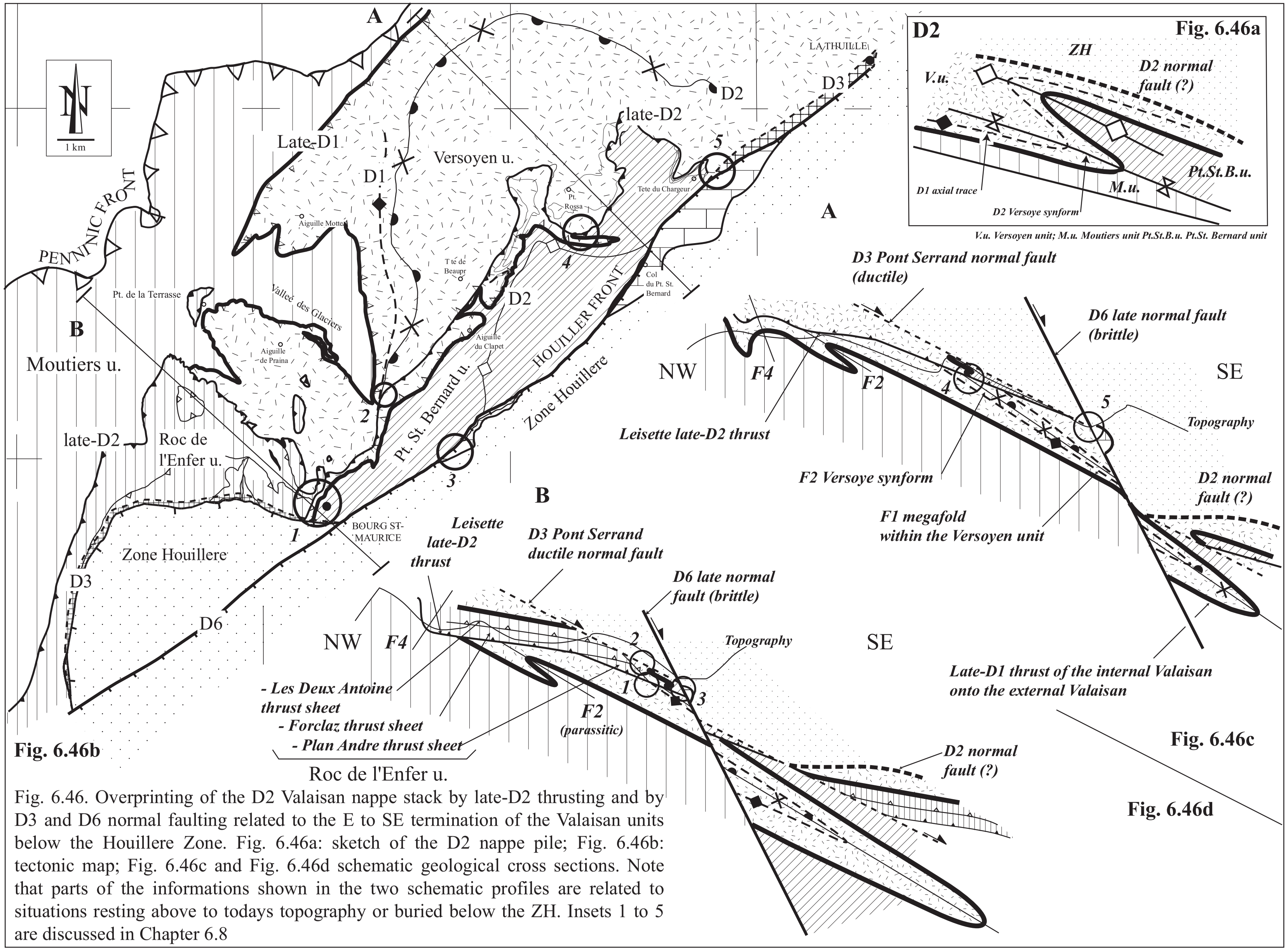


Fig. 6.46. Overprinting of the D2 Valaisan nappe stack by late-D2 thrusting and by D3 and D6 normal faulting related to the E to SE termination of the Valaisan units below the Houillere Zone. Fig. 6.46a: sketch of the D2 nappe pile; Fig. 6.46b: tectonic map; Fig. 6.46c and Fig. 6.46d schematic geological cross sections. Note that parts of the informations shown in the two schematic profiles are related to situations resting above to todays topography or buried below the ZH. Insets 1 to 5 are discussed in Chapter 6.8

It must also be noticed that several pieces of information which have been shown in the hangingwall of the D3 normal fault and in the hangingwall of the D6 normal fault, correspond to geometries which are not exposed but have to be deduced from the available outcrop pattern. Especially D6 late normal faulting plays a mayor role when compared to D2 thrusting and D3 detachment, presently hidden below the Houillère Zone and in the hangingwall of this brittle fault. They are partially exposed in its hanging wall.

In the area of the Pt. St. Bernard pass (Fig. 6.46.c) most of the D3 normal fault is above the present-day topography. Only in a small area close to the Tête du Chargeur (inset 5) are the late-D2 Col des Rousses thrust and the D3 Pont Serrand normal fault exposed. In the area of Bourg St. Maurice the same situation is exposed near inset 1 (Fig. 6.46d). Here the D3 normal fault is postulated to be situated along the contact between the Zone Houillère and the Roc de l'Enfer unit (Chapter 6.3.5). This thrust sheet system (the Roc de l'Enfer unit) is no more exposed further N, probably due to local out of sequence structures. Despite this complication the late-D2 Leisette and Col des Rousses thrust planes are considered as a single thrust plane all the way from Bourg St. Maurice to the Tête du Chargeur area.

Another important geometry is represented in inset 2 (Figs. 6.46b and d) where both the D1 and the D2 Versoye synform axial planes traces are in the footwall of the Col des Rousses thrust. In conclusion, the E and SE omission of the Houillère contact of the internal Valaisan in present-day map view is due to two effects: firstly, late-D2 thrusting which leads to the duplication of parts of the regional scale D2 synform and secondly, SE-directed normal faulting during D3 and D6.

Insets 3 and 5 correspond, in the hanging wall of the D6 normal fault, to the occurrence of Mesozoic cover sequences which rests below the Zone Houillère. The tectonic significance of these sediments is not completely clear. In the following, without alter the conclusion exposed above, we will briefly describe the field evidences in order to propose a simple solution. The inset 3 is shown in the geological map of Fig. 6.45, where these Mesozoic sediments are represented by a hundred meter slice of intensely deformed calcschists separated from the Zone Houillère by the occurrence of gypsum and cargneules. At the Col du Pt. St. Bernard other occurrences of Lower Triassic quartzite and Middle Triassic dolomite show the same situation, i.e. Mesozoic sediments in the hangingwall of the D6 normal fault. In inset 3, these calcschists display a SE dipping mylonitic foliation (47° towards 156) which is subparallel to the main foliation in the gypsum (40° towards 160°) characterised by intense deformation (Fig. 6.45).

