

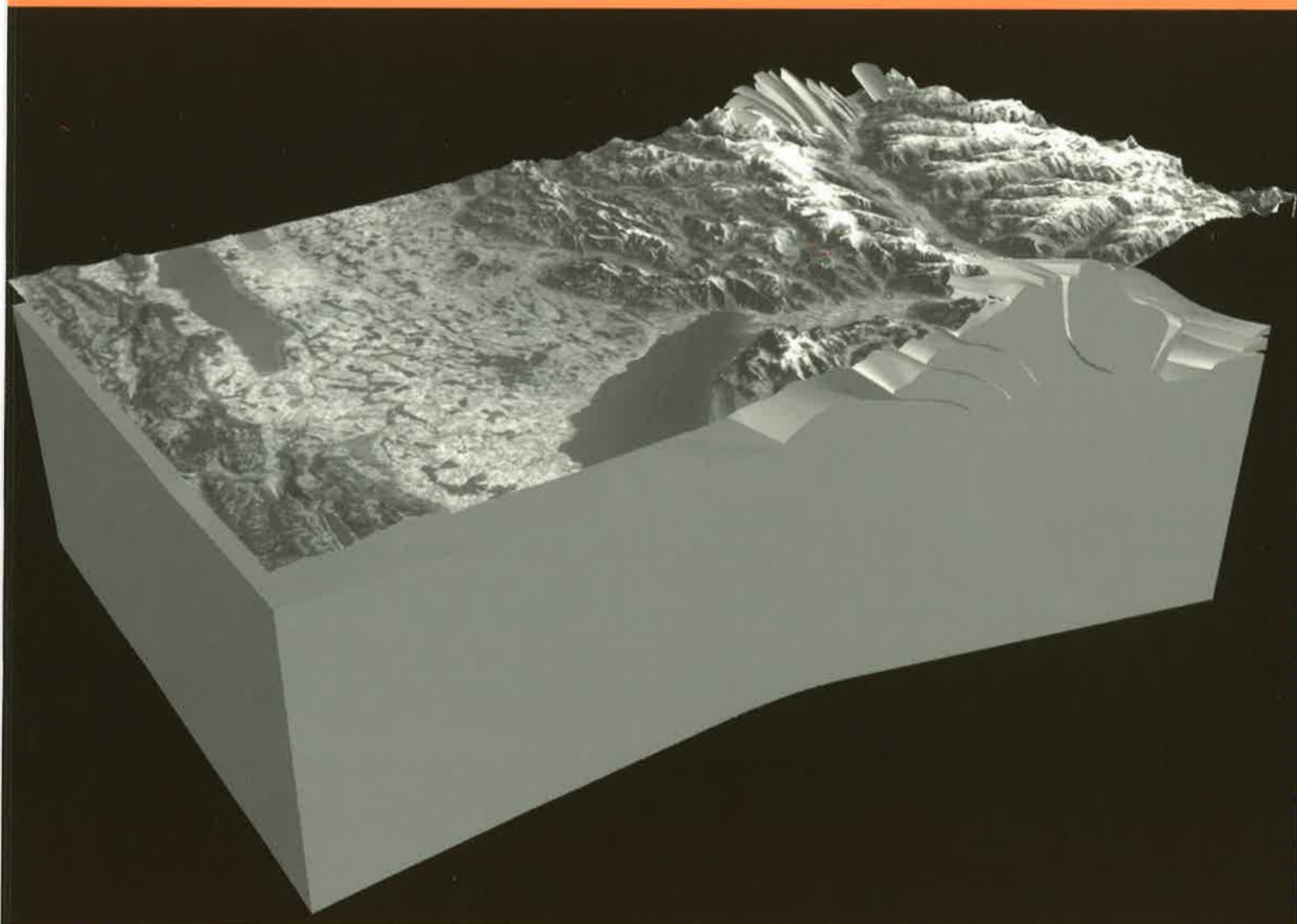
Geology of the western Swiss Alps, a guide-book

edited by: Gérard M. Stampfli

with contributions by:

G. D. Borel, F. Bussy, N. Capuzzo, J-L. Epard, A. Escher, R. Marchant, M. Marthaler, J. Mosar,
Y. Ringgenberg, G. M. Stampfli, A. Steck, P. Thélin & J. von Raumer

technical editor: B. G. Matti



Mémoires de Géologie (Lausanne)

Section des Sciences de la Terre
Université de Lausanne

BFSH-2, 1015 Lausanne, Suisse



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Institut de Géologie et Paléontologie
BFSH-2 Université de Lausanne
CH-1015, Lausanne SUISSE

COMITÉ EDITORIAL

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345 Middlefield Road
94025 Menlo Park, California, U.S.A.

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8006 Zürich

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10125 Torino ITALIE

Mémoires de Géologie (Lausanne)
Section des Sciences de la Terre
Institut de Géologie et Paléontologie
Université de Lausanne
BFSH-2, CH-1015 Lausanne

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A guide-book



Editor

Gérard M. Stampfli

With contributions by

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Technical editor

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Foreword

The contributors and myself gathered their effort and knowledge to produce the present volume when it was decided that IGCP project 453 would have its second field meeting in Sion, in the midst of the western Swiss Alps. Therefore, this guide book presents 6 field trips which are at the programme of the meeting. The field trips are divided in two sections corresponding to the Alpine Externides and Internides, allowing the visitors to get a good overview of the entire western Swiss Alps.

The theme of IGCP 453 is: *Uniformitarianism revisited: a comparison between modern and ancient orogens*, under the leadership of Brendan Murphy and Duncan Keppie. It was obvious then, that a good share of basement geology should be included in the guide, and two field trips (Ft1 and Ft5) are mainly devoted to Variscan and pre-Variscan rocks outcrops.

We felt that it was also necessary to present in such a document the latest ideas on Alpine and pre-Alpine geology of the western Tethys realm. Parts of recent publications were used for that matter, they were updated, and put in a general introduction with numerous figures, mainly devoted to the geodynamics of this area in space and time.

Coloured Alpine cross-section and block diagram, and excerpts of the new 100'000 structural map of Western Switzerland have been put in annex, in a pouch on the back-cover. Similar excerpts of the topographic map of Switzerland, on which stops are located, accompany each field trip.

Making a guide-book for such a geological monument as the western Swiss Alps is quite a challenge... and obviously the present guide is presenting only a small part of what can be seen in this area of the Alps. We hope you will enjoy visiting our mountains, and hopefully get some insights on the fascinating geological history hidden behind those magnificent landscapes.

G rard M. Stampfli

-Part I : General introduction-

Geology of the western Swiss Alps

1 - INTRODUCTION

If the present Alpine structures are quite intricate and the number of structural units certainly too large to be memorized by people visiting the Alps for the first time, it is still possible to enter this geological esoteric field through the concept of continental margin evolution. Alpine geology has moved recently from pure structural analyses and cylindrical paleogeography to modern concepts of plate tectonic and basin evolution, and structural units can be grouped according to geodynamic settings.

After a century and a half of detailed structural work, the general nappe framework of the western Alps is now well understood and a common structural model accepted by most alpine geologists. This structural model benefited from recent deep seismic projects carried out by French, Italian and Swiss teams (*e.g.* Pfiffner et al. 1997); together with tomographic data, these studies allow to understand the Alpine structures also at crustal and lithospheric scales (fig. 1). It is now possible to reconstruct the former geometries of the continental margins involved in the formation of the western Alps, and constrain their evolution through detailed stratigraphic and sedimentological analyses of their deposits. Pre-Alpine palinspastic models can in turn be used to unravel the Variscan and pre-Variscan history of this part of the European basement.

This field trip guide presents this evolution, with an emphasis on the works carried out at the Earth Science section of the University of Lausanne. In such a guide book an exhaustive reference list

cannot be given, mainly considering that for example the Préalpes were already at the center of discussions more than a century ago... and that the Alps extends from the Carpathian domain to the Mediterranean sea. For compilation works on the geodynamics of the Alpine Tethys and the western Swiss Alps the reader is invited to consult reference lists in: Escher et al. (1997), Marchant & Stampfli (1997a), Stampfli & Marchant (1997) and Stampfli et al. (1998) and for the Préalpes: Mosar et al. (1996) and Borel & Mosar (2000); a very exhaustive list of references on the Préalpes can also be found on our web site: www-sst.unil.ch/marge&co/prealpes/referen.htm

Regarding Alpine structures and metamorphism in the western Swiss Alps, most references are found in Steck et al. (2001). For the Variscan basement relevant references can be found in von Raumer & Neubauer (1993) von Raumer et al. (1999) and Stampfli (1996). Regarding the plate tectonic reconstructions presented here for the western Tethyan realm, main references can be found in Stampfli & Mosar (1999), Stampfli (2000), Stampfli et al. (2001a, b, c & d).

A series of field guides of the western Alps have also been published and may be of interest for further exploration of the Alps:

Valais (Steck et al. 1979),

Alpes, Savoie et Dauphiné (Debelmas 1979),

Suisse Lémanique, Pays de Genève et Chablais (Charollais & Badoux 1990)

Alpi et Prealpi Lombarde (Cita et al. 1990),

Le Alpi dal M. Bianco al Lago Maggiore (Dal Piazz 1992)

Géologie de la Suisse (Labhart & Decrouez

1997) and *Carte tectonique des Alpes de Suisses occidentales* (Steck et al. 2001)

1.1 THE GEOLOGICAL BACKGROUND OF THE WESTERN ALPS

When analyzing a mountain belt mainly by means of cross-sections one is easily led to propose palinspastic models corresponding to *in situ* unfolding of structural units. Doing so for the Alps, most authors proposed a multi-ocean model with oceanic strips opening simultaneously and parallel to each other, separating micro-continents whose sizes are often of a few tens of kilometres! Actualistic models have shown that this approach is not valid anymore and that a mountain belt is made of structural elements corresponding to geodynamic stages like divergent and convergent plate tectonic processes responsible for the present structures (Stampfli & Marthaler 1990). Formerly

the complexity of such a mountain belt was regarded as directly related to a complex paleogeography, when actually it is due to a complex structural history. In that respect we can regard the western Alps as an accretionary prism where different geological objects, corresponding to different stages of accretion, can be recognised:

- oceanic accretionary prism of the Piemonte ocean, including crustal elements from the former toe of the southern passive margin (lower Austroalpine elements),
- accretion of the Briançonnais terrain derived from the Iberic plate,
- accretion of the Valais domain,
- accretion of material from the former European continental margin (Helvetic domain).

Then one passes to the formation of the orogenic wedge proper that we place after the detachment of the subducting slab in the Early Oligocene (*e.g.* Stampfli & Marchant 1995). The resulting heat flux allowed some more units to be detached from the

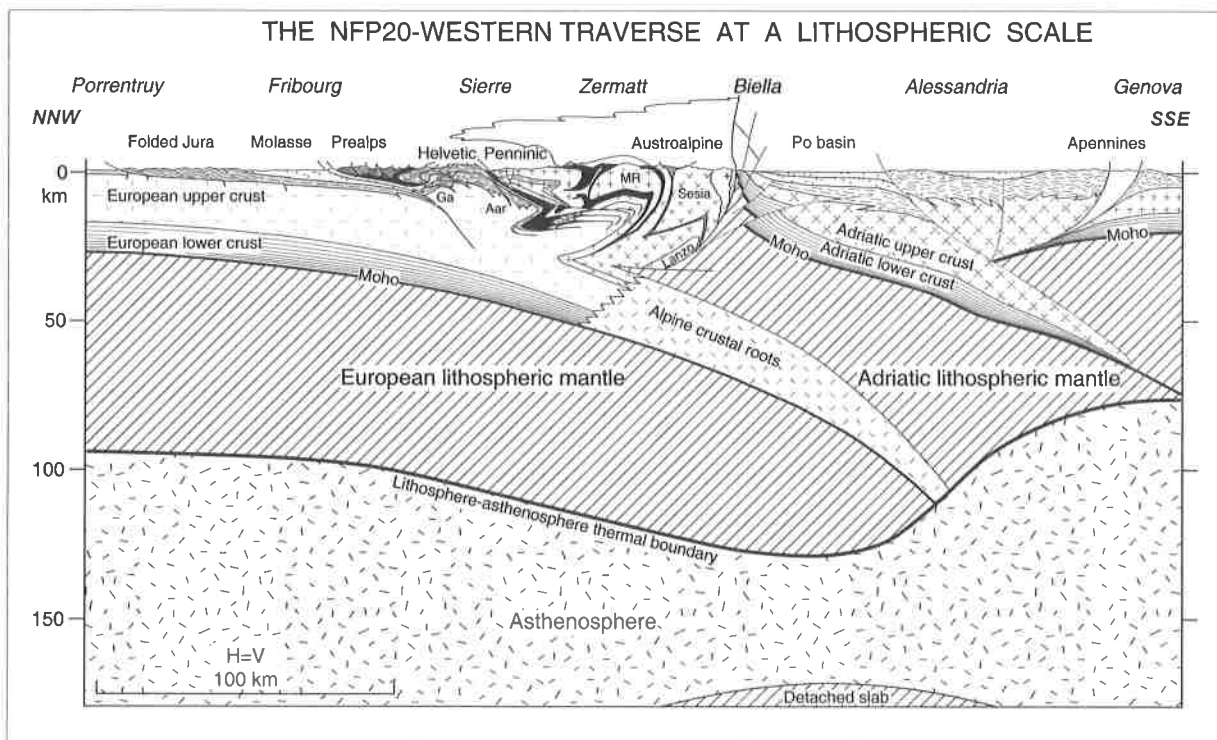


Fig.1 - Lithospheric cross-section of the western Alps, modified from Marchant (1993) and Marchant & Stampfli (1997), location shown on figure 2 (broad white strip).

European continental slab and resulted in large scale subduction of continental material (Marchant & Stampfli 1997b) and obduction of the more external units:

- external Variscan massifs and their cover,
- molassic basin,
- Jura mountains.

To these accretionary events one has to add other tectonic processes such as the “Pyrenean” inversion phase (Late Cretaceous-Middle Eocene) that affected the Helvetic margin, and the flexuration event of the lower European plate (Paleocene to Miocene).

Following this geodynamic scheme, the structural model as presented in figure 1 can be seen as composed of remnants of paleogeographic units whose main bodies were subducted (more than 90%). Each tectonic package is formed of obducted material with a geodynamic signature which allows to replace these structural units in their former paleogeographic domain (*i.e.* rim basin, rift shoulder, passive margin, oceanic sequences, inversion basin, flexural bulge). For a long time structural units have been grouped into relatively cylindrical patterns without producing nonsensical models, but in terms of paleogeography, lateral displacements of terrains should be taken into account. In doing so, the cylindrical structural pattern explodes. Then, its elements have to be grouped together again and paleogeographic and geodynamic markers (rift shoulders, synrift deposits, denuded mantle) can be used to reconstruct the diverging phases, whereas paleo-structures, ages of flysch deposits, of exotic terrain deposits and of metamorphism as well as structural indicators, are used to reconstruct converging processes. This palinspastic exercise eventually leads to a pre-Alpine continental fit which, in turn, can be used to decipher the Variscan and pre-Variscan history. The onset of the Alpine cycle proper could be placed when the Alpine Tethys ocean opened, *i.e.* in Early to Middle Jurassic time (Favre & Stampfli

1992; Bill et al. 1997). However, the Tethyan cycle *s.l.* is older, it is separated from the Variscan cycle by a complex Permo-Triassic geological evolution. The complexity comes from the non cylindrical and non uniform evolution of the Variscan domain in space and time. If the Variscan cycle in the Appalachian domain is terminated by the final collision of Gondwana with Laurasia in Permian times, this is not the case eastward. Particularly, in the Alpine-Mediterranean area, continuing northward subduction of a remnant Paleotethys in Permo-Triassic times led to the opening of back-arc basins. The nearest of these basins to the Alps is the Meliata marginal ocean (*e.g.* Kozur 1991), whose northern passive margin is represented presently by the Northern Calcareous Alps of Austria and Germany (the upper Austroalpine domain is not present in the western Alps - Faupl & Wagneich 1999). These marginal basin openings were preceded by pervasive rifting and collapse of the Variscan cordillera in Late Carboniferous-Permian times (Ziegler & Stampfli 2001). These crustal attenuation processes permitted the sea to slowly transgress former Variscan reliefs since Early Permian times. This Permo-Triassic period also corresponds on the Gondwana side to the opening of the large Neotethys ocean, therefore, the Tethyan *s.l.* sedimentary cycle is usually regarded as starting during the Permian (Stampfli 2000).

From an Alpine point of view and also due to former nomenclature, the lower part of the Tethyan cycle (Permo-Carboniferous rift related sequences) is often considered by Alpine geologists as pertaining to basement, whereas the cover units have their base quite often in Triassic evaporites or younger levels. This is a fundamental difference between Alpine geology *s.str.* and Tethyan geology *s.l.*, or between the Alpine orogen (Alps and Carpathes) and the Tethysides (Dinarides-Hellenides, the Middle-East mountain belts and the Himalayas *s.l.*)

parautochthonous cover represent the substratum of the Subalpine molasse. The Mesozoic sequence is more distal than the Jura sequence, but the lower part is missing (Triassic evaporites, Liassic platform). This area was a high during the Alpine Tethys rifting as a consequence of tectonic unloading to the south of it in the Helvetic rim basin.

The Mont Blanc crystalline massif and its cover the Morcles nappe, represent the northern part of the Helvetic domain. Those crystalline massifs represent the Variscan basement of the northern fringe of the orogen. They are made of metamorphic rocks in amphibolitic facies, Permo-Carboniferous clastics and Permo-Carboniferous granitoids.

The Helvetic nappes (field trips 2 & 3)

This classical domain of the Swiss Alps (Masson et al. 1980) consists of the Morcles (Doldenhorn), Ardon, Diablerets and Wildhorn nappes. The first two were deposited on the Mont Blanc massif, the décollement level being generally the Toarcio-Aalenian shales. The Mesozoic sequence of these nappes is different from the Jura-parautochthonous sequence. The Lias-Dogger sequence is influenced by crustal extension which resulted in the formation of the Helvetic rim-basin to the north of the Piemont rift shoulder. The Late Jurassic and Cretaceous sequence is marked by the progradation of a carbonate platform which never succeeded filling-up the Wildhorn basin and never reached the distal Helvetic margin. Major uplift took place in Late Cretaceous and Paleocene time, related to the flexure of the European margin and to the Pyrenean movements. A general transgression of the fore-bulge took place during the Eocene forming a new carbonate platform which gave way progressively to the flysch deposits (see figure 11). In France the Helvetic domain is called Dauphinois.

The Ultrahelvetic nappes (field trip 2, 3 & 6)

These nappes are found in several locations: first as a sole of the Préalpes massif; as klippen on the Helvetic nappes and in the Helvetic root zone. They represent distal portion of the southern part of the Helvetic rim basin, the rift shoulder area and the attenuated northern rifted margin of the Alpine Tethys. The Mesozoic sequence contains Triassic evaporites, a Liassic-Dogger platform sequence and a condensed, starved Late Jurassic-Cretaceous pelagic sequence.

The lower Penninic nappes (field trips 4 & 5)

These highly metamorphic, mainly crustal nappes represent the former rift shoulder and syn-rift domains (*i.e.* tilted blocks) of the European margin. They consist of the Verampio, Antigorio and Monte Leone crystalline nappes with incomplete Mesozoic cover and the Lebendun nappe composed of thick clastic deposits attributed to former syn-rift deposits (Spring et al. 1992). The Ultrahelvetic and Niesen nappes were certainly partially deposited on this lower Penninic domain. The Niesen nappe contains Jurassic syn-rift deposits at its base and then inversion related Late Cretaceous clastics.

The Valais suture zone (field trips 4 & 5)

Coming from this relatively thin structural domain we find mélanges like the Submédianes zone of mixed elements of the Valais trough and distal Helvetic margin origin, found as exotic elements in the Préalpes. The distal clastics and pelagic Cretaceous deposits of the Sion-Courmayeur zone would have been deposited on the Helvetic distal margin at the ocean-continent transition zone (the Valais trough). Denuded continental mantle units like the Geisspfad massifs represent this transition found within the lower Penninic nappes named Valaisan domain in the paleogeographic nomenclature. The infra-Moncucco ophiolitic zone represents Piémont oceanic crust trapped between the Helvetic margin

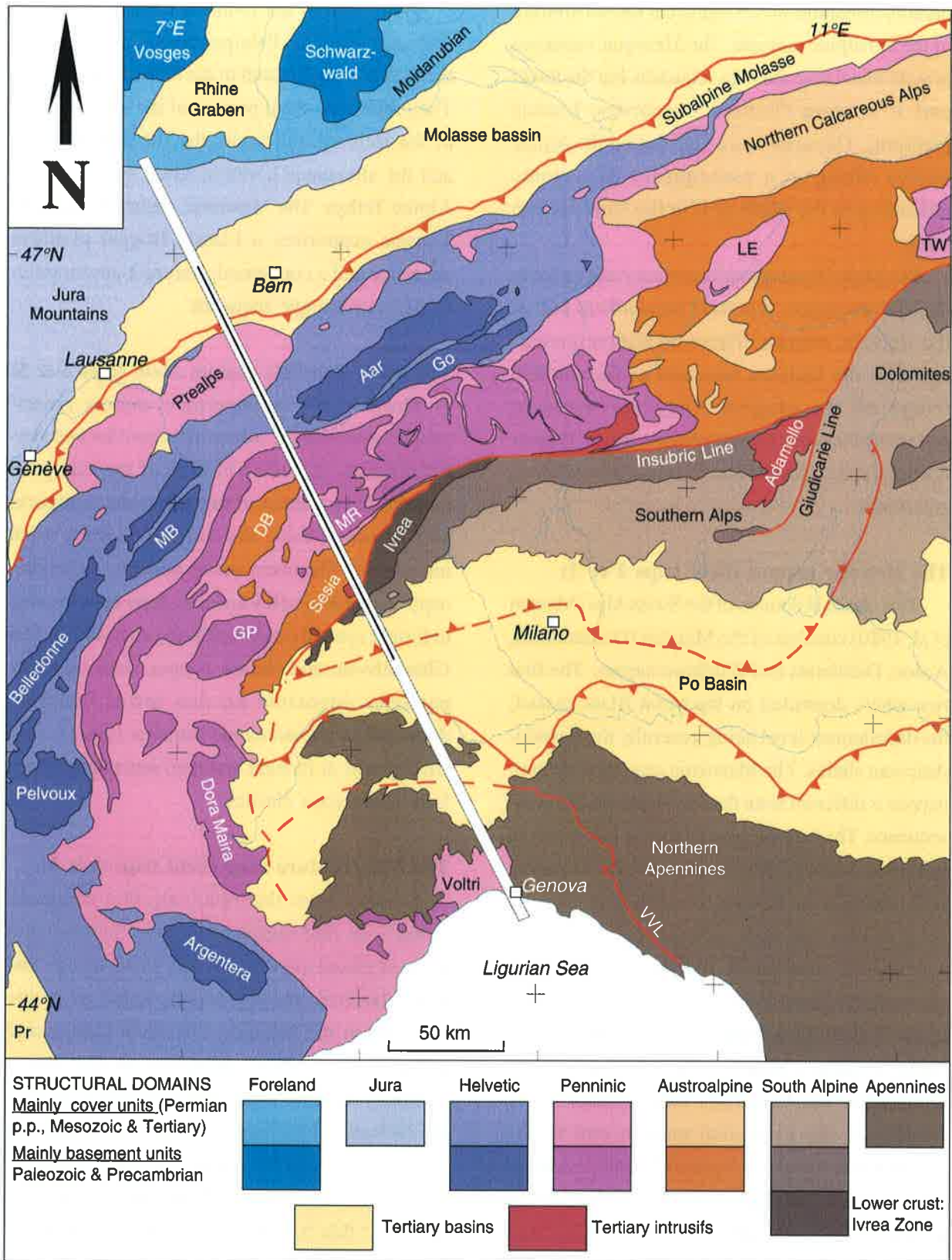


Fig. 2 - Tectonic map of the Western Alps modified from Berthelsen (1992). Ao = Adamello intrusions; DB = Dent Blanche nappe; Go = Gotthard massif; GP = Grand Paradis massif; LE = Lower Engadine window; MB = Mont Blanc massif; MR = Monte Rosa nappe; Pr = Provence basin; TW = Tauern window; VVL = Villalvernia-Varzi-Levanto line. Broad white strip: lithospheric cross-section (fig. 1); thin black line: cross-section of the western Alps (fig. 3)

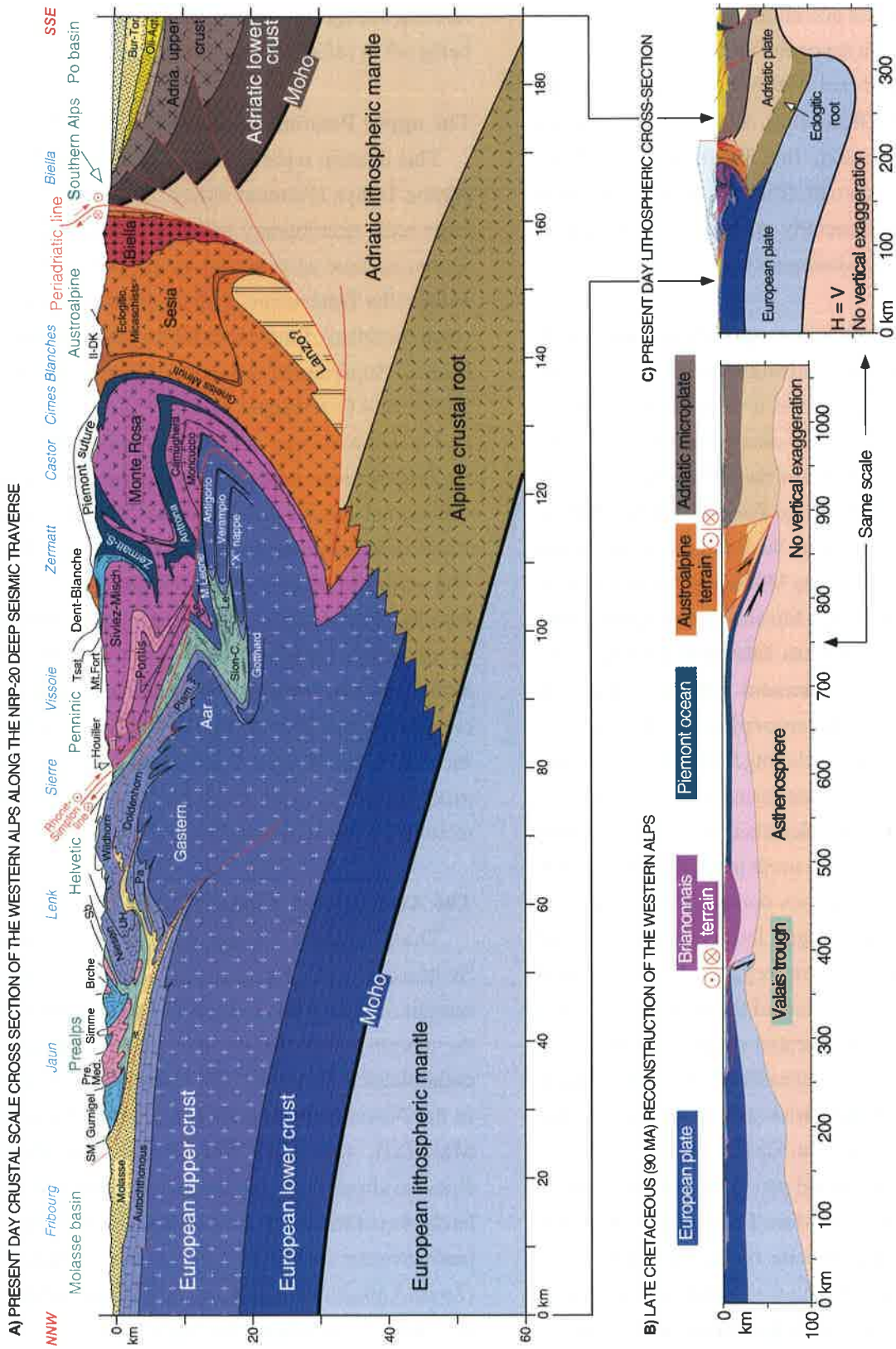


Fig.3 - Cross-section of the western Alps on a western Switzerland transect, and simplified palinspastic model, modified from Marchant (1993) and Marchant & Stampfli (1997). Location shown in figure 2 (thin black line).

and the Briançonnais exotic terrain. This suture is better expressed in central Switzerland where the Bündnerschiefer accretionary prism (between the European Adula nappe and the Briançonnais Tambo nappe; Schmid et al. 1990) contains a fair amount of Piemonte MORB relicts. A Late Eocene mélange is found between the Valais zone and the Briançonnais domain (Pierre Avoi unit, Bagnoud et al. 1998). Generally speaking this domain is poorly dated (Ackermann et al. 1991).

The middle Penninic nappes (field trips 4 & 5)

This domain belonged formerly to the Briançonnais terrain and is exotic in regard to the more external units, an observation already made by Schardt at the end of the 19th century (Masson 1976). It is made of the Zone Houillère (Permo-Carboniferous graben), the Pontis basement nappe on which the Préalpes Médiannes Plastiques were deposited, the Siviez-Mischabel basement nappes on which the Préalpes Médiannes Rigides were deposited. These basement nappes are made of Variscan polymetamorphic rocks with a sedimentary cover including Permo-Carboniferous clastic deposits transgressed by a Mesozoic sequence, most of it detached from its substratum and transported to the north in the Préalpes region. The Préalpes Médiannes domain was a rim basin of the European margin, located north of the rift shoulder. It can be regarded as a lateral southwestern equivalent of the Helvetic rim basin and was formerly located south of France. The duplication of these rim basin/rift shoulder elements on a western Alpine cross-section is a fundamental feature of the western Alps.

In the more internal part of the middle Penninic units are found the Mont-Fort nappe (located in the Valais) and Brèche nappe (located in the Préalpes), representing the former syn-rift part of the European margin on the Briançonnais transect. This type of unit are sometime called pre-Piemontais units as they represent the transition to

the Piemonte ocean (*i.e.* the western Alps part of the Alpine Tethys ocean, its Italian part being referred as Ligurian ocean; its Austrian equivalent being often referred to as Penninic ocean)

The upper Penninic nappes (field trips 4 & 5)

This domain represents the suture zone of the Alpine Tethys (Piemonte suture). It is made of a large scale accretionary mélange comprising the former oceanic accretionary wedge of the Alpine Tethys (the Tsaté nappe, with relicts of blue schist metamorphism), from which the non-metamorphic Nappes Supérieures are derived (the Gurnigel, Simme and Gets nappes of the Préalpes domain) and a zone of (eo-Alpine?) Alpine HP-LT metamorphism made of mixed oceanic and continental crustal units (Zermatt-Saas Fee and Antrona ophiolitic nappes and the Internal Massifs). The eclogitic internal massif (Mont Rose, Gran Paradiso, Dora Maira) are continental exotic blocks of disputed origin (lower Austroalpine or Briançonnais), underplated in the accretionary complex, sometimes at great depth (over 100 km for UHP Dora-Maira eclogites) and subsequently mixed with oceanic elements in a westward tectonic escape movement during their extrusion.

The Austroalpine nappes (field trips 4 & 5)

These mainly crustal nappes in western Switzerland represent the former southern passive margin of the Alpine Tethys. The thinned part of the margin was probably subducted and partially underplated to form the exotic elements now found in the Piemonte suture zone (parts of the Internal Massifs?). The rest of this domain (the Dent Blanche klippe and Sesia zone) formed the former back-stop of the Tsaté accretionary prism and was partly overthrust by the Adriatic micro-continent (South Alpine units). Subsequently this overthrust was largely deformed by back-folding of the orogenic wedge and the indentation of the Adriatic plate. Major lateral movement of these units took

place during the Late Cretaceous and they can be regarded as displaced terrains in regard to Adria. The Sesia zone records Late Cretaceous HP-LT metamorphism.

The South Alpine domain

This domain represents the northern margin of the Adriatic microplate, its northeastern part (Canavese) collided with the displaced Austroalpine elements during their westward escape in Late Cretaceous. So here too, we have a potential duplication of the southern marginal domain and a possibility of trapped oceanic rocks between the Canavese and Sesia units. The Adriatic lower crust is outcropping in the Ivrea zone which forms the present day back-stop of the orogen. This back stop was strongly affected by two events: the back folding of the internal units since the Oligocene and the emplacement of the peri-Adriatic plutonic rocks derived from the detachment of the Tethyan slab more or less at the same time.

2 - THE GEODYNAMIC FRAMEWORK OF THE WESTERN ALPS

Introduction

In this chapter we present first the Tethyan-Alpine history in order to allow the reader to understand the complexity of the Alpine puzzle and see how one can use geodynamic and basin analysis criteria to propose a palinspastic model. This model leads to a Permian Pangea fit which in turn can be used to unravel the Variscan history which is presented in a second part

2.1 The Alpine cycle

G.M. STAMPFLI, J. MOSAR, R. MARCHANT & G. BOREL

2.1.1 The plate tectonics of the western Tethyan regions

The reconstructions shown in figure 4 are based

on a pre break-up fit as well as magnetic anomalies from the Central Atlantic. They have been constructed in a continuous effort to apply plate tectonic concepts to palinspastic model of the western Tethys. The constraints and data bases used in that project can be found in the literature cited above in the introduction. This plate model takes into account the likelihood of a Late Paleozoic rifting and sea-floor spreading of the eastern Mediterranean basin. This opening would be concomitant with the opening of the Neotethys and the drifting of the Cimmerian continents since late Early Permian. This model considers also a late closure of the Paleotethys (Middle to Late Triassic) on a Greek and Turkish transect of the Tethyan realm, accompanied by the opening of back-arc basins (fig 4 A, Meliata, Maliac and Pindos back-arc basins).

The Apulia-Adria problem

A review of paleomagnetic data regarding the Alpine and Mediterranean area suggests that the method cannot sort out paleopositions of small terrains having suffered small amounts of displacement. It is possible, however, to show that the Apulian plate *s.l.* (Italy) suffered relatively little rotation in regard to Africa since the Triassic (Channell 1992; Channell et al. 1992; Channell & Doglioni 1994; Channell 1996). This leaves open the question of an Apulian plate being an African promontory or not. Also, the continuity between the active subduction zone under Greece and the outer Dinarides (Wortel & Spakman 1992; de Jonge et al. 1994) shows that there is a possible plate limit between Apulia *s.l.* and the autochthonous of Greece.

Together with major problems concerning the reassembling of micro-plates in a pre-break-up position (fig. 4), this led us to consider that the Apulian plate *s.l.* is most likely cut into two pieces, an Apulian plate *s.str.* to the south and an Adriatic plate *s.str.* to the north. The Apulian part is definitely

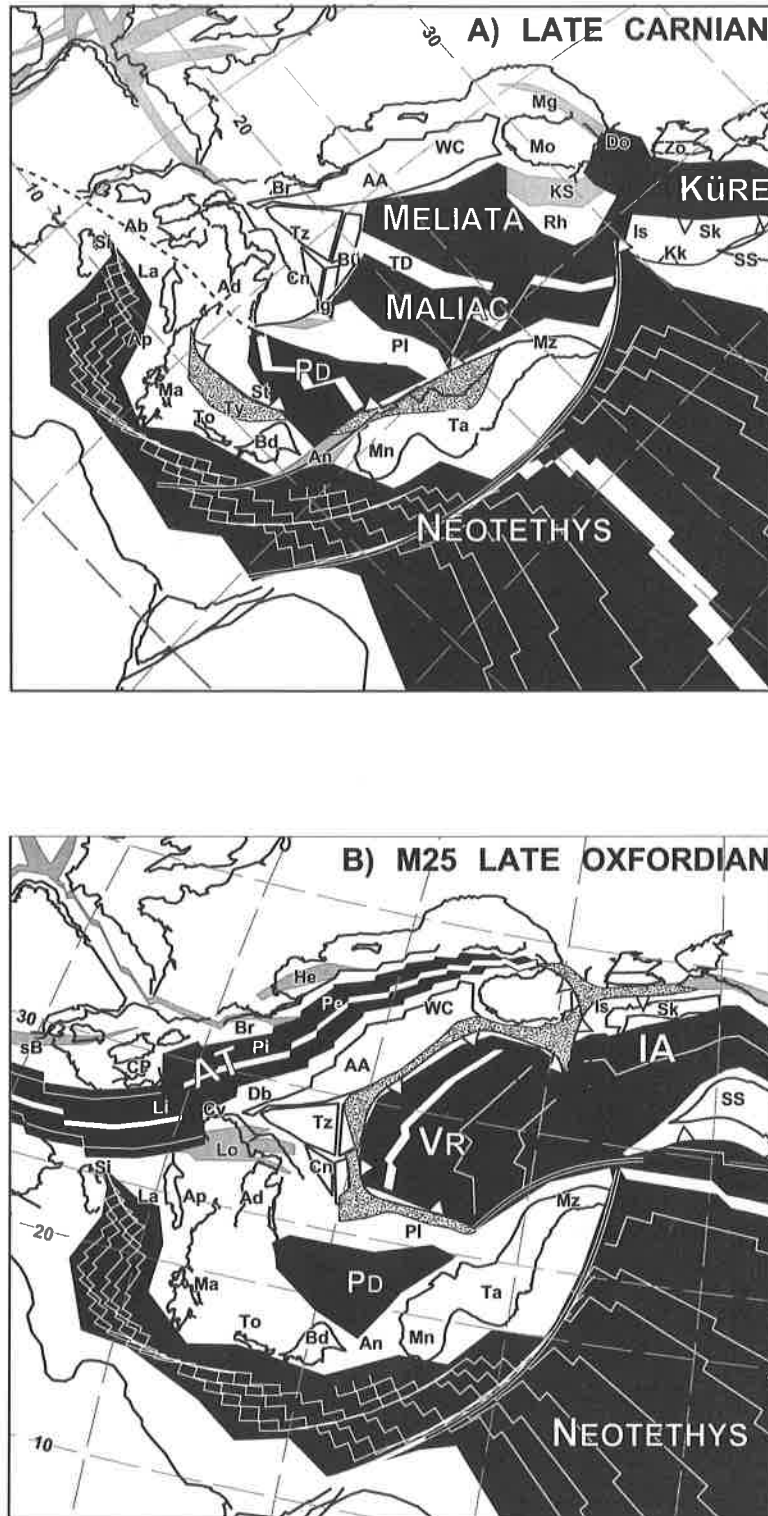


Fig.4 - A) Late Carnian and B) Late Oxfordian (anom. M25) reconstructions of the western Tethyan realm, modified from Stampfli et al. (2001a). AA, Austro-Alpine; Ab, Alboran; Ad, Adria s.str.; An, Antalya; Ap, Apulia s.str.; AT, Alpine Tethys; Bd, Beydaghlari; Br, Briançonnais; Bü, Bükk; CP, Calabria-Peloritani; Cv, Canavese; Db, Dent Blanche; Do, Dobrogea; He, Helvetic rim basin; IA, Izmir-Ankara ocean; Ig, Igal trough; Is, Istanbul; Kk, Karakaya forearc; KS, Kotel-Stranja rift; Kü, Küre ocean; La, Lagonegro; Li, Ligurian; Lo, Lombardian; Mn, Menderes; Mo, Moesia; Mz, Munzur dag; Pd, Pindos rift/ocean; Pe, Penninic; Pi, Piemontais; Pl, Pelagonian; Rh, Rhodope; sA, south Alpine; sB, sub-Betic rim basin; Sd, Srednogorie rift-arc; Si, Sicanian; Sk, Sakarya; SS, Sanandaj-Sirjan; St, Sittia; Ta, Taurus, s.l.; TD, Trans-Danubian; To, Talea Ori; Ty Tyros fore arc; Tz, Tizia; Va, Valais trough; Vr, Vardar ocean; WC, West-Carpathian; Zo, Zonguldak.

Black represents oceanic crust, rift zones are in grey, stippled represents foredeep bassins.

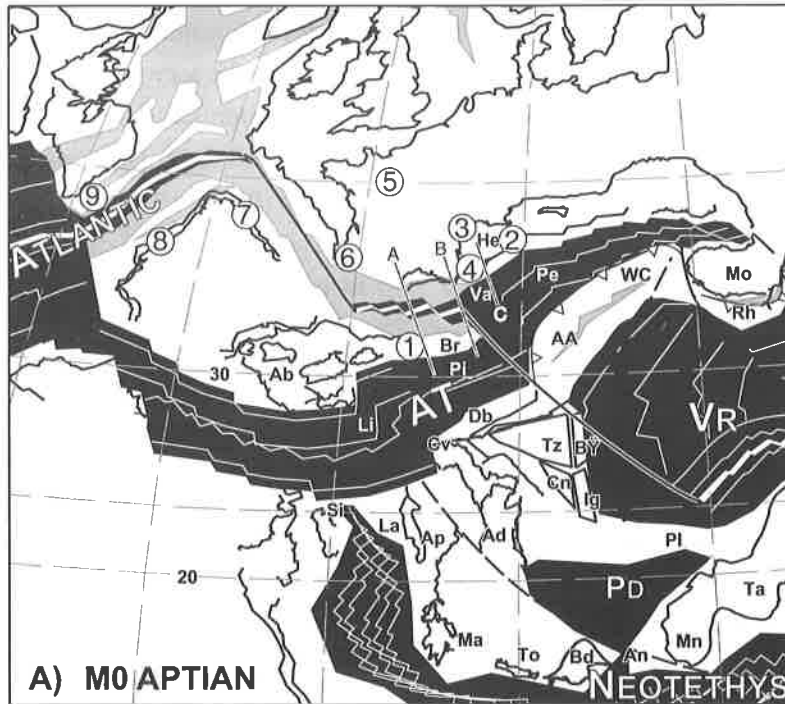


Fig.5 - A) Aptian (anom. M0) and B) Santonian (anom. 34) reconstructions of the western Tethyan realm, modified from Stampfli et al. (2001a). See fig.4 for legend.

an African promontory from Middle Triassic to recent times and represents the eastern most Cimmerian element detached from Gondwana in Late Permian times during the opening of the East Mediterranean basin. The Adriatic and Apulian plates were welded in an eo-Cimmerian collision phase during the Triassic and were part of the African plate until Early Cretaceous. Then, Adria started a left-lateral displacement to reach its final position in the Miocene as a separate entity, it was partially subducted under the Dinarides and the Apennines.

The Adriatic plate *s.str.* can therefore be considered as a displaced terrain like most large tectonic units from the former southern margin of the Alpine Tethys: the Austroalpine, Carpathians (Channell 1992) and Tisia composite terrains. Unlike Adria and Tisia which are still rooted in the lithosphere, the Austroalpine and Carpathian composite terrains were decoupled at upper crustal level and incorporated in the orogenic wedge. Their composite nature comes from the fact that they record the closure, not only of the Alpine Tethys, but also of older oceanic domains such as the Meliata-Maliac and Vardar oceans.

Triassic back-arc oceans of the Paleotethys

The exotic character of the upper Austroalpine terrain (North Calcareous Alps) and the necessity to open an oceanic area between the internal part of the Carpathian domain (Tisia) and the Austroalpine domain is now widely accepted. This Meliata-Hallstatt ocean is now better known and its different parts have been studied and dated in some details (Kozur 1991; Haas et al. 1995). We consider the opening of this Meliaticum domain as a result of continuing subduction along the eastern part of the European margin in the Late Paleozoic (Stampfli et al. 2001c; Vavassis et al. 2000). The northward subduction of a remnant Paleotethys induced the opening of back-arc type basins already in Carboniferous-Permian times (see

below) in a general context of cordillera collapse following the accretion of the Variscan terrains in Early Carboniferous time. The back-arc spreading lasted until Late Triassic in the Mediterranean domain (Kozur 1991; Stampfli et al. 2001b).

The Meliata rift possibly extended westward to the Ivrea zone (Zingg et al. 1990) in the Southern Alps where Permian and Triassic rifting and volcanism is well documented (*e.g.* Winterer & Bosellini 1981). The Meliata rift extension in that region would have been of intracontinental type and is also possibly recorded in the Sesia nappe (Venturini 1995). However, remnant of Meliata oceanic material, and metamorphic rocks related to its closure, are not present in the Swiss Alps. Accelerating slab roll back and even detachment of the Paleotethys slab, induced a jump of the back-arc spreading to the Maliac rift in late Middle Triassic (De Bono 1998), then a new jump opened the Pindos basin in Late Triassic (Vavassis 2001). During the opening of the Central-Atlantic/Alpine Tethys system the Meliata oceanic domain started subducting southward in connection with the southward subduction of the the Küre basin north of Turkey (Stampfli et al. 2001b). The slab roll-back of Meliata was responsible for the opening of the Vardar back-arc basin in Middle Jurassic times, which partially obducted in Late Jurassic onto the Pelagonian micro-continent (in Greece, *e.g.* Baumgartner 1985, Albania, Yugoslavia, fig. 4).

The opening and closing of the Meliata-Maliac domain affected the whole Alpine history. First it created a generalised subsidence starting in the Late Permian responsible for the large scale Triassic transgression in the Alpine domain. Then its closure in the Late Jurassic (Kozur 1991) affected the whole Austroalpine domain and can be seen as a major element in the onset of the closure of the Alpine Tethys itself on an Eastern Alps transect (Froitzheim et al. 1996). Effectively, the southward subduction of the Maliac-Meliata-Küre system is certainly responsible for the (local)

southward subduction of the European plate, a unique feature in the whole Alpine system where generally the southern continents are subducting northward (Africa, Arabia, India).

The Alpine Tethys rifting

Late Triassic ages obtained from shear zones in the Southern Alps (Schmid et al. 1987; Hunziker et al. 1992) witness important transtensive events which eventually developed into the opening of the central Atlantic/Alpine Tethys ocean and the final break-up of Pangea. The areas previously affected by transtension and emplacement of mafic material during the Permo-Triassic period, became subsiding rim-basins of the Jurassic Alpine Tethys (fig. 4). This is the case for the the Médiannes and Helvetic rim basins. Paleozoic faults were also reactivated in these domains to form shallow half-grabens (Borel 1997). The Jurassic rifting is actually cutting in between these zones of thinned and already cold lithosphere. It has to be emphasised here that the thermal subsidence of large parts of the Alpine orogen started in Triassic time and that this subsidence is hardly disturbed in areas far enough from the Jurassic rift (Stampfli, 2000). These different subsidence behaviours can be used to sort out paleogeographic units; they also indicate the importance of the Permo-Triassic lithospheric extensional phase. Mainly it explains why the Alpine Tethys rifting phase started under water, and that main clastic inputs coming from the rift shoulders were locally delayed until the peak of thermal expansion in Late Liassic times (see Ft3 and 6).

Subsidence patterns of the marginal areas of the Alpine Tethys (Loup 1992; Borel 1995) together with stratigraphic and sedimentological records allow to place the onset of rifting in Sinemurian (fig. 7). Spreading in the central Atlantic is placed in the Early Toarcian (Steiner et al. 1998). The former Atlantic rift would have extended first toward the Lombardian basin, which aborted, then the rifting jumped to the Alpine Tethys *s.str.* (the

Liguro-Piemont-Penninic ocean) where the onset of spreading is of probable Aalenian age. Gabbros have been dated in several areas of the Alpine Tethys, the oldest are usually Bajocian (*e.g.* Bill et al. 1997), in Corsica older ages have been found and points to a possible Late Toarcian-Aalenian age (Beccaluva et al 1981) which is in line with a general onset of thermal subsidence of the margin in Bajocian time.

Rim Basins

Thermal subsidence of the Alpine Tethys margins in Bajocian affected large areas around the new oceanic area. The progradation of carbonate platforms toward the rift/ocean was hampered by the presence of rim basins on both sides of the Alpine Tethys in the western Alps. To the north the Helvetic-Dauphinois basin (see introduction to the Externides, p.43) and its SW extension toward the Sub-Briançonnais rim basin, Sardinia and the Sub-Betic rim basin. To the south, the Lombardian rim basin representing a clear aborted arm of the Alpine Tethys (Bertotti et al. 1993) active from Late Triassic to Early Jurassic. Progradation did not succeed infilling these rim-basins, and the former rift shoulders of the Alpine Tethys (the Briançonnais and south Helvetic domains in the north and the Canavese and lower Austro-Alpine domains to the south) became drowned submarine ridges developing condensed sequences. Therefore, a general phase of sediment starvation characterises the distal part of the Alpine Tethys margins. However, small carbonate platforms developed for a short time around the rift shoulders (*e.g.* the Briançonnais platform, Septfontaine 1983). The latter was linked to larger platforms along the rift shoulder as found in Corsica and Sardinia, and was finally drowned in Late Jurassic, whereas the south Helvetic platform (found in the Ultrahelvetic nappes) could never develop as a large entity and must have been made of small carbonate patches already drowned in

Oxfordian times. The progradation of the Briançonnais platform took place towards the rim basin – to the north, away from the rift shoulder, so in an opposite direction of the general progradation on the European platform (see Ft6).

Exotic terrains

In the southern margin

The rotation of Iberia-Africa in the Early Cretaceous as well as the closing of the remnant Meliata domain induced the closure of the Liguro-Piemont part of the Alpine Tethys ocean. This closure is quite different on the Austrian transect and the Swiss transect due to the presence/absence of the Meliata domain. In the eastern Alps transect the subduction of the Alpine Tethys is the consequence of the subduction of the Meliata domain since Jurassic time. The subduction there can be seen as a continuous process during which the Austro-Alpine blocks were subducted and part of their covers was included in the accretionary prism to form the Middle and Upper Austro-Alpine nappe system (Faupl & Wagreich 1999, and references therein). In the western Alps the Adria margin stayed passive for a longer time due to the fact that Iberia and Africa have the same wander path during the Cretaceous (at least between anomaly M0- Aptian and 34 - Santonian – fig. 5). The subduction of the Piemont part of the Alpine Tethys is marked by HP/LT metamorphism (Hunziker et al. 1992) of elements pertaining to the toe of the Austroalpine margin (Internal massifs? Former back-stop of the prism) and to the accretionary prism *s.str.* (Tsaté nappe, Marthaler & Stampfli 1989; Stampfli & Marthaler 1990; Deville et al. 1992).

Most of the eo-Alpine ages seems now to be younger than formerly dated (*e.g.* Froitzheim et al. 1996), however some ages are still older than the onset of collision between the accretionary prism and the Briançonnais micro-continent and can for sure be related to subduction of the oceanic

domain. Younger ages (Middle Eocene or younger) are related to the collision proper (Gebauer 1999). Therefore the onset of oceanic subduction on a western Alps transect could be younger than proposed so far, but not younger than the incorporation of exotic elements in the accretionary prism represented by the Gets and Simme flysch. These exotic elements are made of ophiolitic blocks and blocks derived from the Lower Austro-Alpine margin, their minimum age is Cenomanian (Clément 1986) and the onset of subduction of the Piemont part of the Alpine Tethys can certainly be placed around that time (90-100 Ma). This time corresponds also to the onset of flysch deposition in the Lombardian rim basin (Bichsel & Häring 1981; Bernoulli & Winkler 1990). Before to become an active margin this western portion of the lower Austroalpine margin suffered a westward displacement in regards to Adria (Frank 1987; Trümpy 1992; Froitzheim et al. 1994) related to the closure of the Vardar/Maliac-Meliata oceanic realms and the important rotation of Africa at that time to which Adria was still attached (fig.5). This induced trapping of Piemont oceanic crust or mantle between the Southern Alps domain (Adria) and the Lower Austroalpine units as found between the Canavese zone and the Sesia nappe (Venturini 1995).

In the northern margin

The separation of the Iberic plate from North-America in the Late Jurassic (the onset of spreading is dated as M0, Early Aptian - 110Ma; fig. 5) implies a separation of the Iberic plate from Europe too. This separation is brought about by the Pyrenean pull-apart rift system that we extended eastward into the Valais or North Penninic suture zone (Stampfli 1993). The Briançonnais domain (fig. 4 and 5) was attached to Corsica/Sardinia and therefore to the Iberic plate (Stampfli 1993). Thus, its former position was more to the SE than usually supposed, a proposal already

made by Frisch (1979). The most internal south-Helvetic (Ultraschotivetic) domains, the lower Penninic Simplo-Ticinese nappes and external Valais zone (*i.e.* zone Sub-Médiane and the Bündnerschiefer area of central Switzerland) are therefore considered as former elements of the northern Piemont oceanic margin trapped by the eastward displacement of the Briançonnais terrain in front of them in Late Cretaceous times. This remnant trapped Jurassic oceanic strip with an Early Jurassic margin to the north and an Early Cretaceous margin to the south is called here the Valais trough, but is also sometimes referred as the north Penninic ocean or Valais ocean. This displacement induced a duplication of the Alpine Tethys northern margin in the present day Alpine orogen from the French Alps up to the Engadine window. The eastward escape of the Briançonnais was induced by the opening of the Bay of Biscay and the large scale rotation of the Iberic plate together with Africa. It implies a partial closure of the Valais trough already during the Late Cretaceous. This rotation was closely followed by a general shortening between Africa-Iberia and Europe culminating into the Pyrenean orogen. This orogen can be followed from the present Pyrenees eastward up to southern France (Provence, *e.g.* Arthaud & Seguret 1981), and continues in the Alps in the form of large scale uplifts of the Helvetic margin and local inversion of the Jurassic tilted blocks, well expressed by the deposits of the Niesen flysch (mainly Maastrichtian) and Meilleret flysch (Middle Eocene), sedimented on a structured Mesozoic basement (see field trip 6- stop 6F, and field trip 2-stop 2C). These flyschs clearly predate the Alpine collisional event in the Helvetic region characterised by the deposition of collisional flysch not before the earliest Oligocene. It is during the Alpine collision that the Briançonnais domain became an exotic terrain, obducted in the Alpine accretionary prism.

The Alboran plate (Rif, internal Betic, Kabylies,

Peloritan, Calabria microplates, Wildi 1983) formed the southern margin of the Iberic plate (fig. 4). This margin was affected by deformation processes as from Early Cretaceous (Puga et al. 1995); resedimentation of the Dorsale Calcaire (former rift shoulder) in the flysch basin started in Cenomanian time, grading into major olistostrome deposits in the Maastrichtian (Gübeli 1982; Thürow 1987). However, these deformation are most likely related to local inversion and strike slip displacement between Africa and Iberia rather than to subduction, because no shortening can be demonstrated at that time between the two plates. Subduction started in Eocene/Oligocene times as a result of the Pyrenean orogeny and graded into major terrain displacements during the late Tertiary opening of the Algero-Provençal ocean, liberating the Alboran blocks from their Iberic motherland (fig. 6). These displacements generated duplication of paleogeographic elements, creating pseudo oceanic sutures. We propose a displacement toward the SW of the Internal Betic domains (together with the Rif) subsequently incorporated into the Betic orogen as a terrain.

The Iberic margin is considered as a Tertiary active margin developing back-arc spreading, and the African margin as a passive margin. Terrain displacement was related to the slab roll-back of the remnant Alpine Tethys and the lateral detachment of its slab. The Apenninic accretionary prism developed on the eastern side of Iberia and by Late Miocene time the slab roll-back had reached the Ionian basin (the western most part of the Neotethys) and this led to the opening of the Tyrrhenian ocean as a new generation of back-arc, accompanied by the drifting of the Calabrian micro-plate to its present position (fig. 6).

Vergence of subduction

As we have seen above, the southward subduction of the Piemont part of the Alpine Tethys ocean is related to the history of the Meliata-Maliac

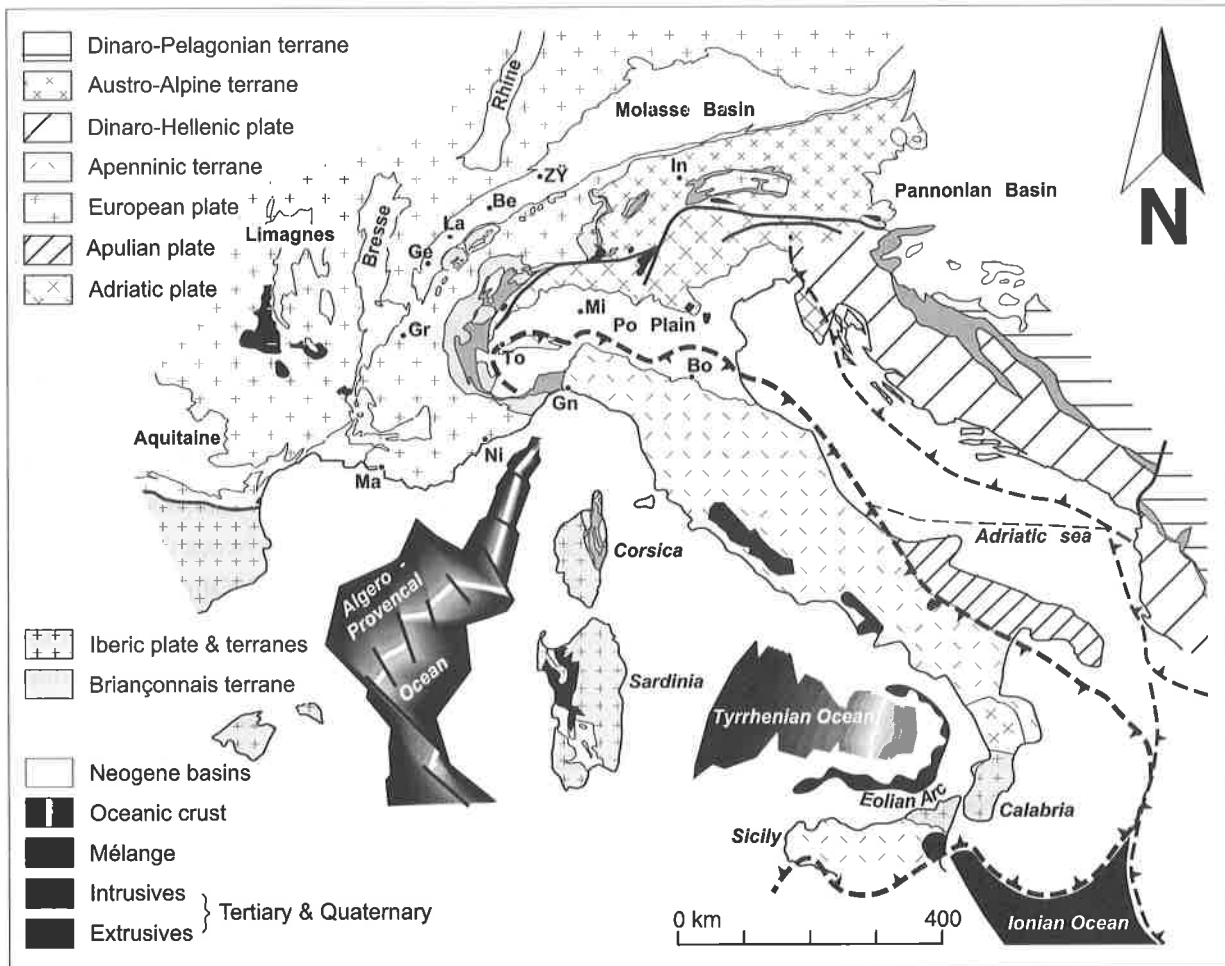


Fig.6 - Simplified tectonic map of the Alpine-Mediterranean area showing major tectonic plates.

domain. The change of vergence of the Alpine Tethys subduction westward (northward, under the Iberic plate) is directly linked to the connection between the Alps and the Apennines. It must be emphasized here that the Penninic prism is older (Late Cretaceous-Eocene) than the Apenninic one (Oligocene- Pliocene), actually one started when the other one stopped. If there is a confusion between the two it is because the Apenninic prism is re-mobilising parts of the Penninic prism as exotic elements (*e.g.* the Braco ophiolitic ridge, Elter et al. 1966; Hoogenduijn Strating 1991). These complex situations of duplication of paleogeographic elements led to many paleogeographic interpretations (most of them done only on 2D cross-sections) not taking into account large-scale lateral displacement of terrains and diachronous subduction events. Most models

propose several parallel small oceans, separating even smaller strips of continental crust. These solutions are not feasible, mainly in regard to rheological constraints on plate boundaries and sedimentological evolution of these margins.

2.1.2 The continental margins of the Alpine Tethys

The pre-rift sequences

Similar pre-rift sequences are found on both sides of the Jurassic rift which gave birth to the Alpine Tethys. Therefore, the pre-rift sequences (basement included) cannot be readily used to say if a structural unit belonged to a margin or its conjugate opposite. So, there is no such thing as a typical Austroalpine basement, a typical European Permian or a typical Briançonnais Triassic.