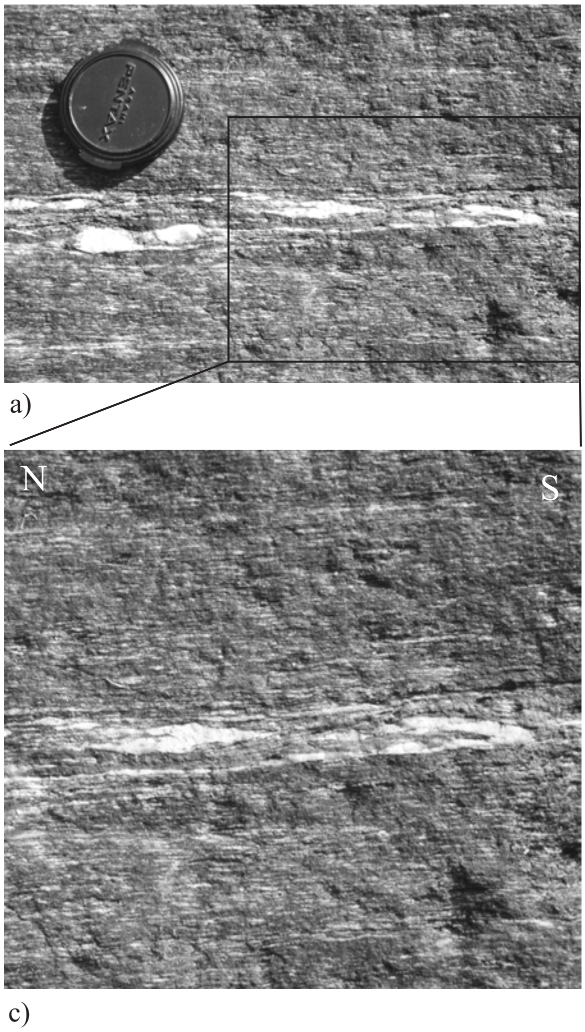
The foliations in the calcschists and in the gypsum are axial planar cleavage of isoclinal folds refolding an older cleavage (Fig. 6.47). Fold axes are NNW-SSE oriented (42° towards 168°) and subparallel to the stretching lineation (48° towards 186°) observed in the calcschists (Fig. 6.45). In the calcschists, the stretching lineation was found within lens shaped aggregates, probably inherited from a former compositional layering (S0//S1), made by calcite and quartz which is subparallel to the older cleavage, presently forming dismembered isoclinal folds as the results of the ongoing deformation (Fig. 6.48). In only very few cases these isoclinal folds are locally overprinted by shear bands roughly indicating a top-to-the N sense of movements (Fig. 6.48). Where these aggregates are missing, the deformed rocks show an extremely homogeneous composition which, generally, makes recognising the shear bands and the isoclinal folds quite difficult.

The fact that the stretching lineation is roughly NS-oriented and contained in the second foliation ascribes this preserved tectonic contact to the D2 phases, because they display the same orientation for the stretching lineation. Since the shear bands deflect the main cleavage, the inferred sense of movement (top-to-the N) could be related to the late stages of this isoclinal D2 folding. According to this study the belonging of the two Mesozoic slices in the hangingwall of the D6 normal fault (shown in insets 3 and 5) remain problematic. The attribution of these sediments to the Roc de l'Enfer unit is unlikely because the Leisette thrust, which define the base of this unit, is prolonged in the Col des Rousses thrust (Fig. 6.46b). On the other hand the attribution of these sediments to the Pt. St. Bernard unit is also unlikely because these sediments rest above the D3 normal fault. The only other possibility is that they probably belong to the Zone Houillère.

A)

B)

Fig. 6.47. Photographs of second phase folds at the contact between the Houillere Zone and a Mesozoic cover sequence which is preserved in the hangingwall of the D6 normal fault (for location see Fig. 6.46, inset 3). (a) Isoclinal dismembered folds in the gypsum which outline the Houiller Front, and (b) tight folds in the mesozoic sediments. Note the different orientations of the planes of the photographs. In both examples, folding overprints an older cleavage (S1) and is therefore interpreted to belong to F2.



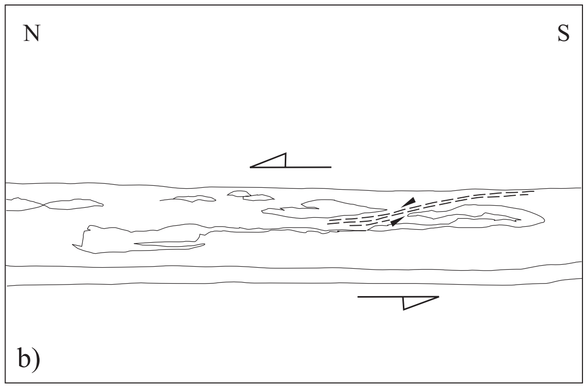


Fig. 6.48. (a) Photograph and (b) line drawning of a shear band in the Mesozoic sediments shown in Fig. 6.47b.
Fig. 6.47 c), enlargement of the same shear band. The inferred sense of shear is top-to-the N.

6.9: Summary of the structural analysis of the Valaisan units

The schematic tectonic map of Fig. 6.49 shows an overview of the main Alpine structural elements recognised in the study area. During pre-D1 phase, the Versoyen and the Pt. St. Bernard units reached the HP-LT peak metamorphic conditions. Therefore all the subsequent deformation phases (D1 to D6) contribute to the exhumation of the Valaisan HP units. The pre-D1 phase is also responsible for the detachment of the Versoyen unit from its oceanic basement.

During D1, the ongoing of thrusting and the onset of D1 folding within the Versoyen unit is related to the formation of the regional scale late-D1 thrust which emplaced the Versoyen unit (internal Valaisan) onto the Pt. St. Bernard and Moûtiers units, respectively, post-dating an F1 mega-fold within the Versoyen unit itself (late-D1 thrust of the internal Valaisan onto the external Valaisan).

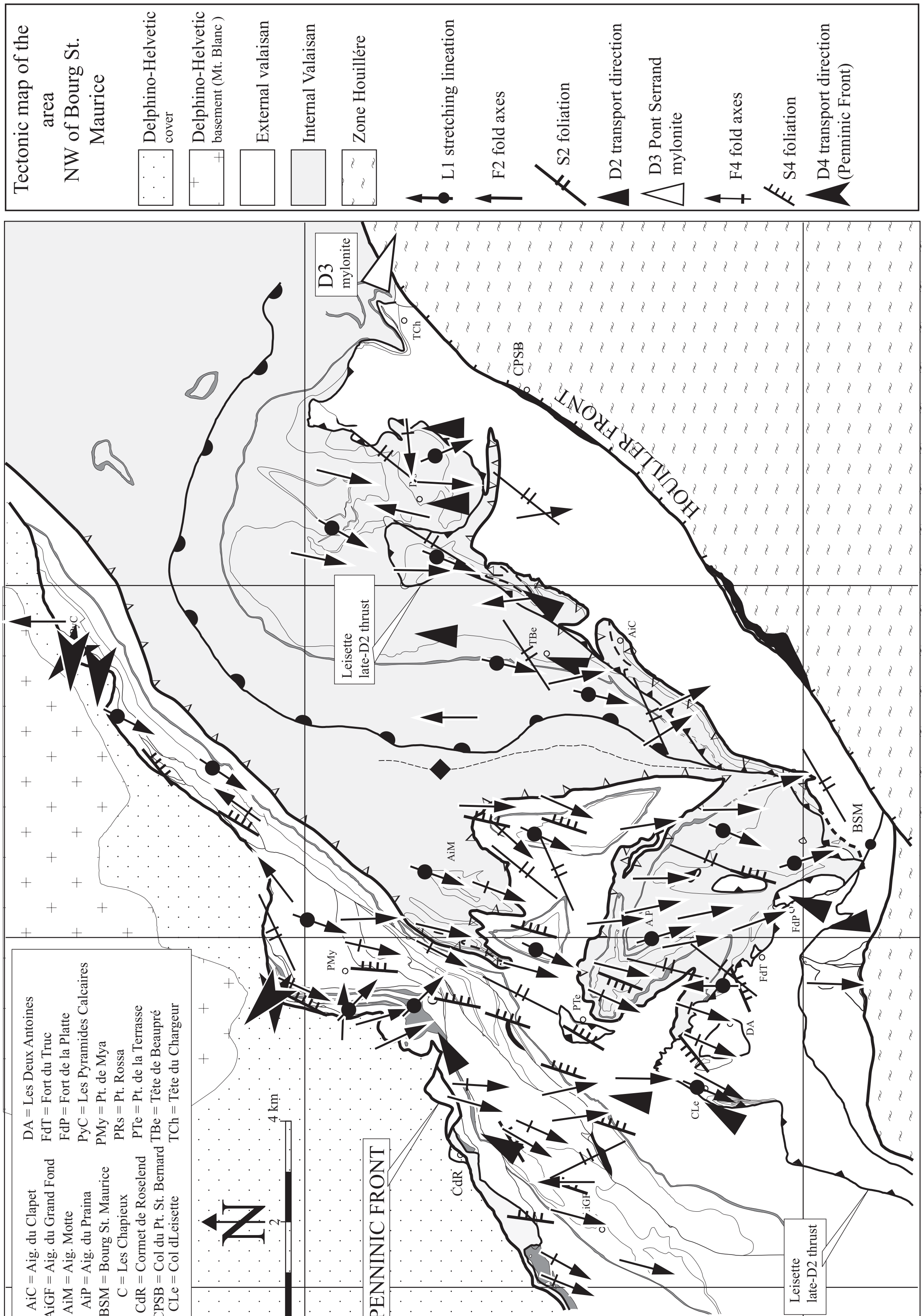


Fig. 6.49 Schematic overview of the structural elements of the Valaisan units. For the simbols not included in the legend, the reader is referred to Fig. 5.1

Associated with the D1 deformation phase is the development of a penetrative cleavage developed under different metamorphic conditions and a generally N-S to NNE-SSW oriented stretching lineation which is interpreted as being parallel to the direction of tectonic transport during D1 (top-to-the N).

During D2 deformation phase the regional scale F2 Versoye synform developed, with elements of the external Valaisan on both limbs (the Moûtiers unit in the lower limb corresponds to the Pt. St. Bernard unit in the upper limb), which envelope and re-fold the F1 mega-fold within the Versoyen unit forming the core of this prominent D2 structure. This large scale E-closing Versoyen synform suggests that the Versoyen unit formerly continued further E. F2 fold axes are generally N-S oriented with gently E dipping axial planes. In a late stage of the D2 deformation phase the late-D2 Leisette thrust is responsible for the individualization of the Roc de l'Enfer unit from the Moûtiers unit leading to the tectonic omission of the Versoyen unit further S. Kinematic indicators related to early D2 folding and late-D2 thrusting indicate a roughly top-to-the N sinistral transpression, which can be invoked for both the D1- and the D2-deformation phase.

A D3 deformation phase consisting in top-to-the ESE directed normal fault overprints the D2 nappe pile in the NE part of the study area partially contributing in the ongoing exhumation of the Valaisan HP units (Versoyen and Pt. St. Bernard units).

A D4 deformation phase, forming NW vergent folds with mostly steep SE dipping axial planes developed obliquely to the trend of the earlier structures (pre-D1 to late-D2) and to the different Valaisan tectonic units. From SE to NW the fourth deformation phase displays a gradient of increasing strain toward NW and toward the Pennine Front (or Roselend thrust) which, in the study area, corresponds to a top-to-the NW directed late-D4 thrust which is demonstrably an out-of sequence structure. However the D4 folding and thrusting do not significantly alter the overall geometry obtained during the preceding deformation phases.

The D5 deformation phase is associated with the exhumation of the Mt. Blanc massif post-dating the activity along the Pennine Front. A last D6 deformation phase consists in a late normal faulting overprinting the Houiller Front.

CHAPTER 7: SUMM A RY

This thesis discussed the evolution of the Valaisan units which represent the frontal part of the Pennine zone of the Western Alps (Savoy, France). The Permo-Mesozoic metasediments of the allochthonous Valaisan nappes underwent a complex polyphase Alpine tectono-metamorphic history, following the Late Jurassic/Early Cretaceous paleotectonic evolution. The subdivision into external and internal Valaisan forms the key to the reconstruction of the evolution Valais basin as discussed in this work. With the aim to combine sedimentological observations with structural analysis an interactive study of outcrops and new detailed field mapping was carried out.

The stratigraphy of the Valaisan units was controlled by a variety of tectonic and sedimentary processes, mainly related to the evolution of a Late Jurassic to Early Cretaceous rifting process. The distinct signature of this rifting process in the stratigraphic record indicates the occurrence of both a thinned continental domain (external Valaisan), and, of an oceanic domain (internal Valaisan). According to the results of this work the external Valaisan was located on the thinned European continental margin while the internal Valaisan represents the newly formed oceanic area (i.e. the Valais ocean), located between the European margin and the more internal Briançonnais domain. Consequently, the paleogeographic arrangement of the Valais units, as proposed here, reflects this subdivision into an external (thinned continental margin) and internal (oceanic) Valaisan. In fact the proposed paleogeographical arrangement strongly differs from previous reconstructions. The external Valaisan comprises (from external to internal): the Quermoz-, Moûtiers-, Roc de l'Enfer- and Pt. St. Bernard units). The internal Valaisan consists of the Versoyen unit as (newly) defined below. The Versoyen unit, as used in this work, only refers to those parts of the Valais domain that contain remnants of oceanic crust (i.e. the Pt. Rossa Complex). All other units, lacking in evidence of ophiolites, represent pre-rift terrigenous to shallow water sediments deposited on continental crust. Applying sequence stratigraphic concepts to the sediments of the Valaisan rift basin led to the recognition of distinct stratigraphic stages, bounded by unconformities. The external Valaisan is subdivided in pre-, syn- and post-rift sequences. The internal Valaisan consists of a) transitional sediments (the Complex Antéflysch Fm. of the Versoyen unit), directly overlying newly formed oceanic crust (preserved within the Pt. Rossa Complex) and b) post-rift sediments. Thus, the post-rift sediments are the only sediments common to both, external and internal Valaisan. An analysis of the facies distribution of the post-rift sediments was performed throughout the entire Valaisan domain in order to highlight its paleotectonic arrangement. The results indicate the presence of a proximal and a more distal area of deposition. The proximal area of deposition was located in the external Valaisan while the internal Valaisan represented the distal depositional environment. Further observations indicate that the clastic material at the base of the post-rift sediments had its source in the external Valaisan (craton-ward). The formation of the Valais ocean therefore predates the deposition of the first post-rift sediments overlying both continental and oceanic domains. Opening of the Valais ocean is tentatively dated as Early Cretaceous.