Permian rift related continental deposits are locally found on both sides. The shallow marine Triassic deposits transgressed these Permian deposits and/or the Variscan basement. As said before the general Triassic transgression is related to the thermal subsidence generated by the Late Permian rifting associated with the opening of the Hallstatt-Meliata ocean.

Triassic deposits are more of German type on the northern side of the Alpine Tethys and more of Alpine type on the southern margin, but there are numerous exceptions. Major changes of facies are noticed when approaching the former Hallstatt-Meliata passive margin as found in the southern Alps, the Northern Calcareous Alps or the inner Carpathian domains, and also when going south towards the westernmost extension of Neotethys in the Ionian sea region (Sicani basin of Sicily, Catalano et al. 1992). For example strong similarities of fish remains have been found between the Middle Triassic of Provence and of the Briançonnais, related to an embayment of Neotethys waters in that direction, an embayment not extending in the Helvetic domain (Kozur pers. comm.).

The syn-rift sequences

Triassic carbonate deposits are still very uniform, of shallow water type, but locally thickness variations and breccias reworking Triassic dolomites (already locally present in the Late Triassic) indicate that crustal extension already started. Thus, it is obvious that rifting started underwater in the Alpine Tethys, it does explain that locally the appearance of breccias was delayed until sufficient reliefs have been created, this is the case in the Helvetic transect (see Ft6). Major breccia deposits are found in many areas, both south and north of the Alpine Tethys. The major one is represented by the Brèche nappe found in the Préalpes, dated as Sinemurian to Bajocian, and related to the major phase of extension of the

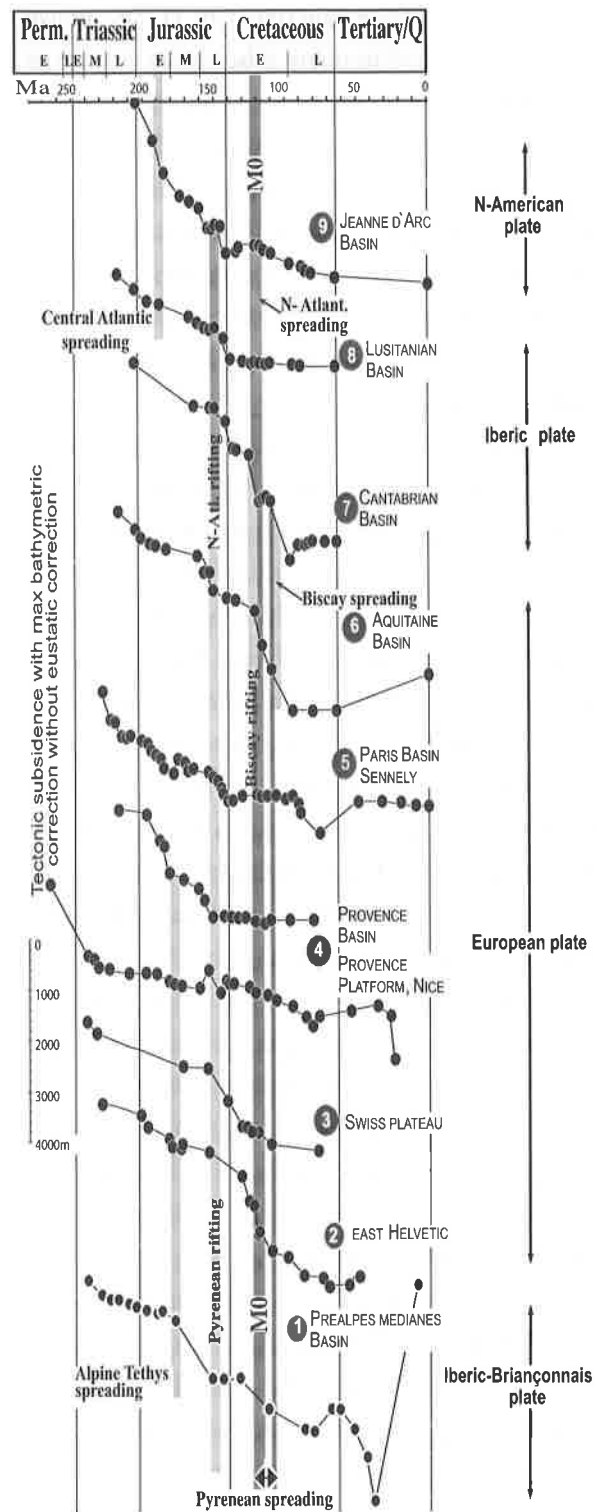


Fig.7 - Synthetic subsidence curves for selected circum-Iberia domains, modified from Borel (1998), location on fig. 5A. The subsidence program used to derived the curves is from R. Schegg. Time scale from Odin (1994)

Alpine Tethys rifting (see field trip 3). This phase can be recognised in many other areas outside of the rift zone (*e.g.* in the rim basins), showing that the area under extension was quite large at the beginning. This Liassic event is certainly related to accelerating extension at lithospheric scale at that time (see fig. 7) leading to sea-floor spreading and thermal expansion of the rift zone in Toarcian-Aalenian time. The thermal uplift, in the order of 500m to a kilometer, triggered the erosion of the rift shoulder, down to basement. This erosion is well recorded in the Brèche nappe for the Briançonnais part of the margin and in the lower part of the Niesen nappe for the Helvetic margin, such breccias are also found in the southern margin (Dolin) but are not well dated. These deposits consist mainly of breccias and conglomerates, but they differ quite significantly due to the nature of the eroded shoulder material. In the Briançonnais, the Triassic sequence was much thicker and the Brèche nappe is dominated by clasts of Triassic limestones and dolomites and rare clasts of Permian; in the Niesen nappe, the “Couches de la Grande Eau” (see field trip 6) are on the contrary dominated by basement clasts with minor Liassic-Triassic carbonates. The proximal nature of these clastic deposits (upper to middle fan) demonstrate their former location in the first tilted block of the margin, directly adjacent to the rift shoulder. The highly metamorphic Lebendun conglomerates of the Simplon region (lower Penninic nappes), have been interpreted as synrift deposits (Spring *et al.* 1992) and could be regarded as equivalent of the Jurassic conglomerates from the Niesen nappe. The subsequent deepening below the CCD in the late Middle Jurassic for the Brèche nappe allows to calculate the overall vertical throw of the major boundary fault of the rift, which was in the order of 4 to 5 km.

Geometry of rifting

If we apply a simple-shear model of rifting,

asymmetric continental margins would be expected, at least at upper crustal level (Stampfli *et al.* 2001c):

- A lower-plate type with large-scale tilted-blocks and denuded upper-mantle at the continent to ocean transition.

- An upper-plate type with a ramp-like or flexural geometry with limited tilted-blocks, trap-like volcanism (not necessarily widespread) and a continent to ocean transition composed of basic intrusions as it is the case for the Arabian side of the Red Sea (*e.g.* Bohannon 1986).

Sedimentological environments and subsidence patterns will be different for both margins as shown for the central Atlantic conjugate margins (Morocco and the US coast; Favre & *al.* 1991). However when dealing with continental margins involved in continental collision, the discrimination between the two types is not so easy. Usually the rift part of the margin has disappeared during the collision. The rheological behaviour of the margin during the collision can still be used to characterise one type or the other. Putting all available information together we proposed an asymmetry for the Alpine margins (Stampfli & Marthaler 1990) with a changing asymmetry somewhere in the eastern Alps for the southern margin (Manatschal 1995) and somewhere between the Briançonnais and Helvetic domains in the northern margin (fig. 4). The main constraints come from the lower Austroalpine Margna unit of the eastern Swiss Alps regarded as a piece of the toe of the southern margin. Trommsdorff *et al.* (1993), Froitzheim *et al.* (1994) and Froitzheim & Manatschal (1996) have demonstrated the lower plate nature of this margin in eastern Switzerland and the presence of denuded mantle (Malenco). Its conjugate margin is represented by the Briançonnais upper plate margin. A change from upper to lower plate can be demonstrated as one goes from the Briançonnais margin to the Helvetic margin (fig. 4 & 8). Denuded mantle is found at the toe of the

South Helvetic margin (Geisspfad; Keusen 1972), it is presently in tectonic contact with the Monte Leone crustal nappe considered as a former tilted block from that margin (see also fig. 12).

This change of asymmetry could correspond to the area where the Pyrenean transform cut through the former Jurassic passive margin (fig. 5 and 8). The Chiavenna denuded upper mantle unit of eastern Switzerland could belong either to the south Valais trough margin, *i.e.* the northern lower

plate margin of the Briançonnais terrain in eastern Switzerland, or to the trapped toe of the Helvetic margin under the exotic Briançonnais terrain.

Denuded continental mantle is also found in the Ligurian accretionary prism and could well explain the change in the direction of subduction (Vescovi 1993). However, it is not sure if these pieces of mantle are Penninic exotics or actually belong to the back-stop of the Apenninic accretionary prism.

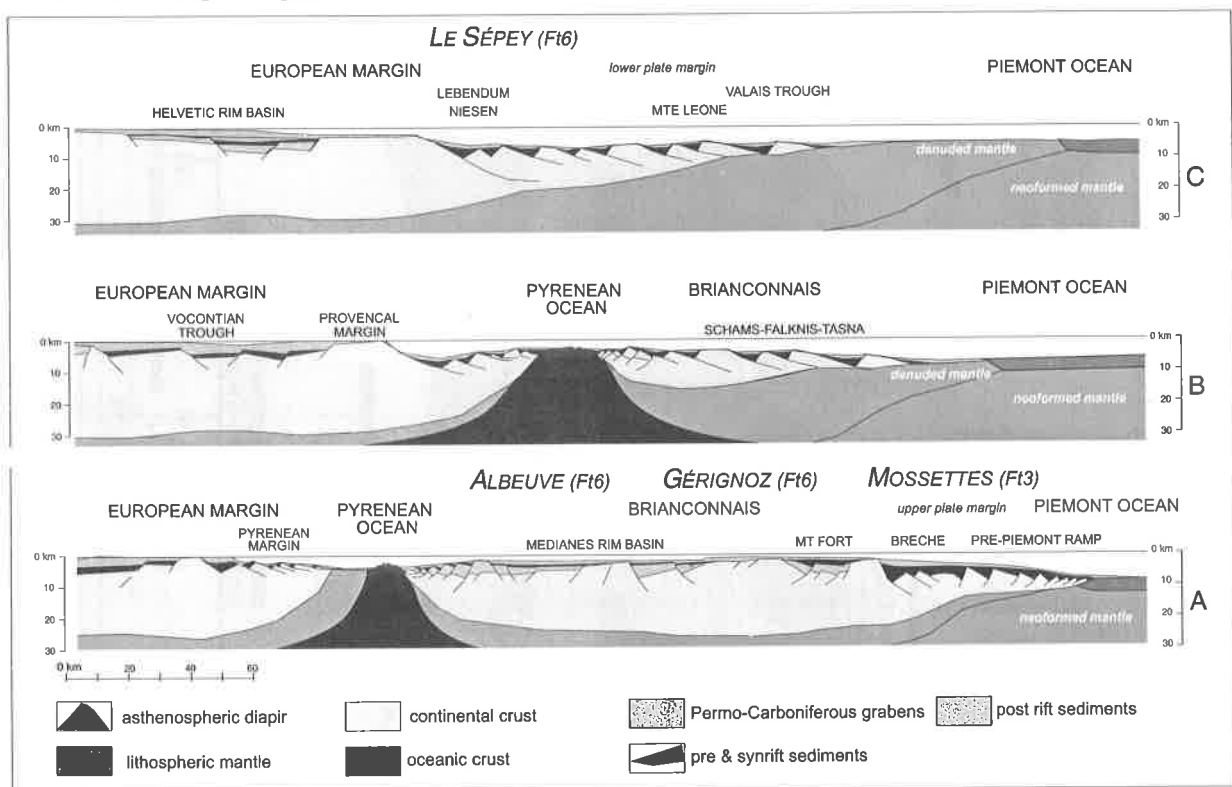


fig.8 - Early Cretaceous reconstruction of the European margin and position of the Pyrenean ocean. Location of cross-sections in figure 5A.

The post rift sequence (fig. 9)

Thermal subsidence was rapid in Middle Jurassic, in most areas, mixed clastic and carbonate are deposited at that time. Around the drowning rift shoulders, small bahamian carbonate platforms developed as in the Briançonnais domain (*e.g.* Septfontaine 1983). During the Callovo-Oxfordian period, the accelerating drowning created a starvation phase mainly registered in the distal margin by the deposition of red clays or even radiolarian cherts.

In areas not affected by the Pyrenean ocean

opening, the pelagic starved sequence graded into pelagic limestone deposits during the Malm and Early Cretaceous.

In the areas affected by the Pyrenean rifting phase, uplift due to unloading along major active faults (mainly transtensive: Schmid et al. 1990; Septfontaine 1995) generated renewed breccia deposits in and around the Briançonnais domain from the Oxfordian to the Early Cretaceous (*e.g.* upper breccia from the Brèche nappe, Falknis breccias in the Briançonnais of eastern Switzerland and Télégraphe breccia in the French Subbriançonnais).

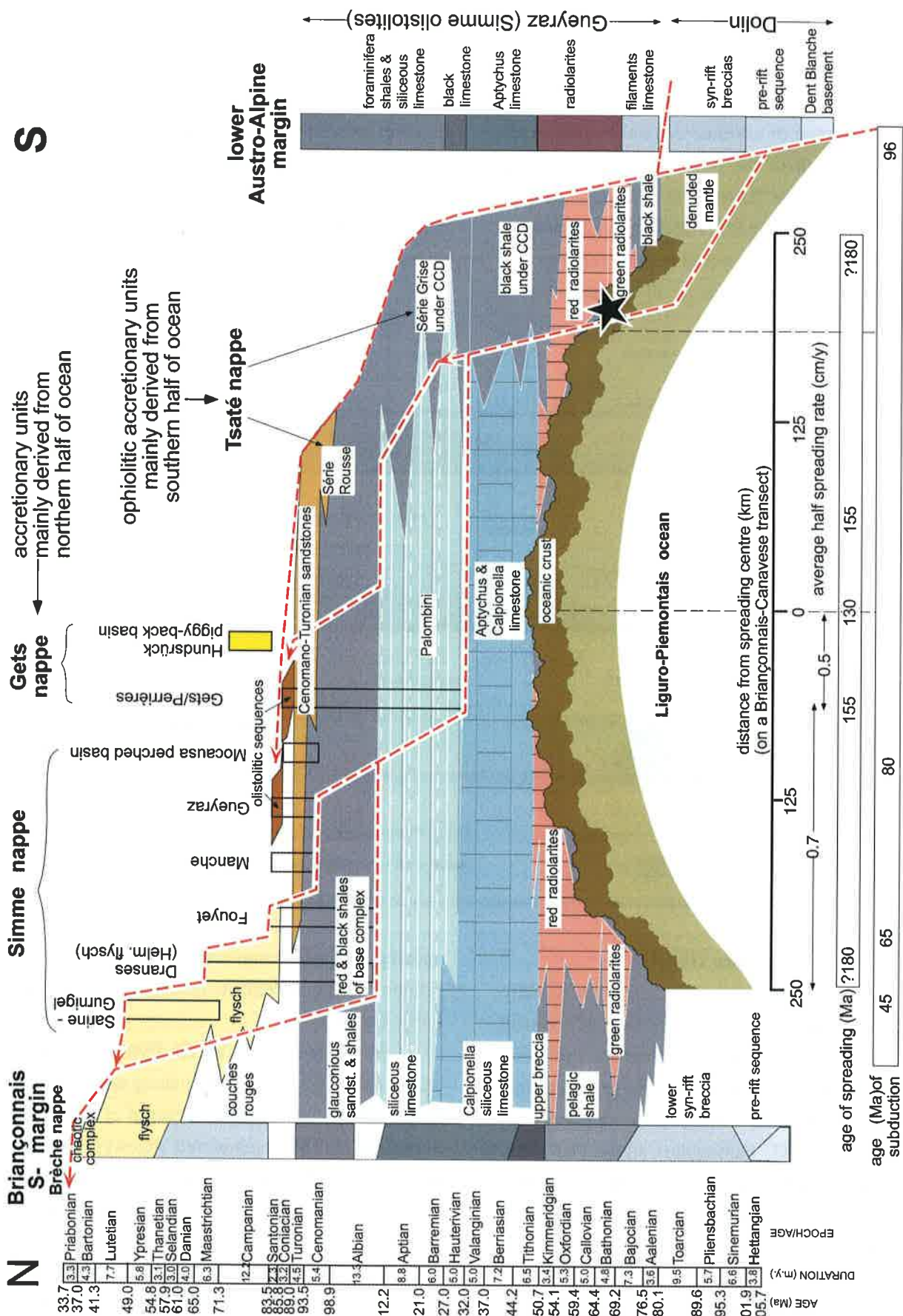


Fig.9 - Stratigraphy and origin of the oceanic series and accretionary sequences found in the western Swiss Alps

The “Valais ocean” question

From much, sometimes contradicting data concerning the Iberian drift (Stampfli 1993; Stampfli & Marchant 1997), it can be determined that the opening of the Pyrenean rift system started in Late Jurassic times when Iberia was detached from New-Ffoundland. The minimal opening would be 200 km following the data from Malod & Mauffret (1990) or 350 km using the data of Sibuet & Colette (1991). Discrepancies stem from the former situation of Iberia with regard to New Foundland. The data from Srivastava et al. (1990) and from Srivastava & Verhoef (1992) allow to narrow down these differences and their proposal was applied for our model where a tight Permian Pangea fit is used (fig. 14).

The portion of the Pyrenean rift between Provence and the Briançonnais was roughly parallel to the central Atlantic rift between Iberia and New Foundland (fig. 5) and therefore, opened at the same time. The oldest magnetic anomaly in that part of the Atlantic is M0 (113 Ma; Srivastava et al. 1990; Sibuet & Collette 1991) which is earliest Aptian; however the sea-floor spreading was inevitably slightly older than the first clear magnetic lineation (for the central Atlantic the discrepancy is in the order of 10 to 15 Ma; Favre & Stampfli 1992).

In the Pyrenean region, the thermal event linked to the emplacement of basic material started around 115 Ma and lasted until 80 Ma (Montigny et al. 1986). The Bay of Biscay opened later, during the rotation of Iberia, after the Valanginian (Moreau et al. 1992) and after the Atlantic opening between Iberia and America, in Late Aptian-Albian times; spreading there stopped in the Campanian (A33 anomaly, see figure 5). Thermal subsidence is active in the Late Cretaceous for the Pyrenees and the Gulf of Biscay region (fig. 7) where the Cenomanian is largely transgressive on the former rift shoulders (Peybernès 1976; Peybernès & Souquet 1984; Simo 1986).

In the Briançonnais domain, the major change of sedimentation is found at the top of the Calcaires Plaquetés (Python-Dupasquier 1990) and can be dated as Barremian (115 Ma). It corresponds to a general drowning and starvation phase of the area at that time (see field trip 6). There is also a major sedimentary gap in the Albian that we relate to the thermal uplift of the Briançonnais. In the French “sub-Briançonnais” domain (mainly ultra-Dauphinois), important re-sediments are also found in the Albian and Early Cenomanian (Kerckhove 1969; Kerckhove & Lereus 1987). This shows that the Helvetic-Dauphinois margin is affected by the Pyrenean rifting phase, which is also quite clear when looking at the subsidence curves for the Helvetic and Swiss plateau domain (fig. 7).

Obviously the timing of sea-floor spreading along this new lithospheric fracture is not synchronous, even more because the northern border of the Iberic plate in the Pyrenean portion is of transform type and, may be, never generated sea-floor spreading.

The main point here is that, the “Valais trough”, as recognised in the western Alps in present days, is actually the remnant of a trapped Piemonte sea-floor and of the toe of the Helvetic margin (fig. 5 and 8), whereas the so-called “Valais ocean” (if ocean there was) was located south of France and we refer to it here as the “Pyrenean ocean” to avoid the confusion between Valais ocean and Valais trough. No direct traces of this ocean have been found so far (excepted Pyrenean peridotites) because its suture was located exactly where the Algero-Provençal ocean opened in Oligo-Miocene times. The southern margin of the Pyrenean ocean was the Briançonnais peninsula (fig. 8), its northern margin was the Corbières-Provençal domain from the Pyrenees to the Maures-Estérel massifs. This area is characterised by the development of Albian basins deepening southward towards the ocean, followed by inversion related basins accumulating thousands of meters of clastic deposits in Late

Cretaceous times in a northward migrating fore-deep type basin (Debrand-Passard & Courbouleix 1984).

Along the European margin, it is interesting to note the similitude of facies between the Rheno-Danubian flysch in eastern Switzerland and Austria and the Valais trough sequence (the Valais trilogy) from Albian to Late Cretaceous (Stampfli 1993, and references therein). The continuity of deep water clastic facies in these two domains allows to assign them to the same position with regard to the European margin (often referred as North Penninic basin). The presence of contourites and strong and changing current directions along the basin (Hesse 1974) suggests a connection with major oceanic domains. In the Ligurian domain (south of the Briançonnais) such turbiditic deposits are absent, the Albo-Cenomanian formation being dominated by anoxic black-shale deposits. So the Valais trough, together with the Pyrenean ocean and the Bay of Biscay ocean, must be regarded as the connection between the Eastern Alpine Tethyan realm and the Atlantic during the Cretaceous.

2.1.3 Subduction and obduction processes

Subduction and obduction processes affecting the oceanic sequences of the Piemont ocean are shown on figures 9 and 10. Spreading rates were obtained from the central Atlantic magnetic anomalies which confirm the opening of the Alpine Tethys at least until Early Cretaceous (Stampfli et al. 2001b). The rates of convergence between Africa and Europe are also derived from the Atlantic magnetic anomalies and reported along a transect perpendicular to the western Alps. We reported the evolution of the flexural bulge in time along this transect and defined its emplacement in regard of the Briançonnais domain and the Helvetic domain. In both areas the flexural bulge can be detected through a good preservation of the sedimentary sequences.

Theoretically, the flexural bulge related to the

southward subduction of the Alpine Tethys affected the Briançonnais domain in Late Cretaceous already and could have been responsible for major hiatuses in the sedimentary record at that time and a very condensed section all together (Couches Rouges; Guillaume 1986). Thereafter it affected the Helvetic margin (fig. 5) creating a general emersion of the Helvetic-Dauphinois (see Ft2) and Subalpine domains since Paleocene time (Allen et al. 1991; Lihou & Allen 1996; Burkhard & Sommaruga 1998). However, Pyrenean inversion within the Helvetic margin and the eastward migration of the Briançonnais peninsula at that time, certainly strongly interfered with the flexural processes creating localised inversions of former tilted blocks (Niesen flysch; Ackermann 1986)

The time of incorporation of the different terrains is well established based on the age of associated flysch deposits (Caron et al. 1989). In the Piemont domain, on a western Alps transect, the Gurnigel flysch was deposited from Maastrichtian until Middle Eocene (Caron et al. 1980a; Caron et al. 1980b) and followed by the chaotic complex of likely late Middle Eocene age (Steffen et al. 1993; Dall'Agnolo 2000), which includes elements from the Briançonnais margin (Couches Rouges, Breccia nappe). On the Briançonnais domain the deposition of the Médiannes flysch lasted until Lutetian time (NP 15, 47 to 43 Ma; Guillaume 1986) it is a rather distal flysch deposit, precluding any deposition on an already detached substratum. Thus the subduction of the Briançonnais domain did not take place before the late Middle Eocene (fig. 11), this is confirmed by metamorphic ages from the Briançonnais basement starting around 38 Ma (Markley et al. 1995, 1999).

Slivers of basement with part of their cover were detached from the subducting slab and underplated; they form presently the bulk of the middle Penninic domain (Escher et al. 1997) (see

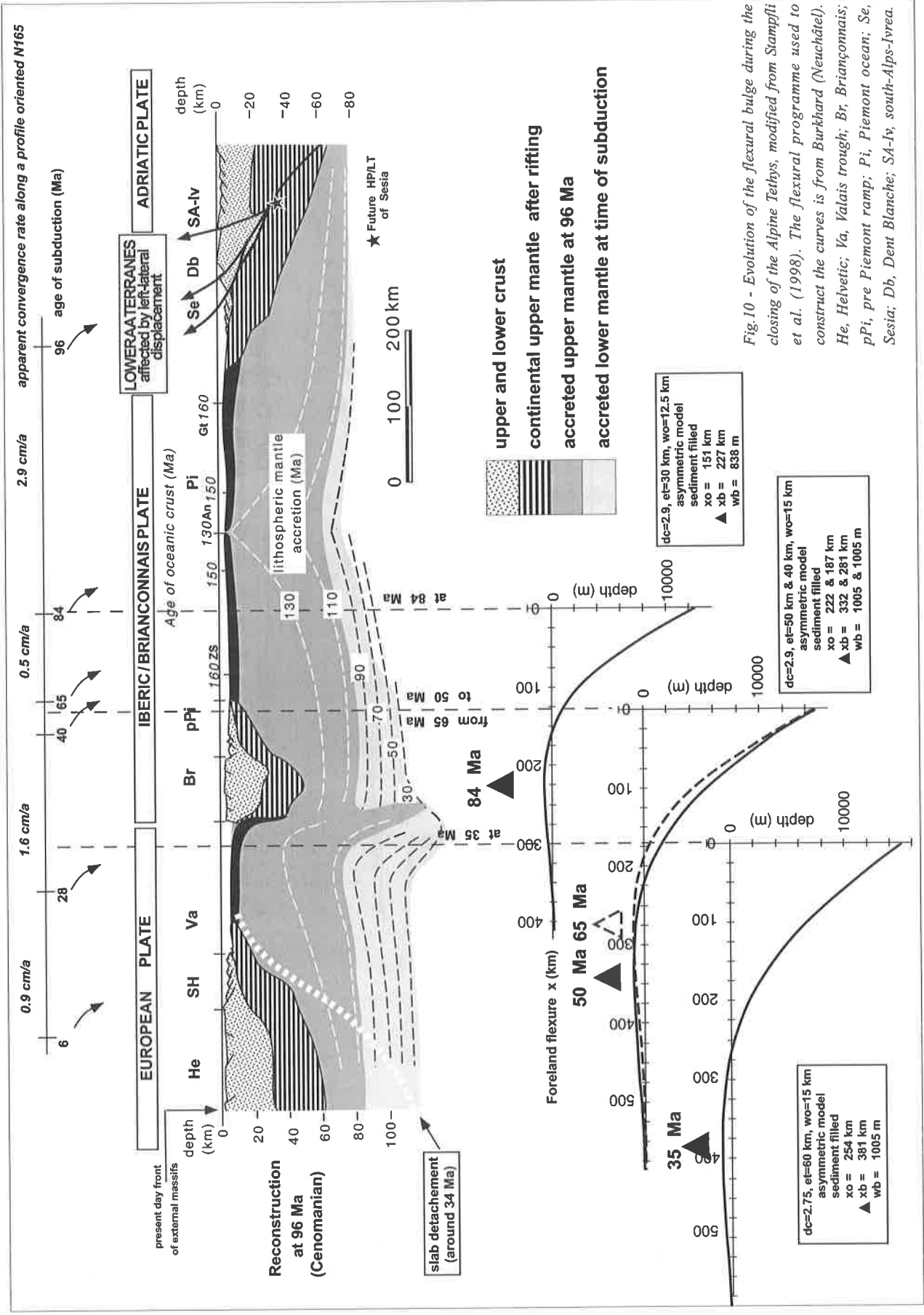


Fig.10 - Evolution of the flexural bulge during the closing of the Alpine Tethys, modified from Stampfli et al. (1998). The flexural programme used to construct the curves is from Burkhard (Neuchâtel), He, Helvetic; Va, Valais trough; Br, Briançonnais; pPi, pre Piemont ramp; Pi, Piemont ocean; Se, Sesia; Db, Dent Blanche; SA-Iv, south-Alps-Ivrea.

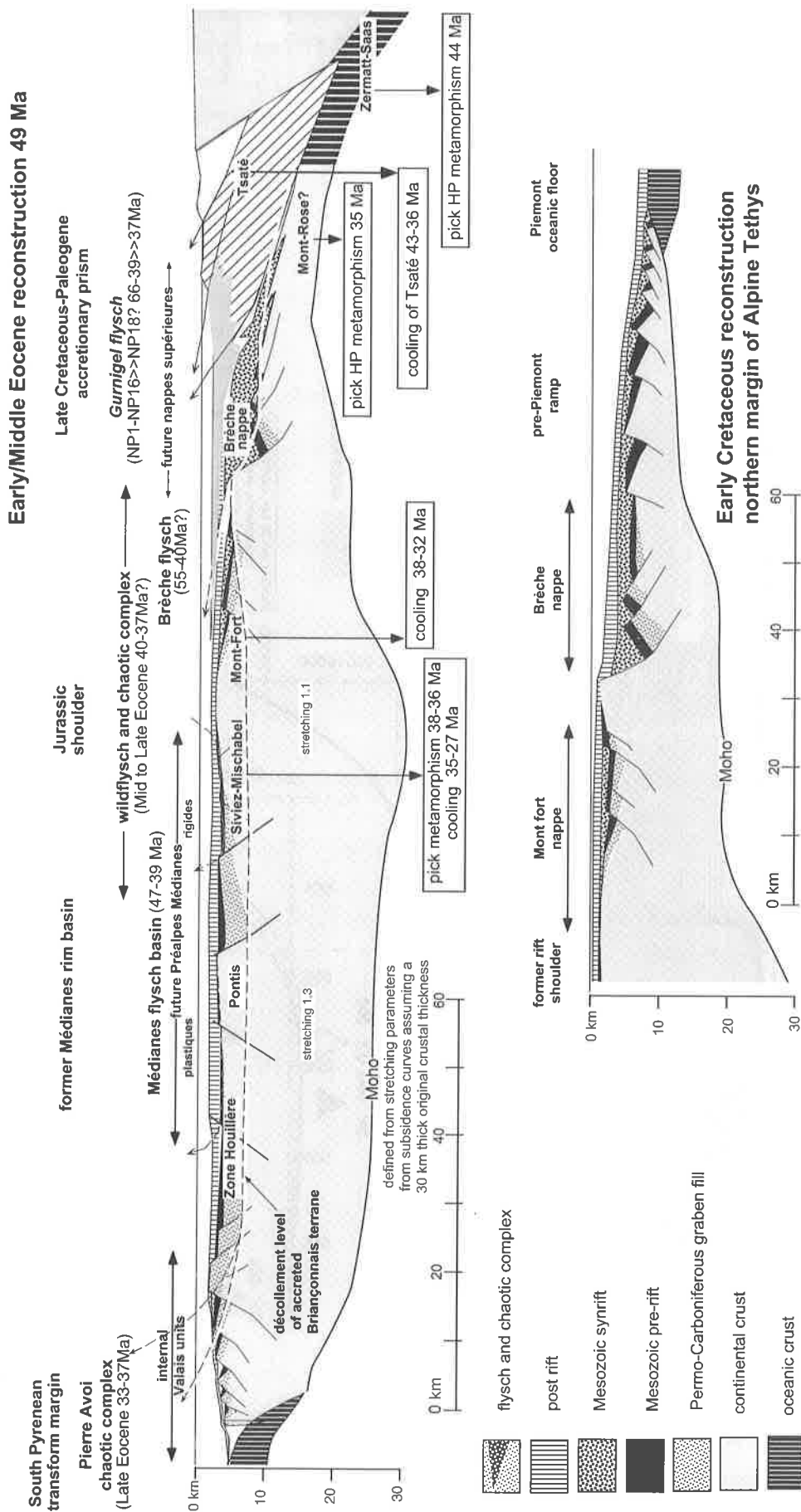


Fig. 11 - Reconstruction of the Briançonnais domain in Early Cretaceous and early Middle Eocene, modified from Stampfli et al. (1998). See text for discussion and references about the timing of events. Stretching factor from Marchant & Stampfli (1997)

Ft4). Part of the cover was detached from the basement and incorporated in the still active accretionary prism to form the future Préalpes Médiannes (Mosar et al. 1996). The exotic Briançonnais sliver was overthrust on the Valais trough and a Late Eocene mélange is found at its base in the Valais (Pierre Avoi unit, Bagnoud et al. 1998). Along the same suture, MORB are found in a few places in the Valais area and in eastern Switzerland (in the Bündnerschiefer; Dürr et al. 1993). They have a signature not different from other Piémont MORB, and as said above are derived

from a piece of the Alpine Tethys trapped between the Helvetic and the exotic Briançonnais terrain.

Elements from the Valais trough were then accreted and are represented by the “Valaisan trilogy” (Aroley, Marmontains, Saint Christophe; Burri 1958) made of poorly dated deep-sea clastic sediments, outcropping nowadays in the Sion-Courmayeur zone (see field trips 4 & 5).

Incorporation in the prism of elements pertaining to the Helvetic margin (fig. 12) is well constrained by the ages of the south Helvetic

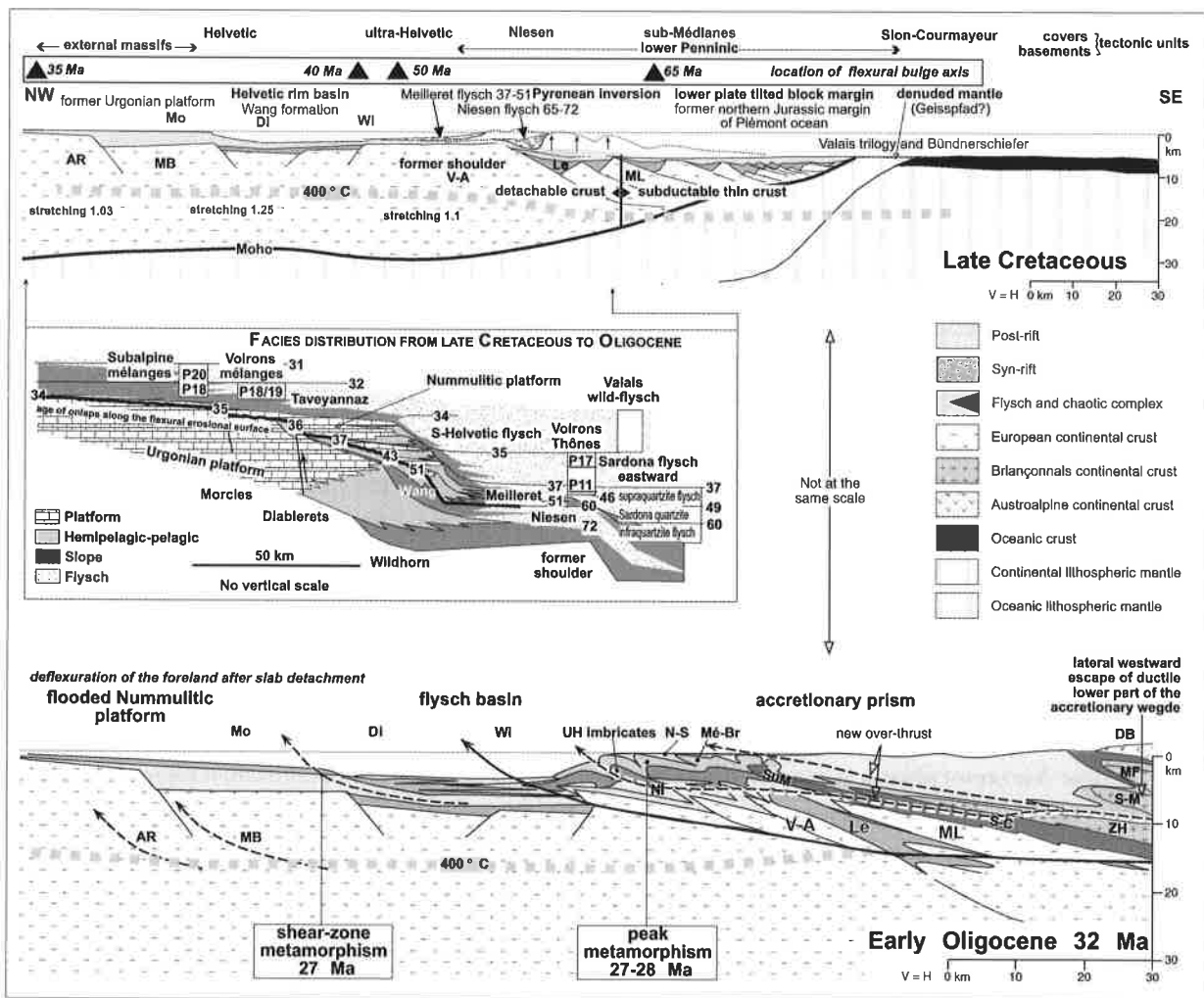


Fig.12 - Reconstruction of the Helvetic margin in Maastrichtian and early Oligocene, modified from Stampfli et al. (1998). See text for discussion and references about the timing of events. Stretching factor from Loup (1992). AR, Aiguilles rouges massif; Di, Diablerets nappe domain; DB, Dent Blanche nappe; Le, Lebendun sequence; MB, Mont Blanc massif; MF, Mont Fort nappe; Mo, Morcles nappes domain; Mé-Br, Médianes and Brèches nappes; ML, Monte Leone nappe; Ni, Niesen nappe; N-S, nappes Supérieures; S-C, zone Sion-Courmayeur; S-M, Siviez-Mischabel nappe; SuM, zone sub-Médianes; Ts, Tsaté nappe; V-A, Verampio-Antigorio nappes; Wi, Wildhorn nappe domain; ZH-Po, zone Houillère, Pontis nappes.

flysch (e.g. Jeanbourquin & Burri 1991; Jeanbourquin 1994) (see Ft3) and Sardona flysch (in east Switzerland; Lihou 1996). It can be placed during or just after the Priabonian since the Ultrahelvetic flysch has been dated as Late Eocene (37 to 33.7 Ma; e.g. Kindler 1988; Charollais et al. 1993). Most of the attenuated crust from the Helvetic margin was subducted (Burkhard & Sommaruga 1998) and is now represented by HT-HP basement nappes of the Simplon-Ticino dome.

General metamorphism of the south Helvetic domain spreads from 30 to 20 Ma (cooling ages)(Steck & Hunziker 1994) and mylonites from the Helvetic nappes have been dated between 32 and 13 Ma (Kirshner et al. 1996) – see also next chapter on Alpine metamorphism.

Further subduction of the thicker part of the margin led to the decoupling of major parts of the crystalline basement represented by the Mont Blanc and Aiguilles Rouges massifs. This decoupling was made possible by the increased heat flux following the slab detachment around 35-33 Ma (Wortel & Spakman 1992; Stampfli & Marchant 1995). This detachment corresponds to the emplacement of the periadriatic intrusives (e.g. Bergell intrusion) and the rapid Early Oligocene transgression on the Helvetic domain following temporary de-flexuring of the lower plate. This phenomenon allowed further subduction of the European plate (Marchant & Stampfli 1997)

2.1.4 Alpine Metamorphism, a review (fig. 13)

A. STECK

The compilation presented in figure 13 is based on original works by Bearth (1958), Burkhard (1988), Bussy & Epard (1984), Chabloz (1990), Colombi (1989), Compagnoni (1977), Frey (1987), Frey & Wieland (1975), Gouffon (1993), Halter (1992), Jaboyedoff & Thèlin (1996), Lardeaux et al. (1982), Lempicka Münch (1996), Niggli & Niggli (1965), Reinhardt (1966), Rubatto et al. (1998), Simic (1992), Steck & Burri (1971), Steck &

Hunziker (1994), Steck & Tièche (1976), Vannay & Allemann (1990), Von Raumer (1974) and Wenk & Keller (1969).

The boundary between non metamorphic and anchi metamorphic rocks in the Helvetic and Prealpine Zones are drawn after Frey et al. (1999). They recognized the transported metamorphism in the Niesen nappe and in the Klippen of the Préalpes médianes rigides of the Gummfluh and the Spillgerten. This limit corresponds also to the reaction kaolinite + quartz = pyrophyllite + H₂O in Al-rich metapelites (Frey 1987), and laumontite = prehnite and/or pumpellyite in the Taveyanne graywackes (Bussi & Epard 1984). In the Morcles nappe, the temperature culmination of the anchi zonal metamorphism is estimated at 260-350 °C and its age is comprised between 28-17 Ma (Kirschner et al. 1995).

We have chosen the disappearance of prehnite and pumpellyite in mafic rocks and the first appearance of biotite in iron rich granitoids and mafic rocks for the boundary between the anchi zone and the greenschist facies (Steck & Burri 1971). From the quartz fabric (Voll 1976) observed in the Mont Blanc, Aiguilles Rouges and Aar massifs, this boundary corresponds to a temperature of approximately 300 °C. The presence of the glaucophane greenschist subfacies in the Pontis, Siviez-Mischabel, Mont Fort and Tsaté nappes indicates an augmentation of the pressure in a SW direction in the greenschist facies zone. The age of the greenschist phengites of the Siviez-Mischabel nappe is comprised between 40-35 Ma (⁴⁰Ar/³⁹Ar method, Markley et al. 1998). The boundary between the greenschist and amphibolite facies is indicated by the appearance of oligoclase An17, replacing albite An0-6 in mafic rocks (Bearth 1958, Reinhardt 1966, Wenk & Keller 1969, Colombi 1989, Steck & Hunziker 1994). It is the limit between the critical assemblages of albite+epidote+hornblende of the epidote amphibolite facies and calcic

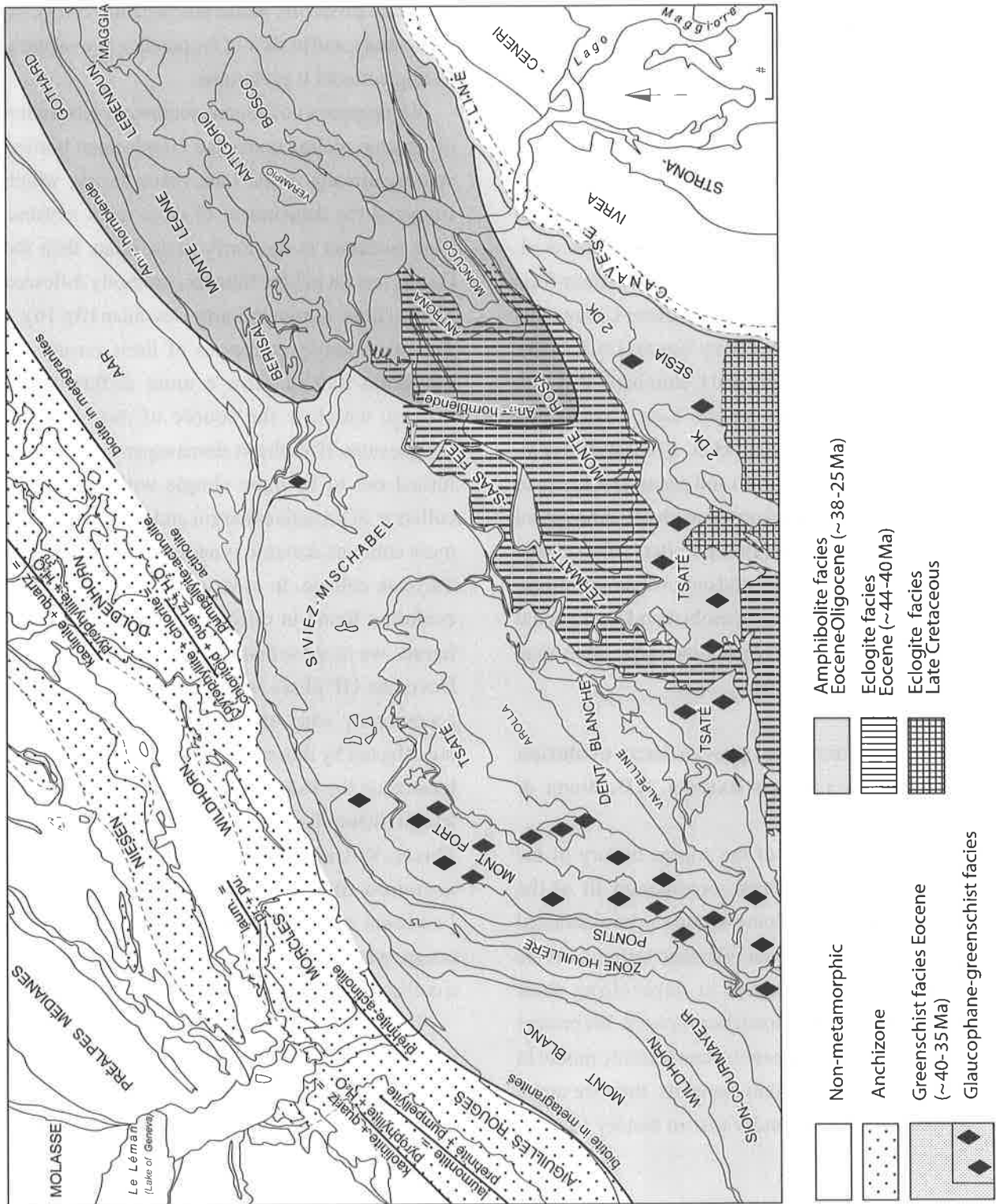


Fig.13 - Map of the Alpine metamorphism of the Alps of western Switzerland. Modified after Steck et al. (2001)

plagioclase+hornblende of the amphibolite facies, from the definition of Eskola (1939). In the Central Alps, the staurolite, index mineral of the amphibolite facies, is unfortunately too rare for being a useful marker of this facies. The age of the temperature culmination event of the metamorphism of the Lepontine gneiss dome is between 38-35 Ma (Hunziker et al. 1992, Steck & Hunziker 1994). The regional occurrence of the Eocene high pressure metamorphism (dated by Rubatto et al. 1998 and Amato et al. 1999) in the Zermatt-Saas Fee, Antrona, Monte Rosa and Etsch-Levaz units is based on the observations by Bearth (1962, 1973), Dal Piaz & Ernst (1978) and Colombi (1989). The regional occurrence of the Late Cretaceous eclogite facies (Hunziker et al. 1992, Ruffet et al. 1997, Rubatto et al. 1999) and greenschist facies in the Sesia Zone is based on the work of Compagnoni et al. (1977), Chabloz (1990), Halter (1992), Simic (1992), Marclay (1999) and our own observations. A rime of 1-2 km of greenschist and anchi zonal rocks follows the Canavese Line in the Ivrea zone (Steck & Tèche 1976).

2.2 The Variscan and pre-Variscan evolution

G.M. STAMPFLI, J. VON RAUMER, G.D. BOREL & F. BUSSY

After the review of the Alpine history of the western Alps, a Permian continental fit of the Mediterranean and Alpine region can be elaborated (fig 14). On this canvas, Variscan terrains can be replaced, and an attempt at unravelling their geodynamic evolution can be proposed. We present first a general geodynamic plate tectonic model in order to define the main elements, then we move to the pre-Variscan and Variscan history in more detail

2.2.1 The proposed model (fig.15) (see Stampfli et al. 2001d, and references therein)

Generally speaking one finds important geodynamic similarities in the Variscan terrains of

Europe. The explanation of the present complexity should not be sought in a complex plate tectonic scenarios involving numerous oceanic realms, on the contrary, and in view of the parallels in evolution, a simple model is preferable.

We propose a continuous southward subduction of oceanic realms under the Gondwanan border, starting already in the Late Precambrian, which triggered the detachment of three main terrains. First Avalonia in the Early Ordovician, then the Gothic terrain in Late Silurian, promptly followed by the Hunic terrain in Early Devonian (fig.16). The accretionary processes of these terrains to Laurentia-Baltica show a more differentiated history, which is the source of the observed complexities. If for the Avalonia super-terrain things turned out to be quite simple with a classical collision of an active margin and a passive one, a more complex scenario is necessary to explain the Variscan collage. In order to respect parallels of evolution found in different parts of the Gothic terrain, we propose that areas affected by the mid-Devonian HP phase were located on the leading accretionary edge of this terrain, whereas areas not affected by this major eo-Variscan event were located on the PaleoTethys passive margin of the terrain (Stampfli et al. 2000; Stampfli et al. 2001d). This eo-Variscan mid-Devonian event is related to accretion of buoyant material derived from Laurussia and subduction of a peri-Laurussian ocean, whereas further east the event is related to a collision with an island-arc system.

To explain the subsequent large scale mixture of these two domains (active and passive margins), large scale lateral displacements have to be evoked. Most hercynologists would agree with the latter proposal (*e.g.* Matte 1991) but most of them would place these translations in a context of collision. What we propose here is a context of displacement of terrains in a still active margin, the translations being accompanied by transtensional and tranpressional events leading to the opening of Gulf

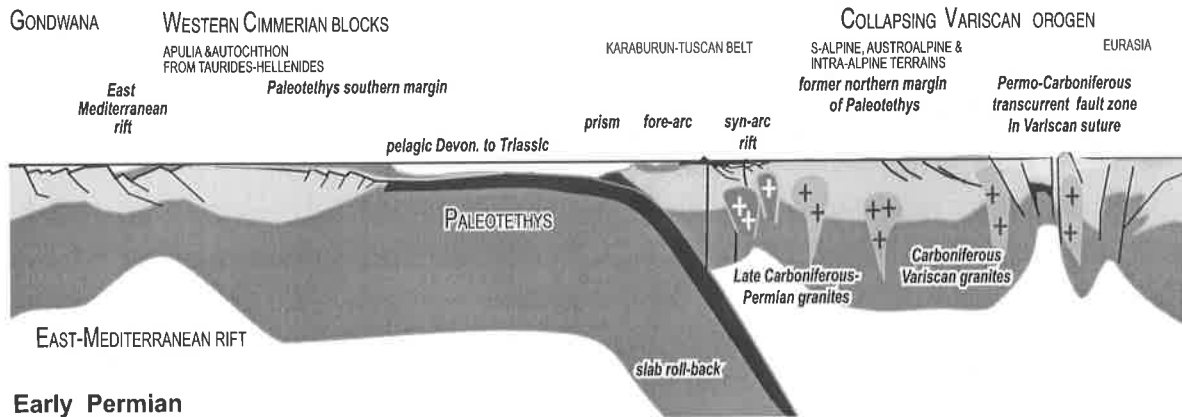
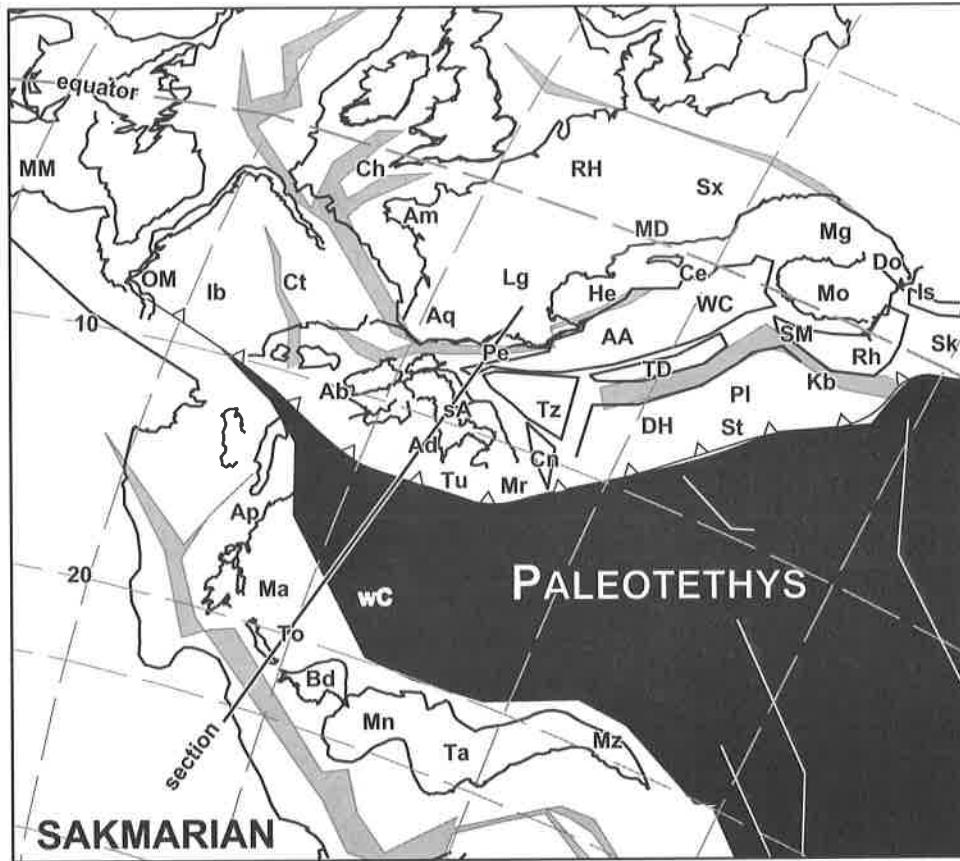
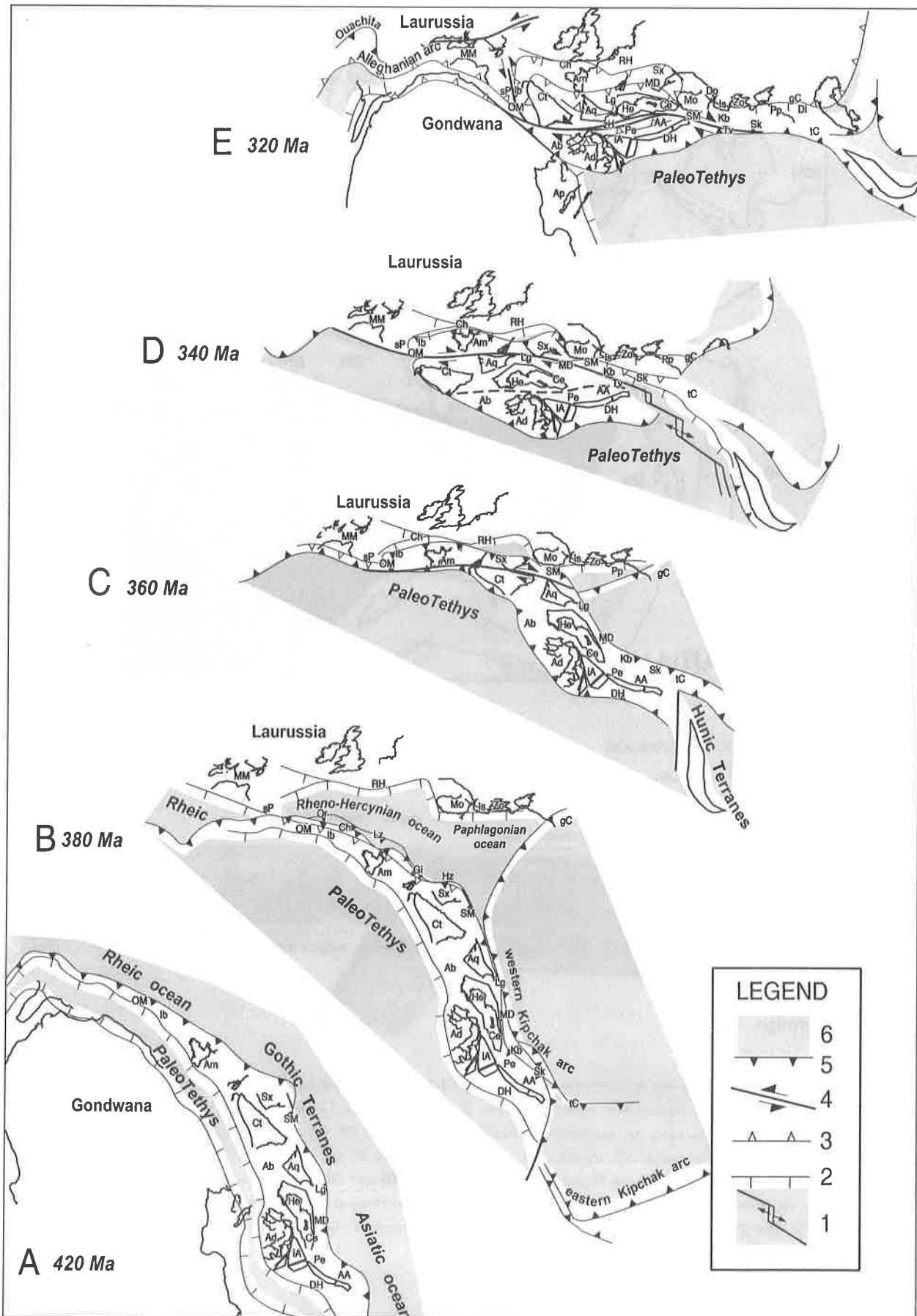


Fig.14 - Early Permian (Sakmarian) reconstruction of the western Tethyan realm modified from Stampfli et al. (2001a), rift zones are shown in grey; and lithospheric cross-section across the western part of PaleoTethys. Legend: AA, Austro-Alpine; Ab, Alboran; Ad, Adria s.str.; Am, Armorica; Ap, Apulia s.str.; Aq, Aquitaine; Bd, Beydaghlari; Ce, Cetic; Ch, Channel; Cn, Carnic-Julian; Ct, Cantabria; DH, Dinarides-Hellenides; Do, Dobrogea; He, Helvetic; iA, intra-alpine terrain; Ib, Iberia, central; Is, Istanbul; Kb, Karaburun; Lg, Ligerian; MD, Moldadubian; MM, Meseta-Meguma; Mn, Menderes; Mo, Moesia; Mr, Mrzlevodice fore-arc; Mz, Munzur Dag; OM, Ossa-Morena; Pe, Penninic; PI, Pelagonian; Rh, Rhodope; RH, Rheno-Hercynian; sA, south Alpine; Sk, Sakarya; SM, Serbo-Macedonian; St, Sitia; Sx, Saxo-Thuringian; Ta, Taurus, s.l.; TD, Trans-Danubian; To, Talea Ori; Tu, Tuscan; Tz, Tizia; wC, western-Crete (Phyl-Qtz); WC, West-Carpathian.



of California type oceans and in other places to the building up of cordilleras.

During the building up of the Carboniferous cordillera two scenarios developed. The westward evolution is towards a continent-continent collision where the accreted terrains got squeezed between Laurussia and Gondwana, this is the prevailing scenario for the Alleghanian regions. The other scenario, eastward, is a continuation of subduction and a generalised roll-back of the Paleotethyan slab. This, in turn, generated the opening of numerous back-arc basins and oceans starting in the Early Permian and until the Middle Triassic closure of the PaleoTethys oceanic domain (see above and fig. 4A).

Post collisional Permo-Carboniferous granites found for example in Morocco (*e.g.* Amenzou & Badra 1996) should be related to slab detachment when major crustal attenuation through generalised extension is not documented. In other western European regions, “post collisional” Permo-Carboniferous granites (as found in the Alps, see field trips 1 and 2) should be related either to slab detachment or the collapse of the cordillera or both, but not really to post collisional processes, the final collision being far away in time and space. Effectively, from Sicily to the Caucasus the final closure of PaleoTethys took place during the eo-Cimmerian cycle, and the closure of back-arc oceans issued from the PaleoTethys slab roll-back finally took place only in Cretaceous times (fig. 4 and 5).

2.2.2 What pieces of the model are to be found in the Alps?

The pre-Variscan elements

Comparing pre-Variscan relics hidden in the Variscan basement areas of Central Europe, the Alps included, large parallels between the evolution of basement areas of future Avalonia and its former peri-Gondwanan eastern prolongations (*e.g.* Cadomia, Intra-Alpine Terrain) become evident (von Raumer et al. 2001, and references therein). Their plate-tectonic evolution from the Late Proterozoic to the Late Ordovician is interpreted as a continuous Gondwana-directed evolution. Cadomian basement, late Cadomian granitoids, late Proterozoic detrital sediments and active margin settings characterize the pre-Cambrian evolution of most of the Gondwana-derived microcontinental pieces. Also the Rheic ocean, separating Avalonia from Gondwana, should have had, at its early stages, a lateral continuation in the former eastern prolongation of peri-Gondwanan microcontinents (*e.g.* Cadomia, Intra-Alpine Terrain). Subduction of an oceanic ridge (ProtoTethys) triggered the break-off of Avalonia, whereas in the eastern prolongation, the presence of the ridge may have triggered the amalgamation of volcanic arcs and continental ribbons with Gondwana (Ordovician orogenic event).

In this general picture the different future Alpine basement units are no exceptions. All four domains, the External, the Penninic, the Austroalpine and the South-Alpine domains, represented by a multitude of outcrop areas,

Fig.15 - Plate tectonic model for the origin and evolution of the pre-Variscan units preserved in the European Variscan mountain chain modified from Stampfli et al. (2001d). 1 spreading ridge; 2 passive margin; 3 sutures; 4 strike-slip fault; 5 subduction zone; 6 oceanic lithosphere. AA, Austro-Alpine; Ab, Alboran; Ad, Adria; Am, Armorica; Ap, Apulia; Aq, Aquitaine; Ce, Cetic; Ch, Channel; Ct, Cantabria; DH, Dinarides-Hellenides; Do, Dobrogea; gC, great Caucasus; Gi, Giessen; He, Helvetic; Hz, Harz; iA, intra-alpine; Ib, central-Iberia; Is, Istanbul; Kb, Karaburun; KT, Karakum-Turan; Lg, Ligerian; Lz, lizard; MD, Moldanubian; MM, Meguma-Meseta Mo, Moesia; OM, Ossa-Morena; Or, Ordenes; Pe, Penninic; Pp, Paphlagonian; RH, Rhenohercynian; Sk, Sakarya; SM, Serbo-Macedonian; sP, south Portuguese; Sx, Saxo-Thuringian; tC, trans-Caucasus; Tv, Tavas; zH, zone Houillère; Zo, Zonguldak.

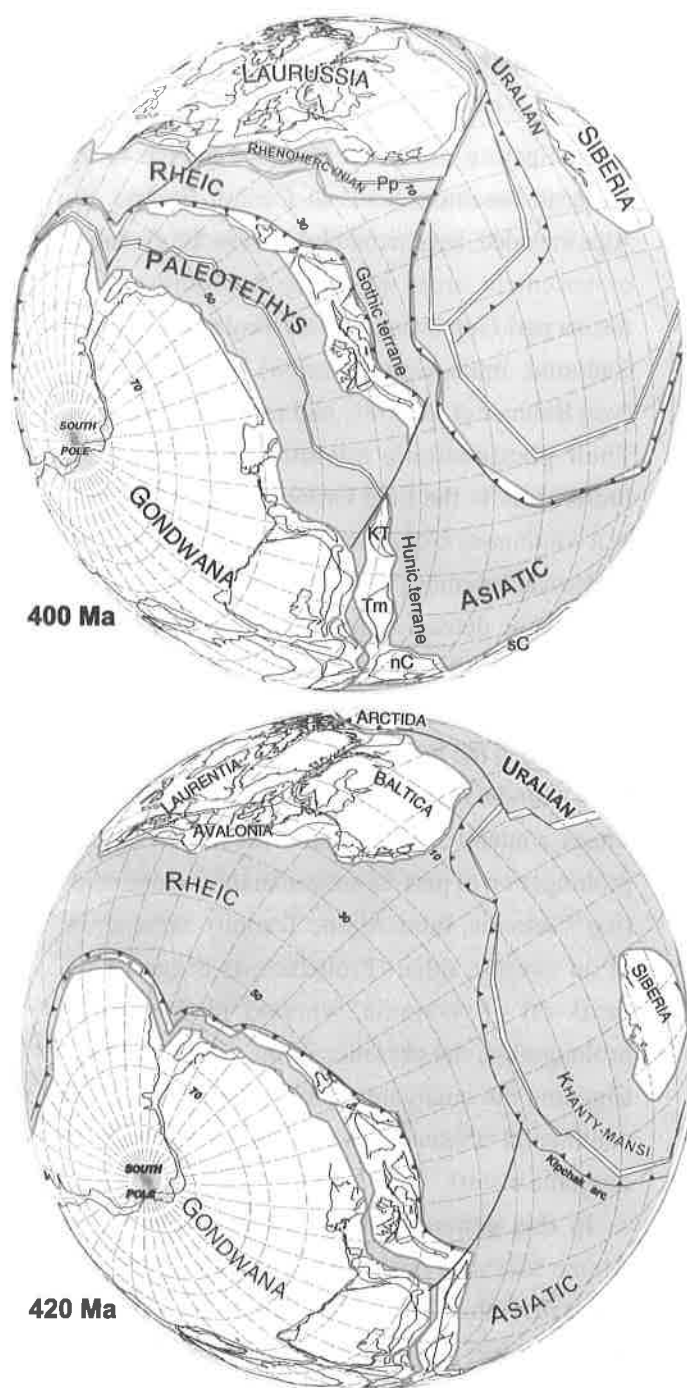


Fig.16 - Global reconstruction for the Late Silurian and the Early Devonian, showing the location of the Hunic and Gothic terrains. The 420Ma projection is centered on present-day latitude 10N, longitude 25E, the 400Ma projection on 10N, 20E. Pp, Paphlagonian ocean; KT, Karakum-Turan; Tm, Turim; nC, North-China; sC, South-China. (Modified from Stampfli and Borel in press).

without distinction, include relicts testifying their former location at the Gondwana margin (von Raumer et al. 1998), but there is no specific basement characterizing one of the four Alpine realms. They have to be seen as puzzle stones including relicts from different plate tectonic situations depending on their former relative location during the plate tectonic evolution from the Late Precambrian to the Ordovician.

Cadomian basement and Late Cadomian granitoids could have been present, but have to be proven, and Late Proterozoic detrital sediments and active margin settings characterize locally the pre-Cambrian evolution of these Gondwana-derived microcontinental pieces. The mostly metamorphic counterparts of detrital sediments characterizing the Gondwana shelf and/or Cambrian rift basins are widely distributed. As said above, subduction of an oceanic ridge (ProtoTethys) triggered the break-off of Avalonia, whereas its presence, in the eastern prolongation of Avalonia – the future basement areas composing the basement of the Alps, may have triggered the amalgamation of volcanic arcs and continental ribbons with Gondwana, thus leading to a short Middle Ordovician orogenic event. Renewed Gondwana directed subduction led to the opening of Paleotethys (fig.16). The great difficulties to understand such pre-Variscan events do not only result from the Alpine juxtaposition of small pieces, but also from the Variscan evolution, transforming most of these pieces to lower to middle crust elements.

The Variscan elements

Introduction

As we have seen above and depending on their former location at the Gondwana margin, the Variscan basement may contain, besides the complex Variscan (and Alpine) overprint, relicts of Cadomian basement, Late Proterozoic volcanic arcs, subsequent stages of accretionary wedges

and back-arc rifting and spreading. The evolution of these terrains was guided by the diachronous subduction of the ProtoTethys oceanic ridge under different segments of the Gondwana margin. This subduction triggered the emplacement of magmatic bodies and the formation of back-arc rifts, some of them becoming major oceanic realms (Rheic, PaleoTethys). Consequently, the drifting of Avalonia is followed since the Silurian and after a short Ordovician orogenic event, by the drifting of middle European and Alpine domains accompanied by the opening of PaleoTethys.

The slab roll-back of the Rheic ocean is viewed as the major mechanism for the drifting of the European Variscan terrains, this, in turn, generated a large slab-pull force responsible for the opening of major rift zones within the passive Eurasian margin (Rhenohercynian ocean). Therefore, the first mid-Devonian Variscan orogenic event is viewed as the result of a collision between terrains detached from Gondwana (Gothic and Hunic terrains) and terrains detached from Eurasia (fig.15). Subsequently, the amalgamated terrains collided with Eurasia in a second Variscan orogenic event in Viséan times, accompanied by large scale lateral escape of major parts of the accreted margin. Final collision of Gondwana with Laurussia did not take place before Late Carboniferous and was responsible for the Alleghanian orogeny.

PaleoTethys evolution

The PaleoTethys is more or less completely ignored by classical hercynology, therefore it is important to present here the main lines of its geodynamic evolution. The opening of PaleoTethys is relatively well constrained on an Iranian transect (Alborz range, North Iran; Stampfli 2000) representing the southern Gondwanan margin of the eastern branch of the ocean. Late Ordovician to Early Devonian flood-basalts, rift shoulder uplift in the Silurian followed by the onset of thermal subsidence in the Devonian, point to a Late

Ordovician/Silurian rifting phase. Sea-floor spreading took place in the Late Silurian or Early Devonian and the rift shoulders were completely flooded by Late Devonian time, following generalised thermal subsidence of the passive margin. From Late Devonian until Middle Triassic, a carbonate-dominated passive margin developed. A similar evolution is found in the Cimmerian part of Turkey (for details see references in Stampfli 1996, and Göncüoğlu & Kozur 1998), and most likely of Apulia, however in Apulia the information does not go deeper than the Permian.

The northern margin of the PaleoTethys ocean is well represented in the middle part of the Gothic terrains (*e.g.* southern Alps: Schönlaub & Histon 1999; Tuscan Paleozoic, Sardinia, Betic: cf. Stampfli 1996) also characterised by a Late Ordovician-Early Silurian clastic and often volcanic syn-rift sequence (Silurian flood basalts are also known in Sardinia). Thermal expansion and related erosion and tilting took place in Silurian time and is often wrongly related to the Taconic event (Tollmann 1985). Open marine conditions are found since the Silurian and are represented by graptolitic facies; a more generalised flooding took place in the Early Devonian and marked the onset of widespread thermal subsidence related to sea-floor spreading. On the northern margin, the Viséan usually marks the onset of generalised flysch deposition, often accompanied by volcanic activity. We regard this major change as representing the general collage of the different terrains to Eurasia to form the Variscan cordillera. It also marks the onset of PaleoTethys subduction and the transformation of the margin from passive to active, shortly followed by subduction of its mid-oceanic ridge certainly responsible for a large part for the high temperature Variscan metamorphism.

Accretionary sequences related to this subduction are little known, most likely because important subduction erosion took place during the cordillera stage as observed nowadays along the South-

American active margin. Potential Paleotethyan accretionary sequences are located in the southern part of the Variscan orogen and in all cases completely metamorphosed and intruded by subsequent Late Carboniferous granites and usually involved in eo-Cimmerian and Alpine deformations. However, pelagic Late Carboniferous to Early Triassic sediments of Paleotethyan origin are found in Sicily, the Dinarides, Hellenides and Taurides (Kozur et al. 1998; Kozur & Stampfli 2000), and, in Chios (Greece) and Karaburun (Turkey), Silurian and Devonian pelagic sequences have been found (Kozur 1997; Kozur 1998). These sequences clearly show that the remnant Paleotethys ocean was located south of the Alpine domain and separated the Variscan cordillera from Gondwana up to the Late Permian opening of Neotethys and the subsequent final closure of Paleotethys in Carnian times (fig. 4).

This closure speeded up through the opening in the former Variscan cordillera of major rift zones, some of them finally leading to the opening of back-arc marginal oceans (Meliata, Maliac, Pindos).

Variscan metamorphic massifs and intrusives in the western Alps

In the general picture given above for the Variscan evolution, the main Alpine realms (Helvetic, Penninic, Austroalpine, Southern Alps) again have to be replaced at their original location after separation from Gondwana and their stepwise approach towards Laurussia. Consequently, they have no specific evolutions characterising their Alpine situation, but they obey to the tectonic evolution from the Silurian to the Late Carboniferous, being laterally aligned until their Alpine juxtaposition. In the Alpine transect (e.g. fig. 3), they appear as basement nappes, separated in many cases by their Mesozoic cover. The general Alpine shortening brought them up to the surface, where they appear as dome like structures surrounded by Mesozoic cover.

Consequently, Variscan metamorphism and magmatism are the mirror of a pre-Alpine tectonic zonation, appearing in many locations as the main overprint, giving to the pre-Variscan lithologies described above the typical aspect of Variscan basement areas, as known all over Europe. Zones of distinct metamorphic grades from lower greenschist facies to high amphibolite facies and granulite facies were tectonically juxtaposed, and different magmatic pulses intruded into the strongly sheared and folded rock units. Such rocks will be presented during the field trips in the Aiguilles Rouges and Mont Blanc domains (Ft 1) and in the Penninic basement (Ft 4 & 5).

Rift related deposits (Houiller, Permian basins)

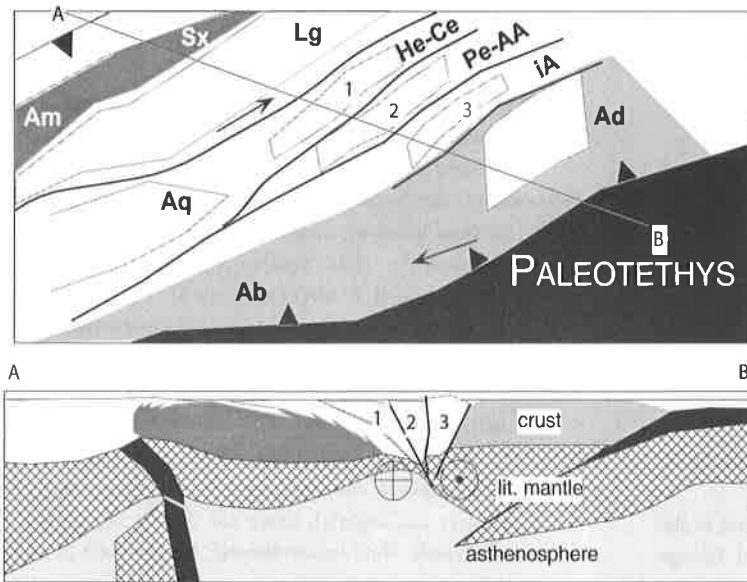
The presence of major transcurrent faults creating intra-mountain basins and the opening of Gulf of California type oceans within the Variscan cordillera influenced the distribution of sedimentary troughs in Central and Southern Europe during the Carboniferous and, again, the Alpine basement areas (Helvetic, Penninic, south Alpine and AustroAlpine domains) cannot be excluded from these tectonic events. The cordillera collapse certainly characterises Late Carboniferous Early Permian times in the Alpine part of the Variscan orogen. Extensional processes initiated already in Middle Carboniferous times and were related to major lateral displacement of the accreted terrains and juxtaposition of units showing quite different metamorphic conditions as found in the Penninic basement of the western Alps (Giorgis et al. 1999) (fig.17).

An example of such a feature are the Salvan-Dorénaz graben in the Helvetic domain (field trip 1 & 2) or the Zone Houillère (Cortesogno et al. 1993) of the Penninic domain (field trip 4 & 5), where extension might have led to local emplacement of E-MORB type extrusives (Cannic 1996) recently dated as Late Carboniferous (Schärer et al. 2000). Continuing extension in Permian times created new

grabens (sometimes with gabbro emplacement) in the Alpine domain locally grading to Early-Middle Triassic clastic/carbonate deposits. Similar mafic emplacement and basin development are known in the Penninic, Austro-alpine (Thöni & Jagoutz 1993) and Adriatic domain (Tuscan Carboniferous

basins, *e.g.* Englebrecht 1997). These basins aborted and never evolved as oceanic domains, but their thermal subsidence or local inversion is responsible for the Triassic paleogeography of these regions as well as for the final location of the Alpine Tethys rift.

Late Devonian



Late Carboniferous

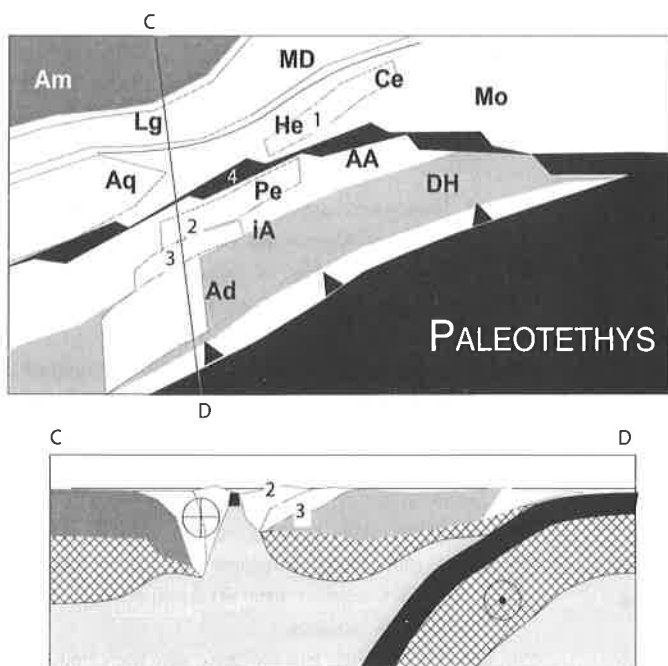


Fig.17 - Late Devonian and Late Carboniferous schematic evolution of the European Variscan domains, modified from Giorgis et al. (1999): AA, Austro-Alpine; Ab, Alboran; Ad, Adria (3); Am, Armorica; Aq, Aquitaine; Ce, Cetic; DH, Dinarides-Hellenides; He, Helvetic (1); iA, intra-alpine; Lg, Ligerian ; MD, Moldadubian; Mo, Moesia; Pe, Penninic (2); 4, zone Houillère rift.

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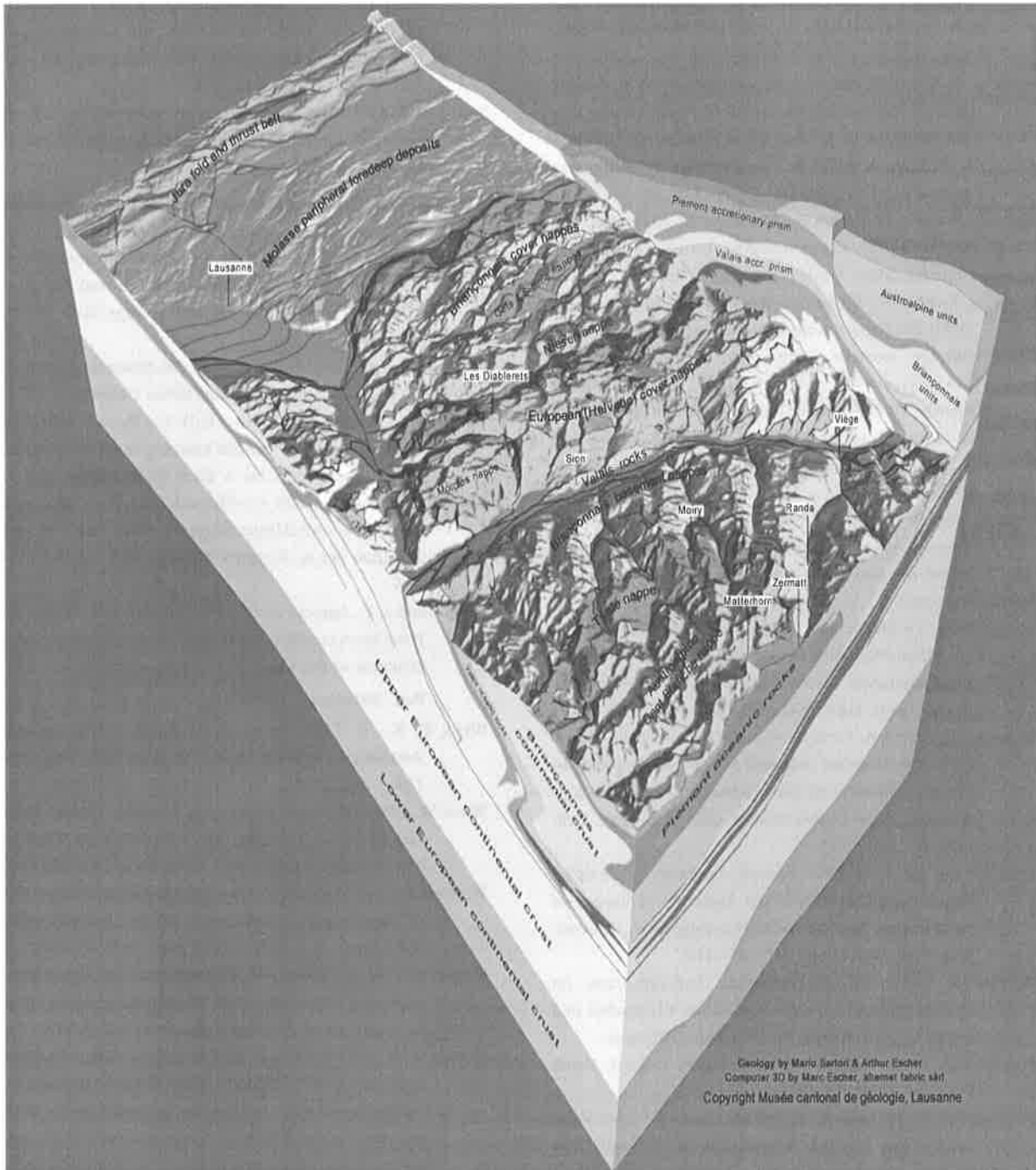
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Block diagram showing the main geologic units in the internal Western Swiss Alps and their relation with the present topography (Escher & Sartori, *Musée de Géologie, Lausanne, 2001*). See larger colored version in the pouch in the back-cover.

The overall tectonic framework of The externides domain

Jean-luc Epard

1 - INTRODUCTION

The area described in this chapter is located between the Prealpine units to the northwest and the Penninic thrust in the Valais region to the southeast (Fig. 18). It is part of a domain known as the Helvetics in the broadest sense. It consists of pre-alpine basement rocks that will be described and discussed in relation to Field Trip 1 and of Mesozoic and Cenozoic cover rocks belonging to several tectonic units. Some of them will be visited during Field Trips 2 and 3. These units are all derived from the European margin, more precisely from a domain which was the SE extension of the

area below the Jura mountains and Molasse basin. The description of this part of the Alpine chain will refer in part to the maps and to the general cross section of the Alps given in appendix. More information can be found also in Steck et al. (2001).

1.1 NAPPE STRUCTURE AND KINEMATICS

Four main tectonic units can be distinguished. They are described from the NW to the SE with respect to their original palinspastic position and from bottom to top from a structural point of view (Fig. 19 and 20).

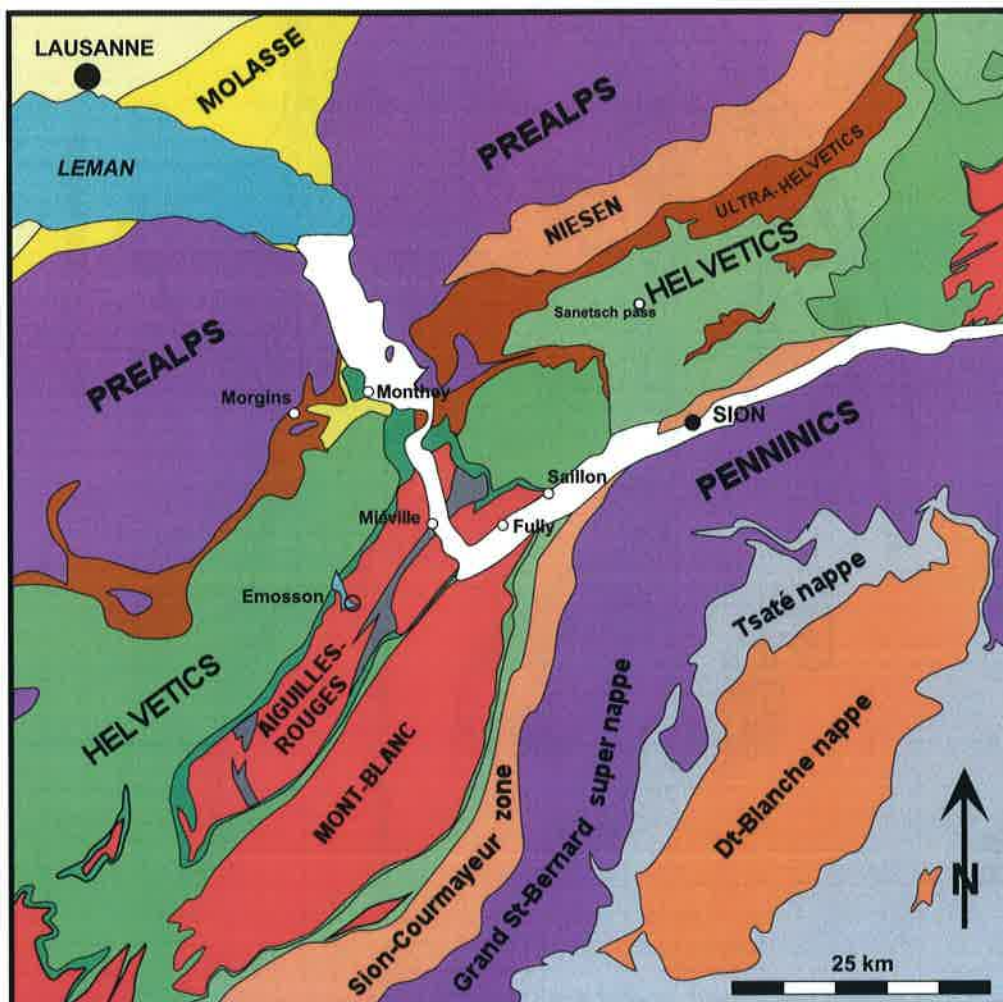


Fig. 18 - Tectonic sketch of the Alps of western Switzerland and bordering area (modified from Spicher, 1980).

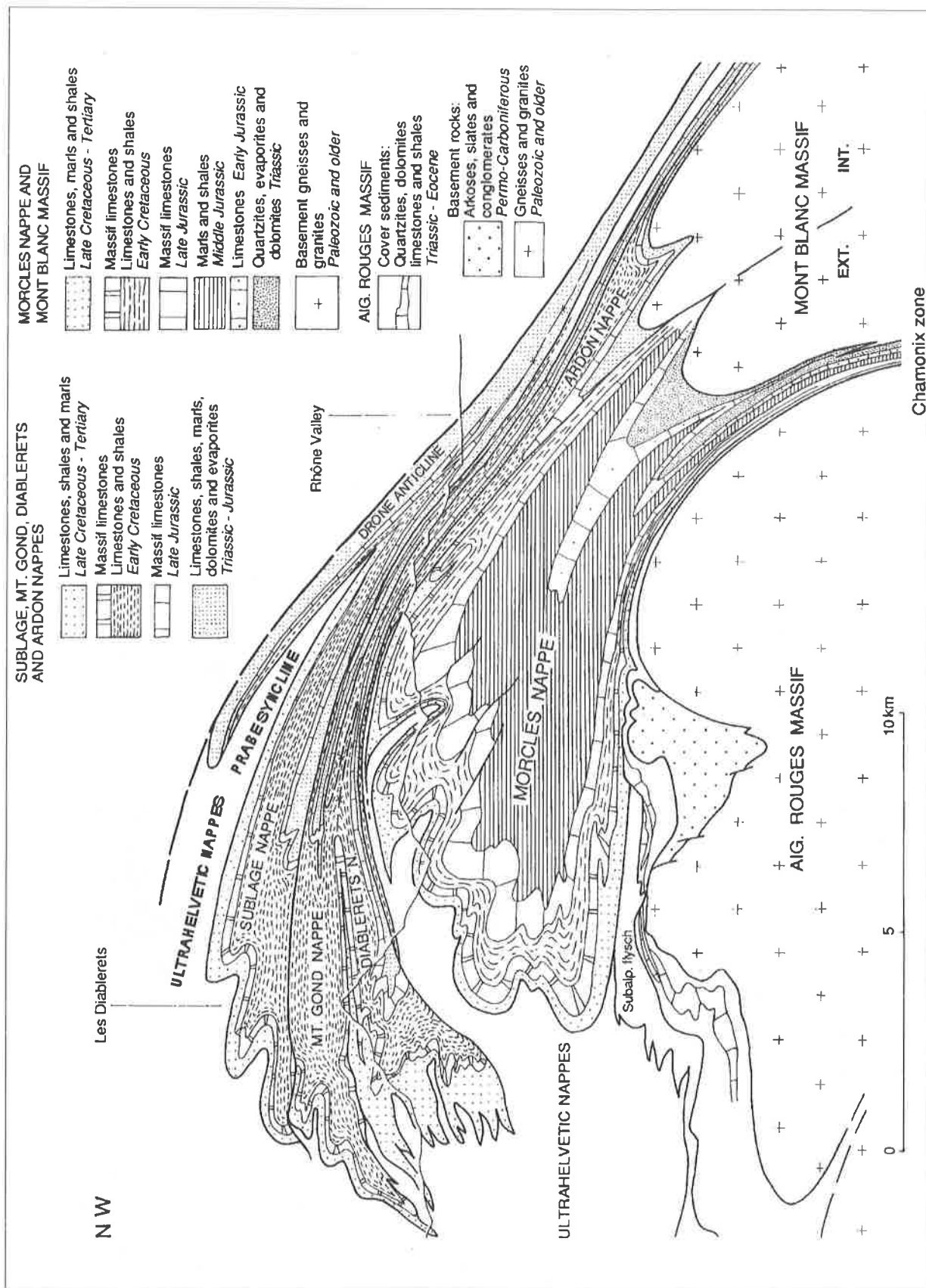


Fig. 19 - Cross section of the Helvetics (slightly modified from Escher et al. 1997).

1) The Aiguilles-Rouges massif and its autochthonous and parautochthonous cover :

The *Aiguilles-Rouges* crystalline massifs represents the more external (NW) part of the European margin outcropping in the area. The Variscan and pre-Variscan history of this massif is described in details in relation with Field Trip 1, including a discussion of the significance of the Late Carboniferous deposits preserved in the structure (trough) of Salvan-Dorénaz. A thin (up to a few hundreds meters) sedimentary cover is associated with this massif. It consists of Late Triassic quartzites and dolomites, Middle, Late Jurassic and Early Cretaceous limestones and Late Eocene breccias, marls and limestones topped by an Early to Middle Oligocene Flysch (Masson et al. 1980). The cover is in some places still in an autochthonous position relative to the Aiguilles-

Rouges massif, in other places, it can be translated a few kilometers northwestward to form parautochthonous thrust slices (Badoux, 1972). Below and in front of the Aiguilles-Rouges massif, some structures such as folds and thrusts have been deduced from seismic data (Steck et al. 1997). The horizontal shortening accommodated by these structures are, at least in part, responsible for the about 30km displacement on the Jura mountain detachment in the Triassic evaporites (Laubscher, 1961). A geometric model for the relationship between fold and thrust nappes is proposed in Epard and Escher (1996).

2) The Mont-Blanc massif, the Morcles and Ardon nappes

The *Mont-Blanc* crystalline basement (c.f. also introduction of Field Trip 1) lies south-east of the Aiguilles-Rouges massif and is separated from the

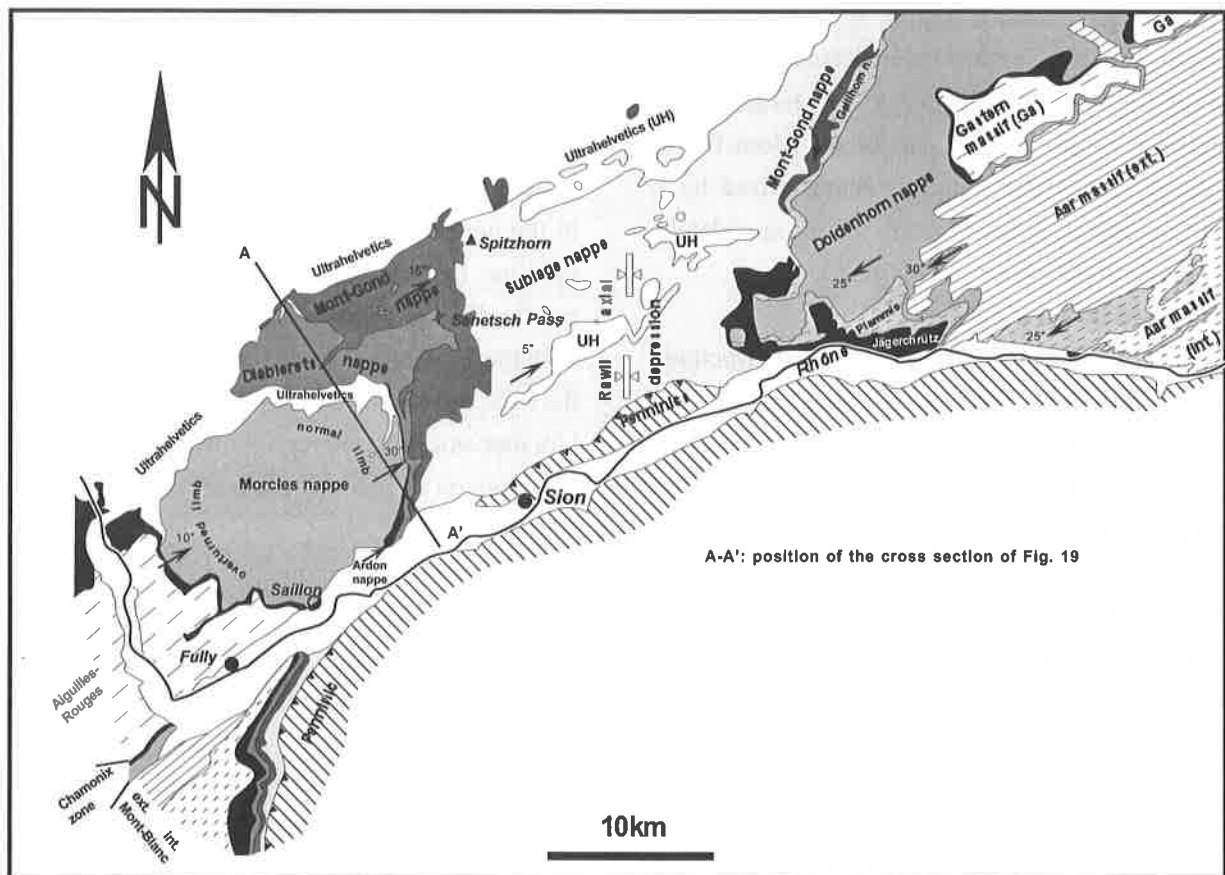


Fig. 20 - Tectonic map of the Helhetics of western Switzerland.

latter by a narrow band of Mesozoic sediments called the *Chamonix zone* (Fig. 19, 20). This massif can be divided in two parts separated by an Alpine fault or mylonitic zone, probably reactivated from older structures (Masson et al. 1980, Epard 1990). In this transect, the *Morcles nappe* is a typical and well developed example of a fold nappe (Escher et al. 1993). It is a large recumbent anticline formed by Mesozoic and Cenozoic sediments with smaller second order folds. It is thrust on top of the Aiguilles-Rouges massif and its cover. The tectonically thinned overturned limb that can be seen in the Saillon area (Field trip 3 Stop3A) contrasts with the rather thick normal limb. The external part of the Mont-Blanc crystalline basement forms the core of this fold nappe.

A thin unit, named the *Ardon nappe* (Masson et al. 1980), was thrust on the Morcles nappe, or more precisely was “dragged along” below the advancing Wildhorn nappe. It is made up of Cretaceous and Tertiary rocks detached from the Mesozoic sediments (Triassic to Late Jurassic) still attached to the internal part of the Mont-Blanc massif. This Ardon nappe is characterized by a very complex internal structure still not completely understood (Masson, comm. pers.).

3) The Wildhorn nappe

The Wildhorn nappe is the highest structural unit of the Helvetics. It is a “super” nappe that groups three different units which are, from bottom to top: the Diablerets, Mont-Gond and Sublage nappes. The base of the lowermost unit, the Diablerets nappe is marked by an important thrust that can be followed from the frontal part to the root zone of the nappes, SE of the Mont-Blanc massif. The last movement along this thrust has been dated by Crespo-Blanc et al. at 15 Ma. The Diablerets and the Mont-Gond nappes are connected by a deep, early formed syncline called “Synclinal de raccord”. The contact between the Mont-Gond and Sublage nappe is mostly a thrust with, locally, remnants of an overturned limb. A fourth impor-

tant structure can be observed in the Wildhorn nappe: the Drôme anticline. It is the innermost important anticline, with a core of Triassic rocks. Its normal, SE limb, is made of Liassic and Early Dogger (Aalenian) rocks. Its overturned, NW limb, displays a more complete stratigraphic sequence. The Drôme anticline is linked to the main part of the Wildhorn nappe by an important syncline- the Prabé syncline, in which the Ultrahelvetic is folded and preserved from erosion.

4) The Ultrahelvetic (UH)

The Ultrahelvetic is a general name that groups together several tectonic units. They have in common the following characteristics: a) their stratigraphic succession is similar to that of the Helvetic, and to the Wildhorn nappe in particular; b) their distinctive tectonic position: in front of or on top of the Morcles and Wildhorn nappes. They are found also in front of, and below the Prealps. These nappes are thought to have originated from the internal and more distal part of the Helvetic basin (Fig. 20). They have been thrust during an early phase on top of the Helvetic, already partly structured. The Ultrahelvetic can be preserved in the core of deep synclines such as the Prabé syncline. In its present position, the Ultrahelvetic can be divided in two groups. One is associated with the Wildhorn nappe (UH-W), the other with the Morcles nappe (UH-M). These two groups of Ultrahelvetic units are represented with a different signature on the general cross section of the Alps in annex.

The present geometric position of the Ultrahelvetic with respect to the Helvetic structures gives interesting information on the kinematic scheme of nappe emplacement. A simplified kinematics can be proposed (Steck et al. 2001):

- a) Thrusting of the Ultrahelvetic on the inner part of the future Sublage nappe domain (on the area of the future Prabé syncline).
- b) Formation of the Wildhorn nappe by the superposition of the Sublage, Mont-Gond and

Diablerets units. At that time, the Ultrahelvetic units had not reached the front of the Sublage nappe, as there are no Ultrahelvetetic units pinched in between the three units constitutive of the Wildhorn nappe. However, as pointed out before, the Ultrahelvetetic units were already thrust on the inner part of the Sublage where they are refolded in the early formed Prabé syncline.

- c) Thrusting of the Ultrahelvetic units on top of the already structured Wildhorn nappe and on the Morcles nappe.
- d) Thrusting of the Wildhorn nappe topped by its associated Ultrahelvetic units (UH-W) on the stack formed by the Morcles nappe and its associated Ultrahelvetic units (UH-M). This last phase produced the present day succession, from bottom to top: the Morcles nappe and its associated UH, then Wildhorn nappe and its associated UH.

1.2 THE RAWIL AXIAL DEPRESSION AND LATE BACKFOLDING

Axial depressions and culminations are typical features of the Alps of Western Switzerland. The external crystalline massifs are outcropping thanks to two important axial culminations. To the west, the crystalline basement is brought to higher elevation to form the Aiguilles-Rouges and the Mont-Blanc massifs. Note that the highest summit of Europe, the Mont Blanc (4807m), lies around this culmination. South-east of the Mont-Blanc massif, a much smaller massif, the Mt-Chétif is outcropping. The second axial culmination consists of the Gastern, Aar, and Gothard massifs. In between these two axial culminations lies the Rawil axial depression. The uppermost tectonic units, the Ultrahelvetic units, are preserved as klippen in this depression on top of the Sublage (Wildhorn) nappe. This typical geometric setting implies that the axial plunge varies along a SW-NE direction (Fig. 20). The axial trend of about 060°-240° is rather cons-

tant in the Helvetic domain. The plunge around the culmination area such as the Dt du Midi, Emosson, or around the Gastern massif is null to moderate, as well, obviously, in the depression around the Rawil pass. High plunge values can be found for example in the Lizerne valley (up to 30° to the northeast) or along the southwestern border of the Aar massif (30° towards the southwest).

Due to this axial depression, the lower tectonic units plunge at depth below the Rawil area. Therefore, excepted for the Sublage nappe, the precise correlation of tectonic units between both sides of the depression cannot be directly observed. However, several connections are widely accepted. For example, the Gastern and Aar massifs are the likely eastern equivalents of the Aiguilles-Rouges and Mont-Blanc massifs respectively. The Morcles and the Doldenhorn nappes are probably connected in the same way. Other units such as the Diablerets nappe have probably no equivalent on the other side of the depression.

The axial depression is not restricted to the Helvetic domain. The Austroalpine Dt-Blanche klippe is located on the western part of the southern extension of the same structure.

The precise origin of the depression is not fully understood. For the external part, it is probably related to a type of "en échelon" setting of late back folds (Steck et al. 1997). The position of the depression could well be related to the structure and composition of the basement at depth. For example, the location of the present day Rawil depression could be inherited in part from an important trough of weak Late Carboniferous sediments in the basement.

These late back folds are responsible also for the refolding of the complete Helvetic nappe stack as seen in the cross section of figure 19. Major thrusts and fold axial planes dip towards the SE in the internal root part of the Helvetic domain in the Rhône valley and dip towards the opposite direction at the front of the nappes. The crest of the fold is

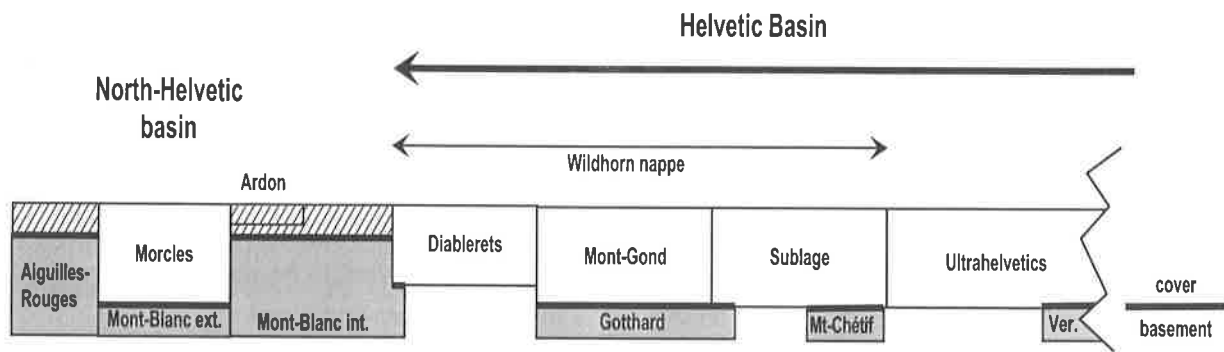


Fig. 21 - Palinspastic scheme and cover-basement relationship.

strongly influenced by the bulk shape of each individual nappe, thinning towards the root. Therefore, the crest position varies from nappe to nappe and is shifted upwards towards the NW

1.3 ORIGINAL POSITION OF THE TECTONIC UNITS AND BASEMENT-COVER RELATIONSHIP.

Figure 21 is a schematic palinspastic section that shows the relative position of the cover unit as well as their possible relations with the pre-alpine basements. This part of the transect of the North-European margin is characterized by two basins (North-Helvetic and Helvetic), and two basement highs with numerous periods of non-deposition or erosion. These two sub-basins form the larger Helvetic rim basin as shown in figures 8C and 12.

From north to south, that is from the external to the internal part, the following sedimentation domains can be observed:

1) The Aiguilles-Rouges basement high. The sedimentation on this area is characterized by several periods of non-sedimentation or erosion and therefore by a rather thin stratigraphic sequence. The typical succession is formed by Late Jurassic limestones resting on Triassic series. Northwestward, this sequence is completed by Early Cretaceous sediments and Oligocene Flysch.

2) The North-Helvetic basin is the area from where the Morcles nappe originated. The Mesozoic and Cenozoic sediments were originally

deposited on the external part of the Mont-Blanc crystalline basement. The sediments thickness can be estimated to about 3km. Its width can be gauged at about 25km (minimum) based on section balancing of the outcropping part of the nappe. The North-Helvetic basin is the NE extension of the Dauphinois basin in France. It narrows towards the NE and closes in the area of the Aar-Gastern transect.

3) The stratigraphy on the sediments deposited on the internal part of the Mont-Blanc massif is also characterized by several gaps and unconformities. It formed a basement high during several periods. The stratigraphic gap increases towards the S or SW. First, Triassic and Liassic rocks are absent and a basal conglomerate of Middle Jurassic (Aalenian) age lies directly on the basement (Grasmück, 1961). Farther to the southwest, near the Italian gate of the Mont-Blanc tunnel, Early Cretaceous (Albian) rocks transgress directly on the Mont-Blanc basement (Compagnoni et al. 1964).

4) The Helvetic basin is the origin of the Wildhorn nappe and the Ultrahelvetics (s.s.) units. The initial position of the Diablerets, Mont-Gond, Sublage and Ultrahelvetics nappes from a proximal to a distal position with respect to the internal Mont-Blanc basement high results in an increasing content of clay minerals in sediments. This tendency is specially well marked in Cretaceous formations (Barremian-Aptian) where the transi-

tion from platform to slope deposits can be observed.

The possible relation between the pre-alpine basement and the cover tectonic units is also summarized in Figure 21. As already mentioned before, the “homeland” of the Morcles nappe is the external part of the Mont-Blanc massif. Part of the Wildhorn nappe can be followed eastward and correlated with the cover of the Gotthard massif. Southwestward, a correlation between the Mont-Chétif can be made with the internal part of the Sublage nappe. Only a part of the pre-alpine basement of the Wildhorn nappe and of the Ultrahelvetic is outcropping. Some parts are still deeply buried below the Penninic nappes and partially known by seismic data (see general cross section of the Alps in annex).

1.4 STRATIGRAPHY OF THE COVER UNITS

Two general stratigraphic sections are given in figure 22. They concern the North Helvetic and Helvetic basins. The series related to basement highs are thinner and incomplete. Facies can vary slightly from one unit to the other, partly in relation to their original distance to detritic sources.

Triassic

The typical Triassic of the Helvetic begins with a rather thin (10m) detritic series (Vieux Emosson Formation, Epard 1989), transgressive on a locally deeply altered basement, followed by a few tens of m of carbonate rocks (mainly dolomites) and, locally, evaporites. In the area of Emosson (cover of the Aiguilles-Rouges massif), the basal detritic formation has been dated as Late Ladinian – Carnian, based on dinosaurs footprints (Demathieu and Weidmann, 1982).

Jurassic

The **Liassic** is made of marls and limestones (Hettangian-Sinemurian) generally followed by sandy limestones rich in echinoderms (Pliensbachian) then by a thick marly formation

(Toarcian). There are two periods of input of detritic material in the basins, the first during the Late Sinemurian (Lotharingian), the second during the Late Pliensbachian (Domerian). These events can be related to the onset of rifting in the Alpine Tethys. This period represents also the first clear distinction between the North-Helvetic and Helvetic basins with respect to the Aiguilles-Rouges and Internal Mont-Blanc basement highs characterized by significantly coarser detritic material.

The **Dogger** begins with a characteristic formation of black shales, very thick (several 100m) in some place. It represents an important (perhaps the most important) detachment horizon in the Helvetic. In the North-Helvetic basin, this formation is followed by siliceous limestone or by shaly siltstones in the Helvetic basin. The top of the Dogger (Callovian) is a relatively thin formation of black shales and siliceous limestone. It turns gradually to the characteristic Late Jurassic white limestones.

The **Malm**, more precisely the latest part of the Malm (Kimmeridgian-Portlandian) is formed by a massive white mostly micritic limestone. Its forms very typical cliffs easily recognizable in the landscape. This formation is very similar throughout the Helvetic domain.

Cretaceous

The early part of the Cretaceous consists of marls and marly limestones of **Berriasian** age. It forms an important detachment level or disharmonic horizon, particularly in the Wildhorn nappe. In the North-Helvetic basin (Morcles nappe), the **Valanginian** consists of platform limestones and the **Hauterivian** of siliceous limestones. The facies in the external part of the Helvetic basin (Diablerets and Mont-Gond units) are comparable. In the more internal part of the Helvetic basin, they turn to more distal and shaly facies. The platform facies of **Barremian-Aptian** age is named **Urgonian**. The Urgonian limestones forms a white

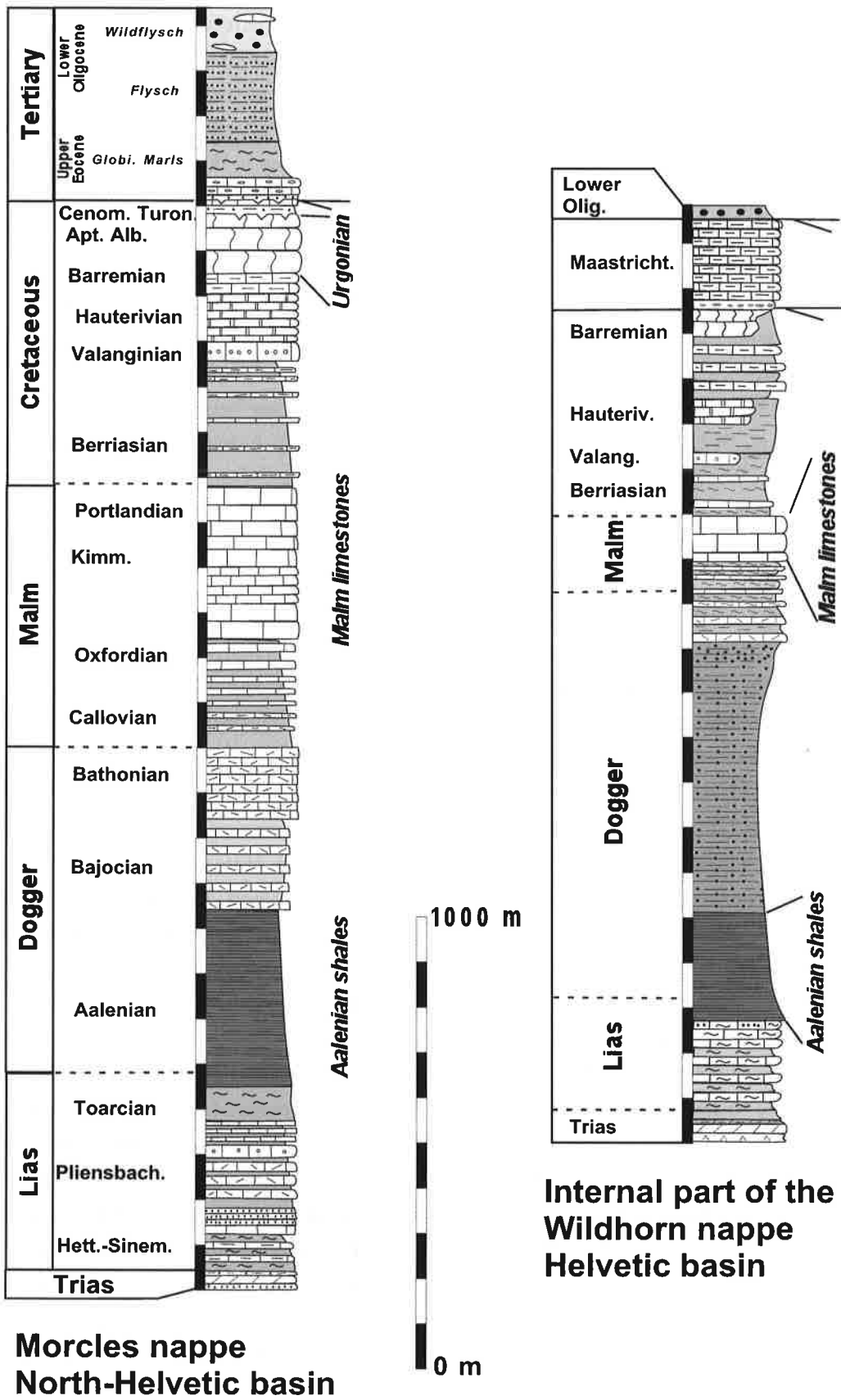


Fig. 22 - Stratigraphic section for the North-Helvetic and Helvetic basin.

cliff and enhanced most of the fold so typical of the frontal part of the Helvetics. Laterally towards the southwest, the platform facies disappears and turns into an alternation of marls and limestones made up of materials derived from the platform. In the internal part of the Helvetic basin, the Urganian platform facies does not exist anymore and all the formation of Early Cretaceous age tends to be much more shaly. Only thin (a few tens of meter) formations of **Late Aptian** to **Turonian** age are locally preserved under the Maastrichtian or Tertiary unconformity. On this transect, Late Cretaceous (**Maastrichtian**) rocks are preserved only in the southwest (internal) part of the Wildhorn nappe (Sublage unit) and in the Ultrahelvetic. It is formed of well bedded limestones and marls (slope deposits). It lies unconformably on older Cretaceous rocks, Turonian in the NW and Barremian in the SW of the Wildhorn nappe, or directly on Late Jurassic limestones in the Sex-

Mort Ultrahelvetic nappe.

Tertiary

In the external part of the Helvetics, Early Tertiary is characterized by an emersion and subsequent erosion. For example in the Morcles and Diablerets nappes, paleokarst structures are filled with continental reddish sandstone called "sidérolithique". The older marine deposits are Early Eocene in age (Ypresian) in the internal part of the Wildhorn nappe and become younger (Late Eocene, Priabonian) towards the NW (Fig. 12). Generally, the Tertiary marine deposits begin with a few meters of sandstones and sandy limestones followed by platform limestones generally rich in Nummulites that gradually turn into marls then Flysch deposits. The Flysch deposits that end the Helvetic sedimentation are Early Oligocene in age. They locally contain basic volcanic detritic material (Taveyenne Sandstone) dated at 32,5 Ma or derived from an ophiolite nappe (Val d'Illiez Sandstone).

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Mont Blanc – Aiguilles Rouges Massifs – (External Massifs)

AN EXAMPLE OF POLYOROGENIC EVOLUTION

François Bussy, Juergen Von Raumer & Nicola Capuzzo

I: INTRODUCTION

1 - PRE-MESOZOIC BASEMENT IN CENTRAL EUROPE

The Pre-Mesozoic basement of Central Europe (Alps included) mostly appears as polymetamorphic domains juxtaposed through Variscan and/or Alpine tectonics (*e.g.* in Iberia, Armorica, Moesia, French Massif Central, Saxothuringian and Moldanubian Domains, External Massifs, Penninic Domain, parts of the southern Alps and the Austroalpine basement). Consequently, Variscan/Alpine structures prevail in most of these basement areas and relicts of former geological events from the Precambrian to the Ordovician are difficult to unravel and to correlate (*e.g.* von Raumer and Neubauer 1993).

The distribution of Cadomian-type basement units and their associated granitoids, detrital sediments, volcanites and Cambrian oceanic crust, as well as provenance studies of detrital zircons and Sm-Nd data (Nance & Murphy, 1994, 1996) all point to a common, Gondwana-derived origin for these relict basement pieces, including the Avalonia microcontinent (fig.16). Identification of subsequent sequences reminiscent of plate-tectonics *i.e.* successive stages of oceanic crust, volcanic arcs, active margin settings and collision zones during the Early Paleozoic led to the geodynamic model proposed in the main introduction of this guide book. This model postulates a rather continuous Gondwana-directed subduction since the Late Proterozoic (von Raumer et al., 2001), and can be summarized as follows (fig.16):

a) a Late Proterozoic active margin setting with formation of volcanic arcs is observed in the

entire length of the future microcontinents at the Gondwanan border, and granites of Late Cadomian age (± 550 Ma) are common in many of the Gondwana-derived basement blocks. Detrital sediments of Late Proterozoic to Early Cambrian age carry the fingerprints of Cadomian/peri-Gondwanan origin. The corresponding sedimentary troughs prepared the future location of the Rheic ocean, which resulted from continuing oblique subduction and rifting in a back-arc situation accompanied by strike-slip movements.

- b) Drift of Avalonia and opening of the Rheic ocean were enhanced after subduction of an oceanic ridge, whereas in the eastern continuation of Avalonia only early stages of the Rheic ocean may have existed.
- c) Drift may have been delayed in the eastern continuation, and the oceanic ridge may have triggered the consumption of the Rheic ocean and the amalgamation of volcanic arcs and continental ribbons with Gondwana in a rather short-lived orogenic event, before the opening of Palaeotethys during the Ordovician, preparing the drift of the composite Hunsu-tertrane (Stampfli 2000).

Depending on their former location, pre-Variscan basement areas hidden in the Variscan belt (Alps included) may thus contain Cadomian elements, Late Proterozoic detrital sediments and volcanic arcs, relicts of the Rheic ocean, Cambro-Ordovician accretionary wedges, relicts of an Ordovician orogenic event and its related granites, as well as volcanites and sediments linked to the opening of the Palaeotethys.

2 - TIMING OF EVENTS IN THE EXTERNAL ALPINE REALM

The so-called "External Crystalline Massifs" of the French and Swiss Alps (*i.e.* Argentera, Pelvoux/Haut-Dauphiné, Belledonne-Grandes Rous-ses, Mont Blanc-Aiguilles Rouges, Aar-Tavetsch-Gotthard) represent pre-Mesozoic basement nappes or slices appearing as Alpine antiform cores among their Mesozoic covers (fig.2). They are located in the Helvetic realm, the external domain of the Alps. As a consequence, they were moderately affected by the Tertiary Alpine metamorphism and preserved most of their Paleozoic features. In particular, the Aiguilles Rouges Massif is known as the first place in the world where superimposed orogeneses (*i.e.* Alpine and Variscan) were clearly identified (Oulianoff 1953, in Ramsay 1967 p.519).

Apart from the low-grade Alpine overprint, the External Massifs recorded several pre-Mesozoic metamorphic episodes (von Raumer et al., 1999a). The main one resulted from the Variscan orogeny, when nappe stacking brought many units to high-amphibolite facies conditions and local anatexis. Despite this major imprint, many relicts testify to an earlier evolution comprising Late Precambrian rifting (sedimentation, formation of oceanic crust), Early Palaeozoic arc formation and subduction, and intrusion of Ordovician granitoids. We are thus dealing with a poly-orogenic evolution, comprising Alpine, Variscan, Ordovician and Neoproterozoic events. It is substantiated by isotopic ages (Tab. 1) and synthetic data from all External Massifs, which can be summarized as follows:

- Late-Proterozoic to Cambrian rifting and oceanization are inferred from paragneisses (metagrauwackes, marbles, metavolcaniclastic horizons, quartzites) hosting detrital zircons older than 600 Ma (Gebauer, 1993), deposited on a slowly subsiding continental shelf, as well as from the 496 Ma old Chanrousse ophiolite

(Ménot et al., 1988b).

-An Ordovician subduction cycle is documented by relicts of MORB eclogites found in all massifs, in particular in the Aar and Gotthard, where eclogitization is bracketed by 467-475 Ma old island-arc type gabbros and the 440 Ma post-HP intrusion of granitoids (Abrecht, 1994; Abrecht et al., 1991, 1995; Abrecht and Biino, 1994; Biino, 1994, 1995; Oberli et al., 1994). Large volumes of S- and I-type granitoids (the so-called younger orthogneisses) intruded between ca. 460 and 440 Ma.

-The Devonian evolution is geochronologically poorly recorded so far (Tab. 1), with the notable exception of trondhjemitic intrusions at 365±17 Ma in Belledonne (Ménot et al., 1988a), suggesting continental rifting. Traces of Early Devonian nappe tectonics might be locally preserved in Belledonne and Aiguilles Rouges.