The recognition of pre- and syn-rift stratigraphic sequences in respect to the opening of the Valais ocean allows to focus the attention to the pre-rift configuration of the Valais domain. The pre-rift sediments preserve evidence of stratigraphic affinities with the Subbriançonnais domain. This includes a) the thick Early Jurassic sediments in the Pt. St. Bernard unit and b) the occurrence of an Upper Triassic (Raethian) potential detachment horizon within all units of the external Valaisan. The Subbriançonnais domain, located further to the S of the study area represents, together with the Briançonnais domain, the passive margin that formed in response to the opening of another oceanic domain, the Piemont-Liguria ocean. This more internal oceanic domain formed earlier, i.e. during the Early Jurassic. Therefore, the evolution of the Valais domain is characterized by the superposition of two distinct rifting events. An older one, the Piemont-Liguria rifting is associated with the formation of the Subbriançonnais and Briançonnais passive margin domain. The younger Valaisan rifting, led to the formation of a new ocean, the Valais ocean, within the former Subbriançonnais domain.

However, in order to reconstruct the paleogeography summarised above the Alpine tectonometamorphic evolution, involving subduction and obduction/exhumation, had to be understood and restored. This is the topic of the second part of this thesis which addressed the Alpine deformation history. Field mapping and the construction of geological cross sections revealed a complex three-dimensional picture of the Valaisan nappe stack. The structural analysis of the Valais domain allowed to distinguish seven (superimposed) deformation phases. Three main phases, involving thrusting and folding (i.e. pre-D1, D1 and D2) led to the formation of the Valaisan nappe stack. The other phases (D3, D4, D5 and D6) modified this geometry and led to the present day map view of the investigated area.

The metamorphic evolution is crucial for understanding of the early orogenic processes. Petrological constraints from published studies were integrated with the structure and, together with new microstructural observations, allowed for a construction of a tectonometamorphic evolution. It is demonstrated that the structural evolution, starting with D1, immediately postdates an earlier HP/LT metamorphic event. However, this pre-D1 HP/LT metamorphism is only recorded in a) the Versoyen unit and b) the Pt. St. Bernard unit. Hence, this pre-D1 deformation phase related to HP/LT metamorphism is responsible for the detachment of the oceanic internal Valaisan sediments and sills (Versoyen units) from its oceanic basement and the early tectonic contact with the Pt. St. Bernard unit.

The most outstanding D1 structures are a regional scale late-D1 thrust which emplaced the Versoyen unit (internal Valaisan) onto the external Valaisan, post-dating the formation of F1 mega-fold within the Versoyen unit itself. During D1 all Valaisan units developed an S1 penetrative cleavage under different metamorphic conditions (metamorphic gradient). The S1 cleavage in the internal Valaisan (Versoyen unit) and in parts of the external Valaisan (i.e. the Pt. St. Bernard unit) formed after HP/LT peak metamorphic conditions. It is demonstrated that chloritoid, replacing carpholite during the exhumation, grown within the first foliation (S1). The S1 cleavage in the other external Valaisan units (i.e. Moûtiers and Roc de l'Enfer) formed under greenschist facies conditions. These differences are explained in terms of a metamorphic gradient, increasing from the external towards the internal Valaisan. Thus looking in map view, the present-day distribution of metamorphic grade within the Valaisan zone is caused by later folding (see below). The increase in metamorphism from the external Valaisan towards the internal Valaisan well matches the paleotectonic reconstruction of the Valais domain as discussed in the first chapter of this work. Furthermore a generally N-S to NNE-SSW oriented stretching lineation (L1) developed during this first phase of deformation, This is indicative for roughly top-to-the-N directed transport during D1.

During the second phase of deformation (D2) the spectacular regional scale F2 Versoye synform developed. This fold has the external Valaisan (the Moûtiers unit) in its lower limb and the Pt. St. Bernard unit in its upper limb. The core of this prominent D2 structure is formed by the Versoyen unit refolded by a F1 mega-fold. This km-scale E-closing Versoye synform suggests that the Versoyen unit formerly continued further towards the E. F2 fold axes are generally N-S oriented with gently E dipping axial planes.

During a late stage of D2 deformation the Leisette thrust formed. This thrust is responsible for the individualization of the Roc de l'Enfer unit from the Moûtiers unit leading to the tectonic omission of the Versoyen unit further S. Structural analysis of the D1 and D2 deformation phases indicates that the present-day outcrop pattern of the Valaisan units does not match the original paleogeographical configuration. This is particularly clear for the Pt. St. Bernard unit. This unit was generally ascribed to the Subbriançonnais domain and considered as a tectonic slice emplaced during a late stage in the overall evolution. In contrast to this interpretation this work demonstrates that its contact with the Versoyen unit is a very old one (pre-D1). Shear sense criteria related to early D2 folding and late-D2 thrusting indicate a roughly top-to-the-N directed transport. In a larger context, these top-to-the-N directed movements (D1 and D2) are interpreted as being related to sinistral transpression between the internal Western Alps (Briançonnais and more internal Pennine units) and the still undeformed European foreland (Dauphinois) within a NNE-SSW-striking corridor (Ricou and Siddans, 1986).

D3 is characterized by top-to-the-ESE directed normal faulting under greenschist facies conditions, overprinting the D2 nappe pile in the NE part of the study area (near the Pt. St. Bernard pass). This normal fault allows for the exhumation of all Valaisan units found in the footwall of the Zone Houillère. Yet it has to be pointed out that D3 normal faulting is not responsible for all the exhumation of the HP rocks of the Valaisan, which by that time have already been exhumed to intermediate depths, related to D1 and D2 deformation.

D4 is characterized by NW verging folds with steeply SE-dipping axial planes. These folds developed oblique to the trend of the earlier structures (pre-D1 to late-D2) and to the different Valaisan tectonic units. From SE to NW D4 displays a gradient with increasing strain towards the NW, i.e. towards the "Pennine Front" (or more specifically the Roselend thrust as defined by Ceriani et al. in press). The study area is the type locality of the Roselend thrust, an Oligocene (late-D4 with respect to the structural scheme of the Valaisan) top-to-the WNW directed thrust which is demonstrably an out-of sequence structure. However D4 folding and thrusting did not significantly alter the overall geometry obtained during the preceding deformation phases. Noteworthy this D4 phase represents the WNW-directed (orogen-perpendicular) shortening, which many authors (e.g. Platt, 1989) regarded as the predominant feature and only significant thrusting direction in the Western Alps. For the southern continuation of the Roselend thrust the reader is referred to Ceriani *et al.* (2000).