-The subsequent Carboniferous evolution is recorded at different levels of a supposedly large nappe pile. Wrench tectonic seems to be active during all this period, either in transpressive or transtensive mode. It is accompanied by (and possibly triggering?) important exhumation processes coupled to vigorous erosion. By Stephanian times, more than 10 km were stripped off all massifs. Very coarse to fine-grained sedimentation mixed with volcanoclastic material is recorded in intramountain basins of Early- and Late-Carboniferous age, respectively. Deep-seated units were affected by a Barrowian-type metamorphism of high amphibolite grade and locally by decompression melting (ca. 320 Ma in Aiguilles Rouges, Bussy et al., 2000). Several short-lived pulses of granitic magmatism are recorded, whose typology reflects progressive readjustment of the Variscan lithosphere (Bussy et al., 2000). Plutons are essentially syntectonic and intruded along transcurrent fault zones. A first pulse of high-K monzonitic to shoshonitic magmas (340-330 Ma) originated in a metasomatized lithospheric mantle

age, methode	rocks	reference
A Early Palaeozoic events		
496±6 Ma U/Pb Zrn	plagiogranite	Ménot et al., 1988a
497±27 Ma Sm/Nd	Chamrousse ophiolite	Pin and Carme, 1987
489±22 Ma U/Pb Zrn	granite, Belledonne	Barfétý et al., 1997
461 Ma Sm/Nd grt wr	eclogite, Gotthard massif	Gebauer et al., 1988
468 Ma U/Pb Zrn	eclogite, Gotthard massif	
467-475 Ma U/Pb Zrn	island arc gabbro, Gotthard + eclogite facies	Oberli et al., 1994
479±3 Ma U/Pb Zrn	island arc gabbro	Abrecht et al., 1995
475-450 Ma U/Pb Zrn	Ext. M., metabasic protoliths	Paquette et al., 1989
445±2 Ma U/Pb Zrn	migmatites, Aar massif	Schaltegger, 1992, 1993
456±2 Ma U/Pb Zrn		
485±50 Ma SHRIMP	Peclet orthogneiss	Bertrand et al. 2000
458±41 Ma, SHRIMP	Modane granite	Bertrand et al. 2000
436±17 Ma Rb/Sr wr	granitoids, Gotthard	Arnold, 1970a,b
>440 Ma U/Pb Zrn	granitoids, Gotthard	Bossart et al., 1986
439±5 Ma U/Pb Zrn	granitoids, Gotthard	Sergeev and Steiger, 1993
453±3 Ma U/Pb Zrn	granitoids, Mt. Blanc	Bussy and von Raumer, 1994
438±5 Ma U/Pb SHRIMP	meta-dacite, Argentera	Lombardo et al., 1997a
424 Ma U/Pb Zrn	eclogite, Argentera	Paquette et al., 1989
B Devonian		
395 Ma U/Pb Zrn	eclogite, Belledonne	Paquette et al., 1989
370 Rb/Sr thin slab	mylonites, Aiguilles-Rouges	Thöni, 1989
352±56, 365±17 Ma U/Pb Zrn	Trondhjemite, Belledonne	Ménot et al., 1988b
350-373 Ma Ar/Ar Ms	Argentera	Monié and Maluski, 1983
356±3 Ma SHRIMP	Cogne granodiorite	Bertrand et al. 2000
C Late Variscan evolution		
<i>shoshonitic-ultrapotassic- monzonitic rocks</i>		
332-334 Ma, U/Pb Zrn	Aar massif	Schaltegger, 1994
333 Ma U/Pb Zrn	Tödi granite, Aarmassif	Schaltegger & Corfu, 1995
332±2 Ma U/Pb Zrn	Pormenaz, Aig. Rouges	Bussy et al., 1997
337±8 Ma U/Pb SHRIMP	Argentera	Lombardo et al., 1997b
332±13 Ma Zrn Pb ev	Sept Laux granite, Belledonne	Debon et al., 1998
315-330±7 Ma Ar/Ar Bt	Argentera, Tinee	Monié and Maluski, 1983
324±12 Ma K/Ar Am	Belledonne	Ménot et al., 1987
324±22 Ma K/Ar Am, Bt	Pelvoux	Vittoz et al., 1987
337-331 Ma Ar/Ar Am	Western Gneiss Unit, w.Aig.R.	Dobmeier, 1998
334 Ma Ar/Ar Am	iron-skarns; Mont-Blanc area	Marshall et al., 1997
330±3 Ma U/Pb Zrn	Aar massif, greenschist to low amphibolite facies	Abrecht et al., 1995
327 U/Pb Mnz	metapelite, therm. peak, Aig. R.	Bussy and Hernandez, 1997
317 Ma; U/Pb Mnz	migmatite Mt. Blanc	Bussy and von Raumer, 1994
321 Ma, U/Pb Mnz	migmatite Aig. Rouges	Bussy and Hernandez, 1997
308-310 Ma, U/Pb Zrn	diorites, granitoids, Aar massif	Schaltegger, 1994
307 Ma, U/Pb Zrn	gabbros, granitoids; Aig. R.	Bussy and Hernandez, 1997
303±3 Ma U/Pb Zrn	Mt. Blanc granite	Bussy et al., 1989
308 Ma U/Pb Zrn	acidic volcanics, Aig. Rouges	Capuzzo and Bussy, 1998, 2000
±300 Ma, U/Pb Zrn	partial melting Aar massif	Olsen et al., 1997, 2000
299-303 Ma U/Pb Zrn	pyroclastic deposits, Aar massif	Schaltegger & Corfu, 1995
298 Ma, U/Pb Zrn	younger granitoids, Aar massif	Schaltegger, 1994
294, 299 Ma U/Pb Zrn	younger granitoids, Gotthard massif	Bossart et al., 1986
		Guerrot and Steiger, 1991
		Sergeev et al., 1995
295±3 Ma U/Pb Zrn	acidic volcanics, Aig. Rouges	Capuzzo and Bussy, 1998, 2000b

Tab.1 - Isotopic ages from the External Massifs (wr = whole- rock age; ev = Zrn evaporation age).

with variable lower crustal contamination. A second pulse (310-306 Ma) mainly consists of peraluminous crustal-derived granitoids associated to non-shoshonitic gabbros and diorites, whereas a third pulse (303-295 Ma) includes alkali-calcic (sub-alkaline) granites of mixed mantellic-crustal origin.

-Permian events are hardly recorded in the External Massifs, where Mesozoic sediments rest directly on any kind of pre-Permian lithology. Late Carboniferous erosion and peneplanation carried on in an extensional tectonic regime up to the transgression of the Mesozoic sea.

3 - GEOLOGICAL OUTLINE OF THE MONT BLANC / AIGUILLES ROUGES MASSIFS

The present-day Aiguilles Rouges-Mont Blanc massifs consist of a complex assemblage of tectonic units with contrasting maximum P-T metamorphic conditions, separated by major, steeply dipping NE-SW faults and/or mylonitic zones (see fig. 18). Most of these tectonic contacts probably formed during the late Variscan strike-slip regime and were reactivated during the Alpine orogeny. These two massifs are essentially composed of polymetamorphic, amphibolite-facies grade rocks and granitic plutons. Low-grade monometamorphic detrital sediments with interlayered volcanites record an Early Carboniferous basin development at the southern end of the Aiguilles Rouges massif (Dobmeier 1996). On the other hand, unmetamorphosed continental, coal-bearing deposits of Late Carboniferous age (308-297 Ma, Capuzzo & Bussy 2000a,b) are preserved in the Salvan-Dorénaz Alpine syncline (northern part of the Aiguilles Rouges massif) and Saint Gervais – Les Houches area (southern part of the Aiguilles Rouges massif).

Age determinations in the polymetamorphic units of the Aiguilles Rouges – Mont Blanc area

(Tab. 1) demonstrate a polyorogenic evolution, comprising Variscan, Ordovician and Late Precambrian events (Von Raumer et al. 1999a). Magmatic ages on zircon of c. 450 Ma have been obtained both for eclogitized MORB-like basic rocks (Paquette et al., 1989) and for S-type and I-type calc-alkaline non-eclogitized metagranites (“Ordovician granitoids” in fig.23) (Bussy & Von Raumer 1994). The latter intruded detrital sequences (now paragneisses) of supposedly Late Precambrian to Ordovician age, composed of sandstones and graywackes, with minor carbonate intercalations and tholeiitic basaltic layers. Flysch-type sediments enriched in Cr and Ni, together with eclogites and ultrabasic rocks (Aiguilles Rouges, von Raumer and Fracheboud, unpublished data) might represent deposits in a former accretionary prism (Von Raumer 1998). These lithologies have been interpreted as evidence for a Late Precambrian to Cambrian rifting/drifted episode with opening of an oceanic domain, followed by an Ordovician subduction, either in an island-arc, or an active continental margin environment, linked to the southward subduction of the Rheic oceanic lithosphere underneath Gondwana (see geodynamic model above). The subsequent high P event (700°C/ >14kbar) recorded by eclogites of the Lake Cornu area (Aiguilles Rouges) is not precisely dated. Paragneisses display a succession of deformation events attributed to the Variscan orogeny, with thrust tectonics (Dobmeier 1998) and nappe stacking, leading to the development of a Barrowian-type metamorphism (Von Raumer et al. 1999a). Metapelites record a typical clockwise P-T path, as commonly found in the internal parts of the Variscides, with a peak T at about 327 Ma (Bussy et al., 2000). Rocks of suitable composition experienced decompression melting during exhumation at c. 320 Ma (Bussy et al., 2000).

Magmatic rocks are widespread; as pre-Carboniferous (450-460 Ma) medium- to high-grade metamorphosed granites (e.g. Bussy & Von

Raumer, 1993; Wirsing 1997; Paquette et al. 1989; Von Raumer et al. 1990; Dobmeier et al. 1999) or as Carboniferous, essentially non- to weakly metamorphosed intrusions. Subvolcanic facies are associated to some of the intrusions, whereas volcanic horizons are interlayered in the Early- (Dobmeier 1996) and Late Carboniferous (Capuzzo & Bussy 2000a,b) detrital basins, respectively.

4 - THE SEDIMENTARY RECORD

Pre-Carboniferous sediments in the External Crystalline Massifs experienced a strong metamorphic overprint and are only crudely datable on the basis of their inherited zircon content (*e.g.* Gebauer, 1993) or crosscutting relationships with dated magmatic intrusions. A notable exception is the Cambrian - Ordovician age of black schists in the Grandes Rousses massif (Huez Formation, Giorgi et al., 1979), based on the discovery of Reitlingerellides fossils. On the other hand, Viséan detrital deposits are better preserved and have been identified by fossils in the Taillefer detrital series of the Belledonne-massif (crinoids, Gibergy, 1968), and in the low-grade metapelitic series of the Aiguilles Rouges massif (acritarchs, Bellière and Strel, 1980). Upper Carboniferous sediments are only affected by Alpine metamorphism and host both datable volcanic layers and abundant plant fossils. In this chapter, we will focus on the Carboniferous sedimentary record of the Aiguilles Rouges/Mont Blanc area, the best preserved of all External Massifs.

4.1 LOWER CARBONIFEROUS DEPOSITS

They outcrop in Servoz-Les Houches, at the southwestern end of the Aiguilles Rouges massif (fig. 23) in two bands on either side of the Montées Pélissier granite. This is where Bellière & Strel (1980) dated Lower Carboniferous sedimentary rocks for the first time using palynology (Late Viséan acritarchs), thus allowing a clear separation

from the nearby upper Carboniferous deposits. These detrital rocks consist of metamorphosed and variably deformed phyllites, graywackes and sandstones, which recrystallized in greenschist facies conditions [chlorite zone, $qtz + Chl + Ser + Pyr$] (Dobmeier, 1996, 1998). Deformation is penetrative, although original sedimentary features are still recognizable. Several fold phases and associated structures (including mylonites) developed during a long-lasting transpressive regime (Dobmeier, 1998).

Interlayered with the metagraywackes are found meter-thick bands of green metavolcanites (*e.g.* at the train station of Les Houches) of basaltic to andesitic composition ($SiO_2 = 49-60$ wt%), consisting of $Plg + Chl \pm Amp \pm Qtz$. Trace-element chemistry points to Fe-basalts of E-MORB affinity (Dobmeier, 1996), possibly recording Early Carboniferous transtension linked to the opening of the sedimentary basin and to the 330-340 Ma high-K magmatic pulse.

4.2 UPPER CARBONIFEROUS DEPOSITS (THE SALVAN-DORÉNAZ BASIN)

They are remarkably exposed in the so-called Salvan-DorénaZ Alpine syncline (northern part of the Aiguilles Rouges massif), one of the best preserved example of intramountain sedimentary basins of the Variscan Alps (Capuzzo 2000, unpub. thesis; Capuzzo and Wetzel 2000). This structure has an asymmetric, half-graben geometry, up to about 4 km wide, filled along its northwestern side with up to 1.7 km of sediments that thin to the southeast (fig. 24). Sediments are exposed for 25 km in a NNE-SSW direction along the eastern margin of the Aiguilles-Rouges massif, and are separated from the Aiguilles Rouges basement units, and from the Vallorcine granite, by a steep SE dipping mylonite zone, which may have been active during the Late Carboniferous as a right-lateral strike-slip, transtensive fault (Brändlein et al., 1994). Other steeply dipping N-S to NE-SW

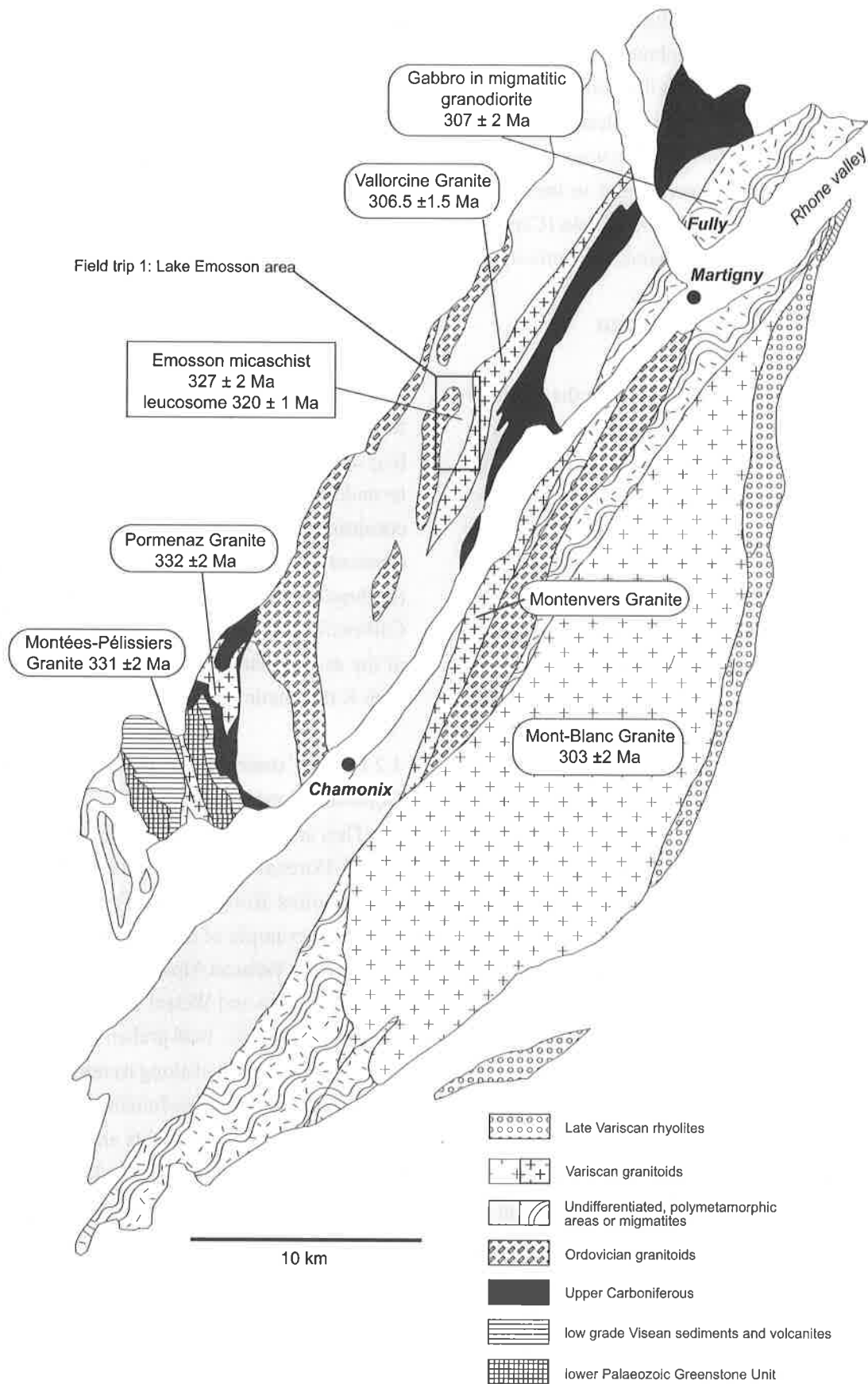


Fig. 23 - Simplified geological map of the Aiguilles Rouges-Mont Blanc massifs, after von Raumer et al. (1999).

oriented faults located near the basin margins seem to have affected the structural evolution of, and the sedimentary facies within, this basin (Pilloud, 1991; Niklaus and Wetzel, 1996). Two Alpine deformations in the brittle-ductile field, resulted in the complex synclinal structure of the basin with fold axes generally dipping 15° - 20° toward the northeast (Pilloud 1991; Badertscher and Burkhard 1998). Consequently, increasingly deeper parts of

the basin are exposed to the southwest. Illite-crystallinity of upper Carboniferous sediments indicate that Alpine metamorphism attained anchimetamorphic grades (Pilloud, 1991; Frey et al., 1999). A low-angle regional unconformity between the upper Carboniferous and onlapping shallow marine Triassic deposits indicates Permian erosion, probably related to moderate inversion of the basin along its flanks (Pilloud, 1991;

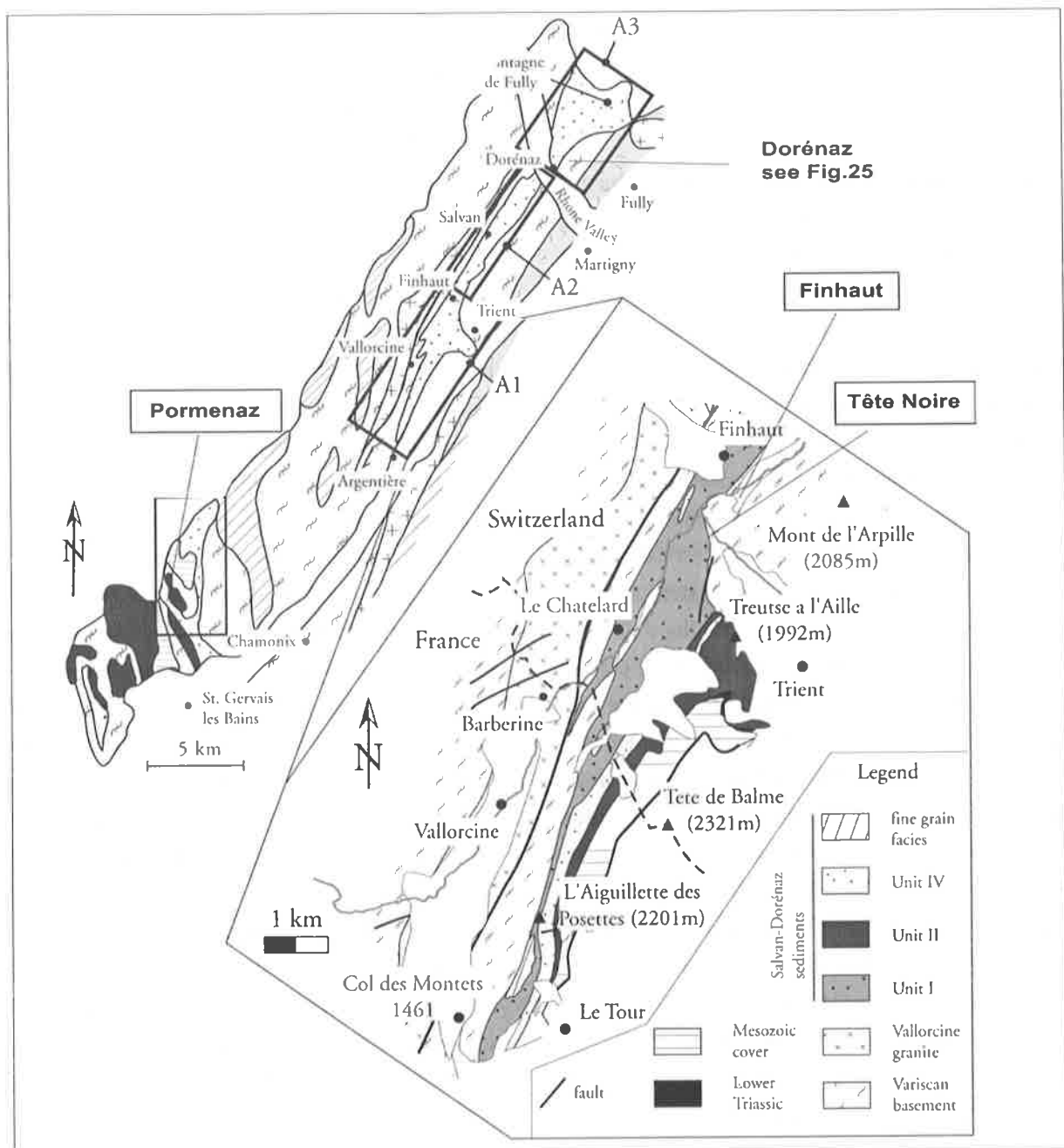


Fig. 24 - (from Capuzzo 2000, Fig. 13) Schematic map of the Aiguilles-Rouges massif (modified after Brändlein et al., 1994). and distribution of sedimentary facies in Late Carboniferous sediments. (with indication of localities).

Badertscher and Burkhard, 1998).

The age of the basin fill was first determined from palaeofloral associations, and later from isotopic dating. Macrofloral determinations indicate Late Westphalian (C-D) at the base of the succession and Stephanian ages further up (Jongmans, 1960; Weil, 1999, unpubl. data). Recent radiometric age determinations on synsedimentary volcanic deposits constrain the basin fill to the Late Carboniferous (Capuzzo and Bussy, 2000a), with ages of 308 ± 3 Ma for basal dacitic flows, and of 295 ± 3 Ma for a tuff layer from the upper levels of the basin.

4.2.1 Evolution of the Salvan-Dorenaz basin

Based on the structural analysis of Pilloud (1991), Niklaus and Wetzel (1996) and Capuzzo (2000), four lithologic units can be distinguished (alluvial fans and braided, anastomosed and meandering river deposits), which record a sedimentary evolution in a strike-slip tectonic regime (fig.25).

Unit I: The evolution of the Salvan-Dorénaz basin started at the end of the Westphalian (308 ± 3 Ma) with mainly coarse-grained clastics forming an alluvial fan system from the western margin, an overall wedge-shaped body thinning to the SE. Intense weathering produced abundant clastic material mainly derived from metamorphic and igneous rocks (Sublet, 1969; Niklaus and Wetzel, 1996). Granitoid boulders of Late Carboniferous age imply rapid uplift and denudation in the source areas. The sediments suggest deposition in an intramountain setting affected by active faulting and probably rapid uplift in the catchment areas. Mass flows and debris flows dominate the proximal areas of the alluvial fan systems close to the footwall slope, whereas the distal parts are characterised by braided distributary channels (figs 25 and 26). All climate indicators, especially a rich flora, point to a humid, seasonal climate. The groundwater table was probably close to the land

surface, as dark coloured, hydromorphic paleosols dominate.

Unit II: A drastic change in facies association occurred as braided river deposits (Unit I) were overlain by mud-dominated floodplain deposits. The fine-grained alluvial plain sediments accumulated in a swampy environment with anastomosed, sand-filled channels (fig.26). They display palaeoflow to the NE and document an axial drainage. Rapid subsidence led to the reduction of valley slope, and preservation of palaeosols and primary-structured volcanoclastic deposits suggest a rapidly subsiding setting. Asymmetric subsidence is indicated by shallow-lacustrine and peat-swamp deposits along the western side of the basin. The spatial association of localised "black shales" and coals deposits with fault zones is ascribed to differential subsidence leading to the formation of shallow, temporary lakes. In the upper part of Unit II laterally persistent, thick, caliche-bearing palaeosols can be considered as evidence for low aggradation rates, during periods of tectonic quiescence under semiarid, seasonal climate. The matrix of the conglomerates and breccias deposited on the alluvial fan at the northwestern side, however, document the onset of climatic changes during early Unit II, when the hydromorphic conditions prevailing during Unit I changed to well drained conditions producing red soils indicating a warm, at least seasonally dry climate which corresponds to the Late Carboniferous-Early Permian climate scenario in the European Variscides, located within the equatorial belt (see fig.14) (Ziegler, 1990; Scotese and McKerrow, 1990).

Unit III: The anastomosed river deposits of Unit II gradually change to meandering river deposits, which reflect the readjustment of the fluvial system to an increasing valley gradient and decreasing accommodation rate. The reversal in palaeoflow from NE (Unit II) to SW (Unit III) may be attributed either to drainage reversals during a

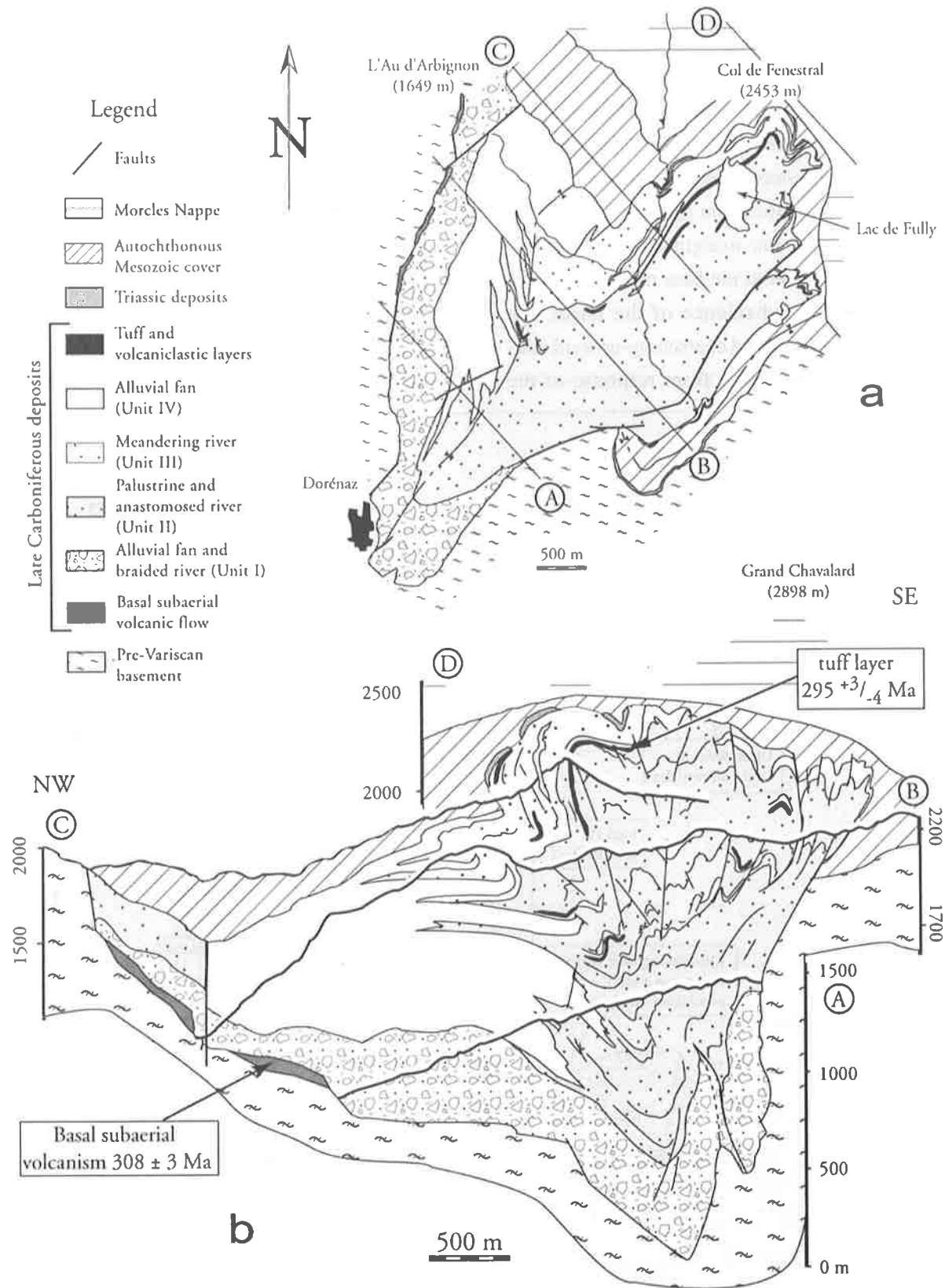


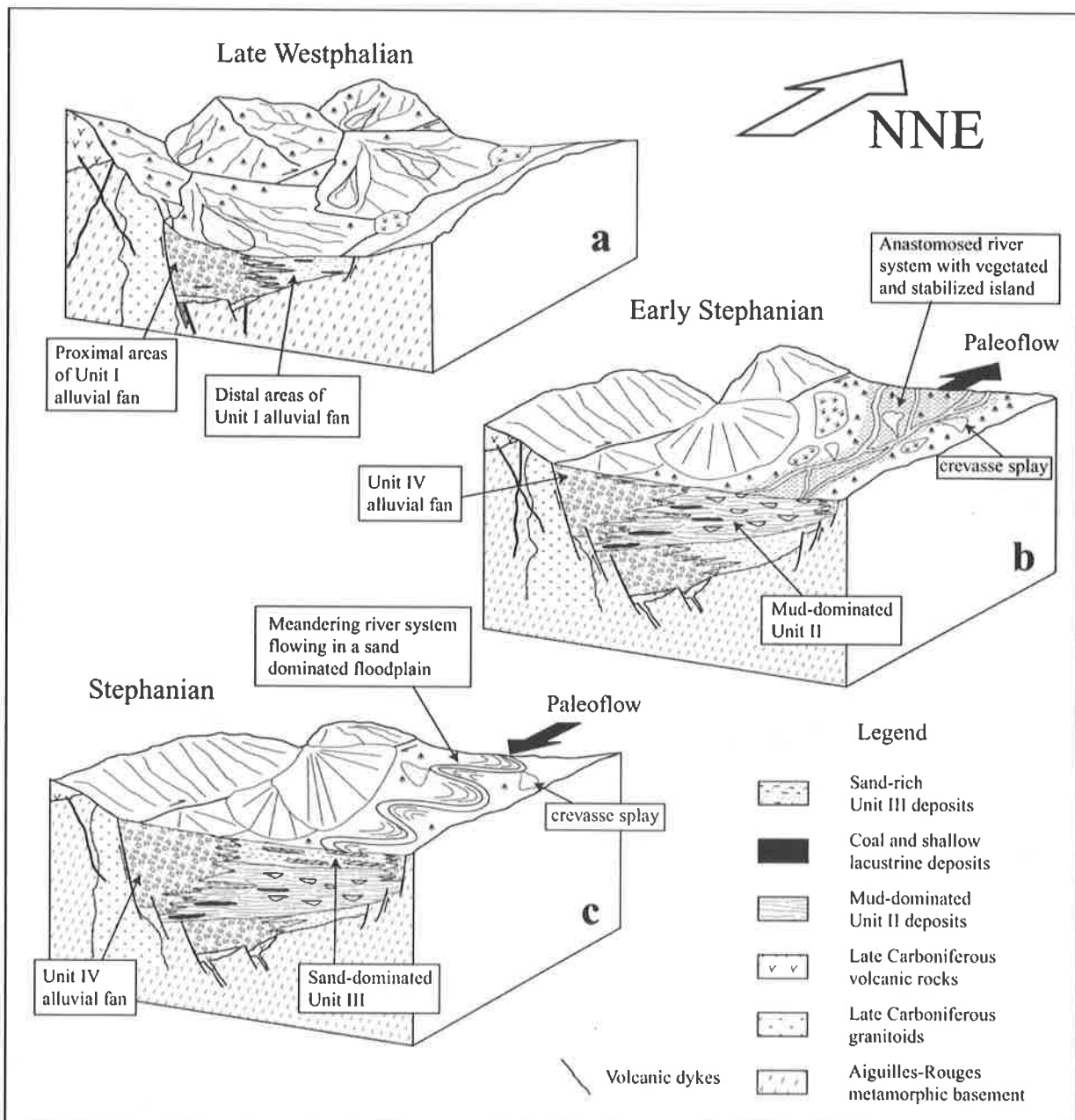
Fig. 25 (Fig 4.3 in Capuzzo 2000) : a) Geological map of the northern areas of the Salvan-Dorénaz syncline. The four lithological units that fill the basin are schematically reported, as the location of the volcanic and volcanigenic layers; b) Multiple cross-sections of the northern areas of the Salvan-Dorénaz syncline. Section lines are reported in the geological map and indicated by capital letters (modified after Pilloud, 1991).

stress field change or to uplift and subsidence of a wide area inducing backward incision of valleys and the capturing of the catchment areas of rivers draining to the opposite direction.

Unit IV: From the western margin of the basin an alluvial fan system repeatedly prograded into and retreated from the basin floor as documented by the migrating fan margin and coarsening-upward cycles. Mud at the base of a cycle documents enhanced subsidence of the basin, the overlying prograding and coarsening-upward alluvial fan sediments result from response of the

catchment area to relief generation. Additionally, the alluvial fan sediments at the western side of the basin document a gradual right-lateral displacement of the fan depositional area relative to the elevated sediment source areas.

Synsedimentary volcanism: The strike-slip movements favoured synsedimentary magmatism. Basal rhyodacitic flows and autobrecciated products, localized along the north-western margin of the basin, were deposited at 308 ± 3 Ma during its initial stage of development. This lower volcanism is probably associated with the



syntectonic intrusion of the nearby Vallorcine granite dated at 307 ± 1.5 Ma. On the other hand, ash-fall and volcanoclastic layers found within sediments of Unit II and III testify for high-explosive volcanic eruptions from distant volcanic centers at $295 \pm 4/-3$ Ma (Capuzzo and Bussy, 2000). Their zircon typology presents a bimodal distribution, which suggests derivation from alkaline magma series contaminated by crustal material. Coeval, highly explosive volcanism is known from the Aar massif in the Central Alpine basement (Schaltegger and Corfu, 1995), and tuff layers associated with this magmatic event have already been described in a Permo-Carboniferous basin located in northern Switzerland (Schaltegger, 1997).

5 - MINERAL ASSEMBLAGES AND METAMORPHIC EVOLUTION

As mentioned above, the Aiguilles Rouges–Mont Blanc massifs consist of a complex assemblage of lithological units with contrasting metamorphic histories, including non-, mono- and polymetamorphic units. Such associations might represent former basement-cover relationships or the tectonic juxtaposition of slices, which experienced different metamorphic paths. Among the polymetamorphic units, most contain mineral assemblages of amphibolite, granulite or eclogite facies grade. Although successive parageneses are observed, a clear attribution to specific orogenic events is difficult. For convenience, Alpine mineral parageneses will be distinguished from late Variscan and earlier relict assemblages, sometimes overinterpreted in the past (von Raumer 1976,

1981), but re-evaluated by von Raumer et al. (1999).

5.1 ALPINE MINERAL ASSEMBLAGES

One of the main effects of the Tertiary Alpine orogenic phase in the Aiguilles Rouges – Mont Blanc area is the formation of large-scale basement folds (see general cross-section) with a locally well developed schistosity (*e.g.* in the Mont Blanc granite), as well as major fault zones, locally of mylonitic type. Alpine structures are often at angle with older ones (*e.g.* they often display a $N45^\circ$ orientation in the Aiguilles Rouges - Mont Blanc area against $N10^\circ$ for Variscan structures), but not always, as pre-existing (especially brittle) structures might be reworked, making interpretation ambiguous. As a consequence of the Alpine compression, the original distance between the Aiguilles Rouges and the Mont-Blanc massif was probably in the order of 20 km, instead of 1 km now. Metamorphic conditions reached only the lowermost greenschist facies in the Aiguilles-Rouges and a slightly higher grade in the nearby Mont-Blanc massif (400°C and 0.25 GPa for fluid inclusions in quartz, Poty et al., 1974) (von Raumer, 1971; Frey et al.; 1999).

In the Aiguilles-Rouges area, pumpellyite, prehnite and laumontite are found in weakly retrograded amphibolites; stilpnomelane is observed in the matrix of nearly undeformed Late Carboniferous rhyolites; orthogneisses yield chlorite-albite mineral assemblages, and quartz shows the first stages of undulation and low angle boundary crystallisation (polygonisation) (von Raumer 1974, 1984). Characteristic healed frac-

Fig. 26 - (from Capuzzo 2000, Fig. 2.10) Schematic block diagrams representing various depositional environments during the evolution of the Late Carboniferous Salvan-Doré basin. a) formation of the basin induced by asymmetric subsidence along western bounding faults, which also favoured the emplacement of basal dacitic flows and controlled the deposition of Unit I alluvial fans derived from western source areas. Sediment production by weathering exceeded fan transport capacity. b) Deposition of the mud-dominated Unit II and establishment of an anastomosed river system with axial drainage to the NE; this change was possibly induced by increased differential tectonic subsidence within the basin; c) Deposition of the sand-dominated Unit III by a meandering river system formed in a relatively steeper fluvial valley presenting a reversal of its axial drainage towards the SW. Schematically, in figures b and c are illustrated prograding and retreating cycles of alluvial fans (Unit IV), and their right lateral displacement through time.

ture patterns appear in specific lithologies and Alpine foliation is expressed as a faint neofoliation of white mica of lowest greenschist facies grade, accompanied by pressure solution of quartz grains in black shales of Late Carboniferous age. In the Mont Blanc granite, a penetrative foliation developed (leading to the so-called protogine of early authors). The mineral paragenesis [green biotite–chlorite–epidote–albite] indicates lower greenschist facies conditions (von Raumer 1963, 1971), stilpnomelane is omnipresent (von Raumer 1968) and neofoliation of chlorite, garnet and/or epidote is observed along joint surfaces.

5.2 VARISCAN MINERAL ASSEMBLAGES

The strongest metamorphic imprint of polymetamorphic units in the Aiguilles Rouges and Mont Blanc massifs reached high amphibolite facies grade and locally anatectic conditions. The few available isotopic ages all point to a late Variscan age of 327–317 Ma (Bussy et al., 2000); but as lower Paleozoic migmatites are documented in other External Massifs, it is suspected that some high grade rocks in the Aiguilles Rouges–Mont Blanc area are pre-late Variscan in age as well. So far, only high pressure rocks (*e.g.* eclogites) can be safely related to a pre-Late Carboniferous metamorphic phase.

The high-T/anatectic late Variscan event is recorded in most lithologies of the Aiguilles Rouges massif (**Stop 1G**), whereas the Alpine overprint destroyed all index minerals in the Mont Blanc area.

(a) Metapelites are characterized by sillimanite-bearing assemblages [Bt + Pl + Qtz + Sil ± Grt ± Crd ± Ms ± Kfs] and by late stage andalusite found in quartz ± Kfs tension gashes (Von Raumer, 1984), pointing to a clockwise decompression path typical of a Barrowian metamorphic evolution (fig. 27). Monazites from such micaschists yielded a U-Pb isotopic age of 327 ± 2 Ma at Emosson (fig. 23 and Tab. 1), interpreted as the peak temperature age (Bussy et

al. 2000).

(b) Metagraywackes and metagranites experienced partial melting of variable intensity, ascribed to isothermal decompression during exhumation processes. Typical mineral assemblages of these migmatites are [Pl + Qtz + Kfs + Bt ± Sil ± Ms ± Grt ± Crd]. Monazite from a leucosome vein of a migmatitic graywacke at Emosson yielded a crystallization age of 320 ± 1 Ma, whereas monazite from a migmatitic granite in the Mont Blanc massif yielded 317 ± 2 Ma.

(c) Paragneisses of various compositions display monotonous assemblages [Qtz + Bt + Pl(olig) ± Ms ± Grt ± Sil ± Kfs] (Schulz & Von Raumer, 1993, Dobmeier, 1998).

(d) Metabasites of basaltic or gabbroic origin found as boudins of variable size record amphibolite facies conditions with mineral assemblages [Pl + Am + Ilm ± Grt ± Ca-cpx ± Zo]. Relicts of earlier high-P events are often preserved.

5.3 PRE-LATE VARISCAN MINERAL ASSEMBLAGES

As stated above, high-pressure relicts are the only mineral assemblages indisputably older than the Late Carboniferous high-T event, although none of them has been dated so far. Such relicts are found in metapelites and paragneisses as remnants of staurolite and kyanite, the latter being sometimes rimmed with reaction coronas of cordierite and spinel. High-P assemblages are better preserved in the retrograded eclogites of Lac Cornu (Liégeois & Duchesne, 1981) and Val Bérard (Schulz & Von Raumer, 1993), where [Grt + symplCpx + Am + Pl] parageneses are found. Relict omphacite Jd₂₆ within garnet or associated with plagioclase yield T-P estimates of 700°C/0.14 Gpa and 700°C/0.08 Gpa, respectively. Leucocratic, equigranular rocks associated with some eclogitic boudins have been interpreted as early decompression melting products linked to the exhumation of high-pressure units (Von Raumer et al., 1996). A preliminary dating of zircons from one of these leucosomes

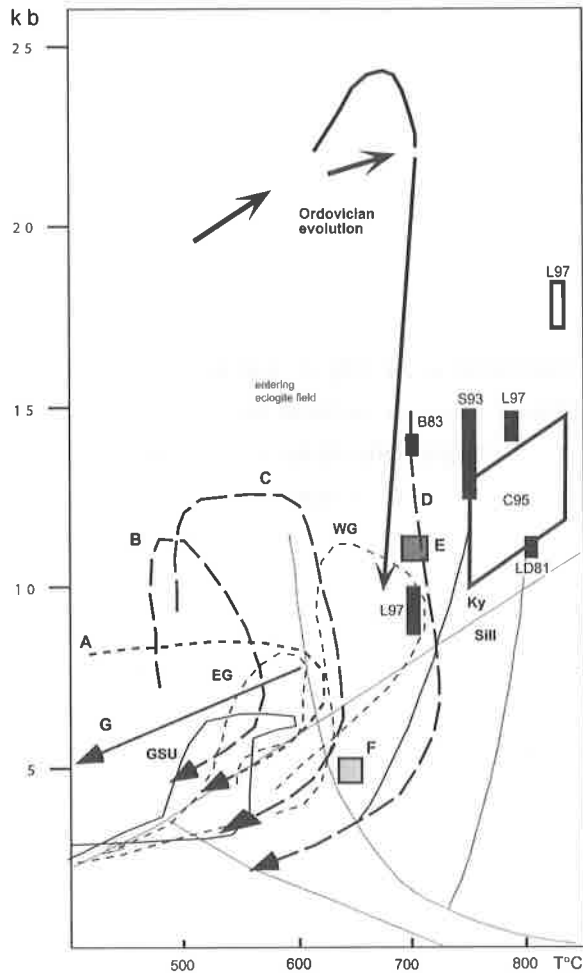


Fig. 27 - Pressure (P) - Temperature (T) evolution discussed for the External Massifs (from von Raumer et al. 1999). Reaction curves: Al-silicate stability fields after Holdaway (1971), granitic solidus and corresponding dehydration melting curves after Thompson (1990).

Pre-Late Ordovician evolution in the Gotthard area (Biino 1995), P - T estimations for retro-eclogites (black boxes) and granulites (white boxes) found in the External Massifs: Argentera - B83 Latouche & Bogdanoff (1987); C96 Colombo (1995); L97 Lombardo et al (1997a,b); Aiguilles Rouges - LD81 Liégois & Duchesne (1981); S93 Schulz & von Raumer (1993).

Variscan evolution in the central Aiguilles Rouges area (from Schulz & von Raumer 1993): A, B, C - garnet bearing gneisses; D - retrograded eclogites; E - kyanite-bearing tonalitic melts from the Lake Cornu area (von Raumer et al. 1996); F - Formation of W-Au skarns, 317 Ma, Salanfe area, Chiaradia 1993; Southwestern part of the Aiguilles Rouges (St. Gervais - Les Houches area) from Dobmeier (1996, 1998): Eg - Eastern gneiss unit; WG - Western gneiss unit; GSU - Greenstone unit; North-Eastern Mont Blanc area: G - Late Variscan P - T path after Marshall et al. (1997).

yielded an age of c. 340 Ma (Bussy & Schaltegger, work in progress).

Assuming that the high-pressure episode preceded the high-temperature phase without major cooling in-between, P - T paths have been established from mineral assemblages of various lithologies (Fig.27). They point to contrasting metamorphic histories, with both counterclockwise (south of the Aiguilles Rouges, Dobmeier, 1998) and clockwise (central part of the Aiguilles Rouges, Schulz & Von Raumer, 1993) paths. The position of the early leucosomes associated with the Lake Cornu eclogites is reported on the isothermal decompression path of the latter. Marshall et al. (1997) proposed a Variscan P - T - t path for the northeastern part of the Mont Blanc massif, which is comparable to those of the Aiguilles Rouges.

6 - MAGMATIC EVOLUTION

Two main groups of magmatic rocks are isotopically dated in the Aiguilles Rouges-Mont Blanc massifs, an Ordovician group at 450-460 Ma and a Carboniferous group at 330-300 Ma. The existence of older intrusive or sub-aerial rocks is suspected on the basis of indirect stratigraphic and/or lithologic arguments, but has not yet been demonstrated.

6.1 ORDOVICIAN MAGMATIC ACTIVITY

(Retro-)eclogitized pyroxenites, lherzolites and N-MORB-type basalts are outcropping in the Lake Cornu area (Aiguilles Rouges). Paquette et al. (1989) interpreted an upper intercept U-Pb zircon age of $453 \pm 2/-3$ Ma as a magmatic crystallization age. These rocks were derived from a depleted mantle source (initial $\epsilon_{Nd} = +6$).

They have been originally thought to be emplaced in a thinned continental crust associated with the initial stages of oceanic rifting (Paquette et al., 1989) and more recently to be emplaced in a back-arc setting associated with an active continental margin (Dobmeier et al., 1999). The latter

interpretation is supported by the contemporaneous intrusion of several granites of calc-alkaline (I-type) or peraluminous (S-type) character, as expected in an active continental (or island-arc) margin. Dated intrusions include the I-type Luisin granodiorite (457 ± 2 Ma, Bussy, unpublished data) in the Aiguilles Rouges and the Lognan S-type orthogneiss (453 ± 3 Ma, Bussy & Von Raumer, 1993) in the Mont Blanc massif. Other I- and S-type orthogneisses in the Aiguilles Rouges (Val Bérard, Wirsing, 1997) are presumably of similar age. **(stop 1G6).**

Among the still undated, and possibly older magmatic rocks, are strings of amphibolite (+ retroeclogites) boudins interlayered with metasediments, which can be followed over distances of several kilometers in the Aiguilles Rouges massif, and locally in the Mont-Blanc as well. They have been interpreted as sills or dykes of continental tholeiites (Von Raumer et al., 1990) intruded in pre-Ordovician sedimentary units, either linked to the Lac Cornu metabasites or to the Late Proterozoic / Cambrian rifting episode mentioned in the geodynamic reconstructions. Another metavolcanic sequence defined as the "Greenstone Unit" (Dobmeier et al., 1999) in the southwestern part of the Aiguilles Rouges Massif consists of an association of basalt to andesite and basalt to rhyolite lava flows of continental tholeiitic and calc-alkaline affinity, respectively. These rocks have been interpreted as additional remnants of the above-mentioned Ordovician active continental margin (Dobmeier et al. 1999).

6.2 CARBONIFEROUS MAGMATIC ACTIVITY

The Carboniferous magmatic rocks consist of syntectonic intrusive granites, locally associated with subvolcanic facies, acidic sub-aerial lava flows and volcanic/volcaniclastic ash-fall deposits. High-precision U-Pb dating evidences three short-lived pulses at 330 Ma, 307 Ma and 303 Ma (Bussy & Von Raumer, 1993; Bussy et al., 2000).

6.2.1 The 330 Ma magmatic pulse

The 330-340 Ma event is well known throughout the European Variscan belt and is characterized by high-K calc-alkaline to shoshonitic plutons. In the southwestern part of the Aiguilles Rouges massif, it is documented by the Pormenaz monzonite (332 ± 2 Ma) and the Montées-Pélessier granite (331 ± 2 Ma). **The Pormenaz monzonite** is a 1.4 by 2.5 km porphyritic funnel-shaped mass in vertical cross section, which intruded amphibolite-facies metamorphic rocks and lower Carboniferous metagraywackes (Délitroz & Fellay 1997). The shape and internal structures of the intrusion suggest a syntectonic emplacement of the magma and post-crystallization mylonitic deformations along a long-lasting dextral transpressive fault. The main facies is a porphyritic to equigranular monzonite with large pink or white K-feldspar megacrysts (up to 4 cm) in a dark gray-green amphibole-rich matrix. Euhedral crystals of brown sphene are clearly visible on hand specimen. Plurimetric bodies of durbachites (Holub 1977; Rock 1991) are found as dark green, equigranular, magmatic enclaves. The Pormenaz monzonite is characterized by high to very high concentrations in LILE like K, Rb, Ba, Sr, in Mg and transition elements like Cr, Ni, and V, and in incompatible elements like LREE, Zr and Th. Conversely, Ca is relatively low in this range of SiO₂ content. REE are strongly fractionated without substantial Eu anomaly. The durbachitic enclaves have the same chemical characteristics as the monzonites, reminiscent of lamprophyres.

The Montées-Pélessier granite is a vertical 3 km-long by 500 m-wide sheet-like pluton in tectonic contact with its country-rocks. A transpressive regime induced subhorizontal movements with extreme elongation, subsequently reoriented into subvertical displacements (Dobmeier 1998). The granite intruded syntectonically along ductile shear-zones during the transcurrent stage and recorded the subsequent

vertical movements during cooling (Dobmeier 1996). Primary magmatic flow fabrics are mostly superimposed by low-T ductile to brittle deformation. The Montées-Pélissier granite is a fine-grained foliated two-mica monzogranite, which hosts rare biotite-rich restites, mafic microgranular enclaves and biotite-bearing lamprophyres. It is peraluminous with A/CNK values between 1.17 and 1.27; but primary muscovite is scarce and the mean representative point of its zircon morphological population plots at the limit between peraluminous and calc-alkaline granite fields in the typological grid of Pupin (1988).

6.2.2 The 307 magmatic pulse

Three peraluminous granites intruded simultaneously in the Aiguilles Rouges-Mont Blanc area. Located in the northern part of the Aiguilles-Rouges massif (**stop 1A and 1G1**), the **Vallorcine granite** is a 15-km-long by 1-km thick sheet-like pluton (see fig. 23), which intruded at 306.5 ± 1.5 Ma along a steeply dipping, NE-SW trending, dextral strike-slip shear zone (Brändlein 1991; Brändlein et al. 1994). Syntectonic intrusion is inferred on the basis of structural analysis, which documents long-lasting strike-slip shearing both in the country rocks (pre- and post-intrusion fabrics) and in the granite. The latter developed an early NE-SW trending magmatic flow structure, evolving locally into a post-solidus foliation. The Vallorcine granite was subsequently affected along its SE contact by an intense low-T ductile shearing leading to the well-known “Miéville ultramylonite” (Kerrick et al. 1980) of Late Variscan age (but reworked during the Alpine orogeny). Alpine relief provides a 2000 m-high natural cross-section through the pluton. The lowermost facies is a biotite-rich monzogranite hosting numerous enclaves up to 30 cm in size; including gneiss xenoliths from the country rocks, hornfelses, micaceous restites with sillimanite and hercynite, early cordierite-bearing leucogranite blocks and mafic

microgranular enclaves. The upper facies is finer-grained with less biotite and almost no enclaves, pointing to an enclave unmixing process during upward motion of the magma. Aplitic and subvolcanic dykes are concentrated within the wall rocks of the upper contact. The Vallorcine intrusion is a typical S-type granite, as confirmed by zircon typology, stable isotopes (Brändlein et al. 1994) and the presence of Al-rich minerals (cordierite, muscovite, andalusite, etc.).

The **Fully granodiorite (stop 1B)** is located in the vineyards of the northern end of the Aiguilles-Rouges massif (Krummenacher 1959). It is a highly heterogeneous, coarse-grained granodiorite of migmatitic aspect with cordierite clots (up to 10 vol.% now greenish pinite), dispersed K-feldspar megacrysts and numerous small biotite-rich restitic enclaves. Schlieren and nebulitic structures are common. Micaschists, gneisses, marbles and amphibolites are found as dm-long xenoliths, whereas mafic magmatic enclaves can reach one meter in diameter. Cordierite-bearing leucogranitic dykes and stocks crosscut the main facies. No systematic internal structure is observed. Despite the migmatitic structure, the whole mass has a clear intrusive character. It is a typical peraluminous, anatectic granitoid with a large restitic component and abundant Al-rich minerals (garnet, cordierite, muscovite, hercynite). Zircons preferentially develop the {211} crystallographic form (low A index) and plot in the expected field of intrusive aluminous granites in the typological grid of Pupin (1988). The mafic magmatic enclaves include typical mafic microgranular enclaves of intermediate composition (quartz-dioritic to granodioritic), as well as angular to rounded pieces of fine- to coarse-grained gabbros, up to one meter in size (Bovay 1988). The latter preserve clear, locally pegmatitic, undeformed igneous textures, but experienced hydrothermal alteration. Pyroxene recrystallized into Mg-hornblende + minor Mg-biotite and plagioclase (An₆₀) is strongly sericitized.

The angular shape and absence of chilled margins indicate that these gabbro enclaves are pieces of a larger mafic body, that were incorporated as solid-state fragments into the mobile migmatitic mass. They have chemical characteristics of calc-alkaline type similar to those found in other Variscan acid-basic associations (e.g. Corsica, Pyrenees). Interactions with crustal material, possibly during hydrothermal retrogression, are evidenced by high contents in LILE like K, Ba and Rb. Zircon and monazite, extracted from granodioritic, leucogranitic and gabbro samples, yielded ages of 307 ± 2 Ma for all rock types. Although solid at time of incorporation into the anatectic mass, the gabbro enclaves are thus contemporaneous with the acid magmatism within errors. In other words, a mantellic basic magmatism was active at the time of enhanced crustal anatexis.

In the Mont Blanc massif, the peraluminous **Montenvers granite** was emplaced syn-tectonically at 307 ± 3 Ma as a sheet-like intrusion, in a similar way and at the same time as the Vallorcine granite. It is now a strongly deformed, often mylonitic leucocratic orthogneiss hosting both microgranular and restitic enclaves (Morard, 1998). Most primary minerals recrystallized in the greenschist facies, but zircon typology and whole-rock chemistry still point to an S-type granite.

Sub-aerial dacitic flows outcropping at the base of the Salvan-Dorénaz sedimentary basin (fig.25, p53) erupted at 308 ± 3 Ma (Capuzzo & Bussy, 2000); they represent the surface equivalent of the nearby Vallorcine granite and associated rhyolitic dykes. They are characterized by variable proportions of coherent and autoclastic scoriaceous volcanic facies interlayered toward the top with upper Carboniferous sediments. They present porphyritic textures with large euhedral and subhedral quartz and plagioclase phenocrysts. These lava flows are presumably related to the emplacement of a rhyodacitic lava dome along the north-western margin of the half-graben

sedimentary basin. In the Mont-Blanc massif, calc-alkaline rhyolitic dykes were emplaced simultaneously (307 ± 2 Ma) at shallow crustal levels, but they derive from different, deeper magma sources.

6.2.3 The 303 Ma magmatic pulse

It is represented by the voluminous 303 ± 2 Ma Mont-Blanc granite, located in the Mont-Blanc massif. It is a foliated, porphyritic monzo- to syenogranite with K-feldspar megacrysts and Fe-rich biotite as the only mafic mineral (Marro 1988; Bussy 1990). It hosts numerous mafic microgranular enclaves, calc-alkaline micromonzodioritic stocks and synplutonic dykes of mantellic origin, which record magma mingling processes (Bussy 1990). The Mont-Blanc granite is a metaluminous, ferro-potassic, alkali-calcic intrusion characterized by high K, Y, Zr contents and Fe/Mg ratios, and a low $^{87}\text{Sr}/^{86}\text{Sr}$ initial isotopic ratio of 0.705 (Bussy et al. 1989). Zircon morphology shows an extreme typology with very high A and T indices, typical of alkaline granites. The Mont Blanc granite is the last magmatic event recorded in the area, apart from ash-fall deposits embedded at different levels of the Salvan Dorénaz basin, which testify a $295 \pm 3/4$ Ma old episode of high-explosive volcanism from distant volcanic centers, possibly located in the Aar-Gotthard massifs (Central Alps).

6.3 TYPOLOGY, SOURCES AND SIGNIFICANCE OF THE CARBONIFEROUS MAGMATISM

6.3.1 The high-K magmatism

The 332 Ma Pormenaz monzonite has all characteristics of Mg-K-rich magmatic suites as defined for example by Rossi and Cocherie (1995). Typical features are the ubiquity and abundance of sphene, K-feldspar megacrysts, the presence of high-Cr and Ni basic enclaves of lamprophyric affinity, high to very high contents in K, Mg, Rb,

Sr, Ba, Cr, Ni, LREE, Th, Zr in all petrographic facies. Mg-K magmatic suites are well represented in the European Variscan belt, including Corsica, the External Crystalline Massifs, the French Massif Central, the Vosges and Bohemian massifs (see review of Debon et al. 1998). All available ages point to almost simultaneous intrusion between 330 and 343 Ma, in keeping with the 332 Ma age of the Pormenaz monzonite. This episode seems to be always short-lived, commonly syntectonic within major shear-zones and early in the local magmatic activity, either accompanied by anatectic peraluminous melts or not. The Pormenaz monzonite intruded at shallow depth along the transpressive border fault of a small Viséan volcano-sedimentary basin (fig. 23), a situation identical to that observed in the Vosges (Schaltegger et al. 1996) and the Aar massif (Schaltegger & Corfu 1995). The development of these basins attest for local extensional zones (first extensional event of Burg et al. 1994) within an overall transpressive regime, favoring the high-level emplacement of magmas. The lamprophyre-type mafic magmatism systematically associated to the Mg-K granitoids, Sr, Nd and Hf isotopic data (e.g. Schaltegger & Corfu 1992; Cocherie & Rossi 1995; Janousek et al. 1995; Schaltegger et al. 1996; Gerdes et al. 1998a) and episodic zircon inheritance all point to an essentially high-K lithospheric mantle source, metasomatized during an earlier subduction event, mixed with variable amounts of lower crustal material.

6.3.2 The peraluminous magmatism

Peraluminous melts were generated throughout the Late Carboniferous magmatic activity of the Aiguilles-Rouges / Mont Blanc massifs, either as *in situ* or short-range mobilized leucosomes, or as larger, syntectonic intrusions along major dextral strike-slip faults. All these melts have an indisputable crustal origin, documented by their Al-rich mineralogy, high restite content, zircon typology

and oxygen isotopes. Cordierite-bearing granites such as the Vallorcine and Fully intrusions, are high temperature melts formed through dehydration melting of biotite. Partial melting would be triggered by the addition of heat supplied by mantle-derived magmas, as documented by the hosted mafic magmatic enclaves (e.g. Fully gabbros). The latter evolve from shoshonitic to normal calc-alkaline type with time, *i.e.* from the 332 Ma Pormenaz durbachites and Montées-Pélissiers lamprophyric dykes to the 307 Ma Fully gabbros.

6.3.3 The ferro-potassic magmatism

The Mont-Blanc granite belongs to the so-called ferro-potassic or alkali-calcic or low-mg-number suites (Debon & Lemmet 1999). These suites are characterized by acid-basic bimodal magmas, high-K contents (alkali-calcic suites), Fe/Mg ratios and “A” zircon indices. Using the Fe/Mg ratio as a discriminating factor, Debon and Lemmet (1999) showed that the Variscan Fe-K granites (e.g. Aar, Gotthard) emplaced within a well-defined period of time between 295 and 305 Ma, in the same way as Mg-K granites did between 330 and 340 Ma. These authors ascribe the change in the Fe/Mg ratio between the two suites to a combination of several interacting factors, including the evolving nature of the magma sources, the physical and chemical conditions of melting (P, T, PH_2O , fO_2), and the geotectonic settings. According to Bonin et al. (1998), Fe-K suites document the onset of the post-orogenic stage of a Wilson cycle. They would ultimately originate in an asthenospheric mantle source, replacing the older orogenic lithospheric mantle. Granitic melts would result from crustal assimilation and contamination of mantle-derived magmas, as well as [biotite + plagioclase] fractionation under relatively high water pressures.

6.3.4 Tectono-magmatic evolution

The Carboniferous tectono-magmatic evolution

of the Aiguilles-Rouges/Mont-Blanc area can be synthesized in the following way (Bussy et al. 2000):

(a) Peak T metamorphic conditions are recorded at 327 Ma (peak P is undated, but is probably older) in the Emosson metapelites and isothermal decompression melting at 320 Ma. This relatively short time span requires fast uplift rates, accommodated through active tectonic exhumation. According to Thompson and Connolly (1995), decompression melting within the sillimanite-andalusite metamorphic facies cannot occur through simple erosional uplift. Exhumation was thus active already in the Late Viséan, shortly after the transpressive episode recorded by the Montées-Pélissier granite. This is confirmed by $^{39}\text{Ar}/^{40}\text{Ar}$ cooling ages as old as 331-337 Ma measured on white micas from gneisses adjacent to the Montées-Pélissier granite (Dobmeier 1998). The former nappe stacking which led to crustal thickening is documented by microstructures (Dobmeier 1996), but its timing is unknown. In the adjacent Belledonne massif, nappe stacking did not occur before the Viséan (Guillot & Ménot 1999) and decompression melting followed quickly during the Westphalian, a timing very similar to that of the Aiguilles-Rouges massif.

(b) The 332 Ma Pormenaz monzonite intruded along a high-strain, NNE-SSW trending mylonitic transcurrent fault, which also controlled the position of the adjacent Viséan detrital basin. Considering the ultimate mantellic origin of these melts, and their usual syntectonic character in the Variscides, it is inferred that this transcurrent fault zone was a major crustal- to lithospheric-scale structure, which tapped deep-seated magma chambers or possibly enhanced melting processes. The nearby and contemporaneous Montées-Pélissier granite intruded in a similar context. Fluids circulating along the transcurrent fault might have favoured melting of crustal lithologies, triggered by heat transfer from the mantellic magmas (presence

of lamprophyric dykes).