D5 is characterised by folding of the Roselend thrust. Along strike continuous steepening of the Roselend thrust plane, as it approaches to the Mont Blanc massif, is regarded as a large scale fold (F5). The geometry (knee fold) and the mechanism of formation (flexural slip) of this fold clearly demonstrate that the uplift of the Mt. Blanc massif postdates the activity along the Roselend thrust. Microstructural observations indicate that during the formation of the F5 fold a normal-fault component (top-to-the SE) took place within the steep limb. This different kinematic framework is regarded as a local and later overprinting of the Roselend thrust.

D6 is associated with a late (post 5 Ma) steeply SE-dipping brittle normal fault, partly parallel to and thus overprinting the former contact between the Valaisan units and the Zone Houillère (i.e. the Houiller Front). This normal fault can be traced further towards the S (Ceriani *et al* 2000) and is partly responsible for the wedging out of the Valais domain in map view south of Moûtiers.

The main conclusions of this thesis, as inferred from the combined sedimentological and structural approach can be summarized as follows:

1) The Valais domain marks the site of a second and more external oceanic domain that extended all along the arc of the Western Alps. The rather abrupt end in map view south of Moûtiers (France), of the Valais domain is due to tectonic rather than to a paleogeographic reasons (see Ceriani et al 2000).
2) The Valais domain marks the site of a second and more external subduction zone, separated from the internal high-pressure units of the Western Alps by the Zone Houillère. Its continuation further S is buried under the Subbriançonnais and Briançonnais units
3) The arc of the Western Alps was pre-structured by a NNE-SSW-striking corridor of oblique collision, associated with sinistral transpression during the late-Eocene, kinematically linked to the head-on collision in the E-W striking Central and Eastern Alps (see also Schmid and Kissling 2000).
4) The final structuration of the present-day “arc” of the Western Alps is mainly related to Oligo-Miocene orogen-perpendicular shortening associated with WNW-directed thrusting along the Roselend thrust.
5) Orogen-perpendicular extension characterises the latest stages of the evolution.

Reference

Alleman, F., Geologie des Fürstentums Liechtenstein (SW-licher Teil), [Dissertation]. Sonderdruck aus dem Jahrbuch des Historischen Vereins für das Fürstentums Liechtenstein, Universität Bern, 1957.

Allen, P.A., and J.R. Allen, *Basin Analysis*, Blackwell, Oxford, 1990.

Andrieux, J., and J.-C. Lancelot, Les plis "transverses" postpriaboniens dans la Zone des Brèches de Tarentaise, *C.R. Acad. Sc. Paris, t.290*, 1980.

Antoine, P., Quelques précisions sur le Trias moyen du massif du Roc de l'Enfer et des Antoines (Tarentaise), *C.R. Acad. Sci., 258*, 1850-1852, 1964.

Antoine, P., Les écailles frontales de la zone des Brèches de Tarentaise entre les Chapieux et le Haut Val Veni, *Trav. Lab. Géol. Grenoble, 41*, 163-175, 1965a.

Antoine, P., Observations nouvelles sur la termination méridionale des zones du Versoyen et du Pte. St. Bernard (Savoie), *C.R. Acad. Sci. Paris, 260*, 1213-1215, 1965b.

Antoine, P., Sur l'existence du Crétace supérieur daté dans la nappe des Brèches de Tarentaise au Nord des Chapieux (Savoie), *C.R. Acad. Sci. Paris, 261*, 3640-3642, 1965c.

Antoine, P., Nouvelles données sur la structure du flysch de Tarentaise entre Doire de Veni et Doire de Verney (Val d'Aoste), *Géol. alpine (Trav. Lab. Géol. Grenoble), 42*, 5-11, 1966.

Antoine, P., Sur la position structurale de la "zone du Versoyen" (Nappe des Brèches de Tarentaise, sur les confins franco-italiens), *Géol. Alpine (Trav. Lab. Géol. Grenoble), 44*, 6-26, 1968.

Antoine, P., Une nouvelle subdivision dans le domaine valaisan, en Tarentaise et dans le Haut Val d'Aoste, *C.R. Acad. Sci. Paris, 270*, 1548-1551, 1970.

Antoine, P., La Zone des Brèches de Tarentaise entre Bourg-Saint-Maurice (Vallée de l'Isère) et la frontière Italo-suisse, *Trav. Lab. Geol. Univ. Grenoble, Mémoire, 9*, p. 367, 1971.

Antoine, P., Le domaine pennique externe entre Bourg-Saint-Maurice (Savoie) et la frontière italo-suisse, *Géologie Alpine, Grenoble,, 48*, 5-40, 1972.

Antoine, P., R. Barbier, J. Debelmas, and S. Fudral, Précisions chronologiques et paléogéographiques sur les brèches du Massif du Grand-Fond (zone des Brèches de Tarentaise, Savoie), *Géol. alpine, 48*, 49-59, 1972.

Antoine, P., J.C. Barféty, G. Vivier, J. Debelmas, J. Desmons, J. Fabre, H. Loubat, and C. Vautrelle, Notice explicative, Carte géol. France (1/50 000), feuille Bourg-Saint-Maurice (727),,, p.110, 1992.

Antoine, P., J.C. Barféty, G. Vivier, Y. Gros, S. Fudral, P. Landry, and J. Fabre, Carte géol. France (1/50000), feuille Bourg Saint Maurice (727),,, 1993.

Barbier, R., Les zones subbriançonnaise et ultradauphinoise au N du Pelvoux., *Trav. Lab. Geol. Univ. Grenoble, Mémoire, 23*, 135-123, 1942.

Barbier, R., Les zones ultradauphinoise et subbrianconnaise entre l'Arc et l'Isère, *Mém. Serv. Carte géol. Fr.*, 291, 1948.

Barbier, R., Sur la decuverte de fossiles aux pyramides calcaires"(Haut Val d'Aoste) et aux Etroits du Saix (Tarentaise) et sur ses conséquences, *C.R. Acad. Sci, 232*, 748-749, 1951.

Barbier, R., and R. Trümpy, Sur l'age du Flysch de la zone des Brèches de Tarentaise, *Soc. géol. France Bull., 5*, 207-216, 1955.

Bocquet, J., and M.H. Delaloye, J.C.; Krummenacher, D., K-Ar and Rb-Sr Dating of Blue Amphiboles, Micas, and Associated Minerals from the Western Alps, *Contributions Mineralogy Petrology, 47*, 7-26, 1974.

Bousquet, R., L'exhumation des roches métamorphiques de haute pression - basse température : de l'étude de terrain à la modélisation numérique. Exemple de la fenêtre de l'Engadine et du domaine valaisan dans les Alpes Centrales, Ph.D. thesis, 279 pp., Université Paris XI - Orsay, Paris, 1998.

Bousquet, R., R. Oberhänsli, B. Goffé, L. Jolivet, and O. Vidal, High pressure-low temperature metamorphism and deformation in the Bündnerschiefer of the Engadine window: implications for the regional evolution of the eastern Central Alps, *Journal of Metamorphic Geology, 16*, 657-674, 1998.

Bucher, S., Strukturgeologische und petrographische Untersuchungen am Kleinen St. Bernard pass (Zone Valaisanne, unite du Pt. St. Bernard), *Diplomarbeit, Univ. Basel*, 90, 1999.

Cannic, S., L'évolution magmatique et tectono-métamorphique du substratum du domaine valaisan (complexe du Versoyen, Alpes occidentales): implications dans l'histoire Alpine, PhD thesis, 215 pp., Univ. Joseph Fourier Grenoble, LGCA, Grenoble, 1996.

Cannic, S., J.M. Lardeaux, J.L. Mugnier, and J. Hernandez, Tectono-metamorphic evolution of the Roignais-Versoyen Unit (Valaisan domain, France), *Eclogae geologicae Helvetiae, 89*, 321-343, 1996a.

Cannic, S., J.-M. Lardeaux, J.-L. Mugnier, and J. Hernandez, Tectono-metamorphic evolution of the Roignais-Versoyen unit (Valaisan domain, France), *Eclog. Geol. Helv., 89*, 321-343, 1996b.

Ceriani, S., B. Fügenschuh, and S.M. Schmid, Multi-stage thrusting at the "Pennine Front" in the Western Alps between Mont Blanc and Pelvoux massifs, *International Journal of Earth Sciences*, in press.

Dalla Torre, F., Petrographische und strukturgeologische Untersuchungen am Kleinen St. Bernard pass (Zone Valaisanne, Punta Rossa Komplex, unite du Pt. St. Bernard), *Diplomarbeit, Univ. Basel*, p. 100, 1998.

Debelmas, J., R. Caby, and J. Desmons, Notice explicative, Carte géologique de la France (1/ 50 000) Ste-Foy-Tarentaise (728)., BRGM, Orléans, 1991.

Debelmas, J., and M. Lemoine, Etat actuel de nos connaissances sur la stratigraphie du Trias dans le Briançonnais sensu stricto. Colloque sur le Trias de la france et des régions limitrophes., *Mém. Bur. Rech. géol. min.*, 232-242, 1961.

Dietrich, D., Casey, M., A new tectonic model for the Helvetic nappes, in *Alpine Tectonics*, vol. 45, edited by M.P. Coward and D. Dietrich & Park, R.,G. (eds), pp. 47-63, Spec. Publ. geol. Soc. London, 1989.

Ellemberger, F., Trias à facies briançonnais de la Vanoise et des Alpes occidentales., *Colloque sur le Trias de la France et des régions limithrophe Montpellier (1961), Mem. Bur. rech. Min.*, 215-231, 1958.

Eltchaninoff-Lancelot, C., S. Triboulet, B. Douboux, S. Fudral, J.P. Rampnoux, and M. Tardy, Stratigraphie et tectonique des unités delphino-helvétiques comprises entre Mont-Blanc et Belledonne (Savoie-Alpes occidentales). Implications régionales, *Bull. Soc. géol. Fr., 24*, 817-830, 1982.

Elter, G., La zona pennidica dell'alta e media Valle d'Aosta e le unità limitrofe, *Mem. Ist.Geol.Min. univ. Padova, 12*, 114, 1960.

Elter, G., and P. Elter, Carta geologica della regione del Piccolo S. Bernardo (versante italiano)., *Mem. Ist. Geol. Min. univ. Padova, 25*, p. 54, 1965.

Elter, P., Observations géologiques dans le Val veni (Versant italien du Mont Blanc), *Arch. sci. Genève, 4*, 427-429, 1951.

Elter, P., and G. Elter, Sull'esistenza, nei dintorni del Piccolo San Bernardo, di un elemento tettonico riferibile al ricoprimento del Pas du Roc., *Rend. dell'Acad. Naz. dei Lincei, XXII*, 181-187, 1957.

Fabre, J., Contribution a l'étude de la zone Houillère en Maurienne et Tarentaise (Alpes de Savoie)., *Mém. BRGM*, 315, 1958.

Florineth, D., and N. Froitzheim, Transition from continental to oceanic basement in the Tasna nappe (Engadine window, Graubünden, Switzerland): evidence for Early Cretaceous opening of the Valais ocean, *Schweiz. Min. Pet. Mitt., 74*, 437-448, 1994.

Franchi, S., Nuove località con fossili mesozoici nelle zone delle pietre verdi presso il colle del Pt. st. Bernardo (Valle d'Aosta), *Bull. del R. Comitato Geologico, 10*, 305, 1900.

Freeman, S.R., R.W.H. Butler, R.A. Cliff, S. Inger, and A.C. Barnicoat, Deformation migration in an orogen-scale shear zone array: an example from the Basal Briançonnais Thrust, internal Franco-Italian Alps, *Geological Magazine, 135*, 349-367, 1998.

Frey, M., J. Desmons, and F. Neubauer, The new metamorphic map of the Alps: Introduction, vol. 77, pp. 1-4, Schweizerisches Mineralogische und Petrographische Mitteilungen, 1999.

Frisch, W., Tectonic progradation and plate tectonic evolution of the Alps, *Tectonophysics, 60*, 121-139, 1979.

Froitzheim, N., Formation of recumbent folds during synorogenic crustal extension (Austroalpine nappes, Switzerland), *Geology, 20*, 923-926, 1992.

Froitzheim, N., S.M. Schmid, and M. Frey, Mesozoic paleogeography and timing of eclogite-facies metamorphism in the Alps: A working hypothesis, *Eclogae geologicae Helvetiae, 89*, 81-110, 1996.

Fudral, S., Contribution al'etude de l'unité de Moûtiers (zone des Brèches de Tarantaise) entre le vallon du torrent du Cormèt d'Arèches e le hameau des Chapieux, *Thèse III cycle, Univ. Grenoble*, 129, 1973.

Fudral, S., Une nouvelle interprétation de l'unité de Salins (zone des Brèches de Tarentaise) au nord-ouest de Bourg St. Maurice (Savoie). Consequences structurales, *C.R. Acad. Sc. Paris, 290*, 1333-1336, 1980.

Fudral, S., *Etude géologique de la structure téthysienne dans les Alpes franco-italiennes nord-occidentales de la doire Ripaire (Italie) a la région de Bourg Saint-Maurice (France)*, 306 pp., Laboratoire de Géologie de l'Université de Grenoble, Grenoble, 1998.

Fudral, S., and P.-L. Guillot, Découverte de clastes à mariposite dans les conglomérats de la base du "Flysch de Tarentaise" (zone valaisane), Savoie, France. Conséquences, *C.R. Acad. Sci. Paris, 306*, 911-914, 1988.