(c) The next recorded event is the simultaneous formation of the Late Carboniferous detrital basin of Salvan-Dorénaz, starting at 308 Ma (Capuzzo & Bussy 1999), and the 307 Ma magmatic pulse of cordierite-bearing granites. The Vallorcine granite intrudes syntectonically along the border fault of the basin, which is again considered as a major crustal-scale structure. Erosion is very active and thick coarse deposits are accumulating in the basin at least until 295 Ma (Capuzzo & Bussy 1999). This is interpreted as an evidence for continued tectonic exhumation in the latest Carboniferous in an essentially transcurrent to transtensional rather than extensional regime. In that context, the sedimentary trough could have formed as a kind of pull-apart structure or half-graben in case of oblique sliding (intermediate between normal and transcurrent faulting) along the border fault. Granodioritic cordierite-bearing melts require temperatures well above 800°C to form (*e.g.* Patiño-Douce & Harris 1999). Such conditions were not reached through regional metamorphism in the lower-middle crust and required additional heat, most probably supplied by the associated mafic mantellic magmatism. The lower crustal magma sources were tapped by the deep transcurrent faults along which progressive restite unmixing and crystal fractionation occurred.

(d) In the nearby Mont-Blanc massif, the larger post-tectonic 303 Ma Mont-Blanc acid-basic association intrudes within a pull-apart structure in a continued transcurrent to transtensive regime. The inferred mantellic source is different from that of the Pormenaz monzonite.

The Carboniferous tectono-magmatic evolution outlined above is virtually identical to that observed in the other External Crystalline Massifs and adjacent areas, with minor differences in the timing of events (*e.g.* Schaltegger 1997; Debon & Lemmet 1999). More generally, it is in good accordance with the overall Carboniferous

evolution of the internal Variscides (*e.g.* Burg et al. 1994; Rey et al. 1997), considered as a period of post-collisional readjustment of a thickened continental crust.

7 - CONCLUSIONS

The complex metamorphic pattern observed in the External Alpine Massifs resulted from the succession of several orogenic events combined with an early evolution in different continental blocks. Among them, the last two, *i.e.* the Alpine and the Variscan cycles, are easily recognized as they left contrasting imprints in the lithologies. Earlier events are more difficult to unravel, although an Ordovician cycle is definitely documented in all polymetamorphic basement areas (*e.g.* by the “pre-Variscan” granitoids), probably hiding relicts of an even older evolution. According to tentative reconstructions, the External Massifs were part of a post-Ordovician Gondwana-derived microplate, which evolved as an active margin (accretionary wedge, volcanic arc) during the Cambro-Ordovician period at the border of Gondwana.

The time period between the Late Ordovician intrusion of granitoids and the Variscan collision is badly constrained. The microcontinent containing the future External Massifs, like the other peri-Gondwanan microcontinents, should have followed a migration path comparable to that of the Eastern Alps before Variscan collision with Laurussia. Traces of sedimentary and magmatic evolution, during this period of rifting and drifting, were only identified in the Belledonne area. Other relicts may be hidden in the so-called monocyclic domains, if not lost during tectonic evolution or through erosion. The subsequent period is documented mostly by Variscan age information. The zonal distribution of different mineral parageneses from Ky + Grt + St to Grt + Sil, eclogite retrogression, large-scale folding and decompression melting at *c.* 320 Ma are ascribed to Variscan nappe tectonics and

deformation of a cordillera followed by exhumation/erosion in an essentially transcurrent tectonic regime. Short-lived and bimodal magmatic pulses reflect successive stages of post-collisional restoration to normal size of a thickened continental lithosphere, in a geodynamic setting of the Variscan cordillera collapse.

Alpine orogenic events in the External Massifs induced brittle to ductile to mylonitic deformation in anchizonal to greenschist facies metamorphic conditions, as well as a late doming, which brought the massifs to their present-day high topographic position.

II: EXCURSION OUTCROPS

Stops are located on the topographic map on the opposite page, a structural map corresponding to the same area is found in annex (MAP 1).

STOP 1A - MIEVILLE QUARRY [568300/111000]: VALLORCINE GRANITE

Topic: Late Variscan magmatism, vertical magmatic differentiation in a syntectonic sheet-like intrusion and ultramylonitisation.

Outline: The Vallorcine granite is part of the 307 Ma magmatic pulse characterized by the syntectonic intrusion of S-type granites along major transcurrent faults. These high temperature melts result from large-scale dehydration melting of biotite in deep-seated crustal levels, probably triggered by a contemporaneous mantellic magmatic activity. Post-solidus ultramylonitisation of the Vallorcine granite testify for long-lasting transcurrent movements at the largest scale at the end of the Variscan orogeny (see details in §6.2.2).

The outcrop is located in an old quarry at the lowermost level of the vertical sheet-like intrusion of the 307 Ma old Vallorcine granite. The dominant facies is a biotite-rich monzogranite hosting numerous enclaves up to 30 cm in size; including gneiss xenoliths from the country rocks, hornfelses, micaceous restites with sillimanite and hercynite, early cordierite-bearing leucogranite blocks and mafic microgranular enclaves. The upper facies of the Vallorcine granite is visible 1600 m above at Emosson (**Stop 1G1**), it is finer-grained with less biotite and almost no enclaves, pointing to an enclave unmixing process during upward motion of the magma. The Vallorcine intrusion is a typical S-type granite, as confirmed by zircon typology, high $\delta^{18}\text{O}$ values 9-10‰, Brändlein et al. 1994) and the presence of Al-rich minerals (cordierite, muscovite, andalusite, etc.).

Walking along the path in the small forest towards the eastern margin of the intrusion, the granite is increasingly and heterogeneously deformed with the occurrence of black stringlets of mylonite (fig. ft1-A1). About 10 meters from the contact with the upper Carboniferous host rocks, the original granitic texture is hardly recognizable, except a few K-feldspar porphyroclasts. The ultramylonite, looking like a silicified black shale at the contact, undistinguishable from a phyllite, has been carefully described by Meyer (1916); Reinhard and Preiswerk (1927) discussed its tectonic significance during the late Variscan period and its reactivation during the Alpine events. Mylonitisation resulted in a strong grain size reduction ($<5 \mu\text{m}$) without substantial mineralogical and chemical changes (see fig. ft1-A1, Steck & Vocat, 1973). Temperature during translation (superplastic flow) is estimated at $250 \pm 30 \text{ }^\circ\text{C}$ (Kerrick et al., 1980), whereas the timing of event is not directly dated, but a Late Carboniferous age is inferred (with Alpine reworking) on the basis of Rb-Sr thin slab dating of neighboring mylonitic bands (Thöni, 1989).

STOP 1B - FULLY VINEYARDS [575460/110830]: MIGMATITIC GRANODIORITE OF FULLY

Topic: bimodal late Variscan magmatism, crustal-derived migmatitic granite and coeval gabbro enclaves (see in §6.2.2).

Outline: The Fully migmatitic intrusion is another example of the large-scale dehydration melting of crustal units in close association with mantellic magmas. It might be considered as a deeper and less evolved equivalent of the Vallorcine granite (see details in §6.2.2).

The Fully vineyard is growing on the granodioritic intrusion of Fully. The latter is best outcropping on both sides of a small road cutting along the road to the village of Eule. It is a highly

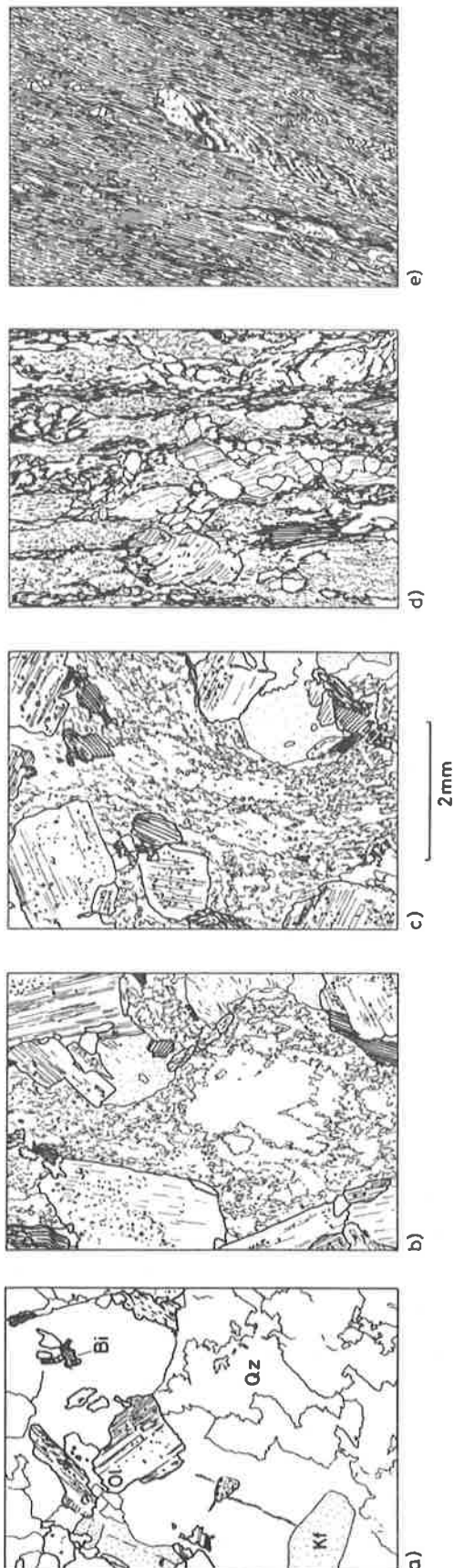


Fig. ft1-A - Progressive transform of the Vallorcine granite into an ultramylonite. Drawn by A. Steck in Masson et al. 1980.

heterogeneous, coarse-grained rock of migmatitic aspect with cordierite clots (up to 10 vol.%, now greenish pinite), dispersed K-feldspar megacrysts and numerous small biotite-rich restitic enclaves. Schlieren and nebulitic structures are common. Dm-long xenoliths are abundant as well as mafic magmatic enclaves, which can reach one meter in diameter. Cordierite-bearing leucogranitic dykes and stocks crosscut the main facies. No systematic internal structure is observed. Despite the migmatitic structures, the whole mass has a clear intrusive character. It is a typical peraluminous, crustal-derived granitoid with a large restitic component.

The mafic magmatic enclaves include typical microgranular enclaves of intermediate composition (quartz-dioritic to granodioritic), as well as pieces of fine- to coarse-grained gabbros (Bovay 1988). The latter preserve clear, locally pegmatitic, undeformed igneous textures, but experienced hydrothermal alteration. The angular shape and absence of chilled margins indicate that these gabbro enclaves are pieces of a larger mafic body, that were incorporated as solid-state fragments into the mobile migmatitic mass. They have chemical characteristics of calc-alkaline type similar to those found in other Variscan acid-basic associations (e.g. Corsica, Pyrenees). Zircon and monazite, extracted from granodioritic, leucogranitic and gabbro samples, yielded ages of 307 ± 2 Ma for all rock types. Although solid at time of incorporation into the anatectic mass, the gabbro enclaves are thus contemporaneous with the acid magmatism within errors.

STOP 1C - MARTIGNY BATAIAZ [571400/105950]: MESOZOIC METASEDIMENTS (CHAMONIX ZONE)

Topic: Multiple deformation in between basement bloks.

Outline: The outcrop area is situated in the Chamonix zone, containing the Mesozoic cover

(middle Jurassic to lower Cretaceous) of the adjacent basement areas. First mentioned by Oulinaoff (1924), the entire zone has been characterized by Ayrton (1980), giving details on stratigraphy and the extreme deformation. In the outcrop area, parautochthonous marly metasediments of Middle Jurassic age suffered high plastic deformation between approaching basement blocks. The main visible structures are intersection lineations l_1 between axial cleavage planes and D_1 -fold-axes with strong changes from vertical to horizontal dip, representing highly plastic creep in such environment. Additionally, vertical D_3 -folds are superposed on the complex structures.

STOP 1D - DORÉNAZ QUARRY [569630/110330]: SALVAN-DORÉNAZ LATE CARBONIFEROUS SEDIMENTARY BASIN

Topic: sedimentary structures in coarse-grained detrital deposits

Outline: The Salvan-DorénaZ basin is the best preserved of the transtensional grabens formed during the Late Carboniferous dismembering of the Variscan belt. It caught huge volumes of detritus from the neighboring eroding relief. Three stops, *i.e.* at DorénaZ, Tête Noire and Finhaut, provide typical lithologic sections in the alluvial fan system of Unit I (fig. 24), (see details in §4.2.1).

The old quarry at DorénaZ shows typical coarse-grained sediments in overturned position as a consequence of the Alpine deformation. The large size of the clasts points to a proximal position in the fan structure. Conglomerates are interlayered with coarse sandstones (fig. 26), and crossbedding as well as channel-structures are visible. A horizon with organic débris is visible in the southernmost section, where Burri (1969) discovered large coalified trunks.

STOP 1E, TÊTE NOIRE [564180/102300]: SALVAN-DORÉNAZ LATE CARBONIFEROUS SEDIMENTARY BASIN

Topic: original contact between the detrital sediments and the metamorphic basement

Outline: see previous stop.

The Tête Noire road cut is the nicest place to observe the original, but tilted (by Alpine compression) contact between upper Carboniferous sediments and the underlying basement. Pinpointing this contact requires careful observation, as basement rocks underwent *in situ* disintegration prior to deposition of the first sediments. The latter are black and fine-grained, but evolve westward to very coarse-grained conglomerate beds of Unit I, interlayered with finer-grained shales hosting plant debris.

Basement rocks are highly micaceous paragneisses, whose sedimentary origin is established by a finely banded layer of calcsilicate marbles, mainly composed of calcite and diopside, visible about 20 meters uproad (east) of the contact. Pegmatitic veins might be related to the 320 Ma anatexis event or to the 307 Ma peraluminous magmatism. The age of the metacarbonates is unknown, they might be of Cambrian age, as inferred for other marble lenses in the area, but a Devonian age is also possible.

STOP 1F - FINHAUT [565000/103860]: SALVAN-DORÉNAZ LATE CARBONIFEROUS SEDIMENTARY BASIN

Topic: mass flow deposits (boulders)

Outline: see stop 1D.

Along the road-cut leading to the village of Finhaut very coarse conglomerates of Unit I can be seen, where large boulders of migmatites represent the proximal situation at the shoulder of

the fan system. The dark coloured fine grained channels underline the quickly changing situation between the deposits in the channel and the very coarse-grained boulder beds.

**STOP 1G - LAKE EMOSSON [561250/101840]:
POLYMETAMORPHIC BASEMENT OF THE AIGUILLES
ROUGES MASSIF.**

Topic: Variscan metamorphism/anatexis and Alpine overprint of sedimentary and igneous lithologies.

Outline: The Lake Emosson area is one of the best sites to observe the polymetamorphic basement of the Alpine External Massifs. The oldest lithologies are upper Proterozoic to lower Palaeozoic sediments and volcanites, which were intruded by Ordovician granitoids (orthogneisses), before all rocks underwent Variscan metamorphism. The latter was of high amphibolite facies grade and locally induced partial melting (see field-trip introduction). Metamorphism was accompanied by long-lasting deformation with superposition of at least three fold generations, the last one being of kilometer scale. The latter is identified by ubiquitous Z-, S- and M-shaped parasitic folds, which allow distinction between adjacent anti- and synforms. Alpine metamorphism reached low greenschist facies grade. All units are unconformably overlain by sandstones of Triassic age, hosting the famous saurian footprints (Demathieu and Weidmann 1982) in the Vieux Emosson Lake area. The Alpine dome-like structure of the massif is underlined by the position of the Mesozoic sediments, which rest horizontally on top of the basement rocks in the middle of the massif (Aiguille de Belvédère, 2600 m), whereas they are steeply dipping on both margins of the latter in the Rhone valley (400 m above sea-level).

A walk around the lake from east to west will give the opportunity to recognize the main lithologies of the polymetamorphic basement. The lakeshore

is a continuous outcrop, but only a limited number of topics have been selected.

Vallorcine granite (stop 1G1)

The upper facies of the Vallorcine granite is outcropping right after the car park, along the small road to the dam. Compared to the lower facies of Miéville (**stop 1A**), it is finer-grained with less biotite and almost no enclaves, which is thought to result from an enclave unmixing process during upward motion of the magma. The intrusive contact of the granite with its gneissic host rocks is visible behind the small chapel facing the restaurant (coord. 561310/102000). It is characterized by a 1 m thick brecciated and silicified band, with fibrous quartz crystallized radially all around the clasts. This spectacular texture has been interpreted as a result of hydraulic brecciation during the shallow level intrusion of a fluid-saturated granitic magma (Genier, 2000).

Mylonite-Zone (stop 1G2)

At the eastern dam edge, a nearly 500 m large mylonite zone separates the Vallorcine granite from the polymetamorphic metasediments located further to the northwest. Despite the very strong deformation, former orthogneisses (Ordovician granitoids) and metasediments, like calc-silicate lenses (former calcsilicate marbles) can be recognized. The strike-slip tectonics producing the mylonites probably facilitated intrusion of the Vallorcine granite. The Rb-Sr-thin-slab method of dating (Thöni 1989) produced an age of 300±20 Ma for the general deformation (± 500°C), which corresponds to the age of the Vallorcine granite, 307 Ma (Bussy et al. 2000). Microstructural observations (Joye 1989) show that the entire zone is dominated by dextral shearing with formation of a subhorizontal stretching lineation (dip 20° NE) produced through a SSW/NNE tangential compression of pre-existing, more horizontal structures (S₂). Joye (1989) interpreted narrow, very fine-grained,

dark veinlets of glassy constitution as probable pseudotachylites.

Metasedimentary units (stop 1G3)

Complex and superimposed tectonic structures exclude any lithostratigraphic reconstruction in the Aiguilles Rouges massif. Only major sequences or units can be crudely identified at the map scale. They consist of:

- (I) a unit of graywackes with metapelitic interlayers;
- (II) a mixed unit composed mainly by metapelites with some thin metagraywacke layers, one quartzite horizon, one layer of carbonates (appearing mostly as large boudins), and hosting one or two amphibolite layers;
- (III) a third unit characterized by finely banded metagraywackes and metapelites with a rusty patina.

Many detailed observations are found in von Raumer (1983), von Raumer and Schwander (1985), Schulz & von Raumer (1993), Dupasquier (1996), Schmocker (1996), Fracheboud (1997), and Marquis (1997). A comparison with other European lithostratigraphic sections brings convincing evidence that these lithologies have a Late Proterozoic to Early Palaeozoic age (see field-trip introduction).

Metaquartzites (stop 1G4) form a recognizable white horizon, which can be followed in the field when mapping strongly boudinized pieces preserved in the highly plastic micaschists. They are rather coarse-grained quartzitic sandstones with tiny garnets and a faint layering underlined by biotite.

Metapelites and metagraywackes (stop 1G5) record a Barrowian-type of metamorphism with early [biotite-staurolite-kyanite-garnet] assemblages evolving towards [biotite-garnet-sillimanite] parageneses. Joye (1989) locally observed sillimanite and cordierite in strongly sheared rocks

among the mylonites. The thermal peak has been dated at 327 ± 2 Ma (U/Pb on monazite, Bussy et al. 2000). Late stages are quartz segregation lenses with K-feldspar and andalusite. Such a sequence of parageneses does not necessarily represent a continuous PT-path, but could reflect two distinct events, *i.e.* an early-Variscan high pressure phase and a late-Variscan, more temperature dominated phase. Geochemical data from different localities indicate that metagraywackes carry the fingerprints of an active margin setting (Bhatia 1983), resulting from the erosion of quartzitic lithologies or acidic volcanites (Roser and Korsch 1988). This agrees well with our general interpretation of shelf sediments located at the Gondwana active margin (see introduction).

Marbles are rare, strongly sheared and completely smeared out among the hosting rock series. Larger lenses are locally preserved as banded calcite-diopside rocks. This lens shape is either the result of boudinage during stretching of the hosting micaschists, or might represent former patch-reefs, a well-known lithology in the Cambrian. These calcsilicate lenses are often hosting scheelite ore deposits, which most probably resulted from metasomatic transfer during late Variscan granite intrusions (Chiaradia 1993).

Metabasites (stop 1G6) appear as boudin-shaped amphibolites, mainly concentrated in the micaschist-series, but also as eclogite bodies, as in the Lake Cornu area, situated a few kilometers to the southwest. Distribution of amphibolites in km-long strings of boudins point to the former existence of one or two relatively thin layers, best preserved in the fold-hinges. They mainly consist of amphibole – plagioclase \pm diopside \pm garnet, and pseudomorphs of former zoisite needles. Von Raumer et al. (1990) distinguished two main groups of amphibolites: plagioclase amphibolites (former spinel-olivine-tholeiites with relatively higher contents in Cr and Ni) and garnet-plagioclase amphibolites (former hypersthene- or quartz-

tholeiites with enriched values of V and P), which were considered to represent a magmatic differentiation series. The fine grain size of the amphibolites and high TiO₂ and V-contents (>0.5% and > 100 ppm, respectively) suggests a volcanic or subvolcanic origin (Pfeifer et al. 1989). The original rocks were interpreted (von Raumer et al. 1990) as transitional MORB volcanites typical for continental rift zones at the onset of ocean floor spreading. Their age could be either Cambrian or Ordovician. Some of the largest amphibolite bodies are accompanied by leucocratic garnet-bearing gneisses, containing large crystals of staurolite and kyanite, the latter with reaction rims of cordierite and hercynite, interpreted as a HT-decompression reaction (Dupasquier, 1996). The origin of these leucocratic rocks remains unexplained, but they could result from dehydration melting of amphibolites comparable to those observed in the Lake Cornu area (von Raumer et al. 1996).

Orthogneisses (stop 1G7) occupy large areas in the Emosson region, as large dike-like bodies, which record two phases of folding, as illustrated by the huge, steeply dipping fold structure adjacent to the former Barberine dam site (now flooded by the Emosson lake) (von Raumer 1984). This fold is clearly visible in the landscape and helped to understand the complexity of the regional structures during mapping of the area. Both I-type (hornblende-biotite-bearing granodiorites) and S-type (biotite-bearing porphyritic granites) intrusions were identified (Wirsing, 1997). The Luisin granodiorite has been dated at 457 ± 2 Ma (U-Pb on zircon, Bussy, unpubl. data), whereas an S-type augengneiss from the nearby Mont Blanc massif intruded at 453 ± 3 Ma (Bussy and Von Raumer, 1994). This magmatic event is ascribed to an active margin context, as also documented by the MORB-type mafic rocks of the Lake Cornu area (von Raumer et al. 2001). The late Variscan metamorphism is well documented in the orthogneisses, which

experienced partial melting. Late Variscan strike-slip deformation is also documented through C-S tectonites.

Migmatites (stop 1G3)

Partial melting affected several of the above-cited lithologies, in particular muscovite-bearing metagraywackes and the orthogneisses. Migmatitic structures are beautifully outcropping along the road west of the dam. Besides local structures due to Alpine metamorphism and deformation, the leucosomes clearly record syn- to post-anatexis dextral shearing. Monazites from one of the thickest leucosome lens yielded an age of 320 ± 1 Ma (Bussy et al. 2000). The leucosomes mainly consist of K-feldspar (50-60 vol%), quartz (30%) and plagioclase (5-30%) \pm muscovite \pm biotite (Genier, 2000). These are mostly local melts, which did not substantially migrate. Anatexis is ascribed to the dehydration melting of muscovite during the high-T isothermal decompression of the rocks. Adjacent metapelites host abundant folded quartz veinlets of metamorphic origin, but no leucosomes at all. Considering that pelites usually melt more readily and at lower temperatures than graywackes, a major tectonic contact is inferred between the migmatitic metagraywackes unit and the non-migmatitic metapelites, for which a lower peak metamorphic temperature is postulated.

Alpine deformation

Alpine metamorphism increases from northwest to southeast throughout the orogen; it is of lowest greenschist facies grade in the Aiguilles-Rouges area (estimated at c. 275°C, on the basis of mineral textures and stability). Most rocks at the dam site show traces of Alpine deformation, where every rock type carries its own characteristics (von Raumer 1974, 1984). Pumpellyite, prehnite and laumontite are found in weakly transformed amphibolites, and stilpnomelane is observed in the matrix of almost undeformed rhyolites of

Permian age. Granitoid orthogneisses yield chlorite-albite mineral assemblages when approaching higher levels to the overlying nappes, where also small drag-folds appear. Quartz shows the first stages of undulation and low angle boundary crystallisation (polygonisation). Depending on the lithology, conjugated shear systems with corresponding tension gashes developed at different scales. Orthogneisses developed a general, closely spaced fracture

cleavage with tiny chlorite-filled tension gashes, whereas micaschists show two sets of larger shearing planes with growth of fiber quartz crystals parallel to the stretching direction. In the slightly deformed overlying Triassic sandstones, tension gashes are up to 10 cm long. All three types of brittle shear probably represent one answer to the same Alpine deformation in the vicinity of the basal Alpine nappe thrust plane.

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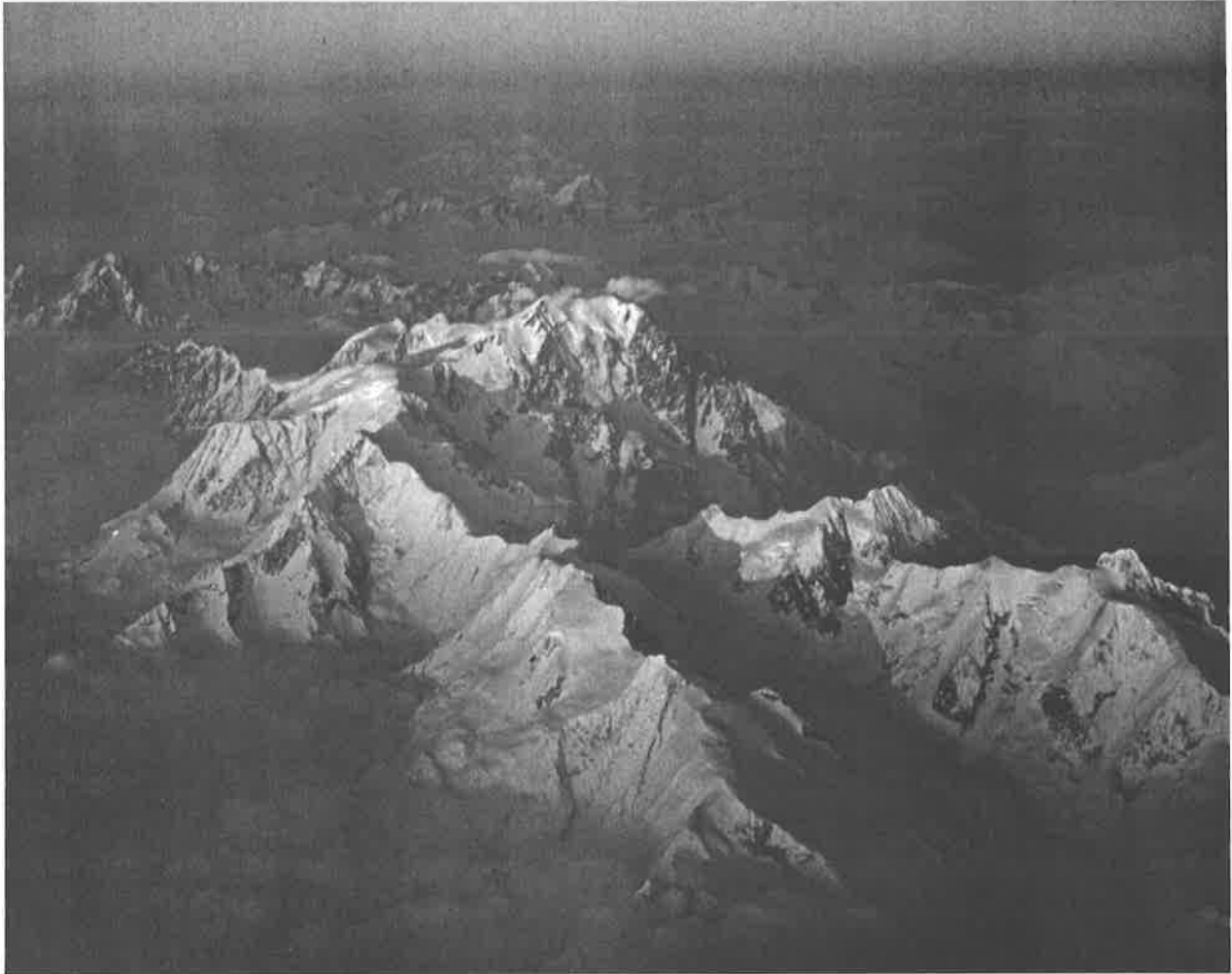
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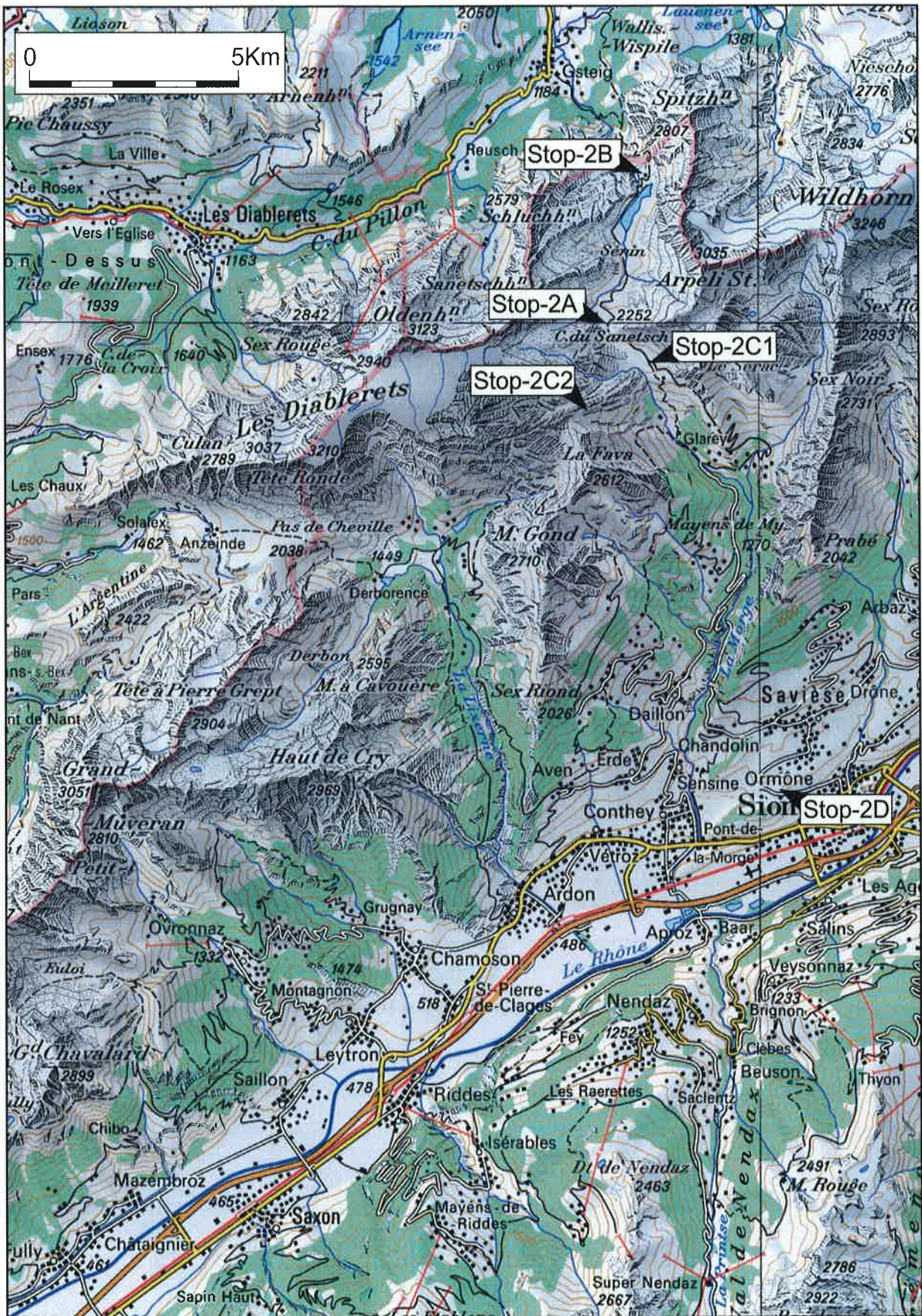
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The Mont-Blanc massif seen from the air, looking east



The Helvetic covers units

Jean-Luc Epard

I: INTRODUCTION

The aim of this field trips is to make a transect from the front to the root part of the Wildhorn “super” nappe, formed by 3 different tectonic units, the Diablerets, Mont-Gond and Sublage nappes. **Stop 2A** (Sanetsch pass) will be devoted to the general setting of the field trip. In **stop 2B**, the frontal part of the Mont-Gond and Sublage nappes will be examined. The stratigraphy and structure of the Urganian and Tertiary rocks of the Diablerets nappe will be illustrated during a few kilometers long walk on the Lapiaz (karstic landscape) (**stop 2C1**). The Sex Rouge hill (**stop 2C2**) offers a good view on the Sublage summit (folds in upper Jurassic) and on the Sex Noir mountains, where Late Cretaceous and Tertiary unconformities can be examined. The last stop (**stop 2D**), will illustrate the contact between the Helvetic nappes and the Penninic domain near the town of Sion. Stops are located on the topographic map in the opposite page, a corresponding structural map is found in annex (MAP 2).

II: EXCURSION OUTCROPS

STOP 2A - SANETSCH PASS AREA, GENERAL SITUATION.

(Geologic map of Switzerland at scale 1:25'000, sheet 1286 St-Léonard, Badoux et al., 1959a and b)

The Sanetsch Pass (588025/131200) is located in the Wildhorn nappe, very close to the limit between the Diablerets and Mont-Gond nappes (fig. ft2-A)

The vast area of lapiaz (typical morphology due to dissolution of limestones) at the west, belongs to the Diablerets nappe (Urganian and Tertiary limestones). The topographic surface is roughly

parallel to bedding therefore the slope of this karstic area gives a good estimate of the general axial plunge of the structures towards the NE (about 10°), below the pass area. The Diablerets nappe is linked to the Mont-Gond nappe by a syncline. The overturned limb of this syncline is highly deformed. It is tectonically thinned down to a few meters and outcrops sporadically in the meadow just southwest of the pass. To the east, this series is topped by a thick sequence of dark marls, Early Cretaceous (Berriasian) in age, belonging to the Mont Gond and Sublage nappes. To the northwest, the crest formed by brown colored rocks (Valanginian and Hauterivian limestones) and plunging towards the S enin lake belongs to the Mont-Gond nappe. The upper part of the north-south oriented ridge (Spitzhorn – Arpelistock) to the right (east) of the lake is made of lower Cretaceous rocks belonging to the Sublage nappe. More details on the Spitzhorn area are given in **stop 2B**. The Sanetsch pass is also located on the structural crest of the late fold refolding the entire Helvetic nappe stack. South of the pass, the general dip of the structures is towards the southeast (towards the Rh one Valley) whereas on the northern side of the pass it is towards the northwest.

STOP 2B - S ENIN LAKE, DAM AREA: FRONTAL PART OF THE MONT-GOND AND SUBLAGE NAPPES.

(Geologic map of Switzerland at scale 1:25'000, sheet 1266 Lenk, Badoux et al., 1962 and Badoux and Lombard, 1962).

The dam (589000/134250) is located north of the pass and is built on lower Cretaceous and Tertiary rocks of the Mont-Gond nappe. On a little hill, located 500m north of the dam area, the transgression over the former flexural bulge of Tertiary

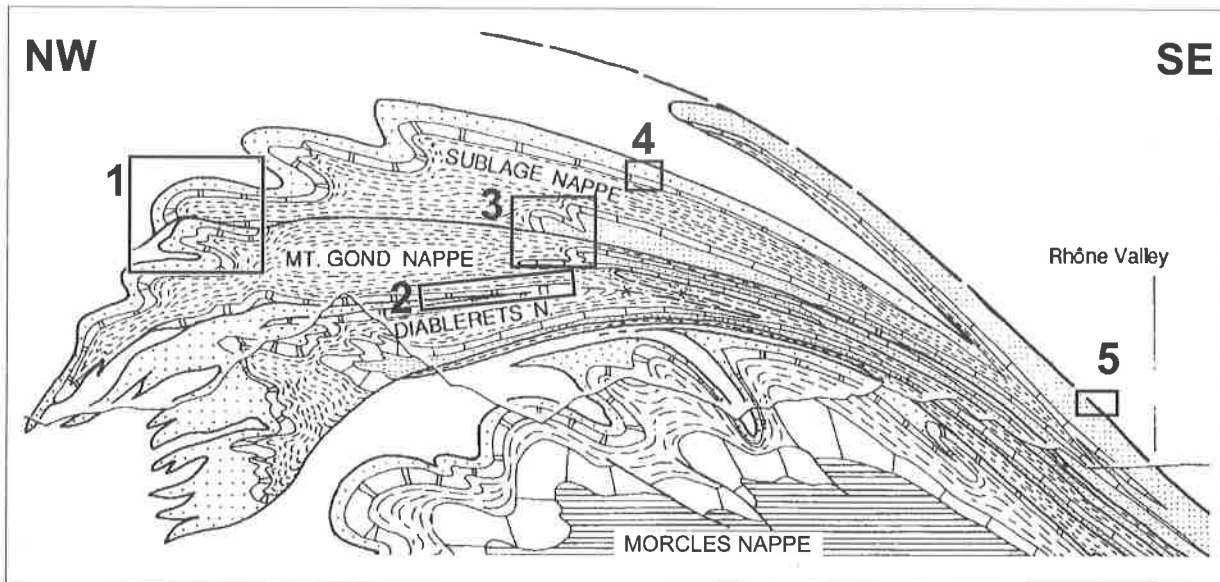


Fig. ft2-A - General cross-section of the Wildhorn nappe (modified from Escher et al. 1997). Boxes refers to the position of specific features described in stops B to D.

rocks on white Urgonian (Barremian-Aptian) limestones begins here with characteristic sandstones, followed by platform limestones locally rich in nummulites and red coralline algae. Both belong to the Sanetsch Formation (Priabonian) (Menkveld-Gfeller, 1994). From this point, the frontal part of the Mont-Gond and Sublage nappes can be observed. (fig. ft2-B). At the lower part of the Spitzhorn massif, the two nice recumbent anticlines (numbered A3 and A4 on fig. ft2-B, A) belong to the Mont-Gond nappe. The core of the synclines is made of Tertiary rocks (limestones and marls). These folds are typical for the frontal part of the Helvetics.

This series is clearly overthrust by the Berriasian to Hauterivian rocks belonging to the Sublage nappe. The thrust is marked out by slices of lower Cretaceous rocks that can be followed southward in the root zone (Masson, pers. com.). These slices show the clear separation between the Mont-Gond and Sublage nappes. An important set of NW-SE directed late faults cut the crest 1.5km south of the Spitzhorn. It is marked by the presence of a wedge of white Urgonian limestones on the crest. A fault plane dipping to the SW, is

clearly visible.

STOP 2C - TSANFLEURON LAPIAZ: URGONIAN AND TERTIARY OF THE DIABLERETS NAPPE; VIEW ON THE SUBLAGE FOLDS AND SEX NOIR UNCONFORMITIES.

(Geologic map of Switzerland at scale 1:25'000, sheet 1286 St-Léonard, Badoux et al., 1959a and b)

1) The Lapiaz of Tsanfleuron is cut in the Urgonian (Barremian-Aptian) and Tertiary rocks of the Diablerets nappe (Box 2, fig. ft2-A). This area is characterized, among other things, by two episodes of karst generation. The recent karstic episodes causes the present day typical morphology with deep narrow groves and caves. The other karstic episode is pre-Middle Eocene in age. It is filled by continental deposits (siderolithic) or palustrine sediments.

A walk from the chalet and cowshed named Tsanfleuron (589075/130200) towards the hill of Sex Rouge (587650/129150) gives the opportunity to observed the following stratigraphic features (fig. ft2-C1).

- Siderolithic (sandstone with reddish clayey

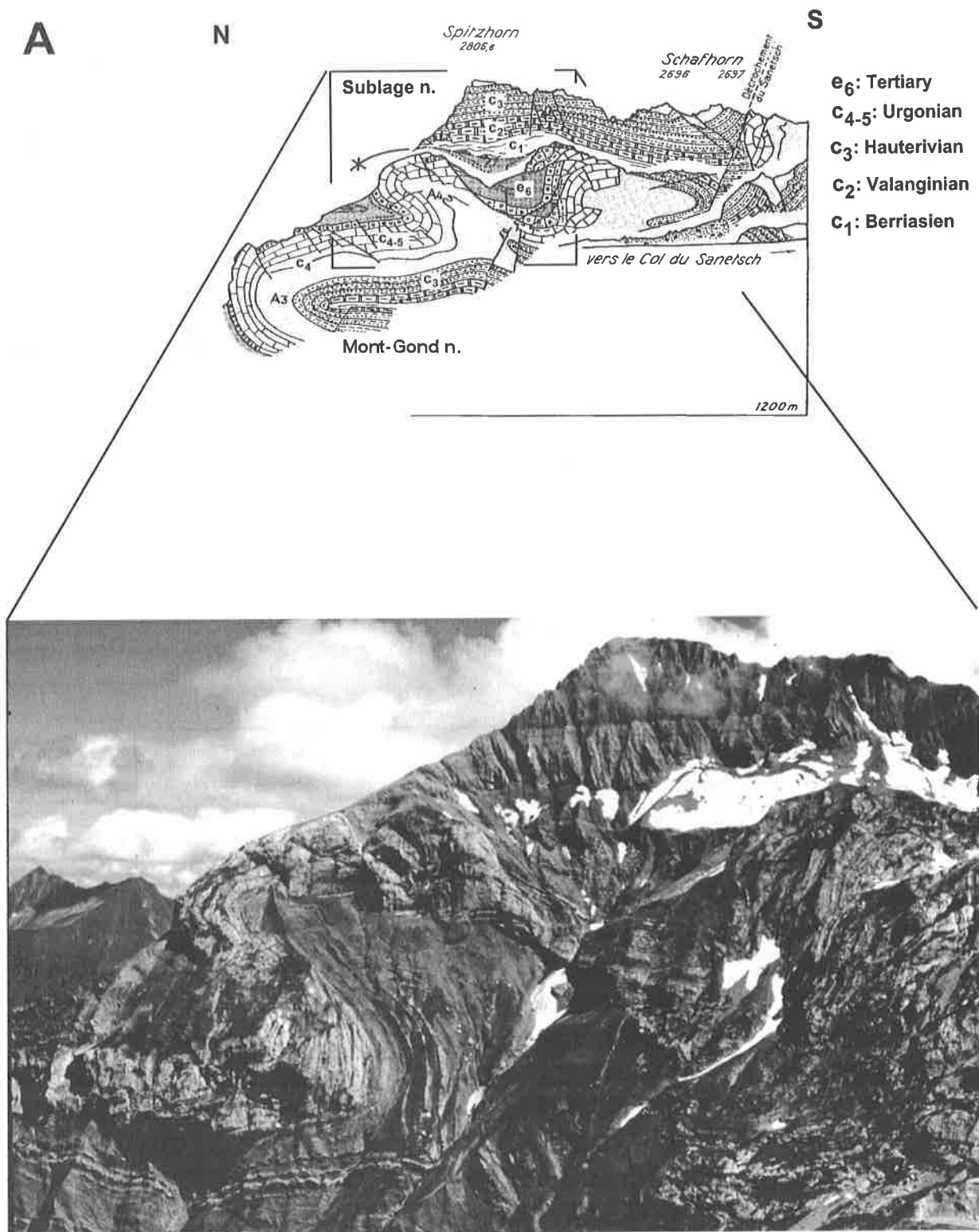
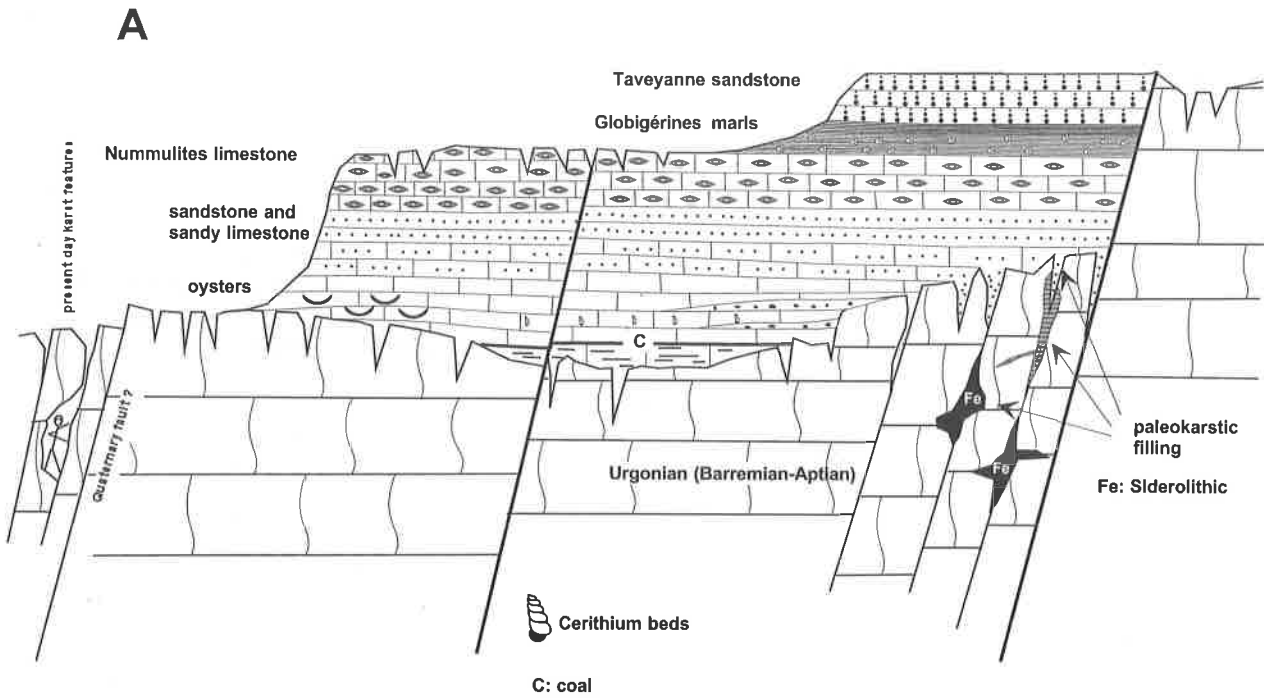


Fig. f12-B - A) Cross section of the frontal part of the Sublage and Mont-Gond nappes (modified from Badoux and Lombard, 1962). The box refers to the position of picture B.



B

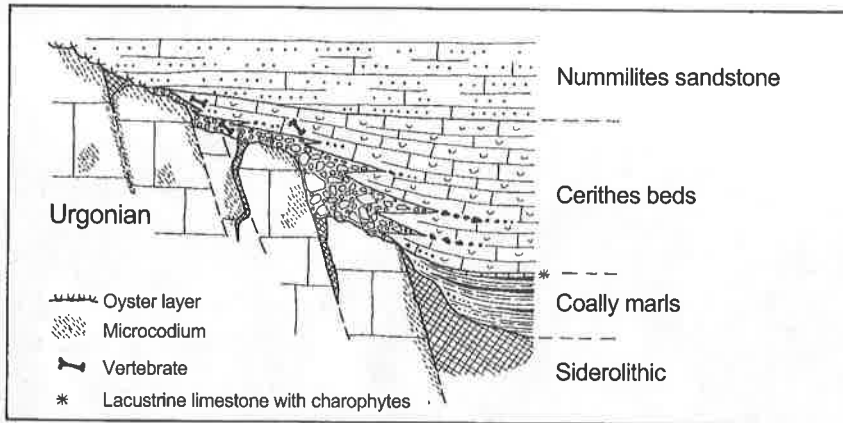


Fig. f12-C1 - A) Sketch of the relationship between transgressive Tertiary rocks on the Urgonian limestone, present day karst and paleokarst, different generation of faults. Not to scale; partly inspired from Kunz (1982). B) Sketch of the stratigraphic relationship of the Tertiary transgressive rocks on the Urgonian. Vertical scale enlarged. (Weidmann et al., 1991).

matrix) filling in paleokarstic cavities near the cowshed.

- Transgression of Eocene rocks represented by limestones and marls with oysters and gastropods (Cerithium beds), sandy limestones and limestone with nummulites and red algae on the karstified Urgonian limestones rich in rudists. Different types of karstic cavities filling in the Urgonian limestones can be distinguished. *Microcodium* (black calcite crystals of supposed bacterial origin) can be locally observed in the Urgonian below the Tertiary transgression.

- Transition from the Nummulitic platform to more shaly and deeper facies (marls with planktonic foraminifers, globigerines) then to Flysch type deposits (Taveyenne Sandstones formation).

Three types of faults can be observed. The older one is sealed by Tertiary deposit, the second is clearly of Alpine age. The third and younger one is more controversial. Some EW oriented faults are compatible with fault-plane solutions from present day micro-earthquakes in the area (Pavoni, 1980) and could well be recent. However, no clear and unambiguous geometric evidences, such as displaced Quaternary deposits, have been reported up to now.

2) The hill named Sex Rouge (587650/129150) is made of dark detritic rocks and represent flysch type deposit. It is a local facies of the Helvetic middle Oligocene flysch characterized by a significant content in detritic andesitic volcanic material. The volcanic material has been dated at 32.5 Ma (Ruffini et al. 1995). That implies a very rapid erosion, transport and deposition from the source rocks, probably located in the south-Alpine domain, to the Helvetic flysch basin. These volcanic rocks have been related to the slab break-off event (see general introduction, part 1). In some places, this formation contains typical minerals of the low grade metamorphism of basic rocks such as

laumontite, prehnite and pumpellyite (Bussy and Epard, 1984). The Taveyenne sandstone are the younger rocks of the Diablerets nappe. They mark out the axial surface of the syncline linking the Diablerets and Mont-Gond nappes.

The Sex Rouge area, offers a good view on several parts 3 of the Helvetics. Looking eastward (fig. ft2-C2A, box 3 in fig. ft2-A), the summits of Serac and Sublage are made up of "Malm" (Late Jurassic limestone) that forms two anticline cored by middle Jurassic sandy limestones, belonging to the Sublage nappe. This unit is thrust on a thick series of dark marls (Berriasian in age) belonging to the Mont-Gond nappe. The tunnels of the road leading to the Rhône valley are open in this thick formation. In the Berriasian dark marls, several "beds", a few meters thick, of white limestone can be observed. These "beds" are in fact isoclinal folds of Malm limestones that can be related to the Mont-Gond nappe. This view illustrates the disharmony between the upper Jurassic Malm limestones and the post Berriasian rocks. That is the reason why middle to upper Jurassic rocks have been distinguished from Cretaceous and Tertiary rocks by different color shades on the tectonic map given in annex (MAP 2).

In the background, to the right (south) of the Sublage summit, a white cliff is topped by dark rocks that forms the Sex Noir summit (fig. ft2-C2B, box 4 in fig. ft2-A). In the cliff, the angular unconformity of the Maastrichtian rocks (dark upper part) on Barremian to Turonian rocks is clearly visible. Locally, on the top of the crest, some brown colored rocks constitute some remnant of Tertiary (Middle Eocene) sandstones. The Maastrichtian angular unconformity cuts deeper towards the south. It is the opposite for the Tertiary unconformity in the Sublage nappe (Schaub, 1936). The Maastrichtian unconformity can be related to the Pyrenean inversion phase (see general introduction, part I).

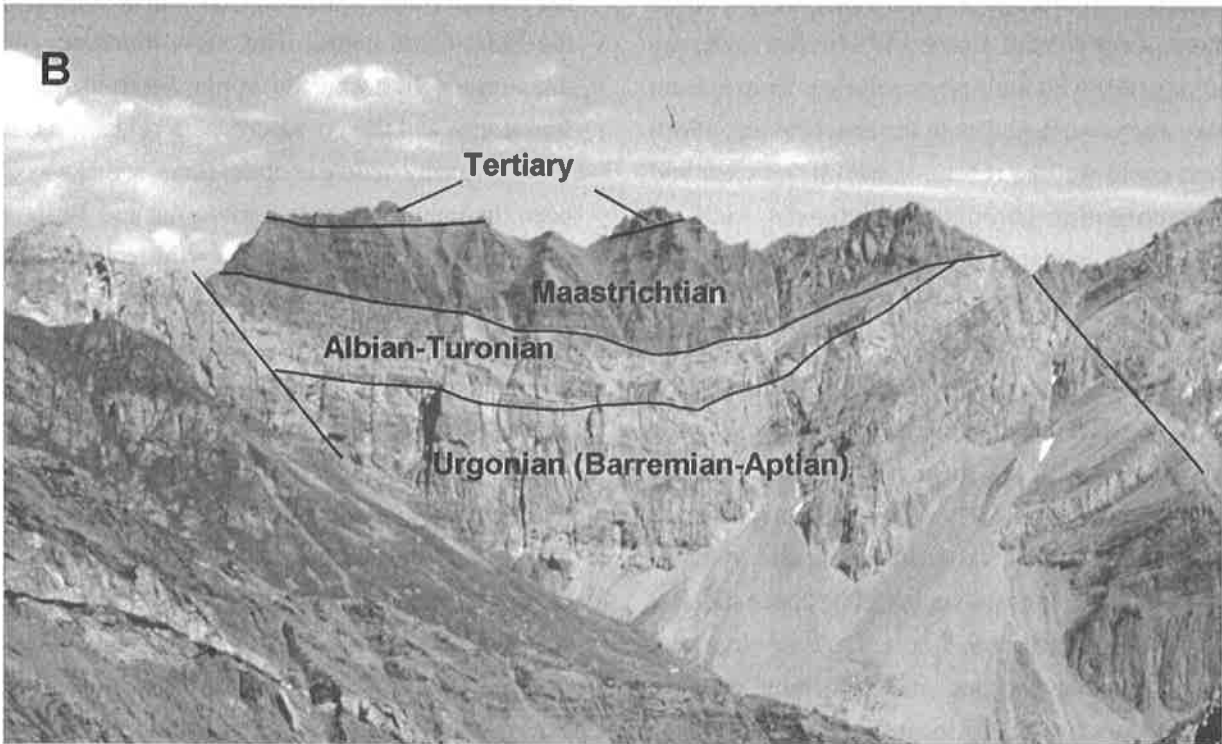
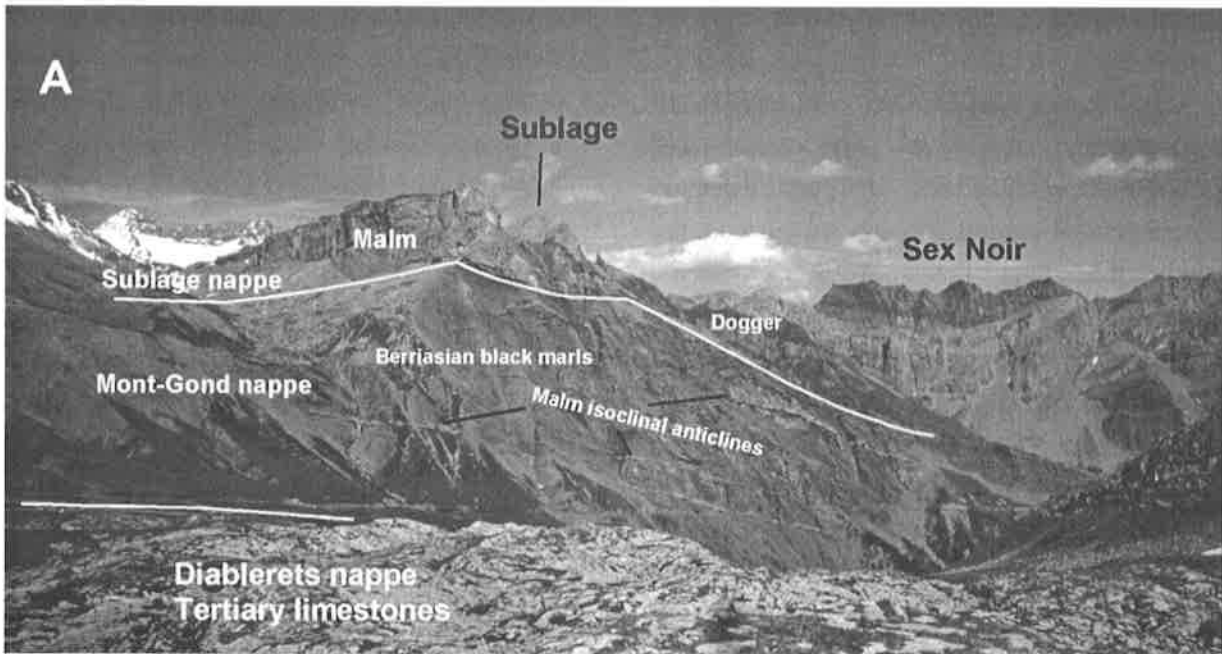


Fig.ft2-C2 - A) Sublage mountain from the west: superposition of the Diablerets, Mont-Gond and sublage nappes. B) Enlargement of the Sex Noir area: Maastrichtian and Tertiary angular unconformities.

STOP 2D - HELVETIC-PENNINIC CONTACT IN THE VICINITY OF SION, RHÔNE-SIMPLON FAULTS ZONE. (Geologic map of Switzerland at scale 1:25'000, sheet 1286 St-Léonard, Badoux et al., 1959a and b)

Between Sierre and Conthey, the contact between the Penninics and Helvetics is located at the right side of the Rhône Valley, in a populated and cultivated area. The Lower Penninic units (Sion-Courmayeur zone, Valaisan domain) outcrops mainly in the hills and vineyard facing the town of Sion whereas the Helvetics (internal part of the Wildhorn nappe) occupies the plateau of Savièse. One of the best outcrop of the thrust zone of the Penninics onto the Helvetics is located along a small road leading to the Diolly village (592325/120450, fig. ft2-D, box 5 in fig ft2-A). The outcrops can be divided in three parts. First, to the west, the black shales belongs to the Helvetics. They are traditionally considered Aalenian in age. The second zone is about 2 meters thick and is made up of orange-yellow dolomitic highly deformed rocks, traditionally considered as Triassic of Penninic origin. The third part is made up of coarse sandstones, sandy limestones and siliceous shales belonging to the Roignais-Versoyen unit of the Sion-Courmayeur zone (see Ft5 § 2.1). This for-

mation is frequently called “flysch valaisan” and it has not be directly and unambiguously dated. It is frequently considered as Cretaceous. New studies in another unit of the Sion-Courmayeur zone (Pierre Avoi unit) resulted in the discovery of Middle Eocene to Early Oligocene microfossils (Bagnoud et al. , 1998).

This outcrop gives the opportunity to observed the appearance of one important thrust zone of the Alps that brings into contact two different paleogeographic domains (see **stop 2D** and **stop 5A** were the same contact is also described). Although the break has to be important, this limit is rather discreet in the field.

In this area, the contact between the Penninics and Helvetics dipping towards the ESE is cut by numerous late strike slip faults. One of them is located in the depression north of the Mont d'Orge hill occupied by a small lake. The heave on this particular structure can be estimated to a minimum of about 1.5 km. This set of strike slip faults is related to the Rhône-Simplon faults zone active between 18 – 10 Ma (Steck and Hunziker, 1994). The contact between the Helvetics and Penninics has been strongly reworked by this important set of late strike-slip faults, and so the primary thrust contact outcrops only in limited area between the faults.

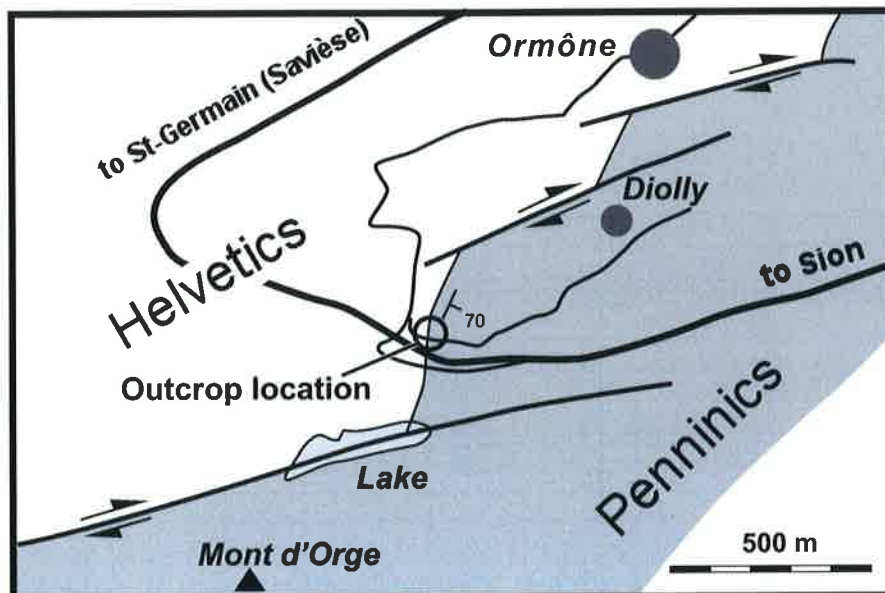
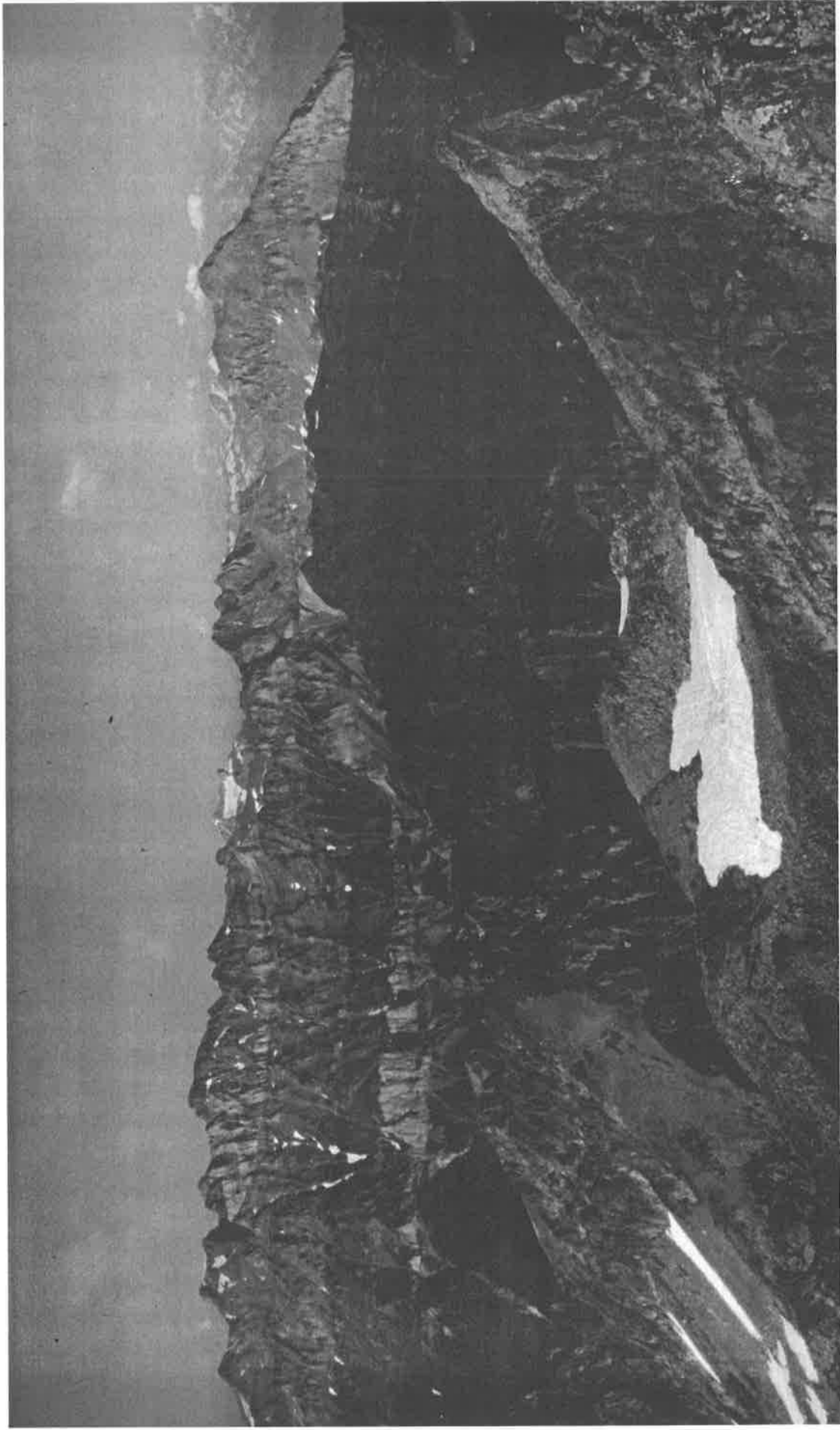


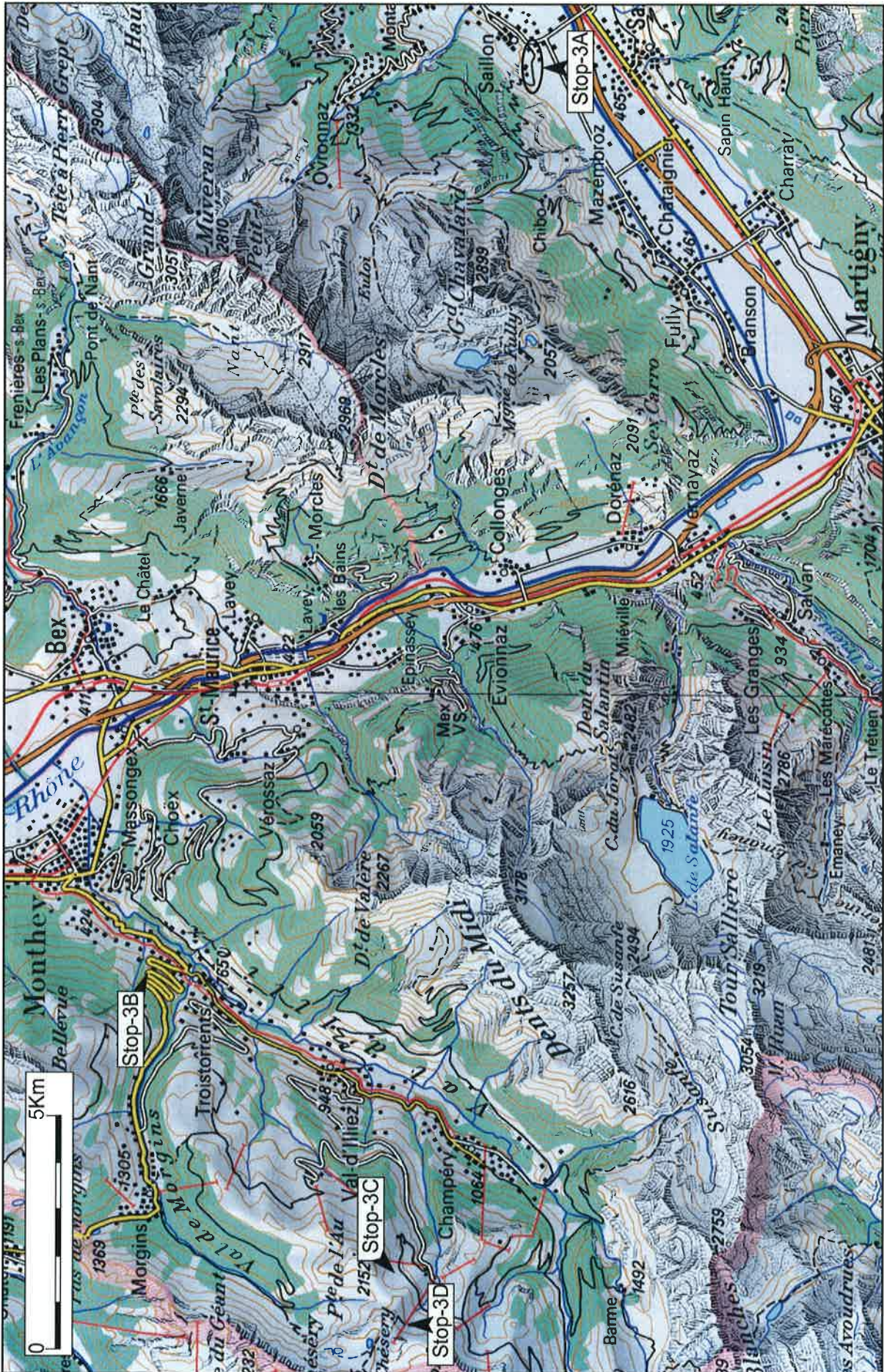
Fig. ft2-D - Tectonic sketch of the limit between the Helvetics and the Penninics in the Mont d'Orge - Ormône area.

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The Dent de Morcles seen from the Aiguilles Rouges massif



The Helvetic root zone, the ultra-Helvetic and transported Penninic suture in the Préalpes, the Brèche nappe

Gérard M. Stampfli & Jean-Luc Epard

I: INTRODUCTION

During this field trip the structural relation between the Helvetics and surrounding domains will be demonstrated. First the root zone of the Morcles nappe is visited at Saillon (**stop 3A**, topographic map on the opposite page, corresponding tectonic map in annex (MAP 1).

We start with a general panorama of the Morcles nappe and its relationship with the crystalline massifs in the area of Saillon (**stop 3A**) and the overturned limb of the nappe will be visited. The effect of a branch of the large scale Simplon-Rhône fault zone will be demonstrated.

Then we move to the front of the Helvetic nappe system by cutting the para-autochthonous Aiguilles Rouges massif along the Rhône valley from Martigny to St Maurice (see Map-1 in annex). There we cut the para-autochthonous cover domain up to Monthey where we enter the autochthonous window and its Red Molasse deposits (**stop 3B**).

The so-called autochthonous window of Monthey is found west of the town of Monthey (Badoux 1960) as one enters the Val d'Illeiez region. Due to the Late Miocene décollement of the whole Mesozoic cover up to the Jura, together with the molassic basin in piggyback position, this autochthonous sequence is a relative autochthon in regard to the Aiguilles Rouges domain only, but it is itself largely transported northwards (see the Alps cross-section in annex).

Due to the clastic nature of the Tertiary sequence and important Quaternary deposits the outcropping conditions are not very good. The sequence starts with the top of the Mesozoic carbonate platform represented by Cretaceous limestone up to the Albian, then there is an impor-

tant gap representing the flexural bulge uplift of the subducting European plate. Onlap of the bulge took place in the Priabonian and is represented by a Nummulitic carbonate platform starting with breccias. As the bulge moved northward, the platform was drowned and covered by Globigerina Shales, which make the transition to the Flysch. The latter passes upward to the Red Molasse through the Quarry Sandstone units (Grès des Carrières) in which volcanic debris can be found, as it is the case in the Helvetic and Para-autochthonous flysch (cf. **stop 2C**). This volcanic debris is derived from volcanoes active along the peri-Adriatic line where intrusives of the same age are also found, this magmatic episode is related to the slab break-off event (see general introduction). The Red Molasse is of Late Rupelian-Chatian age (Late Oligocene); it marks the passage from a deep foreland basin to an overfilled basin due to increasing erosion and clastic input. The window was recognised already in the XIX century and allowed people like Schardt to prove the exotic nature of the Préalpes nappes (Masson 1976), as it precludes to root those nappes in front of the Helvetic domain.

The transition to the Préalpes Médiannes nappe is marked by mélanges mainly of tectonic origin (see also field trip 6). The contacts are characterised by special types of units and rocks such as rauhwackes (cornieules) associated with anhydrites. For the Préalpes Médiannes two groups of mélanges can be distinguished: the supra-Préalpes Médiannes mélanges mostly related with the emplacement of the nappe Supérieure and Brèche nappe, and the mélanges infra-Préalpes Médiannes (they are mostly supra-Helvetic) that are linked to the Préalpes

Médianes emplacement.

As one enters the Val d'Illeiz the Préalpes Médianes Rigides massif of Dreveneuse can be seen towering over Monthey to the NW. But west of the Pointe de Dreveneuse, the Rigides unit disappears and the Brèche nappe is found directly on the Ultrahelvetic (UH) units (**stop 3C**).

The Brèche/Rigides–UH contact represents the transported Valais trough suture (lower Penninic front) in the Préalpes domain, the same contact is also seen in field trips 2 (**stop 2D**) and 5 (**stop 5A**). This contact is well exposed in the Portes du Soleil region above the village of Les Crossets, where very good outcrops of the Brèche nappes will also be visited (**stop 3D**). The Brèche is the best example in the Alps of syn-rift deposits associated to the opening of the Alpine Tethys in Late Liassic times.

II: EXCURSION OUTCROPS

STOP 3A - SAILLON: HELVETIC ROOT ZONE.

(Geologic map of Switzerland at scale 1:25'000, sheet 1308 Dt de Morcles, Badoux, 1971)

The village of Saillon is located 15 km downstream the town of Sion, on the right (NW) side of the Rhône valley (580500/113500).

Parking lot south of Saillon

The Saillon old village and its surrounding wall are mainly built on a hill made up of Liassic rocks. The dip of bedding plane is moderate towards the NE. The series is in an upright position, the older outcropping formation (Hettangian marls and limestones) is located at the southwest end of the hill, the younger rocks, Toarcian marls, are located at the northeastern end. The old tower, named Bayard tower and the SW surrounding wall are built near the top of a thick bedded sandstone and quartzite formation, Lotharingian (Late Sinemurian)

in age. This formation can be observed at the beginning of the road leading to the old village, from the southern side of the hill, and is typical for coarse grained detritic rocks deposited on, or near a basement high (internal part of the Mont-Blanc massif). Description of the Triassic and Liassic series of the area can be found in Badoux (1971, 1972), Epard (1989), Loup (1992) and Corset (2000).

Bayard tower (580225/113250)

From the top of the old tower that can be reached from the old village, the following panorama can be observed:

Towards the SW, the highest summits belong to the Mont-Blanc crystalline massif. Due to the Rawil axial depression, the top of the crystalline rocks plunges towards the NE and reaches the Rhône valley just south-west of the village of Saxon, on the opposite side of the valley.

In the continuation of the Rhône valley in the SW direction, a pass can be observed (Forclaz). This pass is cut in the sedimentary rocks of the Chamonix zone that separates the Mont-Blanc crystalline massif to the left from the Aiguilles-Rouges crystalline massif to the right (see Ft1, **stop 1C**). The Aiguilles-Rouges massif itself outcrops in the slopes at the right side of the valley, downstream of the Saillon village. In the forest, an isolated dip-slope is formed by the Triassic sandstone belonging to its autochthonous cover. Sedimentary rocks that outcrop directly NE and on top of the crystalline massif, form the overturned limb of the Morcles nappe. A cave is visible at mid-slope, at the foot of some little cliffs, it is dug in the lower Cretaceous, Urgonian limestone of the overturned limb of the nappe, here highly deformed and transformed in marble that has been exploited for building stones. This marble reaches the valley level in the Sarvaz quarry located 500m W of the tower. In the vineyard at the lower part of the mountain slope, the Longeraie area, is

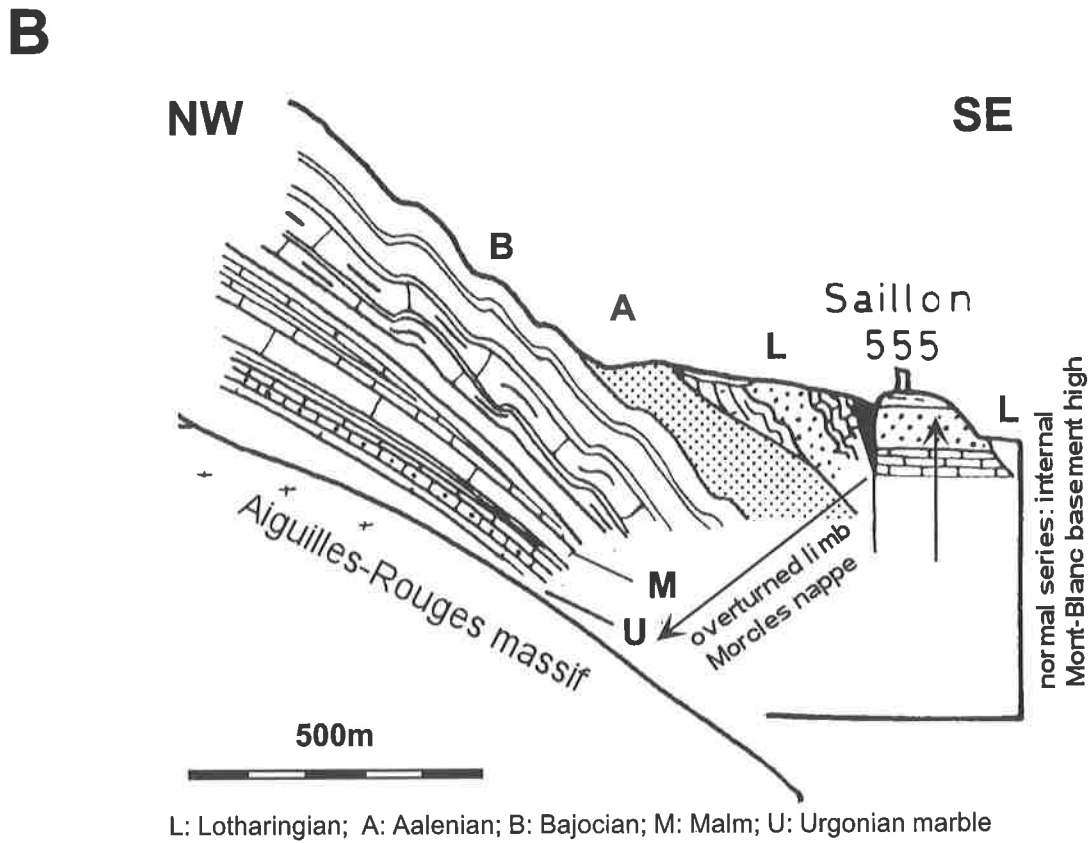
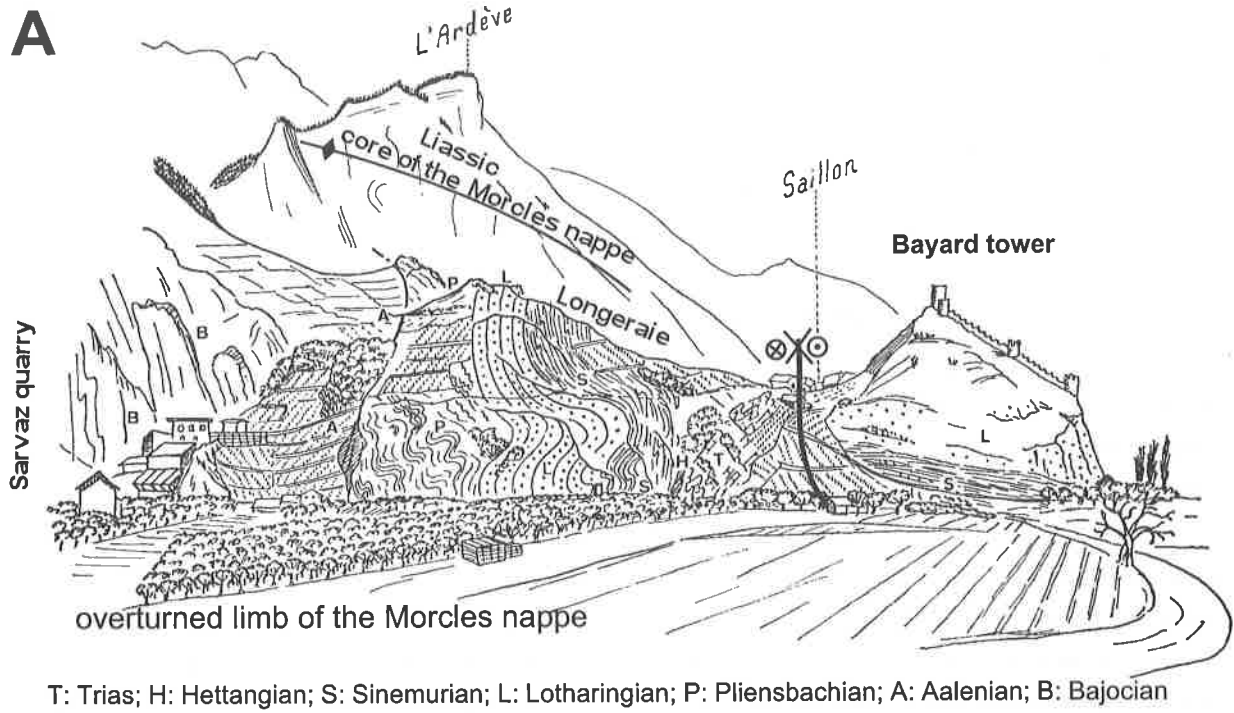


Fig. ft3-A: A) View of the Saillon hills from the SW (modified from Badoux, 1972). B) cross section of the overturned limb of the Morcles nappe in the Saillon area (modified from Badoux, 1971).

occupied by Liassic rocks of the overturned limb of the nappe. Triassic rocks are outcropping at the SW end of the depression located north of the tower and partly occupied by the Saillon village.

In the NE direction, the Ardève forms a characteristic isolated little mountain. Its cliff, facing the Saillon village, is topped by a forested area that covers also the NE side of the hill. This mountain is formed by folded Liassic rocks and represents the core of the Morcles nappe. It is surrounded by a topographic depression occupied by the very thick Aalenian shales. At the background, to the N, a huge cliff of white limestone, Late Jurassic (Malm) in age, corresponds to the normal limb of the nappe (Haut de Cry summit). By clear weather, several folds can be observed in the cliff which is topped by a more shaly formation, in some places pinched in narrow synclines. It corresponds to the lower Cretaceous (Berriasian) alternation of marls and shaly limestones. The foot of the cliff is made of middle Jurassic rocks. This thick upper Jurassic limestone formation (200-300m) can be followed northward in the frontal part then in the overturned limb of the nappe, where it is tectonically thinned down to a few tens of meters. It reaches the valley level in the Sarvaz quarry.

The Saillon fault.

An important fault is located in the SW-NE oriented depression between the Longeraie and Bayard hills. Until recently, this fault was not directly observed but postulated from the following structural argument. The northern part of the depression (Longeraie hills) is occupied by an overturned series of Liassic rocks dipping towards the SE (overturned limb of the Morcles nappe). This Liassic series cannot be directly linked to the Liassic rocks of the Bayard hill which are in an upright position and dip towards the NE. Therefore, the presence of a fault was already postulated by Lugeon and Badoux.

The Liassic stratigraphy of the Bayard hill, in

particular the Lotharingian massive quartzite formation is typical of the cover of the internal part of the Mont-Blanc massif. It contrasts with finer sediments of the same age found in the Morcles nappe in Longeraie or Ardève (North Helvetic basin). The presence of the fault was recently confirmed by direct observation in a little excavation in the Saillon village (Masson, pers. com.) and was cut in a geothermal drill hole. This fault is interpreted as one of the numerous branches of the important Rhône-Simplon fault zone. Based on structural consideration, the amplitude of the dextral strike-slip movement can be estimated from a minimum of about 3 km up to 8km.

Path from Saillon to the Sarvaz quarry.

An horizontal path links the Saillon village to the Sarvaz quarry located 500m to the west (579650/113250). It cuts almost completely the overturned limb of the Morcles nappe. The Triassic rocks and the base of the Liassic are not outcropping directly along the path, but in the vineyard just below. The first Liassic formation easily observed is a thin bedded alternation of quartzite and sandy limestone. It is attributed to the Lotharingian and is the lateral equivalent (North Helvetic basin) of the massive quartzite series (Internal Mont-Blanc basement high) observed in the first outcrop. The series continues with Pliensbachian limestones followed by the Toarcian-Aalenian marls and shales. The beginning of the quarry is cut in Bajocian - Bathonian well bedded limestone. Several small late folds can be seen in this formation. The very deformed upper Jurassic white limestone can be observed in the wall of the quarry. Its thickness (20-30m) is tectonically reduced here to a tenth of the thickness measured in the normal limb of the nappe. Due to the strong deformation the different Cretaceous formations are difficult to distinguish, except for the white marble outcropping at the far end of the quarry, corresponding to highly strained Urgonian limestone

(Barremian - Aptian). Illite-crystallinity studies indicate epizonal metamorphic condition for the Saillon marble, and chloritoid has been found in the black Aalenian shales (Goy-Eggenberger, 1998). Please note that partial filling of the quarry can modify the quality of the marble outcrop.

STOP 3B - MONTHEY TO MORGINS: THE MONTHEY AUTOCHTHONOUS WINDOW

St Maurice geological map (Badoux 1960) at 1:25'000 scale and Samoens-Pas de Morgins map (Plancherel 1998) at 1:50'000 scale.

Taking the road from Monthey to Champéry, then the road to Morgins at Trois Torrents, one crosses the autochthonous window of Monthey, represented along the road by the Red Molasse sequence. There are a few outcrops on the road leading to Morgins, starting in the sharp bend at 953 m of altitude and up to a few kilometres before Morgins. The red colour is characteristic as well as the fine clastic lithology of this molasse sequence.

Tectonic: the location of this molassic window can be seen in figure ft3-B1. It outcrops to the N and NW of the town of Monthey, and consists of the sedimentary cover of the Aiguilles rouges Variscan massif. It is slightly detached from this basement and folded. The Red Molasse forms the top of this cover sequence and is the southern most occurrence of the Molassic basin. Its presence proves the allochthony of the Préalpine nappes.

Sedimentology: the sedimentary section visible in the autochthon window starts with lower Cretaceous limestone series of Valanginian to Early Aptian age (north of Monthey). The top of the sequence is made of reefal white massive (100m) Urgonian (Barremian-Aptian) limestone. Karstic deposits whose ages can range from Albian to Priabonian affect the top of the Urgonian. The

Priabonian is represented by Nummulitic limestones followed by Globigerina marls then by flysch whose age reaches the Rupelian (Early Oligocene). The Red Molasse follows and is of Late Rupelian-Chattian age (Late Oligocene), it is made of red-coloured clastic rocks where fine sandstones with a carbonate cement predominate, green reduction spots are also found and allow to measure the deformation through the degree of their ellipsoidal shape.

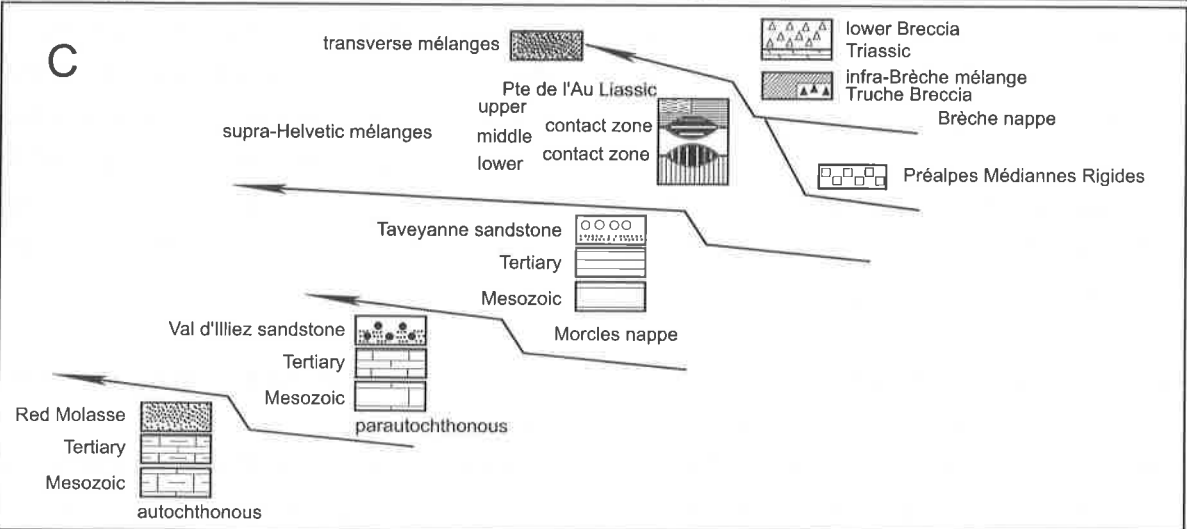
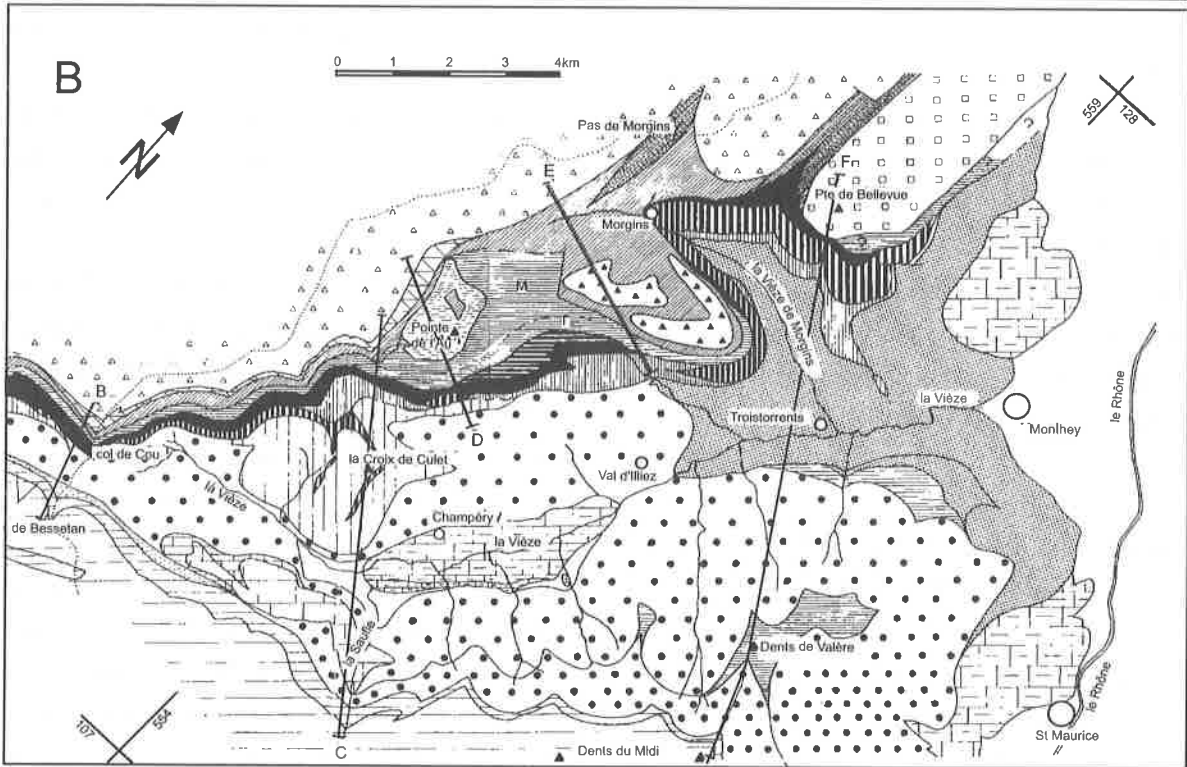
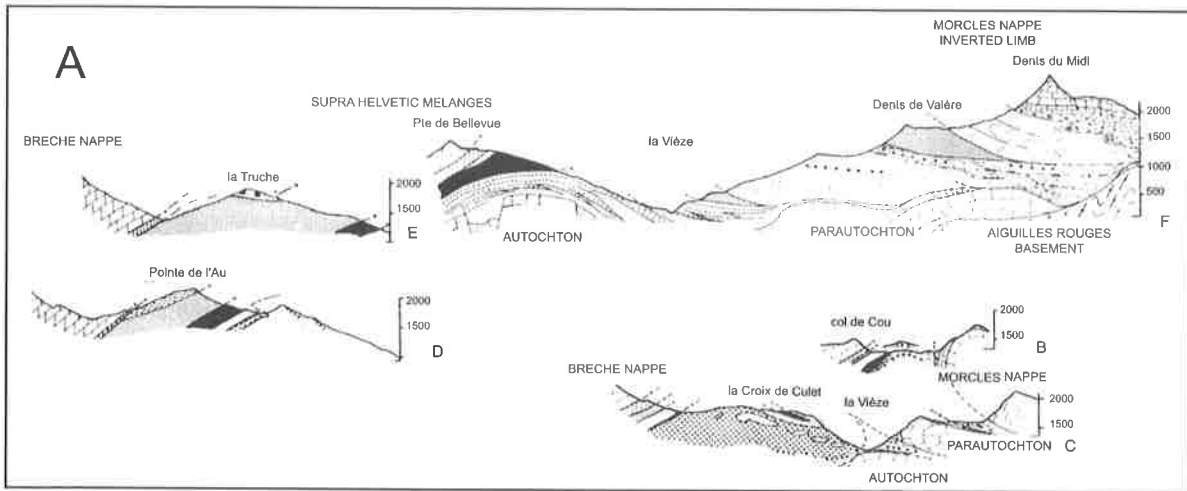
STOP 3C - LES CROSETS TO THE PORTES DU SOLEIL PASS: PANORAMA OF THE HELVETIC FRONT AND ULTRAHELVETIC MÉLANGES

St Maurice geological map (Badoux 1960) at 1:25'000 scale and Samoens-Pas de Morgins map (Plancherel 1998) at 1:50'000 scale.

From the ski-resort Les Crosets we take a small road leading to the Les Portes du Soleil (formerly Portes de Solet) pass (552900–115900) and then to the Lac Vert area, on the right of the Pointes des Mossettes, where there is a cable-car end-station. Along the road before the pass one can see the Ultrahelvetic mélanges zone making the transition to the Brèche nappe (2). From the ridge a few hundred meters from the pass up-slope (552800–115600), a panorama (1) allows to see towards the south, the Helvetic nappes (fig ft3-C1) and towards the north, the Ultrahelvetic mélanges (fig. ft3-C2 & -C3) and their relation with the Brèche nappe. The following description follows the geological order, not the order in which the different units can be seen along the road. The panorama on the Dents du Midi is seen all along the trip, weather permitting... but the spot where the picture was taken is after the pass.

1–Panorama of the front of the Helvetic nappes (fig. ft3-C1)

The autochthonous: the Monthey window is just seen from this spot, it is far down towards the



Rhône valley on the left and represented by a cliff dominating the Vièze river.

The para-autochthonous: the En Barme parautochthonous Mesozoic window (section C, fig. ft3-B) is seen on the right, under the Dent de Bonavaux and behind the Croix de Culet down in the valley. Large outcrops of parautochthonous flysch occupy the northern flank of the Dents du Midi. A cryptic thrust plane separates this flysch from a similar flysch of the Morcles nappes (cross-section F, fig. ft3-B)

The Helvetic nappes: The Dents du Midi massif and its continuation westward to the Dent de Bonavaux and the Dents Blanches massif, represent the inverted (overturned) limb of the Morcles nappe (see Alps cross-section in annex). In the back-ground the Tour Salière and parts of the Mt Ruan massif, represent its normal limb, westward this normal limb finally predominates over the inverted limb. Several folds affect the inverted limb and are well visible in the Dents du Midi massif. The base and the top of the Dents du Midi are made of Urgonian (Barremian-Aptian) massive limestones forming cliffs, the rest of the massif is dominated by Hauterivian series.

2-Transitions from the Brèche nappe to neighboring structural units (fig. ft3-C2)

The Ultrahelvetetic nappes and mélange zones

Tectonics: the Préalpes Médiannes rest on top of the Helvetic nappes in the S and the Subalpine Molasse and flysch in the N. All along this contact we observe the “Ultrahelvetetic” and the “Zone Submédiane” units which have been interpreted in recent publications as mélanges similar to those formed in accretionary prisms. They were

subsequently strongly overprinted by Alpine tectonics, especially during nappe emplacement (Jeanbourquin 1994).

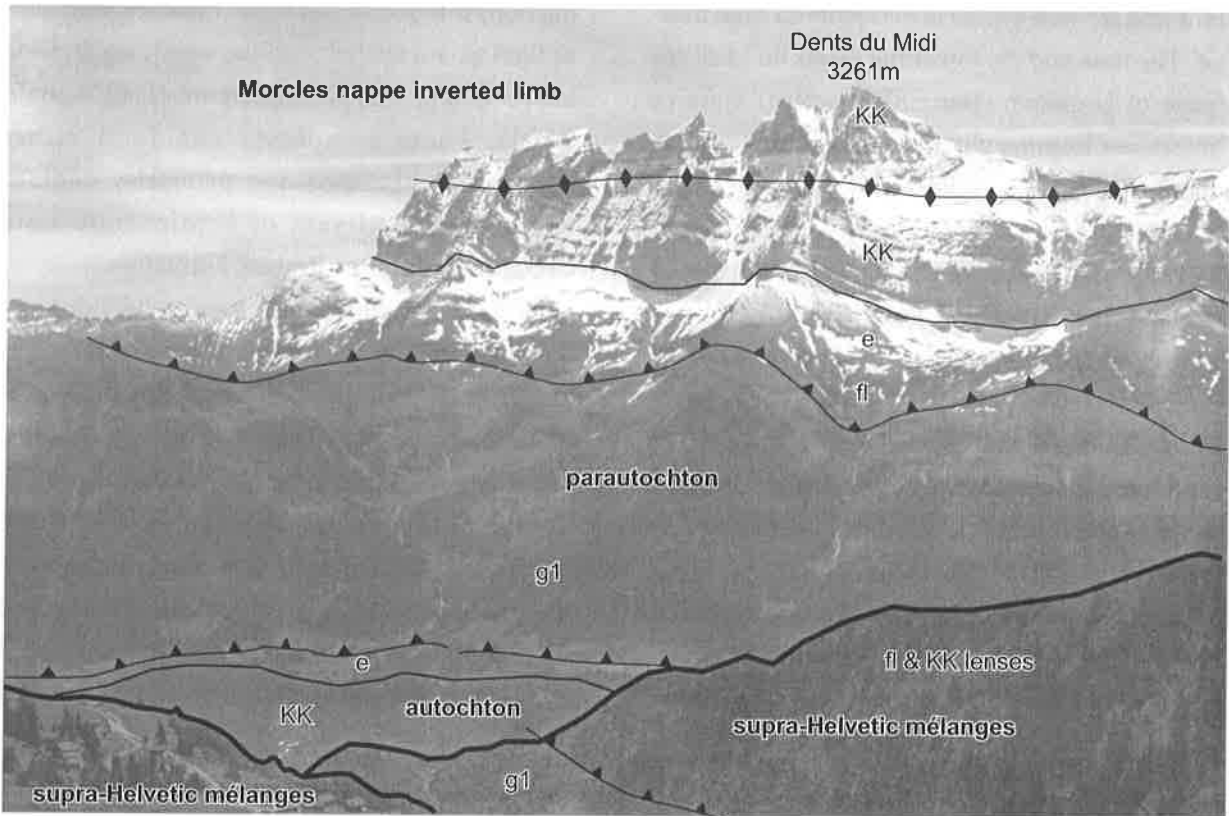
- The Zone Submédiane (Weidmann et al. 1976) is found below the basal décollement of the southern part of the eastern lobe (Romandes) of the Préalpes Médiannes. It forms the contact with the underlying Niesen flysch nappe in the central and eastern Préalpes Romandes and the Ultrahelvetetic units in the western part of the Préalpes Médiannes along the Rhône valley. West of the Rhône valley, in the Chablais part of the Préalpes, where the present field-trip takes place, this zone and the Niesen nappe are not found. Some isolated outcrops along the Val d’Illiez region have been regarded as potential equivalents of these units, but further studies would be necessary to confirm it.

- The “Flysch with Couches Rouges lenses” (Chaotic complex in fig. 11 of the general introduction) is found on top of the Préalpes Médiannes as well as associated with the overlying Breccia nappe and/or Nappe Supérieure (Dall’Agnolo 2000). These complexes can form rather continuous horizons and primarily contain characteristic slivers of hemipelagic Late Cretaceous Couches Rouges Formation.

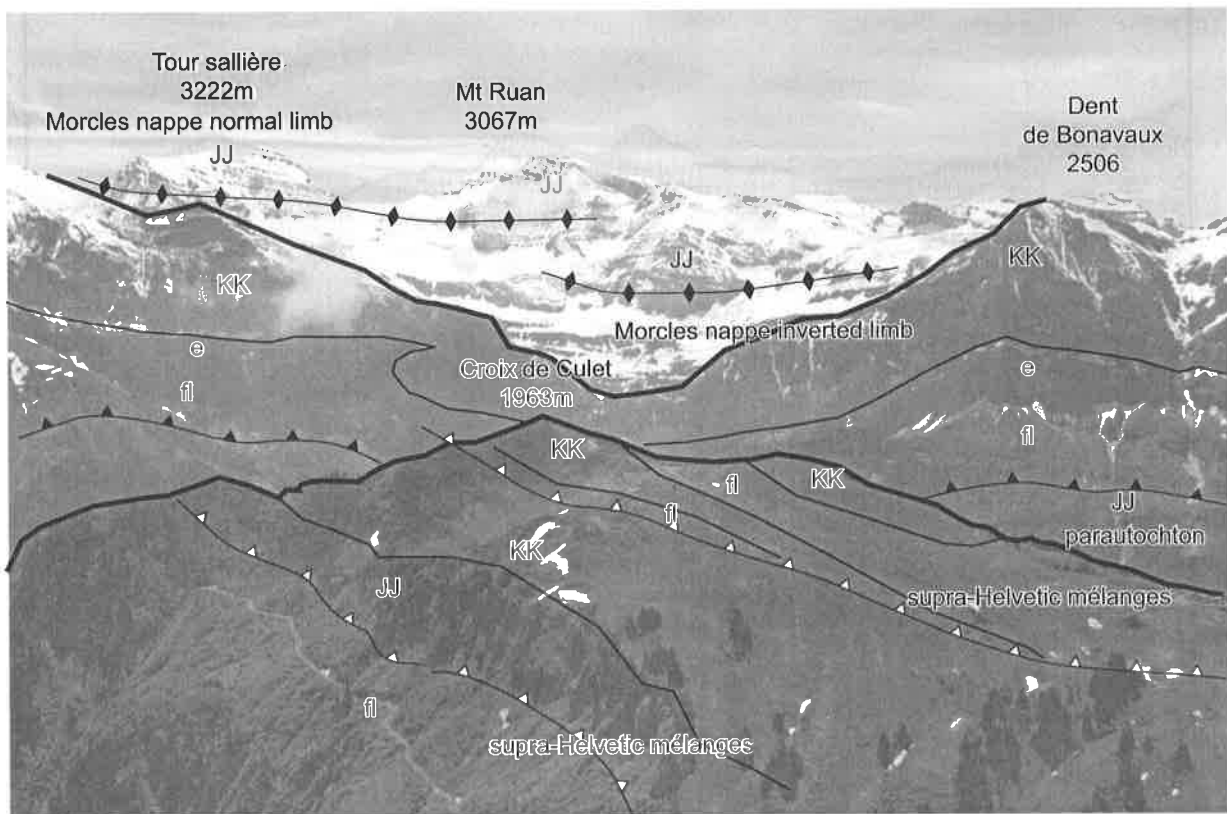
- The “Ultrahelvetics” domain was subdivided into an upper, middle and a lower supra-Helvetic mélanges in the Val d’Illiez region (Jeanbourquin et al. 1992) (fig. ft3-B1). These three units are labelled mélanges infra-Prealpine by Plancherel (1998). These supra-Helvetic mélanges are locally covered by the infra-Brèche mélanges, the limit between the two being not always very clear. The origin of the tectonic slivers found in these supra-Helvetic mélanges is placed in paleogeographic domains of

Fig. ft3-B (p102) - A, Cross-sections of the Val d’Illiez region, location on B, map of the Val d’Illiez region, presenting the relationship between the Helvetic domain and Préalpes domain, legend on C, showing the superposition of the different structural units. Modified from Jeanbourquin et al. (1992)

Fig. ft3-C1 (p104-105) - Panorama of the Dents du Midi massif, seen from the Portes du Soleil area (552600– 115800, alt.c. 2040m)



- | | | | |
|---|----------------------|-----|-----------------|
| — | stratigraphic limits | ▲▲▲ | major thrusts |
| — | horizons | △△△ | minor thrusts |
| — | | ◆◆◆ | anticlinal axis |



fl flysch
 e Eocene
 KK Cretaceous
 JJ Jurassic

g1 parautochthonous flysch

The Dents du Midi massif
 panorama from the Portes du Soleil pass
 coord. 115600-552800

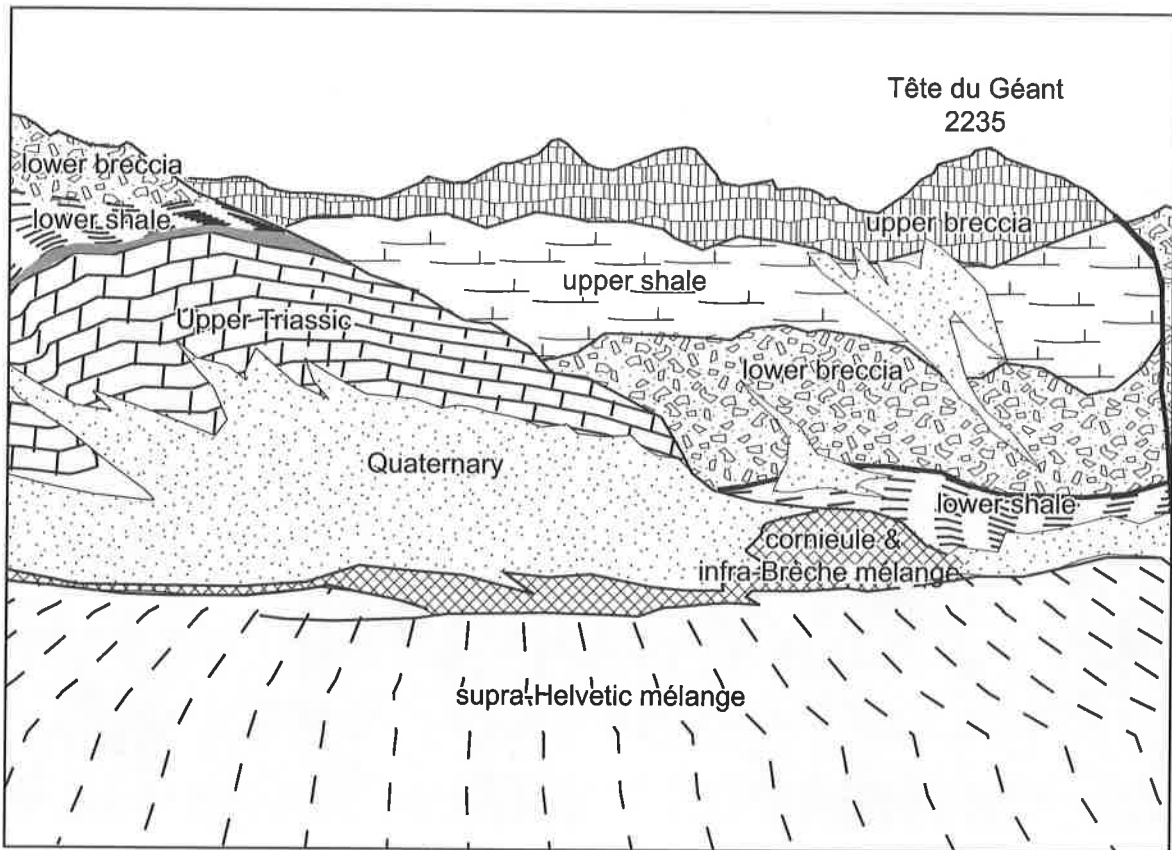


Fig. f13-C2 - View of the contact between the Brèche nappe and the supra-Helvetic mélanges, a few hundred meters from the Portes du Soleil pass.

the whole Helvetic margin and Valais trough, they range from basement and Carboniferous slivers to Tertiary flysch which locally plays the role of matrix.

Locally, as in the slope leading to the pass, block like limestone outcrops can be seen (fig. ft3-C3), however one is not dealing here with olistoliths, but tectonic slivers of composite and contrasting sedimentary nature excluding an olistolitic origin. In the Crosets area and all along the southern border of the Brèche nappe in the Val d'Illicz region, these mélanges are in direct contact with the Brèche nappe (fig. ft3-C2). Therefore, there is a local tectonic omission of the intervening Préalpes Médianes Rigides and Submédiannes zone usually found just above the Ultrahelvetic nappes in the eastern (Romandes) lobe of the Préalpes (see Ft6).

Sedimentology of the UH mélanges: walking along the track from les Crosets to the pass one can see a few lithologies related to the UH mélan-

ges. Starting from the end of the asphalt road (between 1700 and 1800 m of altitude) a few outcrops of marly shales are found along the track corresponding to a ribbon of mid Jurassic outcrops from the middle mélange unit. Around the sharp bend leading to the final straight portion of the road, outcrops of Tertiary UH flysch can be seen along the track, represented by turbiditic immature sandstones and clayey shale. Then one reaches the first block-like limestone outcrop on the right hand side of the track (553400–116100). At its base a sliver of immature mixed carbonate and quartzitic sandstones is found. The sandstones are locally very coarse and contain many clasts of Triassic dolomites, this type of sandstone cannot be regarded as a Tertiary UH flysch. It could be an equivalent of the middle Jurassic Grande Eau sequence (field trip 6F) and therefore an equivalent of the Niesen nappe, a proposition already made by Gagnebin (1934). Similar middle Jurassic sandstones have been reported from other equivalent areas (Kindler

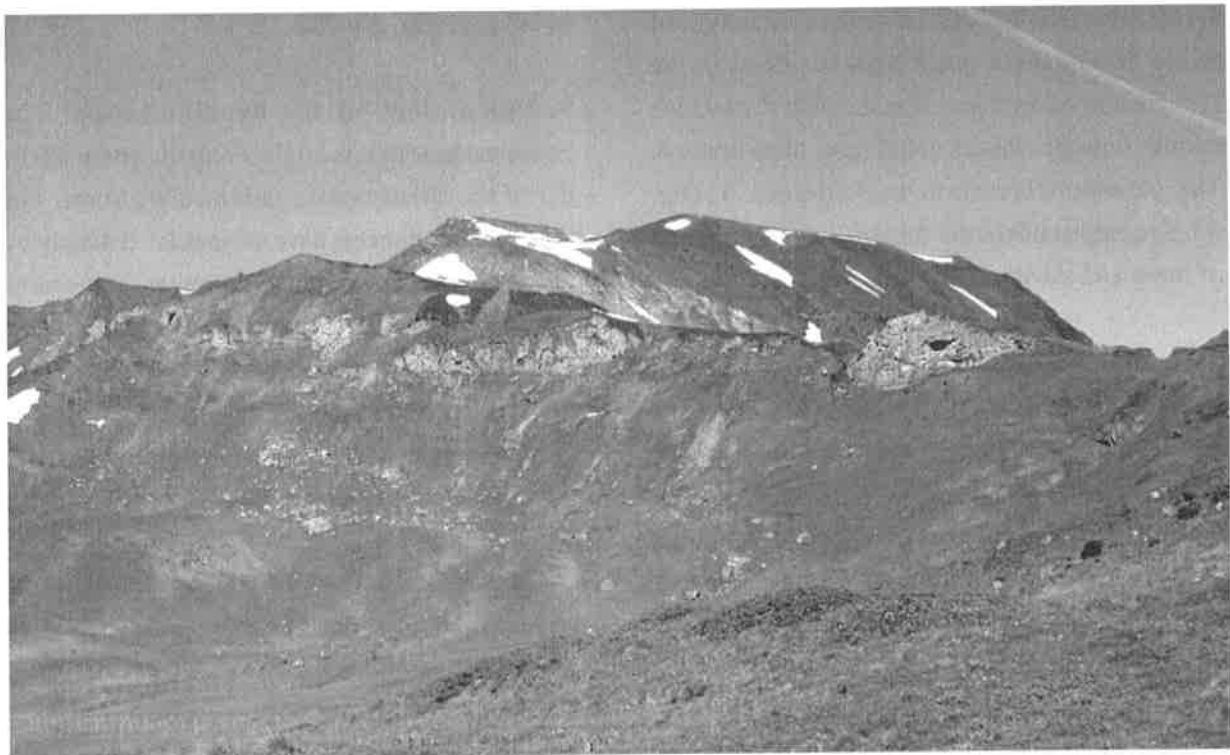


Fig. ft3-C3 - View of the tectonic lenses in the supra-Helvetic mélanges, seen from the road leading to the Portes du Soleil pass, above Les Crosets village. The mountains in the background belong to the Brèche nappe.

1988).

The limestone block is made of micritic and echinodermic dark coloured limestone attributed to the Middle Liassic. Then just before reaching the pass (1964m), the limestone grades into dark marly shale attributed to the Rhaetian (Gagnebin 1934). Therefore the sequence here, if not completely dismembered, would be inverted. This seems to be confirmed by the next outcrop a few hundred meters from the pass up slope, made of light coloured dolomites attributed to the Late Triassic, followed by isolated outcrops of sandstones and black-siltstone and sometime red shales attributed to the Late Carboniferous (Plancherel 1998).

Interpretation: As proposed by Plancherel (1998), the whole sequence could well be the equivalent of the Jurassic sequence found at the bottom of the Niesen nappe in the Le Sepey-La Forclaz region (see field trip 6, **stop 6F**). It would have been in a former syn-rift position, south of the Helvetic rift shoulder, however, similar type of facies development could also be found in the Helvetic rim-basin close to the shoulder from which middle Jurassic clastics could have been derived. The presence of breccia in the Pointe de l'Au (fig. ft3-B) could confirm the former syn-rift position of these UH slivers.

STOP 3D - FROM LES PORTES DU SOLEIL TO THE POINTE DES MOSETTES: BRÈCHE NAPPE: TRIASSIC PRE RIFT SEDIMENTS AND LIASSIC SYNRIFT SEDIMENTS

St Maurice geological map (Badoux 1960) at 1:25'000 scale and Samoens-Pas de Morgins map (Plancherel 1998) at 1:50'000 scale.

Going from the Portes du Soleil pass (552900–115900, alt. 1964m) to the Pointe des Mosettes (551900–115800, alt. 2271m) following the road leading to the cable-car end-station, one can see

the complete lower part of the Brèche nappe (infra-Brèche mélanges, cargneules, Triassic dolomites, Rhetian-Early Liassic shales, lower Breccia). To see the continuation of the sequence, one has to take the track down to the Lac Vert then to the col de Chésery (550800–116500) and take the direction to France. There, along the western flank of the Pic de Chésery, the upper shale sequence (with green and red radiolaritic shales) and upper Breccia are clearly visible and form large outcrops. For the location of the post Upper Breccia formations see Dall'Agnolo (2000).

Tectonics : the Brèches nappe is composed of series of large-scale folds plunging to the NW. They overlie the Préalpes Médiannes Rigides and are overlain by the Nappes Supérieures (see field trip 6). In les Croset area the basal part of the nappe is made of massive Triassic dolomites transformed in rauhewackes (cornieules). The contact between the Brèche and UH units is clearly seen a few hundred meters after the Portes du Soleil pass (fig. ft3-C2).

Sedimentology of the Breccia nappe: The Breccia nappe has its origin along the internal border of the Briançonnais carbonate platform. The bulk of the sediments have accumulated mainly by gravitational deposition from active fault scarps along the rift shoulder. The paleogeographic domain of the Breccia nappe corresponds to a syn-rift position in a continental margin (fig. 11). The Breccia nappe thus offers the possibility to study the sediments associated with the rifting of the Alpine Tethys.

The most important formations of the Breccia nappe can be described as follows (Steffen et al. 1993; Dall'Agnolo 2000)(fig. ft3-D1):

Pre-rift sequence

- Permo-Carboniferous clastics are found locally at the base of the Brèche nappe but are

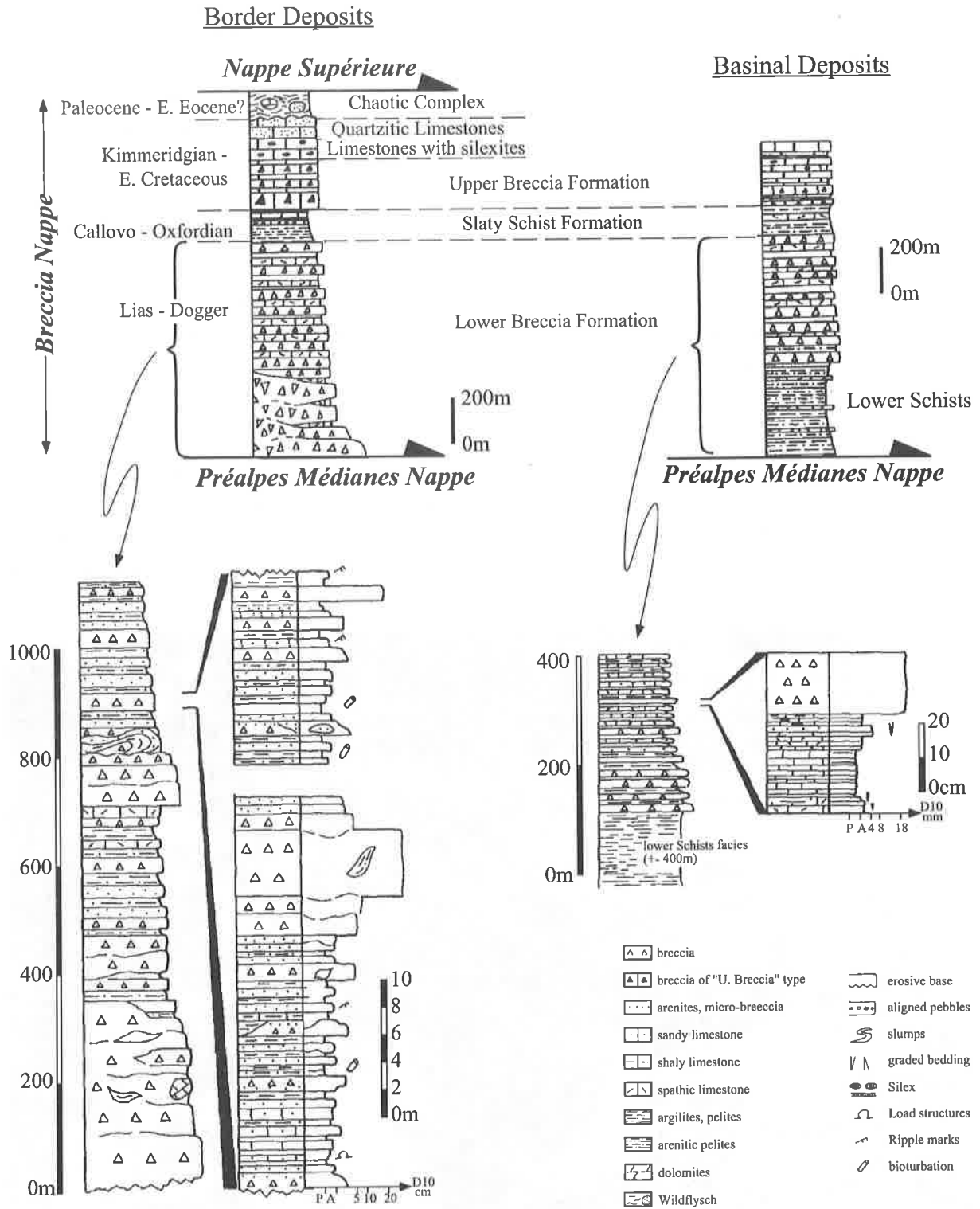


Fig. f13-D1 - Stratigraphic correlation chart between the Médiannes rim basin and the Brèche basin (modified from Steffen et al. 1993).

usually considered to belong to the infra-Brèche mélange or UH mélanges. Usually the base of the nappe is made of upper Triassic carnageules and dolomites, followed by fossiliferous Rhaetian limestones and marls grading into black marls with some limestone beds or lenses (Lower Shale Formation) of Early Liassic age. Along the road going to the Pointe des Mosettes, a few beds of breccia are found already within these dark marly shales, reworking the Triassic dolomites. The pre- to syn-rift limit should be put somewhere within the Lower Shale Formation

Syn-rift sequence of Alpine Tethys ocean

- The Lower Breccia and Lower Shale Formation (Early-Middle Jurassic): these two formations are considered lateral equivalents with gradual

transitions. The basinal deposits of the Lower Shale Formation are formed by an alternation of marly shale and limestone with minor breccias and turbidites rich in crinoidal fragments. The Lower Breccia Formation consists of large breccias beds of debris flow type (fig. ft3-D2) containing carbonates derived from the Briançonnais shoulder, alternating with turbidites and minor shale. The two formations form a sedimentary prism with breccias dominating northward – *ie.* toward the former rift shoulder (fig. ft3-D3).