Fügenschuh, B., A. Loprieno, S. Ceriani, and S.M. Schmid, Structural analysis of the Subbriançonnais and Valais units in the area of Moùtiers (Savoy, Western Alps): paleogeographical and tectonic consequences, *International Journal of Earth Sciences, 88*, 201-218, 1999.

Galbraith, R.F., and G.M. Laslett, Statistical models for mixed fission track ages, *Nuclear Tracks and Radiation Measurements, 21*, 459-470, 1993.

Gely, J.P., Statigraphie, tectonique et métamorphisme comparés de part et d'autre du Front pennique en Tarentaise (Alpes de Savoie, France), PhD thesis, 343 pp., Chambéry, 1988.

Gely, J.P., and Y. Bassias, Le front pennique: implications structurales d'un métamorphisme transporté (Savoie, France), *C.R. Acad.Sci, 310*, 37-43, 1990.

Goffé, B., and R. Bousquet, Ferrocarpholite, chloritoïde et lawsonite dans les métapelites des unités du Versoyen et du Petit St Bernard (zone valaisanne, Alpes occidentales), *Schweizerisches Mineralogische und Petrographische Mitteilungen, 77*, 137-147, 1997.

Gourlay, P., La déformation du socle et des couvertures delphino-helvétiques dans la region du Mont-Blanc (Alpes occidentales), *Bull. Soc. géol. France, 2*, 159-169, 1986.

Guellec, S., J.-L. Mugnier, M. Tardy, and F. Roure, Neogene evolution of the Western Alpine foreland in the light of Ecors-data and balanced cross sections, in *Deep Structure of the Alps*, edited by F.H. Roure, P.; Polino, R., pp. 165-184, 1990.

Hubbard, R.J., Age and significance of sequences boundaries on Jurassic and early Cretaceous rifted continental margins, *American Association of Petroleoum Geologists Boullettin, 72*, p.49-72, 1988.

Jeanbourquin, P., and M. Burri, Les métasediments du pennique inférieur dans la region de Brigue-Simplon. Lithostratigraphie, structure et contexte géodinamique dans le bassin valaisan, *Eclogae Geologicae Helvetiae, 84*, 463-481, 1991.

Jullien, M., and B. Goffé, Cookeite and pyrophyllite in the Dauphinois black shales (Isère, France): implications for the conditions of metamorphism in the Alpine external zones, *SMPM, 73*, 357-363, 1993.

Hesse, R. Flysch-gault and Falknis-Tasna-Gault (Unterekreide): Kontinuierlicher Übergang von der distalen zur proximalen Flyschfazies auf einer penninischen Trogebene der Alpen. Geologica e Paleontologica, SB 2, 1-91, 1973.

Kelts, K., A comparison of some aspects of sedimentation and translational tectonics from the Gulf of California and the Mesozoic Tethys, Northern Pennine Margin., *Eclogae geologicae Helvetiae, 74*, 317-338, 1981.

Lancelot, J.C., Etude tectonique de la zone des brèches de Tarentaise dans la région de Bourg-Saint-Maurice, Savoie, Ph.D thesis, Paris, 1979.

Lasserre, J.L., and C. Lavergne, Le volcanisme tholéiitique de la zone du Versoyen (Alpes franco-italienne) : minéralogie, pétrographie et géochimie, Thèse de doctorat thesis, 252 pp., Grenoble, 1976.

Leikine, M., J.R. Kienast, C. Eltchaninoff-Lancelot, and S. Triboulet, Le métamorphisme polyphasé des unités dauphinoises entre Belledonne et Mont-Blanc (Alpes occidentales). Relation avec les épisodes de déformation., *Bulletin de la Société géologique de France, XXV*, 575-587, 1983.

Lomas, S., Submarine mass-flow conglomerates of the Tarentaise zone, Western Alps: sedimentation processes and depositional setting, *Sedimentary Geology, 81*, 269-287, 1992.

Loubat, H., Sur la présence de laves en coussins dans les phiolites du Versoyen (Savoie), *Arch. sci. Genève, 18*, 131-133, 1965.

Loubat, H., Etude pétrographique des ophiolites de la "zone du Versoyen"; Savoie (France), province d'Aoste (Italie), *These, Acr. Sci. Genève, 21*, 265-457, 1968.

Loubat, H., La zone du versoyen témoin possible d'une intersection entre dorsale oceanique et marge continentale, *Arch. Sci. Genève, 28*, 101-116, 1975.

Loubat, H., Considération preliminares sur la configuration horizontale de l' édificiesubvolcanique du versoyen (alpes franco-italiennes) analogue aux bassins en distension du type Golfe du Californie, *Géol. alpine (ex-Trav. Lab. Géol. de l'Université de Grenoble), 60*, 37-44, 1984.

Loubat, H., and P. Antoine, Sur l'extension des laves en coussins dans la région du Petit-Saint-Bernard (frontière franco-italienne), *Arch. sci. Genève, 18*, 689-692, 1965.

Loubat, H., and M. Delaloye, La zone du Versoyen (Alpes franco-italennes): le témoin due océanisation mésozoique circonscritte constituant un milieu hybride, subvulcano-sedimentaire, avec mobilisats et adinoles, *Géol. alpine (ex-Trav. Lab. Géol. de l'Université de Grenoble), 60*, 45-76, 1984.

Mennessier, G., J. Rosset, J. Bellière, R. Dhellemmes, N. Oulianoff, P. Antoine, and F. Carme, Carte géol. France (1/50 000) feuille Saint-Gervais-les-Bains (703). Notice explicative par Mennessier, G.; Rosset, J.; Bellière, J.; Dhellemmes, R.;Oulianoff, N.; Antoine, P.; Carme, F. (1976), 38 p., BRGM, Orléans, 1976.

Mosar, J., Géologie structurale à l'est de Montreux (Préalpes médianes plastiques, Suisse). Eclogae geol. Helv. 87, 11-32, 1994.

Mugnier, J.-L., Bergerat, F., Damotte, B., Guellec, S., Nicolas, A., Polino, R., Roure, F., Tardy, M., and Truffert, C.,, Crustal structureof the western Alps and their forelands,, *Mem. Soc. Geol. Fr.,, 170,*, 1996.

Mugnier, J.-L., S. Guellec, G. Ménard, F. Roure, M. Tardy, and P. Vialon, Crustal balanced cross-sections through the external Alps deduced from the Ecors profile, in *Deep Structure of the Alps*, edited by F.H. Roure, P.; Polino, R., pp. 203-216, Soc. Géol. France, Soc. Géol. Suisse, Soc. Geol. Italiana, 1990.

Mugnier, J.-L., H. Loubat, and S. Cannic, Correlation of seismic images and geology at the boundary between internal and external domains of the western alps, *Bulletin de la Société géologique de France, 164*, 697-708, 1993.

Nicolas, A., R. Polino, A. Hirn, R. Nicolich, and E.-C.w. group, ECORS-CROP traverse and deep structure of the Western Alps: a synthesis, in *Deep structure of the Alps*, vol. 156, edited by F. Roure, P. Heitzmann and R. Polino, pp. 15-28, Mémoire de la Société géologique de France, Paris, 1990.