Post-rift sequence

-The Upper Shale Formation (Schistes ardoisiers; Callovo-Oxfordian): this formation is mainly composed of distal turbidites and dominating pelagic dark grey, red or green argilites (with



Fig. ft3-D2 - middle fan channel in the lower part of the lower Breccia formation

radiolarian) sometimes interlayered with calcarenites and breccias and Mn-rich sandstones.

Syn-rift sequence of Pyrenean ocean

-The upper Breccia (Kimmeridgian-Tithonian): this formation is characterised by calcilutites interlayered with breccias. The breccias contain reworked clasts, not only from the Jurassic limestones and the Triassic dolomites, but also from the Paleozoic basement (chloritic shales and quartzites). At the top of the formation are found cherty limestones of the Bonave formation (Dall’Agnolo 2000).

Post-rift sequence

-The post-Breccia formations (de Bonave Fm., de la Joux Verte Fm., Couches Rouges and Flysch):

they are made of cherty limestone of Tithonian to Barremian age, black and dark green siliceous limestones with glauconites, argillaceous black (sometimes red or green) shale and limestones of Barremian to Turonian age, followed by Couches Rouges (Red Beds) of Campanian to Paleocene age, then by a flysch sequence and supra-Brèche mélanges (Dall’Agnolo 2000). This sequence is very similar to the Préalpes Médiannes plastique basin succession (fig. f3-D1), with similar hiatuses, but also with significant differences like an earlier apparition of the flysch and a more complete Couches Rouges sequence.

Interpretation: the Brèche nappe is a witness of the Alpine Tethys rifting active from the Liassic to

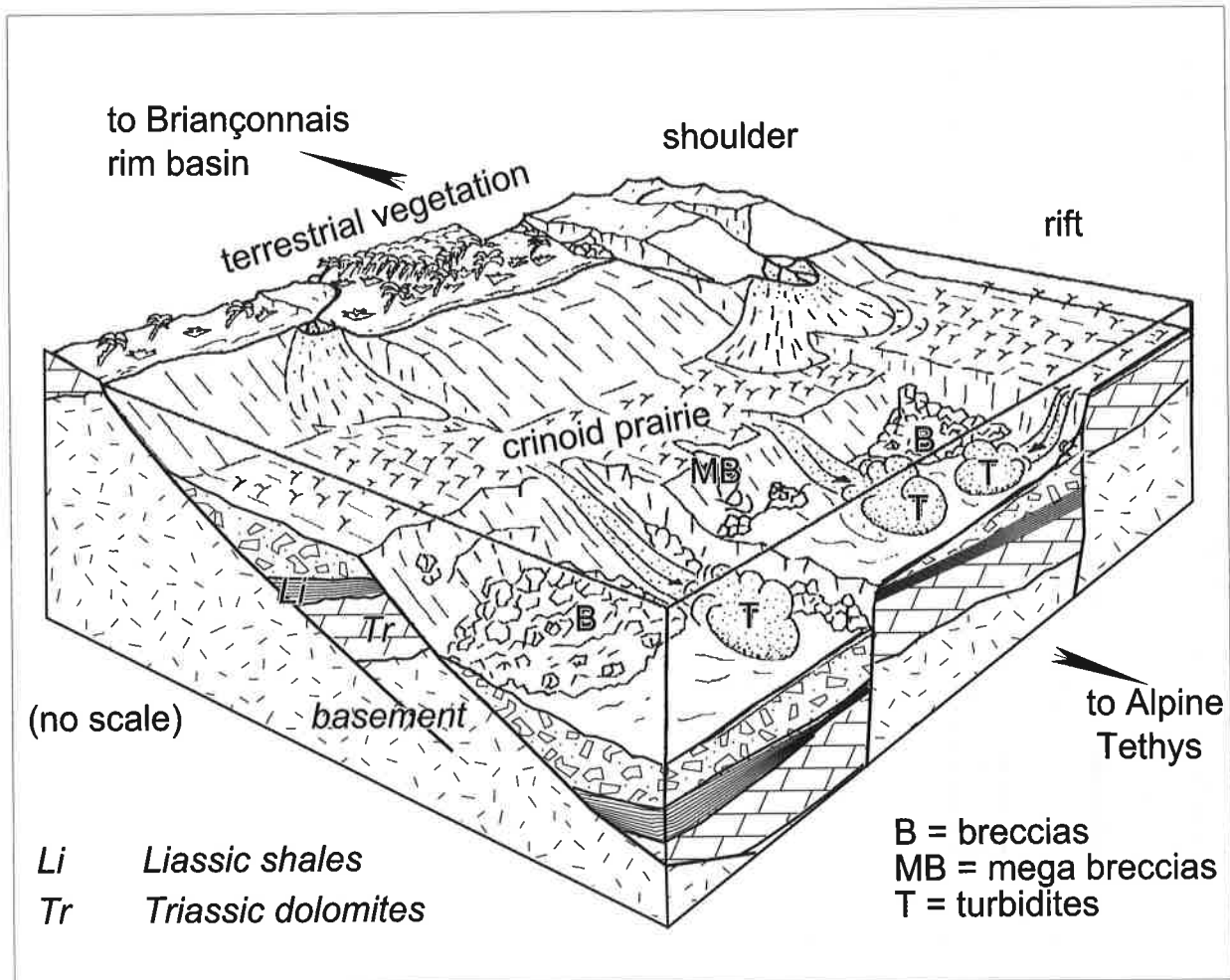


Fig. f3-D3 - Depositional environment of the Lower Breccia (Middle Liassic-Dogger), modified from Steffen et al. (1993)

the Early Dogger, deepening of the basin below the CCD in Callovian times, shows that we are dealing here with a major fault-bound basin close to the rift shoulder. Similar Cretaceous basin evolution between the Brèche domain and the Préalpes rim-basin allows to locate the Brèche tilted block basin on the southern border of the Briançonnais peninsula, both domains being separated by the rift shoulder from which most of the clasts are derived. The latter was drowned in Early Cretaceous times, after a renewed episode of breccia deposits. This event is regarded as a

witness of the detachment of the Briançonnais peninsula from Europe, together with the Iberic plate (see general introduction chapter 2.1.2). Important thickness variations are found between the Brèche nappe in the western Chablais lobe and the eastern Romandes lobe of the Préalpes (fig. ft3-D4), most of the Early Cretaceous missing in the latter. This could be related to the opening of the Pyrenean ocean and differential transportation/erosion of sediments in what is becoming an isolated Briançonnais peninsula between two oceanic domains (fig. 5).

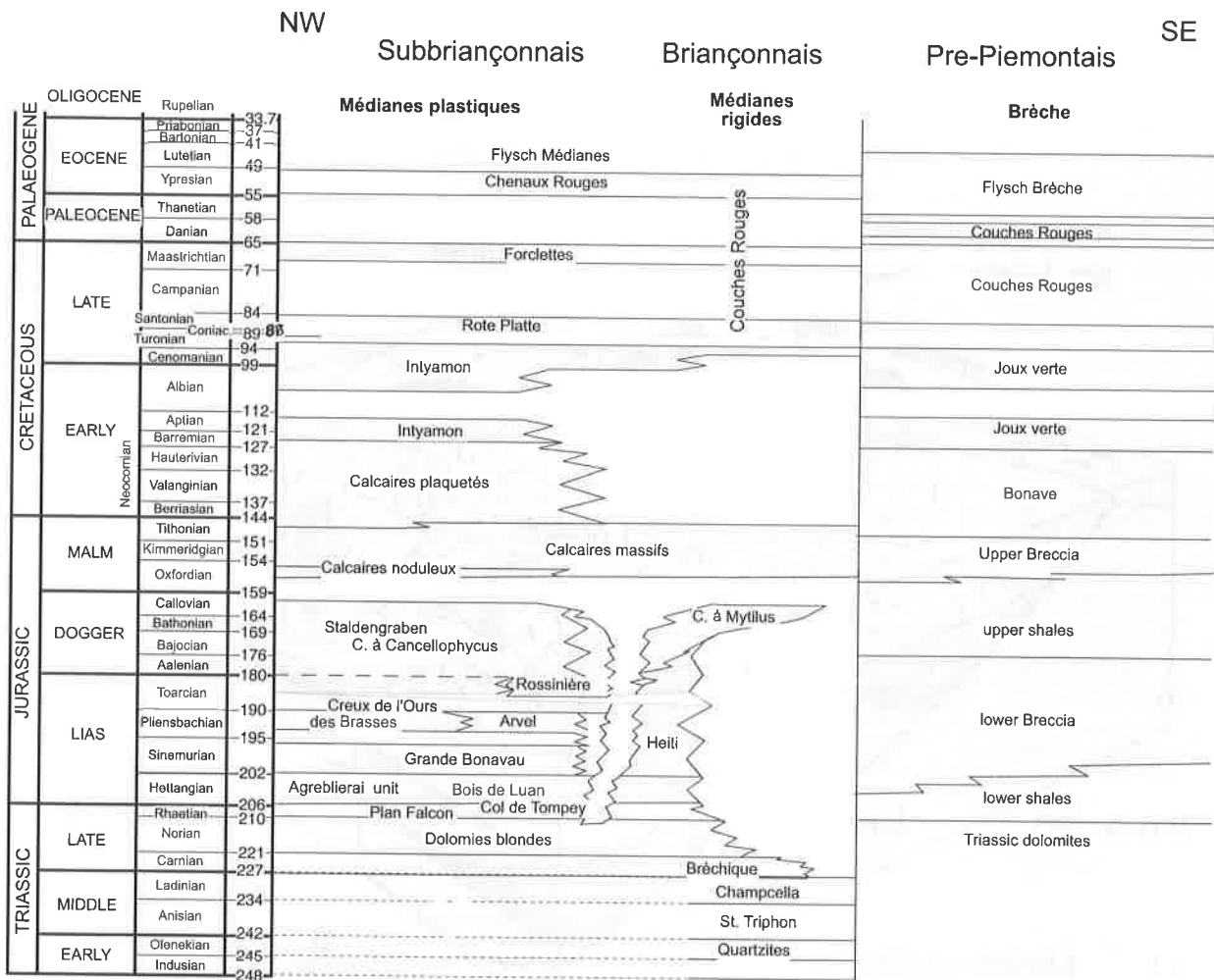


Fig. ft3-D4 - Stratigraphic chart across the Briançonnais peninsula, white represents erosion non deposition.

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I - The overall tectonic framework of the Penninic domain

Arthur Escher

The Swiss Alpine "Internides" are the result of a collision and subduction to the SE between the following 4 main paleogeographic continental and oceanic domains, from the NW to the SE:

- 1) The Valais trough oceanic crust.
- 2) The Briançonnais continental crust (possibly attached to the Iberic plate).
- 3) The Piemont oceanic crust.
- 4) The Austroalpine continental crust.

As shown in figure 28 and 29, remnants of each of these domains can be recognized in the Western Swiss Alps. The continent-derived units are mostly well preserved as basement and cover nappes, while for the oceanic crusts only ophiolite sutures and associated sediments remain. As can be expected, the latter often display accretionary prism characteristics.

The superposition and deformation of the first 3 of the above mentioned domains produced tectonic units which were defined as Lower (Valaisan) Middle (Briançonnais) and Upper (Piemontais) Penninic.

Remnants of the Valais trough oceanic crust and their associated sediments are mainly represented in the Sion-Courmayeur zone and in the Rosswald series. These units are all placed exactly at the boundary between the nappes derived from the European crust and those of Briançonnais origin. They must therefore contain the only remnants of the Valais trough oceanic realm. This is moreover suggested by the probable Cretaceous to Eocene age of the sediments, which could well correspond to that of the subduction and closure of the Valais trough (see general

introduction).

The basement nappes derived from the Briançonnais continental crust form a well individualized central unit in the Western Swiss Alps. It is limited to the NW by the Valais derived flysch sequences and to the SE by the Piemontais ophiolite nappes. This large basement unit is composed of the following nappes (from NW to SE) : The Zone Houillère, the Pontis nappe, the Siviez-Mischabel nappe, the Mont Fort nappe and the Monte Rosa nappe. Most of the Briançonnais Mesozoic and Tertiary sedimentary cover was separated from its original basement during early Alpine deformation, and translated to the NW. It actually forms the bulk part of the Prealpine nappes. Detachment from their basements took place along major ductile layers of middle or upper Triassic evaporites and dolomitic breccias (Masson 1972, Jeanbourquin 1988). These horizons permitted the "décollement" of the cover nappes toward the NW leaving behind them "basement nappes" which often include a remnant of Triassic cover rocks (see fig. 11 in general introduction).

Remnants of the Piemont oceanic domain are relatively well represented in the Western Swiss-Italian Alps: the Zermatt-Saas Fee and Antrona zones are probably both remnants of the NW part of the Tethyan oceanic crust, thrust on top of the Monte Rosa and Grand Paradis (Briançonnais) continental crust. The Antrona zone forms the main filling of an early synclinal core which separates the Monte Rosa from the Siviez-Mischabel nappe. The Tsate nappe is probably the remnant of an accretionary prism formed during the Late Cretaceous subduction of the Piemont oceanic domain (Marthaler & Stampfli 1989, Stampfli &

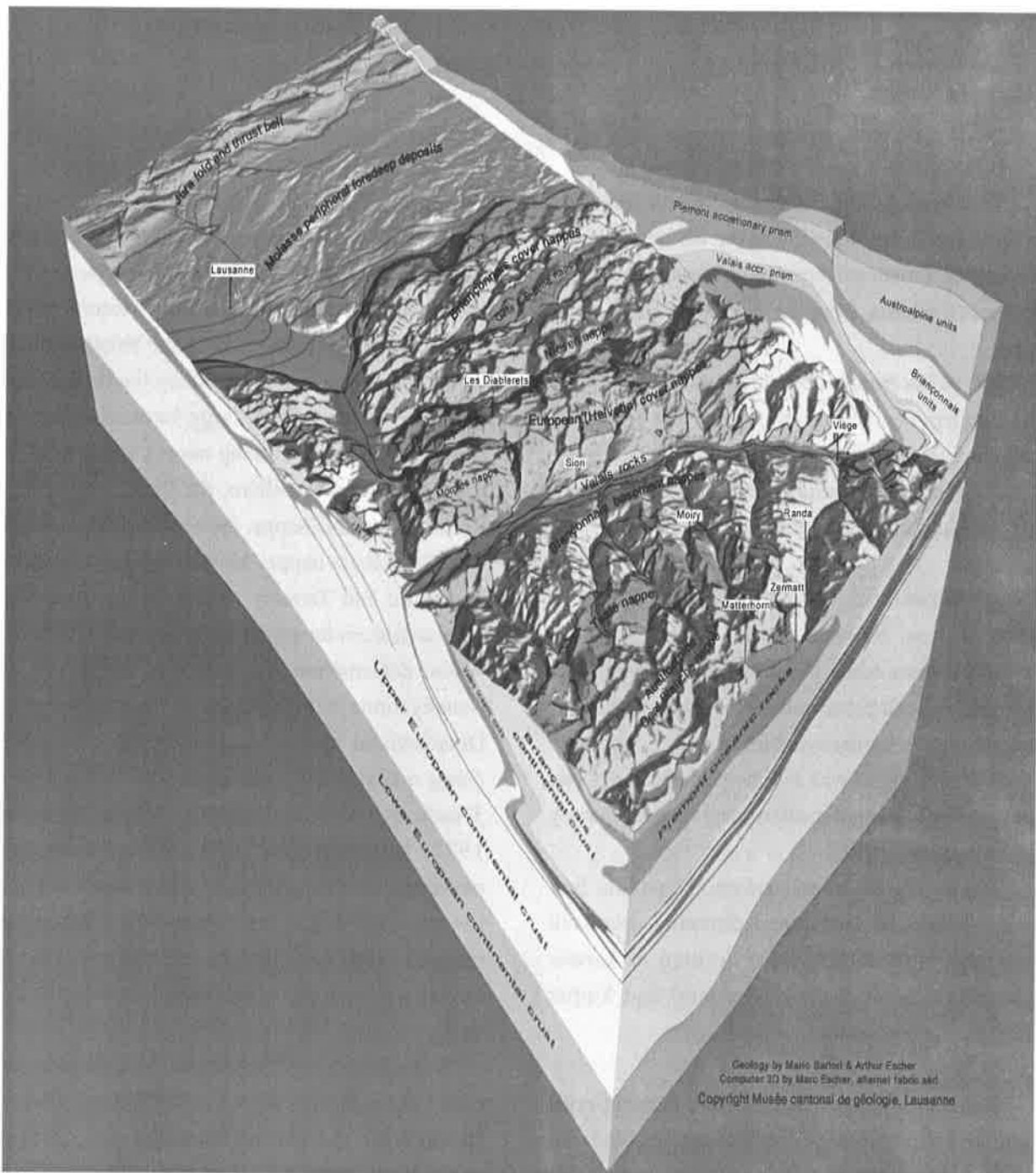


Fig. 28 - Block diagram showing the main geologic units in the internal Western Swiss Alps and their relation with the present topography (Escher & Sartori, Musée de Géologie, Lausanne, 2001). (A larger colored version is presented in the annex).

Marthaler 1990). Near its base it contains a thin but continuous unit of Triassic to Jurassic rocks, the Frilhorn nappe. Separating the Zermatt-Saas Fee zone from the Tsate nappe over a long distance, another thin Mesozoic sedimentary unit occurs, the Cimes Blanches nappe. During the Middle to Late Cretaceous, two main flysch units, the Gets and

Dranses nappes were deposited in front of the active accretionary Tsate prism. They form now the nappe Supérieure of the Prealpine domain (see field trip 6, and fig. 9 in general introduction).

South of the Piemont ophiolite suture zone, the Lower Austroalpine continental crust is

represented by the Dent Blanche "nappe" which is in reality a klippe, and the internal Sesia zone. Both are made of a superposition of two huge basement thrust nappes:

- a) The lower nappe is composed of the Arolla series, the Gneiss Minuti and the Eclogitic Micaschist Complex which are probably equivalent.
- b) The upper nappe comprises the Valpelline zone and the II-DK zone or "Seconda Zona Diorito-

Kinzingitica".

The superposition of the two main basement nappes probably took place during the Early Cretaceous (around 120 Ma), before the subduction of the Piemont oceanic crust.

All these units are considered as corresponding to the Western part of the Lower Austroalpine nappe system (see next chapter).

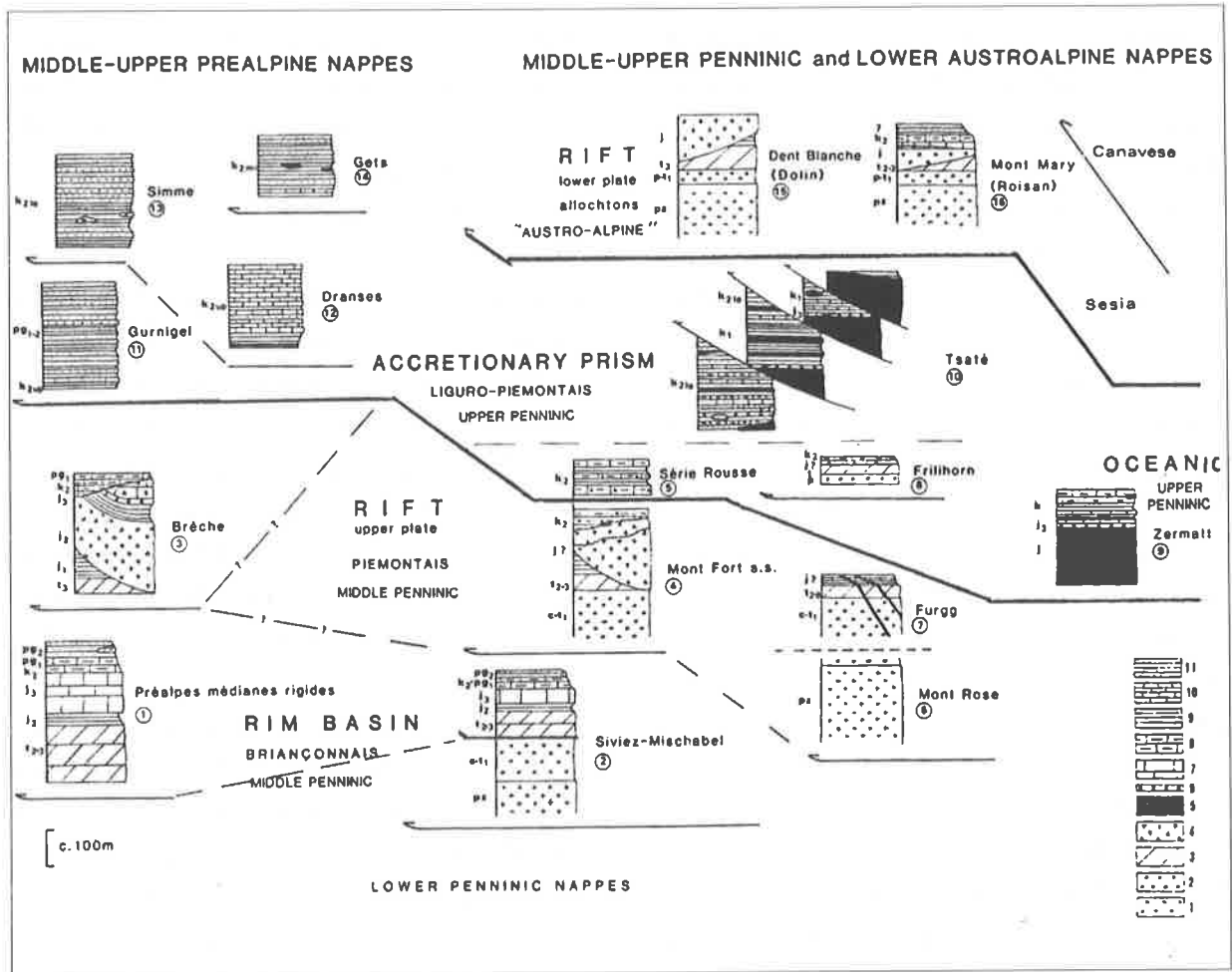


Fig. 29 - Stratigraphy and lithostratigraphy of the major units of the Internides modified from Stampfli and Marthaler (1990). Rock symbols: 1: continental crystalline basement (gneiss and amphibolite); 2: micaschist, prasinite, quartzite; 3: dolomite, limestone; 4: breccia; 5: oceanic basement (ophiolite); 6: radiolarite; 7: limestone or marble; 8: marlstone or phyllitic marble; 9: shale, calcschist; 10: calcareous flysch; 11: gritty flysch. Age symbols: pz, Paleozoic; c-t1, Carboniferous to Early Triassic; p-t1, Permo-Triassic; t2-t3, Middle to Late Triassic; j1,2,3, Early, Middle, Late Jurassic; kl, Early Cretaceous; k2lo, lower Late Cretaceous; k2up, upper Late Cretaceous; pg1, Paleocene; pg2, Eocene.

II - The Austroalpine system in the western Alps

François Bussy

The Austroalpine system represents the uppermost element of the Europe-vergent Alpine nappe pile (fig. 29 and Alps cross section in annex) and formerly the southern passive margin of the Alpine Tethys. It comprises several polycyclic basement units of pre-Alpine and/or Alpine age, and minor monometamorphic Permo-Mesozoic cover units. The Austroalpine system recorded a widespread acid-basic bimodal magmatism of Late Carboniferous-Early Permian age, ascribed to incipient stages of continental rifting, related to the Variscan cordillera collapse (see general introduction). The Alpine metamorphism is polyphased. P-T-t paths are characterized by classical clockwise loops with early eclogitic assemblages (550°C/16kbar) dated between 65 Ma (U-Pb on zircon and sphene) and 69 Ma (Lu-Hf on garnet) (see review in Bussy et al., 1998), followed by rather steep decompression paths. Peak temperatures reached <350 to 550°C depending on the tectonic units under consideration (see review in Venturini, 1995). Ar/Ar cooling ages on phengites from greenschist facies rocks yielded ages between 45 and 50 Ma (Venturini, 1995).

The Austroalpine system can be subdivided into an eastern and a western sector, the latter being present in the Aosta valley (Italy) and southwestern Valais (Switzerland). Many subdivisions have been proposed for the western Austroalpine system (see e.g. Venturini et al., 1994, Venturini, 1995 and references therein); it basically comprises the **Sesia zone**, the **Dent Blanche nappe** and a few minor klippen related to the latter (e.g. Mont Mary, Etirol Levaz, Pillonet, Monte Emilius, etc...). The following descriptions are essentially derived from Venturini (1995).

THE DENT BLANCHE NAPPE

The Sesia zone and the Dent Blanche nappe

s.l. broadly consist of similar lithologies, which are traditionally subdivided into a lower and an upper element. The lower element of the Dent Blanche nappe (*i.e.* the Arolla series) is composed of a polycyclic metamorphic basement and monometamorphic cover sequences. The basement rocks are derived from pre-Alpine lithologies (paraschists and gneisses) into which intruded several calc-alkaline stocks of diorite, granodiorite and granite, subsequently transformed into greenschist facies orthogneisses during the Alpine orogeny. One of these intrusions yielded a U-Pb zircon age of 289 ± 2 Ma (Bussy et al., 1998). The Arolla series also hosts large bodies of continental gabbroic rocks (Matterhorn and Mont Collon/Dent de Bertol, Dal Piaz et al., 1977) in mylonitic contact with the other lithologies. The primary mineral assemblages of these tholeiitic mafic masses are amazingly well preserved in the Mont Collon area, where olivine-bearing layered facies can be observed. Zircons from a pegmatitic gabbro and from an anorthositic dyke yielded U-Pb magmatic ages of 284.2 ± 0.7 and 282.93 ± 0.59 Ma, respectively (Monjoie et al., 2001), related to the opening of the Paleotethys back-arc basins (see general introduction, chapter 2.2).

Monometamorphic cover sequences of the lower element are represented by the Mont Dolin series. The latter is composed of Triassic and Liassic marbles, overlain by Late Liassic monogenic breccias (Ayrton et al., 1982), related to the opening of the Alpine Tethys.

The upper element of the Dent Blanche nappe (*i.e.* the Valpelline series) consists of a pre-Alpine high-grade basement only partially reworked by the Alpine metamorphism. It is made of granulite to high-amphibolite facies paragneisses (kinzigites), impure marbles, metabasites and rare mantle peridotites. Partial melting occurred locally, but

granitoid intrusions are typically absent.

The above-mentioned klippen or tectonic slices related to the Dent Blanche nappe essentially comprise the lower element (with some lithologic variations) with the exception of the Mont Mary klippe, which seems to include both tectonic elements.

THE SESIA ZONE

The Sesia (or Sesia-Lanzo) zone is an elongated and complex tectonic entity located between the Penninic domain to the north and the southern Alpine Ivrea-Verbanò zone, from which it is separated by the Insubric line. It can be subdivided in the same way as the Dent Blanche nappe in a lower element consisting of a polycyclic basement plus a monometamorphic cover, and an upper element consisting of high-grade rocks.

The polycyclic basement complex of the lower element mainly consists of high-grade metapelites (kinzigites) with minor bodies of metabasites and marbles, metamorphosed under upper granulite facies conditions. As in the Dent Blanche nappe, large granitic bodies intruded these lithologies and mafic masses are also found. Zircon U-Pb dating yielded an age of $293 \pm 1/-2$ Ma for the Monte Mucrone granite, 293 ± 3 Ma for the Monte Emilius granite and $288 \pm 2/-4$ Ma for the Anzasca gabbroic mass (similar to the Mont Collon gabbro of the Dent Blanche nappe) (Bussy et al., 1998). The lower element has been pervasively affected by the high-pressure Alpine metamorphism and variously retrogressed during subsequent greenschist facies conditions. Eclogitic micaschists are locally well preserved, whereas greenschist mineral assemblages without relicts of high-pressure minerals are characteristic of the so-called *Gneiss Minuti* complex (albite-two mica-epidote gneisses).

The monometamorphic cover complex of the lower element outcrops as a long and narrow band in the central Sesia zone. It consists of a sequence

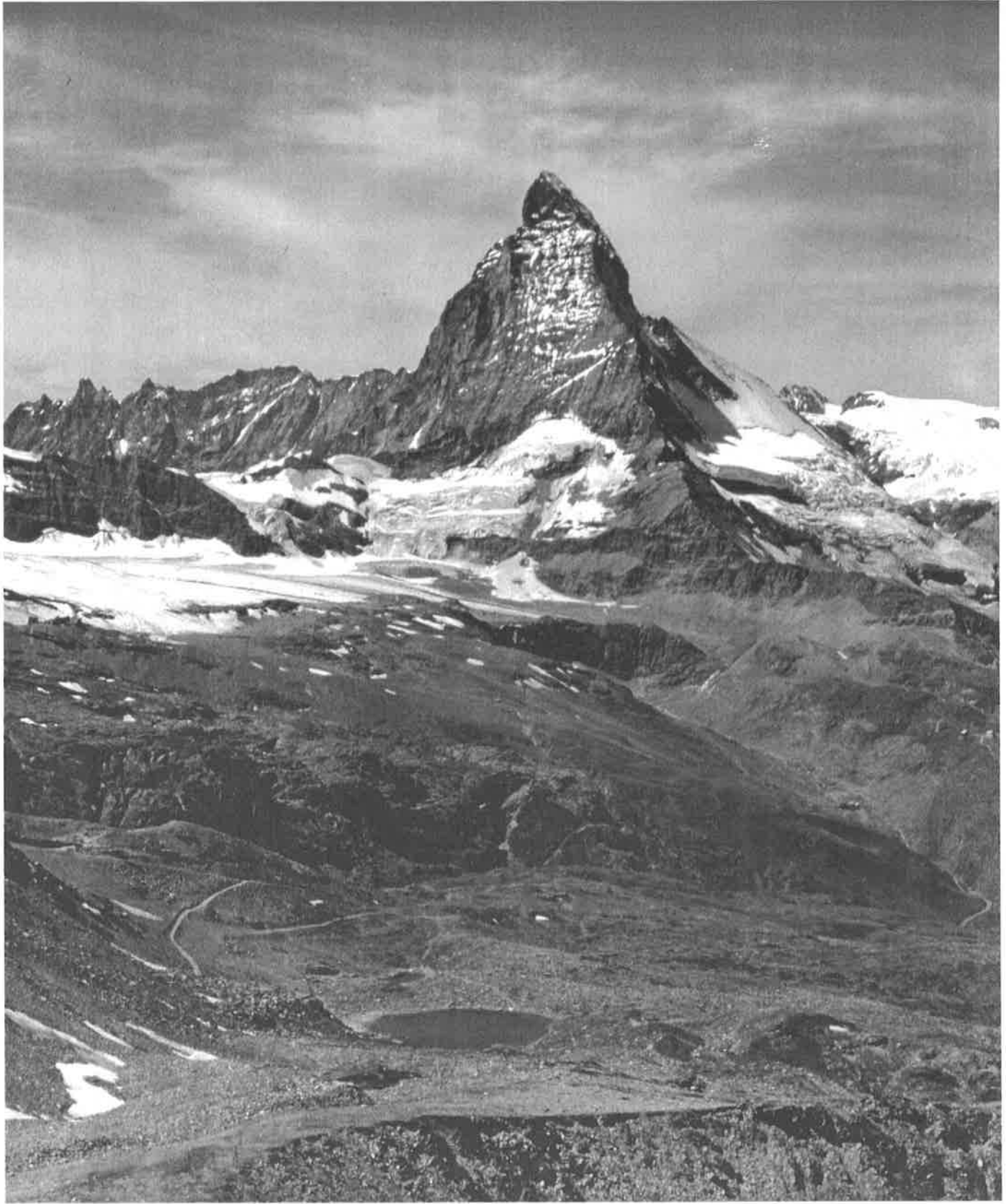
of extrusive basaltic rocks and associated sediments (Bonze unit, Venturini et al., 1994), and a sequence of supposedly Permo-Triassic to Triassic age (quartzites, dolomites, marbles, metapelites, Scalaro unit, Venturini et al. 1994).

As in the Dent Blanche nappe, the upper element is made of high-grade metamorphic rocks, essentially represented by kinzigites (biotite-garnet-sillimanite-plagioclase-Kfeldspar-bearing paragneisses), minor boudins of mafic granulites/amphibolites and impure marbles. As a consequence, this tectonic element is often referred as the *II DK* (*i.e.* the Second Diorite-kinzigite zone).

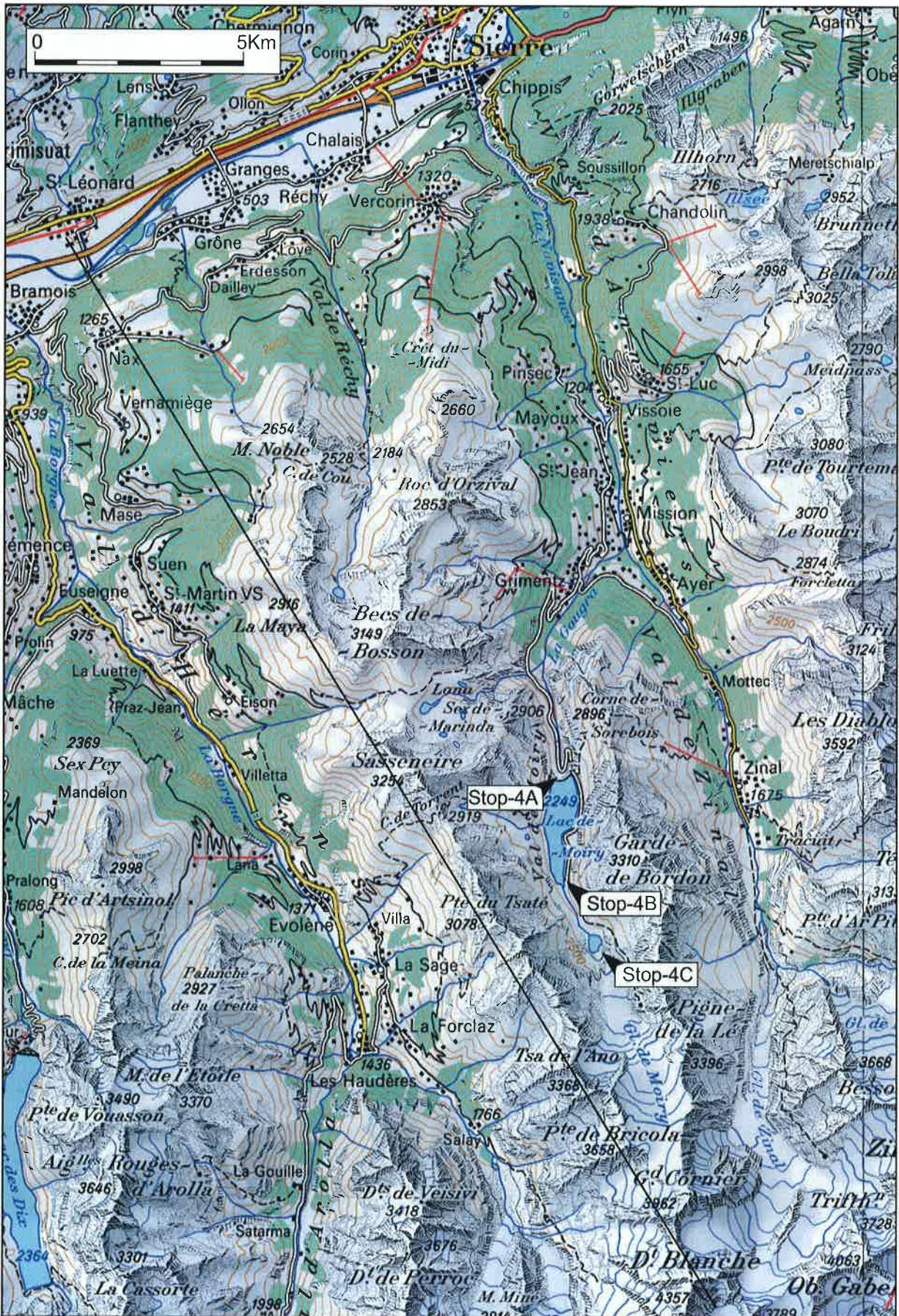
The lithologic associations observed in the Dent Blanche – Sesia system, in particular the succession in time of a Permo-Carboniferous bimodal magmatism, mesozoic basic volcanism, Liassic breccias and Mn-rich calcschists, then high-pressure metamorphism around 65-70 Ma record the breakdown of the Variscan continental crust, the subsequent evolution of this area as the southern passive margin of the Alpine Tethys, and finally its involvement in the Alpine orogenic processes.

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The Matterhorn



Cross section from the Briançonnais Siviez-Mischabel nappe, through the Piemont Tsate nappe, to the Austroalpine Dent Blanche unit (Moiry region)

Arthur Escher and Michel Marthaler

I: INTRODUCTION (fig. 30 and 31)

1 - THE SIVIEZ-MISCHABEL NAPPE

The Siviez-Mischabel nappe forms, south of the Rhone valley in Valais, a huge nappe structure representing the central unit of the Grand Saint-Bernard super-nappe (Escher 1988). During the past 20 years this unit has been mapped and investigated in detail from the Zermatt region to the Val de Bagnes (Bearth 1980, Burri 1983, Marthaler 1984, Thélin 1987, 1989, Sartori 1987, 1990, Sartori & Thélin 1987, Thélin et al. 1993). The results of these investigations can be summarised as follows:

The Siviez-Mischabel nappe has the geometry of a very large recumbent fold with an amplitude of more than 35 km. After its formation and emplacement it was backfolded in its internal part, which resulted in the spectacular Mischabel backfold, analysed by Milnes *et al.* (1981) and Müller (1983).

The core of the nappe is made of polymetamorphic paragneisses and micaschists of Paleozoic age, containing a wide variety of metamagmatites (calc.alkaline granitoids, gabbros, pyroxenites and volcanites). Eclogites and retroeclogites are locally present, usually associated with banded amphibolites. They show HP mineral assemblages which are overprinted by an amphibolite facies event of probable Variscan age (Thélin et al. 1990).

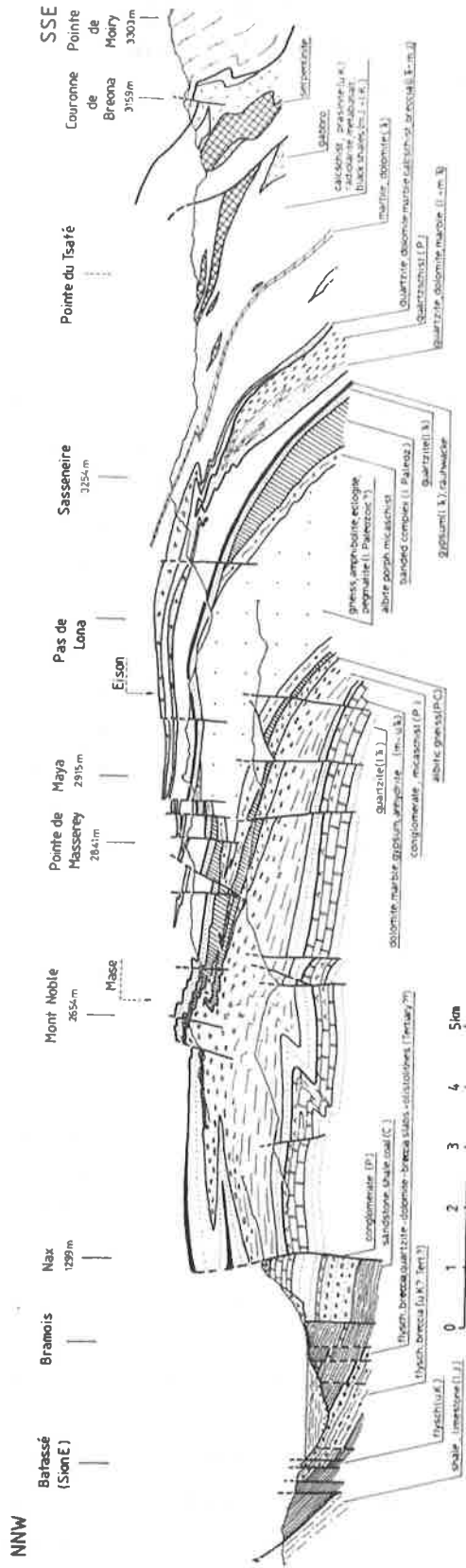
Surrounding the Siviez-Mischabel basement core, a metasedimentary cover can be followed from the normal flank throughout the front of the nappe to its inverted limb (fig. 30). This cover is made in most parts of Late Carboniferous, Permian

and lower Triassic schists, conglomerates, quartzites and evaporites (**stop 4A**). Only in its Eastern and internal part, in the Barrhorn zone, the normal flank displays a complete Briançonnais-type cover sequence up to the Eocene (Ellenberger 1953, Bearth 1980, Sartori 1990).

The part of the cover rocks (middle Triassic to Eocene) missing in most parts of the normal and inverted limb, was stripped from its basement before or at the same time as the nappe was formed, and was tectonically translated towards the NW in the Prealps (see general introduction, fig.11). This separation was made possible by the presence of an extensive Triassic evaporite layer, resting on lower Triassic quartzites, which covers Paleozoic gneisses and micaschists.

2 - THE TSATE NAPPE

The Tsate nappe forms a large unit, corresponding to the oceanic part of the Combin zone *s.s.* as defined by Dal Piaz (1965) and Bearth (1967). It consists of a thick pile of imbricated metasediments and ophiolites in an apparently chaotic pile. Metasediments consist mostly of Cretaceous calcschists and black shales, with locally Jurassic quartzites (meta-radiolarites ?) and lower Cretaceous siliceous marbles and dolomitic breccias. Two types of basaltic rocks can be recognized, strongly deformed and flattened pillow lavas of probable Jurassic and Cretaceous age and basaltic sands metamorphosed into prasinites (**stop 4B**). The latter are probably of Cretaceous age and occur as stratigraphic layers within the



VAL D'HERENS CROSS-SECTION

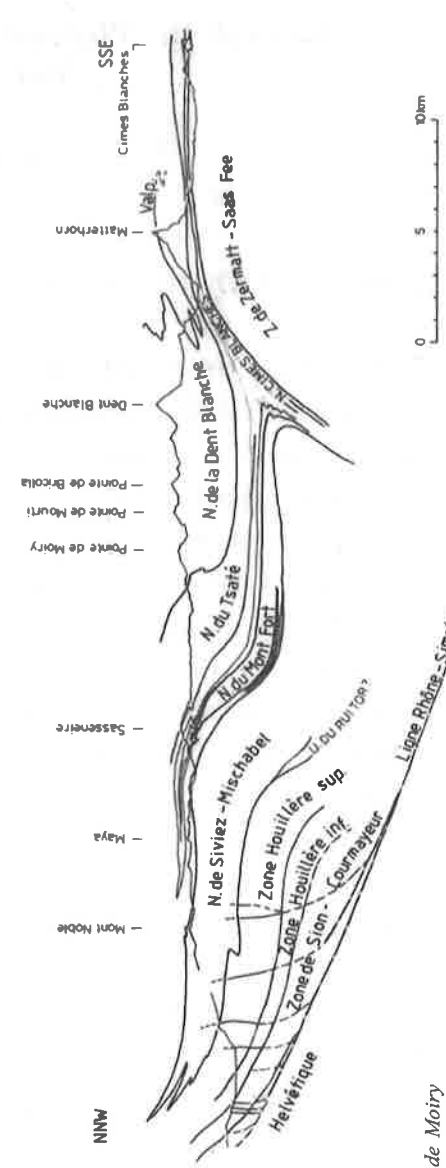


Fig. 30 - Geologic profile from Bramois(Rhône Valley) to the Pointe de Moiry showing the structure of the internal Swiss Alps (by Sartori 1993).

calcschists. The ophiolitic rocks are represented by meta-basalts, meta-gabbros and meta-peridotites. An intense tectonic imbrication took place during subduction and corresponds logically to the internal structure of an accretionary prism (Sartori 1987, Marthaler & Stampfli 1989). The metamorphic history of the Tsate nappe is characterized by an early, relatively high pressure event resulting in the formation of greenschist-blueschist metamorphic assemblages (Dal Piaz 1976, Caby 1981, Ayrton et al. 1982, Pfeiffer et al. 1991). It was followed by a strong pervasive greenschist facies episode.

3 - THE DENT BLANCHE NAPPE

Large parts of the Dent Blanche nappe consist of pre-Alpine continental basement rocks (Arolla series) represented mainly by orthogneisses derived from Late Hercynian granitoids and minor amounts of siliceous paragneisses. Layered gabbros of Late Permian age occur in the Arolla series. The latter forms the lower part of the Dent Blanche and the Pillonet klippen, while the Gneiss Minuti occur in

the external part of the Sesia zone. Both units underwent extensive greenschist facies crystallisation during Tertiary Alpine metamorphism (Hunziker 1974, Lattard 1974, Mazurek 1986). Well preserved relics of high-pressure parageneses of presumed Cretaceous (eoalpine) age are only present in the Gneiss Minuti. It can however not be excluded that part of the greenschist facies metamorphism in the Arolla series could be of eoalpine age as the Dent Blanche klippe represents the frontal part of the eoalpine main nappe (Ayrton et al. 1982, Oberhänsli & Bucher 1987, Pognante et al. 1988, Pognante 1989, Canepa et al. 1990). According to Dal Piaz et al. (1972), Lardeaux et al. (1982) and Vuichard (1989), a high pressure mylonitic zone forms the contact between the Arolla gneisses and the overlying Valpelline series in the Dent Blanche klippe. This most likely implies that the tectonic superposition of both units was an early Alpine (Cretaceous) event. Recent radiometric dating by Cosca (in prep.) gives an Early Cretaceous age (120 Ma) for this mylonite contact.



Fig. 31 - Aerial view of the Val d'Hérens and the Sion region. (Marthaler, 2001).

A-B : Trace of the profile of Fig. 30.

- 1) Sion-Courmayeur zone
- 2) Zone Houillère and Pontis nappe
- 3) Siviez-Mischabel nappe
- 4) Tsaté nappe
- 5) Dent Blanche nappe

II: EXCURSION OUTCROPS

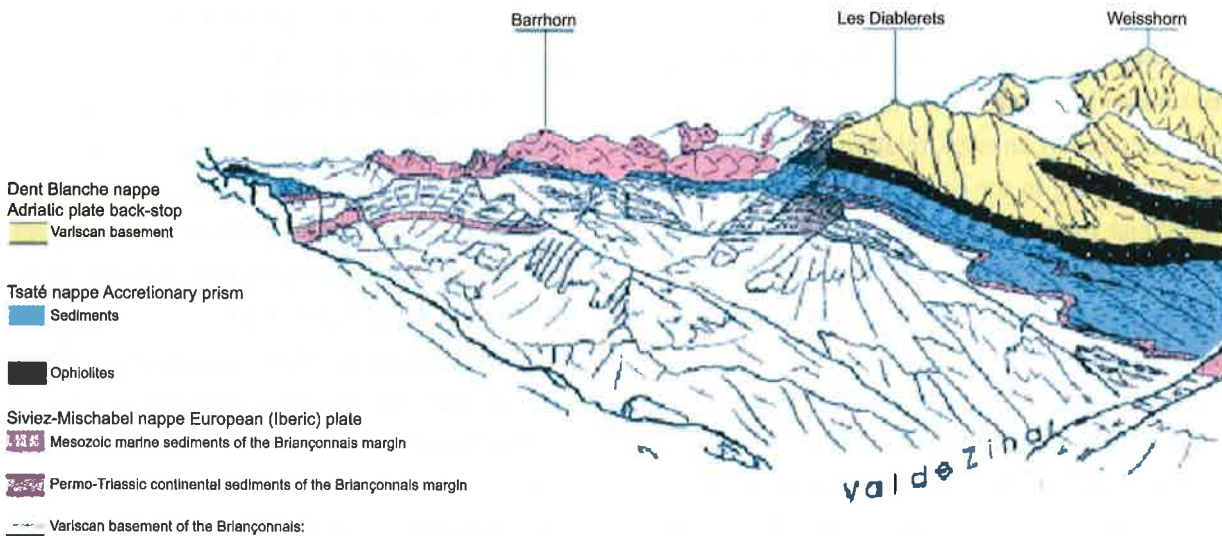


Fig. ft4-A - Geologic panorama of the Val de Moiry and surrounding mountains. Note the position of the dark oceanic Tsate rocks separating the underlying Briançonnais units from the overlying Dent Blanche rocks (Drawn by Sartori, in Marthaler, 2001).

II: EXCURSION OUTCROPS

Stops are located on the topographic map on the opposite page, a structural map corresponding to the same area is found in annex (MAP 3).

STOP 4A - WEST OF THE MOIRY-LAKE DAM (FIG. FT4-A)

Upper part of the normal limb of the Siviez-Mischabel nappe showing the Triassic evaporites in stratigraphic continuity with the Permian and older basement rocks. The originally overlying Mesozoic rocks were tectonically translated towards the NW in the Prealps. The Triassic evaporites are tectonically overlain by the Cretaceous calcschists belonging to the overriding Tsate nappe. This nappe contain refolded slices of the Cimes Blanche nappe (Internal Briançonnais), visible 200 m toward the south along the lake pass (dolomitic marble and breccia of Late Triassic and

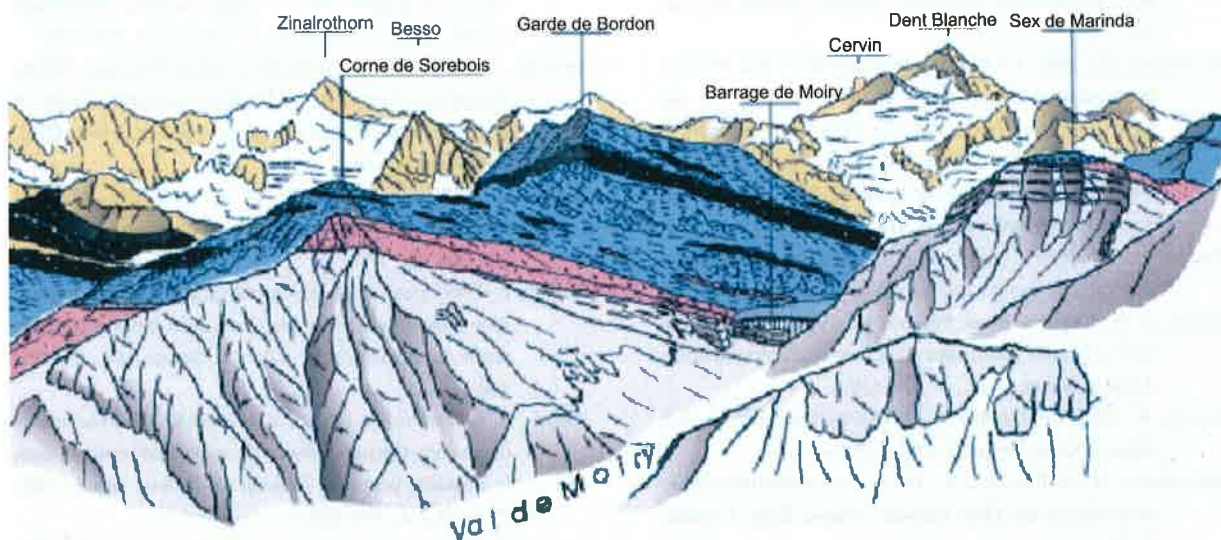
Early Jurassic ages).

STOP 4B - EAST OF MOIRY LAKE (FIG. 30)

The outcrops show the complex imbricate mixture of calcschists, metabasalts and prasinites belonging to the oceanic Piemont Tsate nappe. Locally strongly deformed pillow structures can be recognised.

STOP 4C - SOUTH OF MOIRY LAKE

A large block of meta-peridotite containing breccias and dykes can be observed outcropping just below the front of the Moiry glacier. Towards the SE, one can observe the thrust contact between the Dent Blanche Arolla gneisses and the underlying Tsate rocks. It has been backfolded and tectonically reactivated along a younger normal fault.



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Passing through the pre-alpine basement of the Briançonnais from Visp to Randa (Middle Penninic Units, Eastern Wallis, Switzerland)

I: INTRODUCTION

1 - THE PRE-ALPINE BASEMENT OF THE BRIANÇONNAIS FROM VISP TO RANDA (MIDDLE PENNINE UNITS, EASTERN WALLIS, SWITZERLAND)

Philippe Thélin

1.1 INTRODUCTION

The Grand Saint-Bernard tectonic unit constitutes a “super-nappe”, which belongs to the Western Middle Penninic Alps (fig.13) and corresponds paleogeographically to the Briançonnais domain (Argand 1911 and 1934; Bearth 1963; Müller 1982; Escher 1988; Escher et al., 1988 and 1993; Sartori 1990; Thélin et al. 1993; Stampfli 1993; Sartori and Marthaler 1994; Chessex 1995; Gouffon and Burri 1997; Steck et al. 1999 and 2001). This “super-nappe” consists of four nappes from NW (external) to SE (internal): the “Zone” Houillère, the Pontis nappe, the Siviez-Mischabel nappe and the Mont Fort nappe (fig. 13). During the present field trip, from Visp in the Rhone valley to Randa in the Matter valley (fig. 32), we will mainly cut through the basements of the Pontis nappe (Ahorn outcrop described under **stop 5C**) and of the Siviez-Mischabel nappe, especially its inverted limb (Randa outcrop described under **stop 5D**). The main goal of this field trip is to familiarize participants with the lithology of these pre-Alpine basements, so that the complex lithostratigraphy of the Middle Penninic domain in Valais may become clearer and so that specific topics such as pre-Alpine magmatism, pre-Alpine sediments, pre-Alpine metamorphism can be approached. For all these questions, the main reference remains the book “Pre-Mesozoic

Geology in the Alps” by Von Raumer and F. Neubauer (1993a).

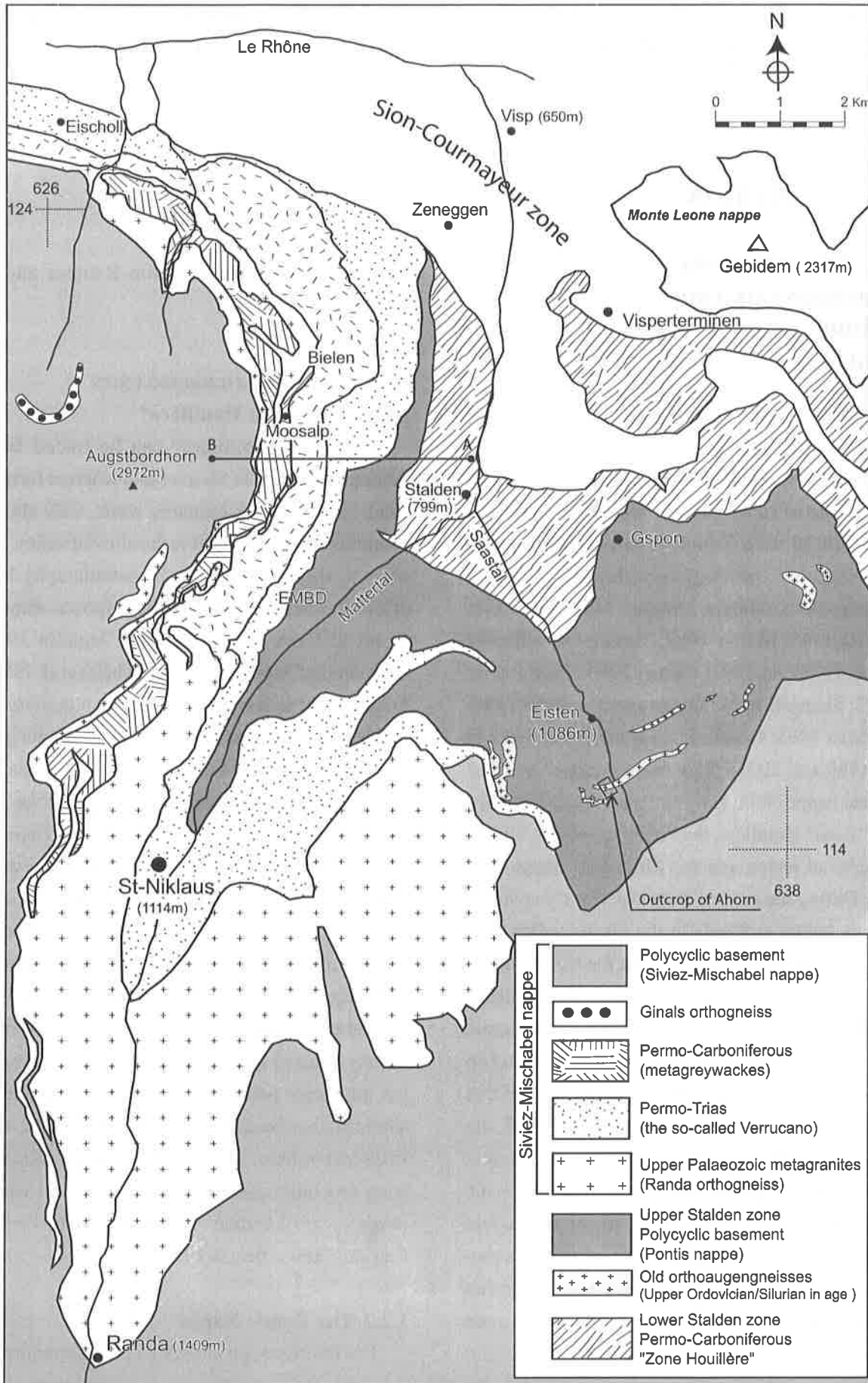
1.2 TECTONOSTRATIGRAPHIC UNITS

1.2.1 The “Zone Houillère”

The “Zone Houillère” can be traced from France and Italy into Switzerland where it forms a band, one to two kilometers wide, with almost continuous outcrops right to the alluvial valley. We will only describe shortly its lithostratigraphy East of the Turtmantal (general descriptions may be found in Burri 1975, Burri and Jemelin 1975, Desmons and Mercier 1993 and Thélin et al. 1993). In fact, in its eastern part, the “Zone Houillère” is divided into two parts. The external part, with grey quartzites in metre-thick layers and black shales, decreases from a thickness of 200m in the Turtmantal to a trail of lenses in the Vispental which ceases to be distinguishable from the shaly levels of the Valais domain (Burri 1979). A thin Triassic band separates this external part from an internal one which thickens eastwards (the “lower Stalden zone”, Escher 1988 and figures 32 and 33). The latter is comprised of thin level of graphitic schists and grey quartzites, locally conglomeratic, but the major part of this series is composed of albitic gneisses and chloritico-sericitic schists, with little carbonate, and light-coloured, locally conglomeratic quartzites. This series, in which prasinities are common, is considered to be Permo-Carboniferous (Bearth 1972 and 1978).