Oberhänsli, R., B. Goffé, and R. Bousquet, Record of a HP-LT metamorphic evolution in the Valais zone: Geodynamic implications, *Boll. Museo Regionale Sci. Nat. Torino, 13*, 221-239, 1995.

Perello, P., Piana, F., Martinotti G., Neo-Alpine structures at the Pennine-Helvetic boundary, *Eclogae Geologicae Helvetiae, 92*, 347-361, 1999.

Platt, J.P.B., J.; Cunningham, P.C.; Dewey, J.F.; Helman, M.; Parish, M.; Shepley, M.G.; Wallis, S.; Weston, P.J., Kinematics of the Alpine arc and the motion history of Adria, *Nature, 337*, 158-161, 1989.

Ramsay, J.G., *Folding and fracturing of rocks*, McGraw-Hill, 1967.

Ramsay, J.G., Fold and fault geometry in the western Helvetic nappes of Switzerland and France and its implication for the evolution of the arc of the western Alps, in *Alpine Tectonics*, vol. 45, edited by M.P. Coward and D. Dietrich & Park, R.,G. (eds), pp. 33-45, Spec. Publ. geol. Soc. London, 1989.

Reading, H.G., *Sedimentary environments and facies*, Blackwell, Oxford, 1989.

Ricou, L.E., and A.W.B. Siddans, Collision tectonics in the Western Alps, in *Collision Tectonics*, edited by M.P.R. Coward, A.C., pp. 229-244, Geological Society Spec. Publ. N.19, 1986.

Roure, F., P. Heitzmann, and R. Polino, *Deep structure of the Alps*, 367 pp., Mémoires de la société géologique de France, Paris, 1990.

Saliot, P., La jadéite dans les Alpes françaises, *Bulletin de la Société française de Minéralogie et de Cristallographie, 102*, 391-401, 1979.

Schmid, S.M., and E. Kissling, The arc of the Western Alps in the light of new data on deep crustal structure, *Tectonics, 19*, 62-85, 2000.

Schoeller, H., La nappe de l'Embrunais au nord de l'Isère avec quelquelques observations sur les régions voisines: bord externe de la nappe du Briançonnais, zone dauphinoise, lambeaux de recouvrement de Sulens (Feuille de Bourg St. Maurice à 1/50 000)., *Bull. Serv. Carte gèol. Fr., 33*, 422, 1929.

Schürch, M.L., Les ophiolites de la zone du versoyen: témoin d'une bassin à évolution métamorphique complexe, PhD thesis, 157 pp., Univ. de Genève, Genève, 1987.

Schwizer, B., Die Tristel-Formation. Vergleichende Untersuchung in Graubünden, Liechtenstein, Voralberg und Bayern. Ph. D. Thesis, Bern University, 186 pp, 1983.

Seward, D., and N.S. Mancktelow, Neogene kinematics of the central and western Alps: Evidence from fission-track dating, *Geology, 22*, 803-806, 1994.

Simpson, C., and S.M. Schmid, An evaluation of criteria to deduce the sense of movement in sheared rocks, *Geol. Soc. Am. Bull., 94*, 1281-1288, 1983.

Sodero, D., Sull' età Barrêmiano-Aptiano delle formazioni basali del "Flysch" della Zona delle Breccie di Tarentasia in Valle d'Aosta., *Boll. Soc. Geol. It., 87*, 223-231, 1968.

Spencer, S., Nature of the North Pennine Front: French Alps., PhD. thesis, Imperial College, London Univ., London, 1989.

Stampfli, G.M., Le Briançonnais, terrain exotique dans les Alpes?, *Eclogae geologicae Helvetiae, 86*, 1-45, 1993.

Stampfli, G.M., and M. Marthaler, Divergent and convergent margins in the North-Western alps confrontation to actualistic models, *Geodynamica Acta, 4*, 159-184, 1990.

Stampfli, G.M., J. Mosar, D. Marquer, T. Baudin, and G. Borel, Subduction and obduction processes in the Swiss Alps, *Tectonophysics, in press*, 1998.

Steinmann, M., Ein Beckenmodell für das Nordpenninikum der Ostschweiz, *Jb. Geol. B.A., 137*, 675-721, 1994.

Steinmann, M., and P. Stille, Geochemical evidence for the nature of the crust beneath the eastern North Pennine basin of the Mesozoic Tethys ocean, *Geologische Rundschau, 87*, 633-643, 1999.

Tardy, M., E. Deville, S. Fudral, S. Guellec, F. Thouvenot, and P. Vialon, Interprétation structurale des données du profil de sismique réflexion profonde Ecors-Crop Alpes entre le front Pennique et la ligne du Canavese (Alpes occidentales), in *Deep Structure of the Alps*, edited by F.H. Roure, P.; Polino, R., pp. 217-226, Soc. Géol. France, Soc. Géol. Suisse, Soc. Geol. Italiana, 1990.

Trümpy, R., Sur le racines helvétique et les "schistes lustres" entre le Ròhne et la Valleé de Bagnes, (Région de la Pierre Avoi), *Eclogae Geologicae Helvetiae, 44*, 338-347, 1952.

Trümpy, R., La zone de Sion-Courmayeur dans la haut Val Ferret valaisan, *Eclog. geol. Helv., 47*, 315-359, 1954.

Trümpy, R., Remarques sur la corrélation des unités penniques externes entre la Savoie et le Valais et sur l'origine des nappes préalpines, *Soc. géol. France Bull., 5*, 217-231, 1955.

Trümpy, R., Du Pélerin aux Pyrénées, *Eclogae Geologicae Helvetiae, 69*, 249-264, 1976.

Trümpy, R., A possible Jurassic-Cretaceous transform system in the Alps and the Carpathians, in *Processes in Continental Lithospheric Deformation*, vol. 218, edited by S.P. Clark, B.C. Burchfiel and J. Suppe, pp. 93-109, Geological Society of America Special Paper, Boulder, 1988.

Walker, R.G., Turbidite sedimentary structures and their relationships to proximal and distal depositional environments, *Journal of Sedimentary Petrology, 37*, 25-43, 1967.

Walker, R.G., Facies models 2. turbidites and associated coarse clastic deposits, *Geoscience Canada, 3*, 25-36, 1976.

Williams, G.D., Tectonics and sismic sequence stratigraphy: an introduction, in *Tectonics and Seismic Sequence Stratigraphy*, vol. 71, edited by G.D. Williams and A. Dobb, pp. 1-15, Geological Society Special Publication, London, 1993.

Yamada, R., and T.N. Tagami, S.; Ito, H., Annealing kinetics of fission tracks in zircon: an experimental study, *Chemical Geology (Isotope Geoscience Section), 122*, 249-258, 1995.

Zulauf, R., Zür geologie der tiefpenninischen zonen nördlich der Dora Baltea im oberen Val d'Aosta (Italien), *These, Zürich*, 150, 1964.