1.2.2 The Pontis Nappe

The Pontis nappe includes a polymetamorphic



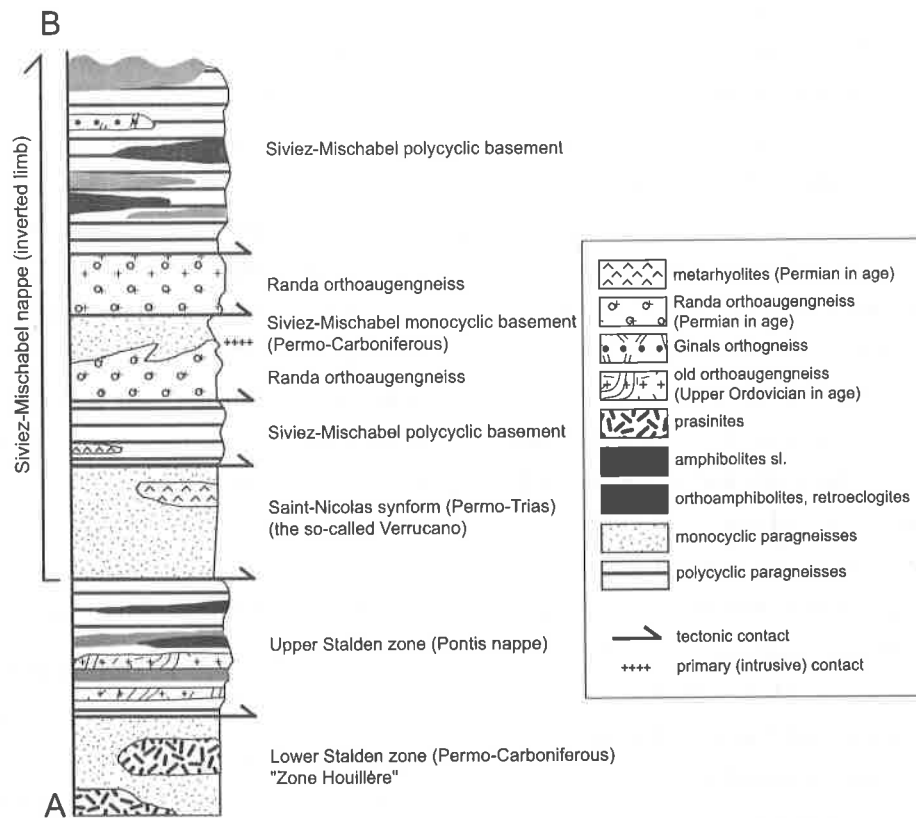


Fig.33 - Synthetic lithostratigraphic cross-section of the Middle Penninic units between ca. Stalden (799m) and Augstbordhorn (2972m). The vertical scale is arbitrary.

basement, a monometamorphic Permo-Carboniferous basement and Triassic fragments of a Subbriannonais cover. The polymetamorphic basement which forms the main part of the nappe is fragmented, due to Alpine tectonics, into three distinct units: from SW to NE, the Ruitor Zone (Grand Saint-Bernard area), the upper Stalden Zone (Mattervalley and Saasvalley) and the Berisal Unit (Simplon area)-fig. 32. (for more geological details, see Burri 1983b; Stille and Oberhänsli 1987; Thélín et al. 1993; Gouffon and Burri 1997; Giorgis et al. 1999; Steck et al. 1999; Steck et al. 2001). Metasediments are mainly characterized by very thick, monotonous, strictly siliceous, detrital rocks. Different varieties of paragneisses may however be distinguished, in the upper Stalden zone for instance, expressing contrasting conditions of detrital sedimentation of unknown age, predominantly

composed of sandstones, greywackes and conglomerates. The pre-Alpine assemblage includes muscovite, brown biotite, staurolite, almandine with prograde zoning and plagioclase (almandine-amphibolite subfacies). A special attention is devoted under part 1.3.2 to the Mont Mort metapelites (Ruitor zone) which offer an exceptional metamorphic window for perfectly preserved Variscan assemblages.

This polymetamorphic basement exhibits a considerable abundance and a wide variety of ultramafic and mafic magmatic rocks, essentially massive amphibolites with large amphiboles (pargasite-hornblende), garnet and/or plagioclase bearing amphibolites, and banded amphibolites (fig. 33). They form lenticular boudins or continuous concordant layers with a sill-like aspect, occasionally of kilometric extent. The host rocks

fig.32 - Simplified geological sketch map of the southern Visp area (Mattervalley and Saasvalley). Modified after Thélín 1987. A-B: location of the lithostratigraphic cross-section (Fig. 33).

are generally metaclastics (sandstones, greywackes) and they are often closely associated with the "old" augengneisses discussed under 4.1. Some boudins (metric in size) reveal an eclogitic metamorphism, retrogressed in amphibolite facies as discussed under 1.3.1. Abundant data are to be found in Stille (1980), Stille and Tatsumoto (1985), Stille and Oberhänsli (1987), Thélin (1989) and Sartori et al. (1989). From radiometric data (Rb-Sr, Sm-Nd) and REE analyses, Stille and co-workers suggest the following petrogenetic interpretations: (a) the mafic and ultramafic rocks are of magmatic origin and constitute a series with a calc-alkaline trend, (b) the banded and the plagioclase-bearing amphibolites are derived from a tholeiitic to dacitic sequence, emplaced in Proterozoic times (ca. 1020-1070Ma.); (c) the massive amphibolites with large amphiboles could represent pyroxenites or picritic-derived volcanics ($\text{SiO}_2 < 48\text{wt.}\%$), emplaced around 475Ma.

1.2.3 The Siviez-Mischabel Nappe

The Siviez-Mischabel nappe constitutes the main part of the Grand Saint-Bernard nappe (fig. 13, 32 and 33). It forms a fold-nappe of more than 40 km extent with a polymetamorphic core, a monometamorphic basement (Carboniferous ? - Early Triassic) and a Mesozoic to Paleogene cover. Investigations by Bearth (1963), Thélin (1987), Thélin et al. (1990 et 1993), Marthaler (1984), Sartori (1990), Sartori and Thélin (1987), Burri (1983a et b), Escher (1988) and Escher et al. (1988) demonstrate the existence of a normal limb, a frontal zone and an inverted limb. All these works confirm the lateral continuity (from the Aostvalley to the Saasvalley, *ie.* from the Grand Saint-Bernard pass area till nearly the Simplon pass area) of its polymetamorphic basement. It could be divided into two distinct lithostratigraphic entities (for more details, see Fig.4 in Thélin et al. 1993):

a) The Ergischhorn unit (fig. 33) is the main component of the basement and is predominantly

composed of monotonous paragneisses whose composition and aspect vary gradually, without any clear-cut boundaries. This unit structurally forms the homogeneous core of the nappe. It is composed of siliceous clastic metasediments with various types of associated amphibolites. Most typical are grey-greenish micaceous gneisses and schists with thin quartz bands or nodules, aphanitic, massive gneisses and well banded gneisses. The pre-Alpine more or less retrogressed assemblage includes zoned almandine, muscovite, brown biotite, plagioclase and K-feldspar. These very thick metasediments (ca. 1000m) are derived from immature, often coarse clastic sediments such as micaceous sandstones, arkoses and conglomerates.

b) The "stratiform" Barneuz unit overlies the latter and exhibits contrasting lithologies with marker horizons of very great regional extension, from bottom to top: (1) banded amphibolites; (2) micaschists; (3) banded complex. The metasediments of Barneuz unit are essentially pelitic. They are especially well developed in its middle part, forming augen-micaschists with albite porphyroblasts (Sartori and Thélin, 1987). Complete Alpine recrystallisation/neof ormation of greenschist grade is pervasive. The marker horizon reaching a maximum thickness of 250 m and an east-west extension of at least 100 km could be derived from semipelites. Ultramafic and mafic rocks have been investigated, in particular in the Turmanntal by Sartori (1990), Thélin et al. (1990), Rahn (1991), Eisele et al. (1997). It is worthwhile to mention that lenses of eclogites are to be found within the lower level, the so-called banded amphibolites. These eclogites form lenses envelopped in retroeclogites and in amphibolites. The eclogitic assemblage with omphacite-garnet-phengite, namely 3T-polytype, predates an amphibolite facies assemblage. This fact leads to the conclusion that this eclogitization must be attributed to a pre-Alpine HP metamorphism.

c) Monometamorphic rocks, mainly metasedimentary, envelop the polymetamorphic basement and have been described to some extent by Trümpy (1966), Sartori (1990) and Thélin et al. (1993). Presumably Carboniferous and Permian strata are distributed in distinct and discontinuous units, while the Permo-Triassic quartzites form a regular, continuous envelope (fig. 32 and 33). Three distinct sedimentary series are defined by Thélin et al. (1993) as follows:

1. Pelites, sandstones and greywackes (Carboniferous?): these form a characteristically rust-coloured series within the eastern part of the normal limb of the nappe; graphite, pyrite and ankerite-bearing schists, albitic schists and quartzites are the main components. There are abundant mafic rocks (prasinities, metagabbros) and acid rock (quartz metaporphyries).
2. Arkoses, conglomerates, greywackes (Late Carboniferous-Permian?): this series is composed of micaschists, chloritic schists, albitic gneisses and metaconglomerates. Scarce and thin layers of quartz porphyries are recognizable.
3. Conglomerates, arkoses and siliceous sandstones (Late-Permian to Early Triassic): the classic association of Briançonnais Verrucano type conglomerates (Trümpy 1966) and white tabular quartzites form a virtually continuous horizon around the nappe. The conglomerates with pink quartz pebbles grade into the white quartzites. Thin levels of quartz porphyries appear within the conglomerates. This series lies unconformably on all the other components of the basement.

1.3 PRE-ALPINE METAMORPHISM

Alpine greenschist facies overprint, essentially of Tertiary age, affects all the visited units (for a review, see Part I; Thélin et al. 1993; Markley et al. 1998; Steck et al. 1999 and 2001). The most representative Alpine minerals are:

Felsic lithology: Grt-Gln-Qtz-Ab-Bt/Chl-Phe-Ep-

Stp

Pelitic lithology: Grt-Gln-Cld-Phe-Bt/Chl-Grt-Qtz-Ab-Ep-Stp

Mafic lithology: Grt-Gln-Hbl/Act-Grt-Bt/Chl-Ab-Ep-Sph-Chl-Stp-Cal.

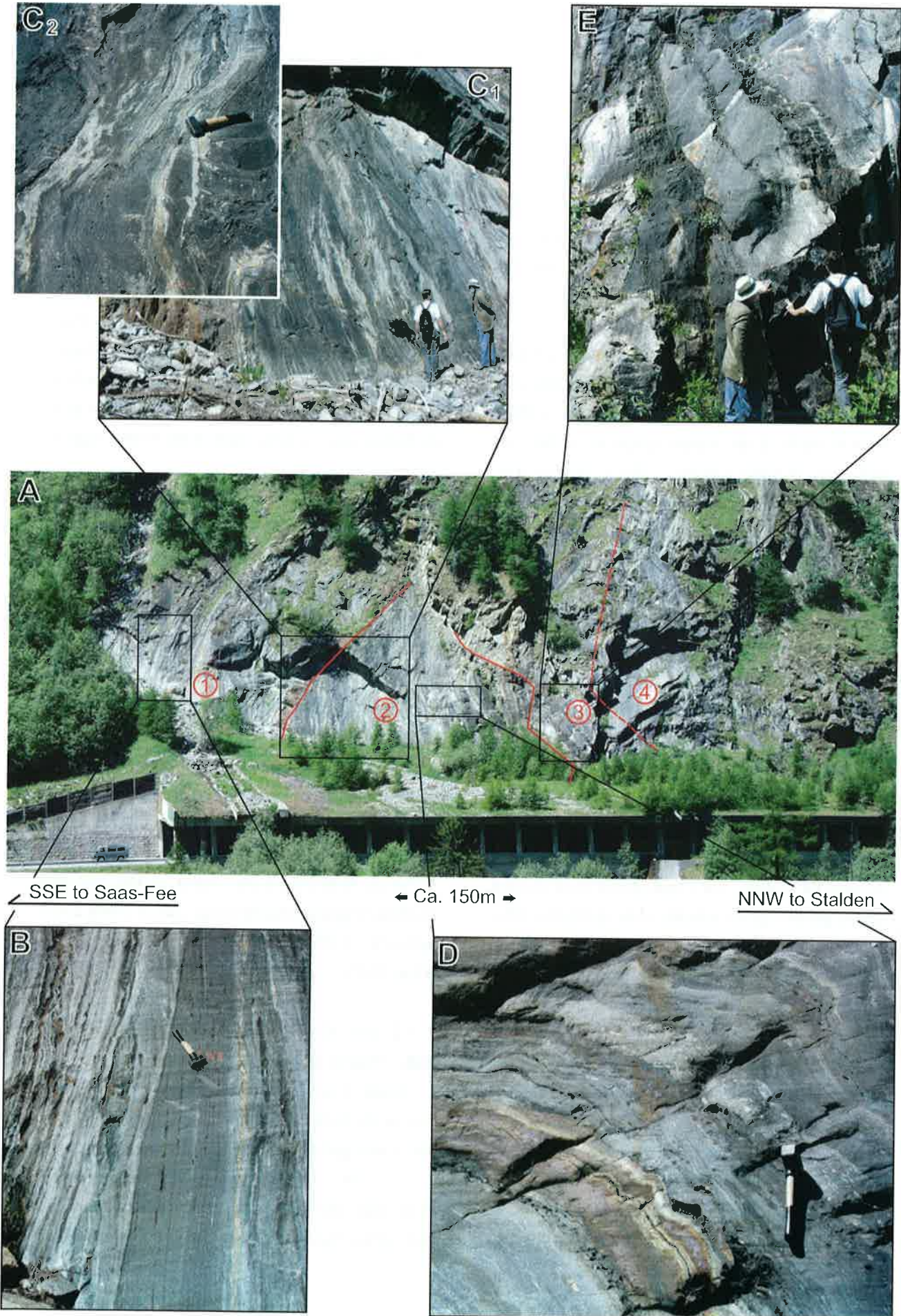
Pre-Alpine almandine amphibolite facies can be uniformly recognized (Bearth 1963; Thélin et al. 1993) within the polymetamorphic rocks (amphibolites, paragneisses, orthogneisses). Two spectacularly preserved metamorphic windows allow to observe the pre-Alpine P-T events. They are the eclogites of the Siviez-Mischabel nappe (Thélin et al. 1990 and Eisele et al. 1997) and the Mont Mort metapelites (Giorgis et al. 1999), within the Ruitor zone (western part of the Pontis nappe).

1.3.1 The eclogites of the Siviez-Mischabel nappe

These eclogites provide the following data on the polymetamorphic history. The eclogitic omphacite-almandine (pyrope-rich)-phengite predates an amphibolite facies paragenesis with pargasite-oligoclase-zoisite, itself followed by a greenschist facies retrogression (Tertiary Alpine event). The chemistry of the garnet, omphacite and phengite assigns the P-T conditions to a bracket of ca. 650°C and 15-20kb. An analogy can be proposed with the Ordovician subduction cycle documented by relicts of MORB eclogites in all External Massifs, in particular in the Aar and Gotthard massifs, where eclogitization is bracketed between 470Ma and 440Ma (Abrecht 1994; Biino 1995).

1.3.2 The Mont Mort metapelites (Ruitor zone, Pontis nappe)

Located near the Grand Saint-Bernard pass, within the Ruitor zone (Pontis nappe), the Mont-Mort metapelites are one of the best preserved relic of the Variscan unit in the Briançonnais basement. These micaschists crystallized during a poly-phase metamorphic cycle, under amphibolite



facies conditions. Mineral parageneses and geothermobarometric calculations indicate a two-stage evolution. Stage (1) (550-600°C and 5-8kb) is documented by assemblages of zoned garnet, staurolite, kyanite (?), biotite, muscovite, quartz and plagioclase. Stage (2) (550-600°C and 2kb) is illustrated by assemblages of andalusite, sillimanite, muscovite, biotite. This metamorphic evolution is characterized by a nearly isothermal decompression path, terminating with the formation of andalusite-bearing veins. Monazites yield a U-Pb age of 330Ma and muscovites yield a $^{40}\text{Ar}/^{39}\text{Ar}$ age of 290-310Ma corresponding to Variscan metamorphism, and yield an estimate of the time interval between the thermal peak and the retrogression stage.

1.4 “OLD” AND “YOUNG” ORTHOGNEISSES

1.4.1 The “old” augengneisses (Pontis nappe)

The polycyclic Pontis basement contains numerous parallel bands of augengneisses of remarkable thickness (ca. 30-300) and continuity, probably repeated tectonically. Within the upper Stalden zone (fig. 32; fig. 33; fig.34A and 34E.), these “old” orthoaugengneisses outcrop very well as bands or lenses at Ahorn (coordinates: 635.575/115.300/1225m) and on top of the Ochsenhorn (Beauregard 1972; Thélin et al. 1993). They can also be traced equally well in the Berisal unit and in the Rutor zone (Burri 1983 a et b; Baudin 1987, Gouffon 1983; Thélin et al. 1993). These authors conclude that they represent porphyric granitoids of calc-alkaline granodioritic and monzogranitic

composition. Zircon typology (Alkalinity index, I.A. = 483; Temperature index, I.T. = 362 according to Pupin's coordinates, 1980) confirm the hybrid origin of the protolith. The megacrysts are single crystals (chessboard albite, low microcline, quartz) or augen-shaped polycrystalline aggregates of the same minerals. These augen are dark grey or even black, egg-shaped and centimetric in size. The polymetamorphic history of these rocks is evidenced by a relic assemblage with plagioclase (An. Max 20 wt.%) - zoned garnet-brown biotite-muscovite (Ti-rich). The geodynamic context in which they were emplaced, on the basis of Nb, Ti and Zr contents, is compatible with that of an active continental margin or of an island arc. Preliminary U-Pb ages on zircons yield ca. 450 Ma., *i.e.* Late Ordovician age. Comparison is warranted by example with similar augengneisses within the External Massifs (von Raumer 1987; Bussy and Von Raumer 1994; Morard 1998; Dobmeier et al. 1999). These authors considered these granites as revealing calc-alkaline (I-type) or peraluminous (S-type) character, as expected in an active continental (or island-arc) margin. A fruitful analogy is founded when making a comparison with the Blasseneck porphyries in Austria which are considered to be of Late Ordovician age (Loeschke, 1989).

A short mention must be done here about the Thyon metagranite which is located in the frontal part of the Siviez-Mischabel nappe (Bussy et al. 1996). It is intrusive in a polymetamorphic banded volcanic complex as leucocratic concordant sills with pseudoaplitic rims. A distinct metamorphic schistosity is defined by dark-green Fe-rich biotite. Abundant mesoperthites, chess-board albite and

Fig. 34 - A: Outcrop of Ahorn (coordinates 635.575/115.300/1225m) along the road Stalden - Saas-Fee. The outcrop occurs within the upper Stalden zone, *i.e.* within the polycyclic Pontis basement. 1->4 refer to descriptions in text. B. Garnet-staurolite polycyclic gneiss cut by a discordant metabasaltic dyke, likely Mesozoic in age. C1. Banded complex in a boudinage structure composed of alternate amphibolites, lenses of retroeclogites and felsic “exsolution”; C2. Boudin of retroeclogite (detail). D. Narrow level of augengneiss, slightly folded, within the banded complex. E. “Old” orthoaugengneisses, Upper Ordovician in age (ca. 450 Ma.). Phyllonitic mylonites locally replace these orthogneisses.

low microcline are presumably related to magmatic stages and/or greenschist-facies metamorphic retrogression. Major, trace elements and REE geochemistry, zircon typology, Y- and Nb-bearing accessory minerals such as fergusonite and euxenite, all point to a metaluminous to peraluminous alkaline A-type granite. High-precision U-Pb zircon dating yielded a sub-concordant age of 500 \pm 3/-4 Ma. The Thyon metagranite is the third record of a Cambro-Ordovician alkaline magmatic activity in the Alps. As A-type granitic magmatism is common in post-orogenic to anorogenic extensional tectonic regime, the Thyon intrusion could mark the transition from the Cadomian to the Variscan cycle (see general introduction, chapter 2.2).

1.4.2 The „young“ augengneisses (Siviez-Mischabel nappe)

The Randa augengneisses (RA) occur in the inverted limb of the Siviez-Mischabel fold-nappe (fig. 32 and 33), in the Middle Penninic domain (Beauregard 1964; Th  lin 1987; Th  lin et al. 1993; Th  lin and Girod 2000). The RA derive from a subalkaline porphyritic granite, Permian in age (U/Pb zircon age: 269 Ma \pm 2 Ma after Bussy et al. 1996), metamorphosed during the Alpine Tertiary event in the upper greenschist facies. This crystallisation age of the Randa granitic protolith is in almost perfect agreement with the surrounding lithostratigraphy of Late Palaeozoic age. Culmination ages of metamorphism and cooling history are given by Steck and Hunziker (1994). The RA bedding (mainly due to a first order schistosity) generally dips 20° to 30° towards the westward. The main body is a pseudo-laccolith with sill-like extensions within the Permo-Carboniferous metagreywackes. It presents all the megascopic characteristics of an intrusive granite: primary discordant contacts, veins of aplite, microgranitic border facies. Mafic inclusions have never been observed. Effects of strong Alpine cataclasis,

concomitant to the Alpine metamorphism, are well developed. Deformations took place from a mylonitic regime under ductile conditions till cataclastic and kakiritic (= gouge) regimes under brittle conditions (fig. 34B).

The lithologic facies of the Randa orthogneisses varies from place to place; most representative is an augengneiss. The large pale crystals of this augengneiss (phenoclasts of 1 to 10 cm in size) are mostly K and Na feldspar (originally: perthitic orthoclase/microcline transformed by deuteric alteration and/or metamorphism and/or cataclasis to chessboard albite) embedded in a coarse-grained (? 0.3 - 1 cm) quartzo-micaceous matrix. The gneiss can also be homogeneously fine-grained or mylonitic; it may contain some concordant fine quartz lodes too. In certain cases cataclasis may lead to mechanical layering. The essential minerals of this rock are the following: 30% quartz ; 10-15% microcline; 45-50% albite; 10-15% white micas (mostly phengite and primary muscovite). The main accessory minerals are chlorite, calcite, green biotite, epidote, zircon, sphene, apatite, pyrite and garnet. There is no significant chemical difference between fine-grained gneiss and coarse grained gneiss (augengneiss). The large augen are essentially replacement feldspars, containing abundant magmatic relics (primary inclusions of plagioclase crystals, corroded dihexagonal quartz grains, primary red-brown biotite, pseudo-rapakivi mantles-Th  lin 1987). Ba-quantification and a thermometric study indicate multistage growth of the original alkalifeldspar, from the orthomagmatic to the subsolidus and deuteric stages. Geochemical data confirm the mantle-derived origin of the primary magma, affording evidence on the original aluminous subalkaline liquid. (Desmons and Ploquin 1989). Zircon typology gives I.A.= 618 and I.T.= 353, which also confirms a deep origin within the field of subalkaline granites. It must be emphasized at this point that many quartz porphyries and metarhyolites were described within the basements

of the Middle Penninic units (Thélin et al. 1993). One of them, the Laget meta-ignimbrite (Siviez-Mischabel nappe, Bagnes valley, western Valais), yielded a crystallization age between 267 to 282 Ma. (Bussy et al. 1996).

2 - CROSS SECTION THROUGH THE INTERNIDES FROM SION TO ZERMATT

Arthur Aescher and Michel Marthaler

The following units will be visited: The Sion-Courmayeur zone, the Pontis basement (Upper Stalden zone), the orthogneiss core of the Siviez-Mischabel nappe and the Zermatt-Saas Fee zone.

2.1 THE SION-COURMAYEUR ZONE

The Sion - Courmayeur zone is by far the best preserved unit derived from the Valais trough domain. It crops out from the Simplon Pass, following the Rhône valley, to Moutiers, parallel to the Alpine belt over more than 200 km. In a transverse section, from the NW to the SE, the following subdivisions can be recognized (fig. f5-A):

- 1) The Ferret and St Christophe calcareous and siliciclastic flysch (Late Cretaceous to Tertiary)
- 2) The Marmontains unit, made of black shales interlayered with quartzites (Middle Cretaceous).
- 3) The Arolay unit, composed of calcschists and breccias with massive limestone beds (Middle Cretaceous or younger)
- 4) The Versoyen and Pierre Avoi chaotic complex, made of a mélange of black schists with blocks and slivers of metagabbro, metabasalt and serpentinite, locally with large slices of gneiss, quartzites, dolomites and marbles, and of Mesozoic breccias. Microfossils discovered in the hills of Sion by Bagnoud et al. (1998) give a Late Eocene age for the siliceous matrix of the Pierre Avoi unit.

These 4 divisions of the Sion-Courmayeur zone were defined by Trümpy (1952, 1955) in the

Western Swiss Alps and by Antoine (1971, 1972) in the French part of the Sion - Courmayeur zone. Later work by Burri (1967, 1979), Ackermann et al. (1991) and Jeanbourquin & Burri (1989, 1991) suggest that the Versoyen - Pierre Avoi unit is a remnant of an accretionary prism developed between the European and Briançonnais domains. It probably represents the most internal and upper part of this accretionary prism which was constructed during the subduction of the Valais trough oceanic crust from the Late Cretaceous to the Late Eocene (see general introduction fig.5).

2.2 THE ZERMATT - SAAS FEE ZONE AND ITS STRUCTURAL POSITION

The Zermatt-Saas Fee zone consists mainly of metaperidotites, metagabbros and metabasalts (Bearth 1967, Colombi 1989). Chemical studies reveal a tholeiitic composition for the mafic rocks and a close similarity with transitional to normal MORBs (Beccaluva et al. 1984, Pfeiffer et al. 1989). The peridotites are relatively undepleted lherzolites, quite similar to those of the Lanzo zone. Some oceanic metasediments are found, consisting mostly of siliceous marbles, garnet and Mn-bearing quartzites (metaradiolarites) and calcschists (Vannay & Allemann 1990).

All these observations indicate that this unit was part of an oceanic lithosphere formed in a slow-spreading environment (Lagabrielle & Cannat 1990). The presence of relics of eclogite facies parageneses show that it underwent a HP metamorphism, probably during the Eocene. A later, Oligocene greenschist overprint is well documented.

MAP FIELD TRIP 5



II: EXCURSION OUTCROPS

Stops are located on the topographic map on the opposite page, a structural map corresponding to the same area is found in annex (MAP 3 and 4).

STOP 5A - COLLINES DE SION (FIG. FT5-A): THE PIERRE AVOI CHAOTIC COMPLEX.

(Arthur Escher and Michel Marthaler)

The Tourbillon castle is built on top of a siliceous and calcareous flysch sequence. It forms the matrix of the Pierre Avoi chaotic mélange. In a nearby situated hill (Mont d'Orge) Eocene foraminifers have been discovered recently (Bagnoud et al. 1998). A important dextral strike-slip fault separates two of the Sion hills. The Valère church is built on a large quartzite bloc (Early Triassic). It was probably sheared off the northern Briançonnais margin during the subduction of the Valaisan accretionary prism.

STOP 5B - VIEW FROM VIÈGE TOWARDS ZENNEGEN: THE VERSOYEN MÉLANGE WITH LARGE OPHIOLITE BLOCKS (SEE FIG. 32).

(Arthur Escher and Michel Marthaler)

The small Village of Zennegen is built on the Southern branch of the Rhône-Simplon fault system which dips 40° to the SW (strike-slip and detachment movements). South of this fault the Briançonnais units are here sub-horizontal but refolded as a recumbent synclinal structure. Its core is made of Triassic dolomitic marbles, breccias and evaporites and is surrounded by Permo-Triassic quartzites and a thick bulk of various older paragneisses (Paleozoic).

To the North, the Valaisan series are strongly refolded. In their central part the Versoyen chaotic complex contains several large blocks: Paleozoic gneisses, Triassic dolomitic marbles, Liassic limestones, oceanic metabasalts and serpentinites of controversial age (Carboniferous or Cretaceous?).

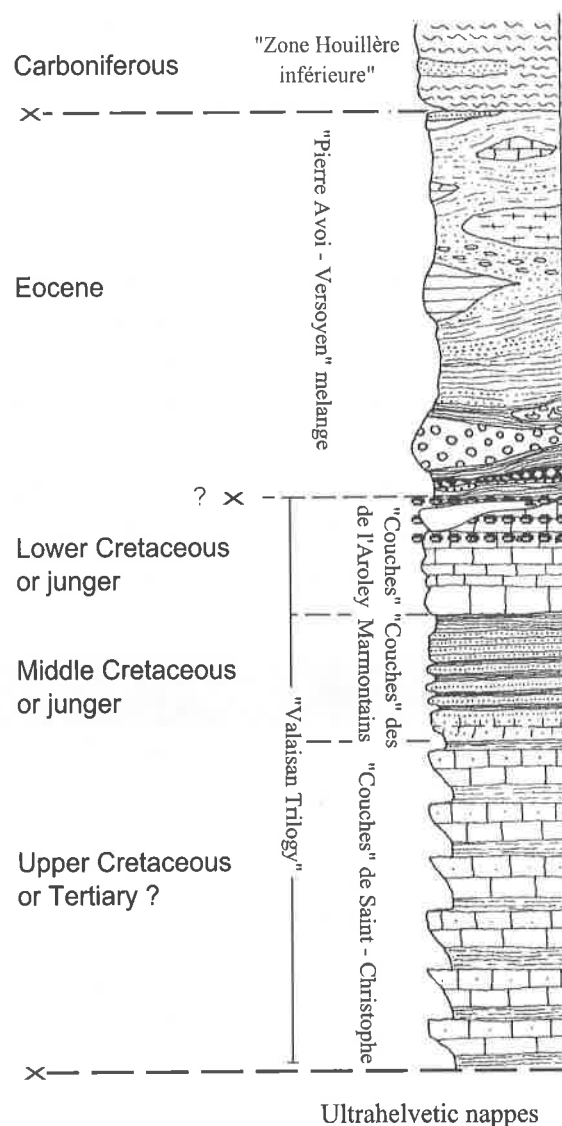


Fig. ft5-A - Synthetic Lithostratigraphy of the Sion-Courmayeur zone (modified from Jeanbourquin and Burri 1991).

STOP 5C - AHORN OUTCROP

(Philippe Thélin)

Swiss coordinates (635.575/115.300/1225 m, see figs.32 and 34, with details on Figs. 34B->E)

This nice outcrop is located opposite the hamlet of Ahorn (Saas valley), above a protection tunnel of the Stalden to Saas Balen road. This slab of polycyclic rocks within the upper Stalden zone (Pontis nappe) allows to see a rather exhaustive summary of rocks occurring in the Middle Penninic basement, such as garnet-staurolite bearing paragneisses, banded amphibolites with boudins of retroeclogites, "old" ortho-augengneisses (Late

Ordovician in age). The section is quite fresh because of the glacial polish and from S to N, one can observe the following lithological succession - (1->4 refer to the red numbers plotted on fig.34):

(1): Two-mica and staurolite-garnet bearing paragneisses (ca.22m thick). Garnets strongly retrogressed into epidote-chlorite. Presence of at least two “dykes” of prasinitic greenstone (ca. 0.2-0.4m thick), slightly discordant related to the main schistosity (fig. 34B). These “dykes” yield an Alpine assemblage actinolite - epidote - chlorite - albite - quartz - calcite. They are of probable Permo-Carboniferous or Mesozoic age. Locally, presence

of boudin-like amphibolite with orthoamphibolite nodules and epidote-rich layers.

(2): Banded amphibolites in a boudinage structure (ca. 9m thick) composed of alternate amphibolites, lenses of retroeclogites and felsic “exsolutions”. Alternation with the two-mica paragneiss (+/- garnet,+/-chlorite)-fig. 34C and 34C1. Locally thin layers of augengneiss within the banded amphibolites (fig. 34D).

(3): Sharp contact with “old” ortho-augengneisses (ca. 25m), locally replaced by mylonitized bands, such as phyllonites; as described

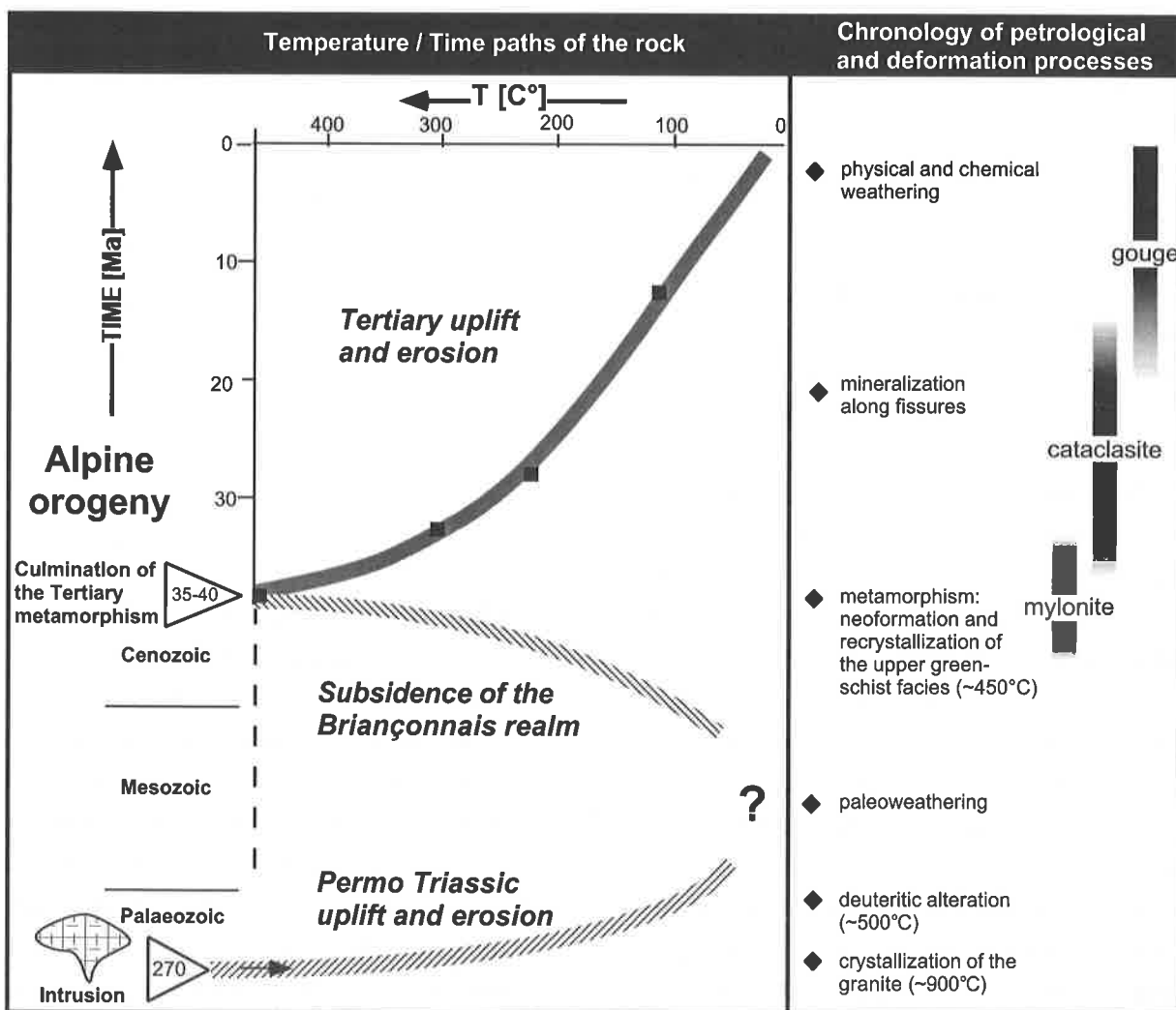


Fig. 15-D - Thermal and petrological history of the Randa granite related to the main geological events and associated with the deformation processes (ductile to brittle). Slightly modified after Girod, 1999 and Th  lin and Girod, 2000.

under 1.4.1. (fig. 34E).

(4): Sharp contact with two-mica and staurolite-garnet bearing paragneisses.

STOP 5D - RANDA VILLAGE (1409M)

(Philippe Thélin)

The Randa unit is characteristic of the sharpest relief on the left side of the Mattervalley, downstream the village of Randa (alt. ca. 1400m). The morphological features of the valley are typical of Quaternary glacier retreat; it is the deepest and longest valley lateral to the large Rhône valley. The Randa stop will allow from one hand to look at the Randa ortho-augengneiss (I-granite, Permian in age, described under 1.4.2.), intrusive within Permo-Carboniferous metagreywackes of the inverted limb of the Siviez-Mischabel nappe (fig. 32 and ft5-D) and from another hand to consider the geomorphology resulting from the 1991 Randa multi-stage rockfall (Jaboyedoff et al. 1998; Girod 1999; Girod and Thélin 1998; Thélin and Girod 2000).

In fact, during spring 1991, several rock falls affected the sector of Randa: 1°) on April 18th, 20·10⁶ m³, mainly of orthogneiss, collapsed at the Grossgufer location; 2°) on April 23rd, and on May 9th 10·10⁶ m³, mainly of paragneiss, collapsed in the same sector. Each phase of rock fall was associated with a significant release of dust, giving place to deposits of more than 15 cm in thickness in a radius of 800m. The volume of dust released can be estimated between 200'000 to 500'000 m³ for each phase. This total volume of about 10⁶ m³ corresponds approximately to 3% of the total fallen volume and equivalent to 400'000 m³ of rock. The first phase of collapse caused an earthquake (3 on the Richter's scale). The transportation channels, train and road, were cut and the course of the Matteredvispa river was stopped by the talus slope. In order to avoid the rise of a lake that would have, in the long term, embedded the village of Randa, it

was decided to dig a deviation channel.

The case study of Randa shows the complexity in evaluating the triggering factors that have led to the collapse of the cliff. As shown by the "Centre de Recherche sur l'Environnement Alpin-CREALP" studies (Jaboyedoff et al. 1998), water table fluctuations within the rock mass are probably the major triggering factor of the rock fall, especially high water pressure due to heavy rainfalls and snowmelt within a connected system of joints. But the study of the weathering processes within the rock mass and its fractures reveal that even slow mineralogical transformations can play a role in the landform evolution. Despite the cold climate of this mountainous area, chemical weathering is active and must be considered as a non-negligible parameter in the development of instabilities of rock masses. Mainly it allows to increase microporosity within the rock mass - but not the permeability - and it leads to the dissolution of primary minerals such as albite and to the precipitation of swelling clays on fault gouge which, if present in sufficient quantities, can modify the geomechanical properties along the fault plane by reducing its friction angle. Geomechanical, structural (Alpine tectonics and recent displacements), hydrogeological (highly fissured aquifers) and geomorphological (post - glacial steep valley) factors do play the leading factors in such a case; as far as the mineralogical factor is concerned, it plays a sly game, dissolving primary minerals, helping to connect structural joints, precipitating swelling clay minerals, all phenomena contributing to weaken the rock mass.

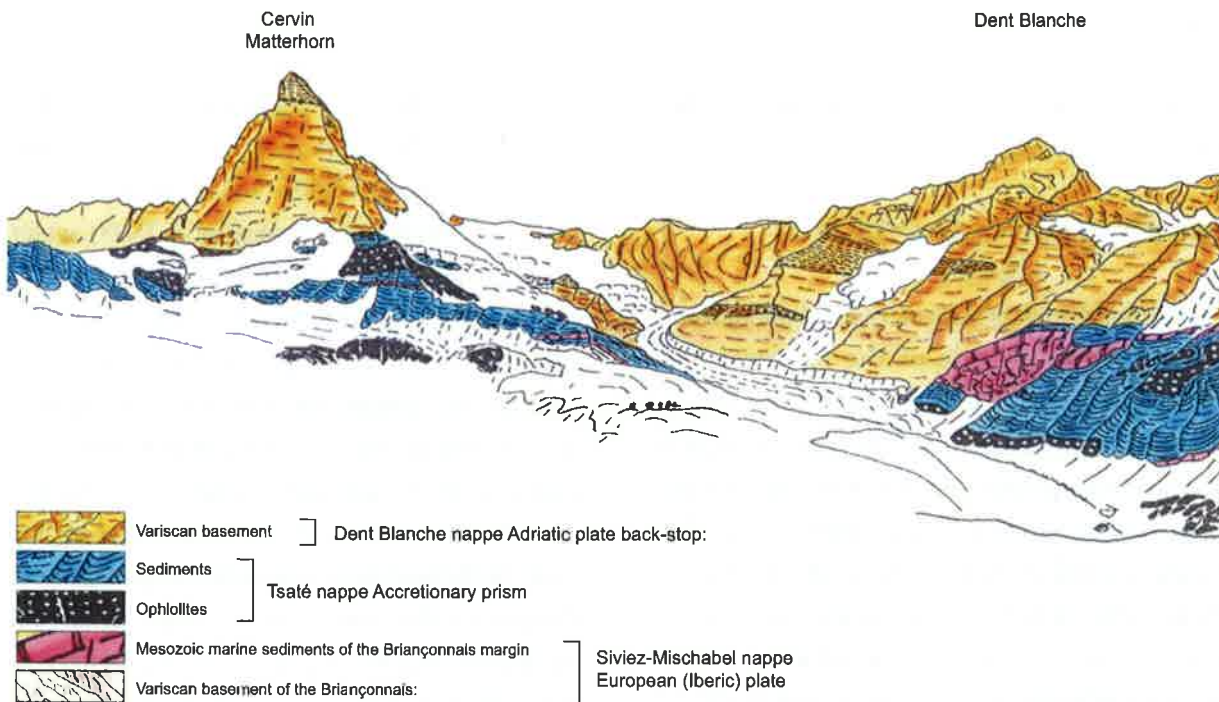


Fig. ft5-E1 - Panoramic view of the Siviez-Mischabel nappe and its large-scale back fold, the Tsate and Zermatt-Saas units and the overriding Dent Blanche nappe (Sartori 1992 and Marthaler, 2001).

STOP 5E - CROSS-SECTION THROUGH THE MATTERHORN (fig. ft5-E2) GEOLOGICAL PANORAMA (2288M) ABOVE ZERMATT (fig. ft5-E1).

(Arthur Escher and Michel Marthaler)

The view towards the West shows the relation between 3 major Alpine units:

- a) The lower Briançonnais continental unit of the Siviez-mischabel nappe with its well developed large-scale backfold.
- b) The middle Piemont oceanic units represented by the Tsate accretionary prism and the Zermatt-Saas-Fee ophiolites.
- c) The overlying Dent Blanche nappe of Austroalpine origin.

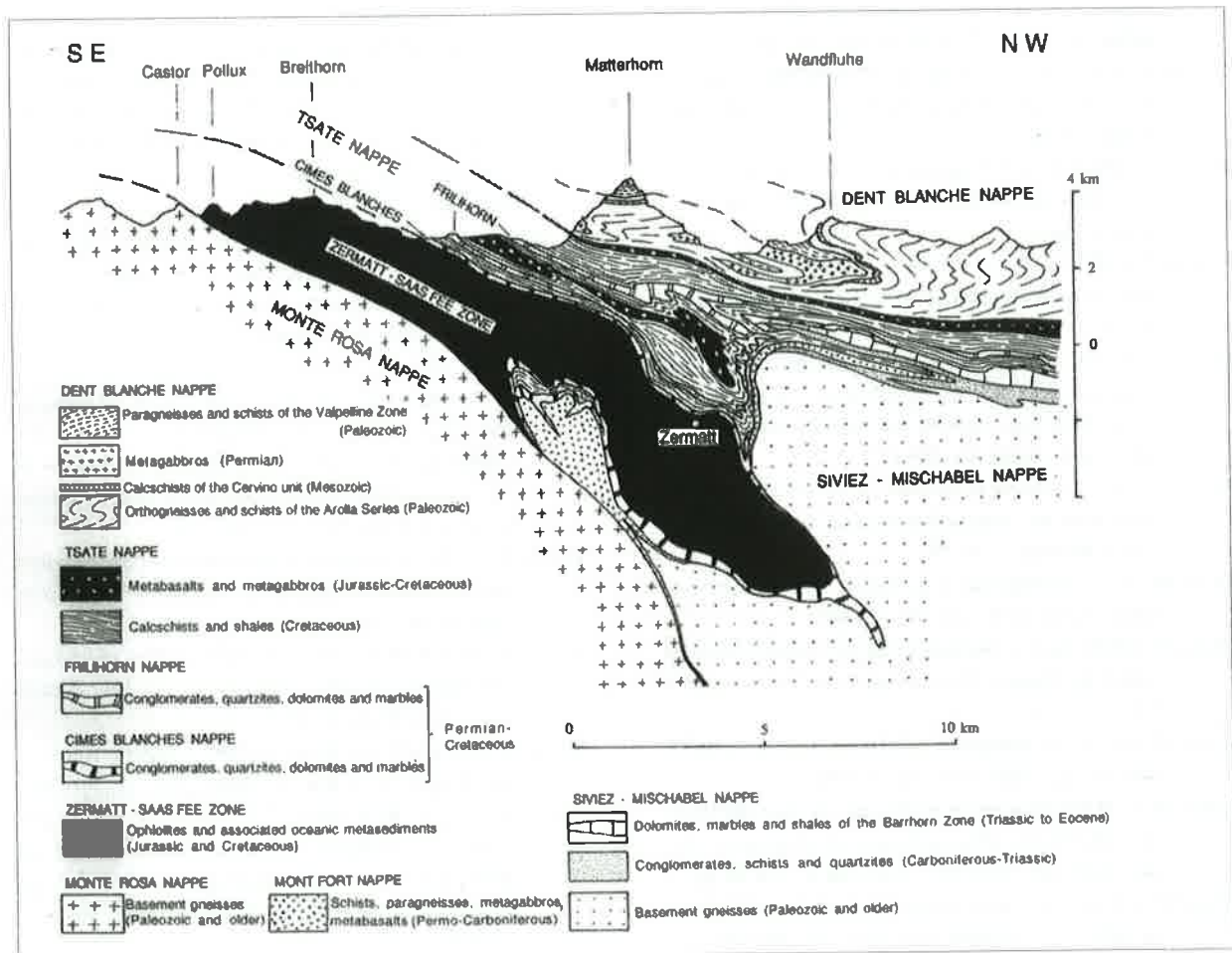
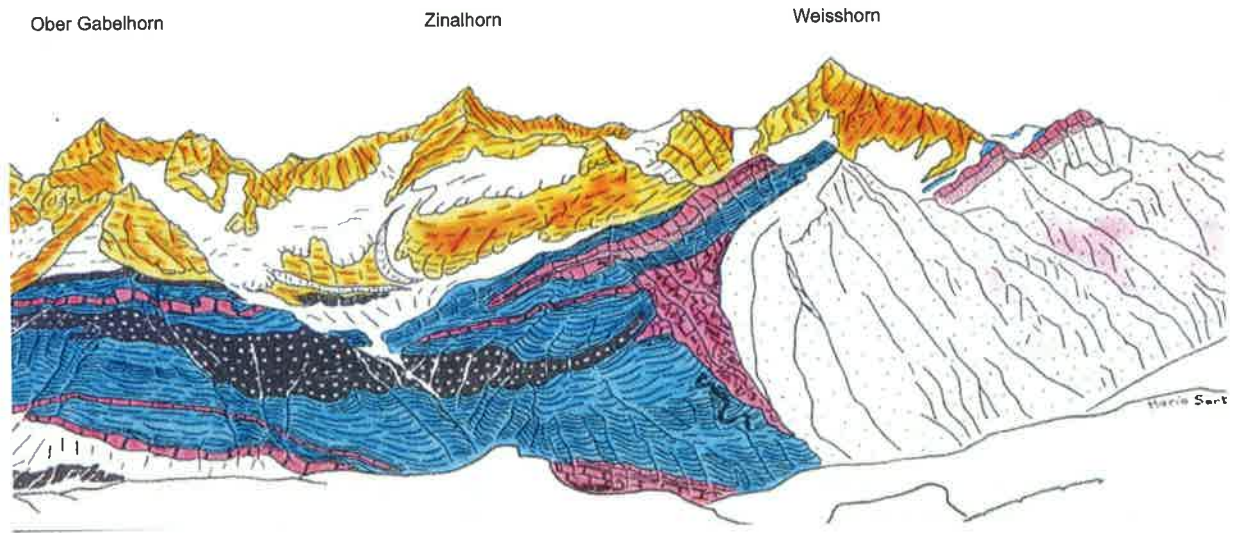


Fig. j15-E2 - Simplified geologic section showing the relation between the Monte Rosa, Siviez-Mischabel, Zermatt-Saas Fee, Tsate and Dent Blanche nappes in the Zermatt region (Escher et al, 1997).

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The internides klippen belt - the Préalpes Romandes domain

Jon Mosar, Gilles D. Borel, Yann Ringgenberg & Gérard M. Stampfli

I: INTRODUCTION

1 - GENERAL INTRODUCTION ON THE PRÉALPES

The Préalpes consist of several klippen along the northern front of the Swiss and French Alps (fig.35) from east of the Mythen near Luzern (Switzerland) to the Klippe des Annes near Annecy (France). The two major lobes are formed by the Chablais Préalpes south of Lake Geneva (lac Léman) and the Préalpes Romandes between Lake Geneva and Lake Thun. The Préalpes Médianes are the most important of several allochthonous structural and paleogeographic units now superposed in the Préalpes fold-and-thrust belt (fig.36, 37). Amongst these different tectonic nappes one can differentiate from top to bottom:

[i] the Nappe Supérieure, which itself can be subdivided into four different units: the Gets Nappe, the Simme Nappe, the Dranses Nappe and the Gurnigel Nappe; [ii] the Breccia Nappe resting on the trailing part of the Préalpes Médianes only; [iii] the Préalpes Médianes nappe and [iv] the Niesen nappe which exists in the meridional part of the Préalpes only and which today forms the southernmost structural unit. The Préalpes Médianes are separated from the Niesen nappe by a structural unit designated "Zone Sub-médiane" of paleogeographic origin located between the Valais trough and the Briançonnais s.l..

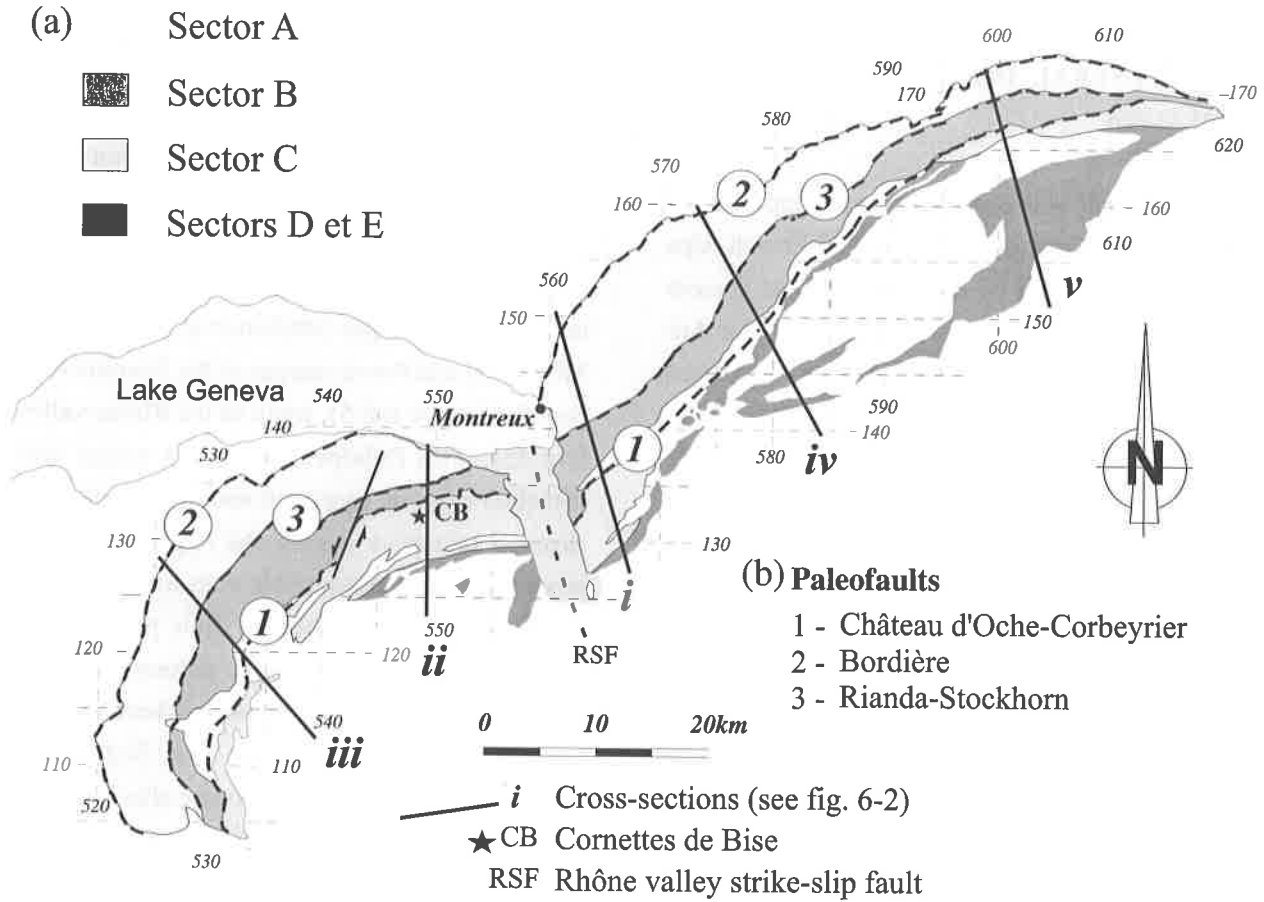
Between the original position and their present-day location as klippen, the Préalpes underwent a complex history of paleotectonics and Alpine tectonics. Due to the opening of the Alpine Tethys ocean the Briançonnais sedimentation realm of the Préalpes Médianes evolved as a rim basin of the

northern passive margin during the Jurassic (see general introduction). Then, with the opening of the Pyrenean rift, the Briançonnais portion of the sedimentary basin and its basement evolved into a micro-continent.

Equivalent stratigraphic units have been found both in the Préalpes Médianes and the Siviez-Mischabel and Pontis nappes of the Pennine Alps (see field trip 4 and 5), south of the Rhône valley (fig.35). The Préalpes Médianes nappe was detached from its basement and underwent thin-skinned tectonics during the Alpine orogeny, whereas similar stratigraphic units to the east and south remained attached to their pre-Triassic basement and were intensely deformed during thick-skinned tectonics. It was Schardt (1884; Schardt 1893; Schardt 1898) who first clearly demonstrated that the Préalpes were allochthonous (he even used the word exotic).

Different paleotectonic features (normal faults, synsedimentary growth structures, inversion structures) developed and were active above a basal detachment in evaporitic layers. Isolated from the Iberic continent at the end of the Late Cretaceous, the Briançonnais exotic terrain was incorporated into the accretionary prism of the closing Piémont ocean and the incipient Alpine orogeny during the Lutetian-Bartonian. The Préalpes Médianes were detached from their basement during the Bartonian-Priabonian and were transported onto the foreland and now show the typical characteristics of a foreland fold-and-thrust-belt (fig.36). The foreland propagating thrust sequence that developed above a basal décollement resulted in a series of large fault-related folds. The fold style is one of the fault-

Simplified geological map of the Préalpes Médiannes



(c) Sectors length along the different sections for latest Liassic

Profiles	Sector A (km)	Sector B (km)	Sector C (km)	Sector D (km)
i - Caux-Tours d'Aï	15	10	10	-
ii- Locum-Cornettes de Bise	<i>10</i>	2.5	10	6
iii - Monts d'Hermone-Roc d'Enfer	15	8.5	9	non-existent
iv - Moléson-Gummfluh	15	6	-	7
v - Hohmad-Turnen	12	5	2	-

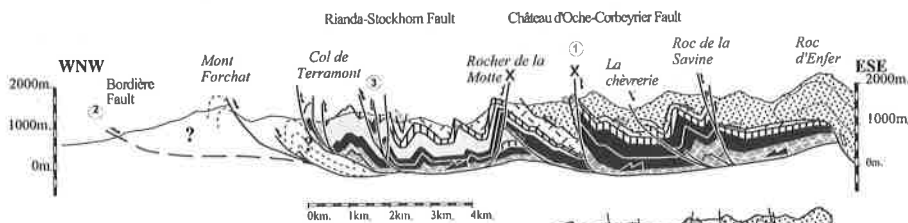
bold = maximum sector length in reference section

italic = estimated sector length, plain = measured sector length

Fig. 35 - Simplified location-map of the Chablais and Romandes Préalpes.

Fig. 36 - Cross-section i to v of the Préalpes domain (see fig f16-1 for location).

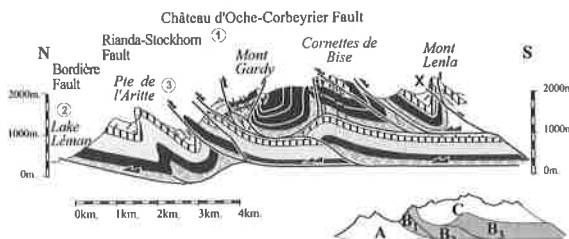
Section i Monts d'Hermone - Roc d'Enfer



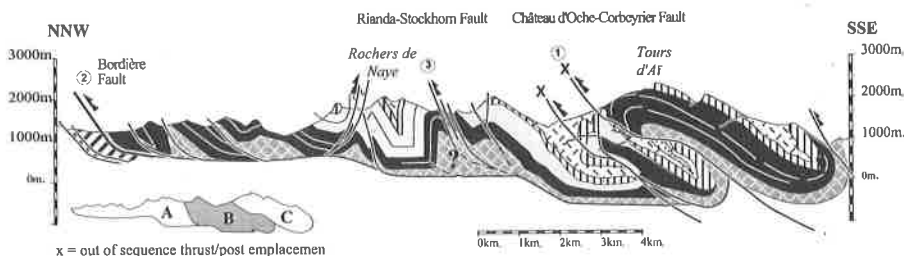
Préalpes Médiannes Nappe

- Cretaceous
- Malm (Calc. Massif Fm.)
- Dogger
- Liassic
- Triassic
- Nappe Supérieure
- Nappe de la Brèche
- Subalpine Flysch
- Thrusts

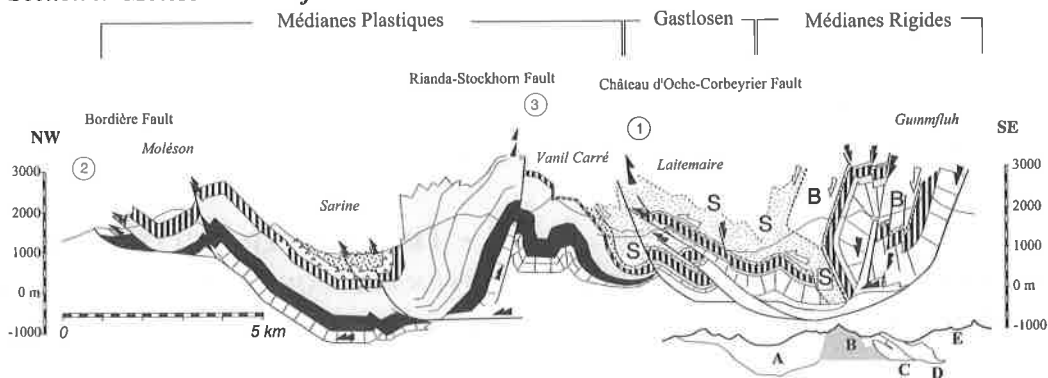
Section ii Locum - Cornettes de Bise



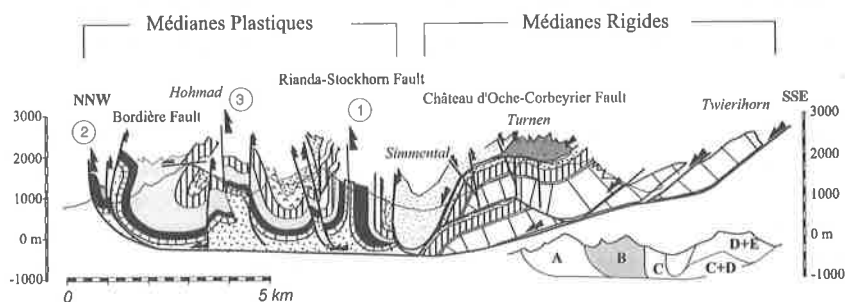
Section iii Caux - Tours d'Al



Section iv Moléson- Gummfluh



Section v Hohmad - Turnen



bend and fault-propagation type, but also of the detachment fold type (box folds). Backthrusts developed at the ramp-flat transitions as well as by inversion of former normal synsedimentary faults. Periclinal closures, en echelon relay structures, tear faults, lateral ramps and fold interference structures are commonly observed features. The general transport direction towards the N-NW, is perpendicular to the general fold axial trend. The very low-grade metamorphic conditions (see below) have their origin in the heat flux induced by tectonic burial by overriding nappes in the accretionary prism. After having been transported on top of the developing Helvetic nappes, the Préalpes were emplaced in their present-day position in front of the Alpine mountain belt during Oligocene times. Post-emplacment and out of sequence thrusting, possibly younger than Oligocene, is observed and can be related to thrusting in the sedimentary substratum and the basement.

The reader can find more detailed description and discussion on the Préalpes Médiannes, as well as references in: Masson 1976; Plancherel 1979; Baud & Septfontaine 1980; Trümpy 1980; Mosar et al. 1996 and a quite complete bibliography (more than 950 references) on the World Wide Web site: www-sst.unil.ch/marge&co/prealps/REFEREN.htm

2 - REGIONAL STRUCTURAL GEOLOGY OF THE PRÉALPES MÉDIANNES

The Préalpes Médiannes nappe is subdivided into the Médiannes Plastiques, forming the frontal (NW) part of the nappe and Médiannes Rigides, forming

the trailing (SE) part of the nappe (fig.35, 36 & 37, Lugeon & Gagnebin 1941). In the Swiss Préalpes, a domain with intermediate characteristics, both structural and sedimentological, exists between the Médiannes Rigides and Plastiques, the Gastlosen range.

The Médiannes Plastiques are composed of a succession of large-scale fault-related folds, whose trends vary from E-W in the eastern part of the nappe to NNE-SSW, and even N-S, in the western part of the nappe. Folds and associated thrust planes die out along strike, resulting in the lateral transfer of displacement to other "en échelon" fold-thrust structures.

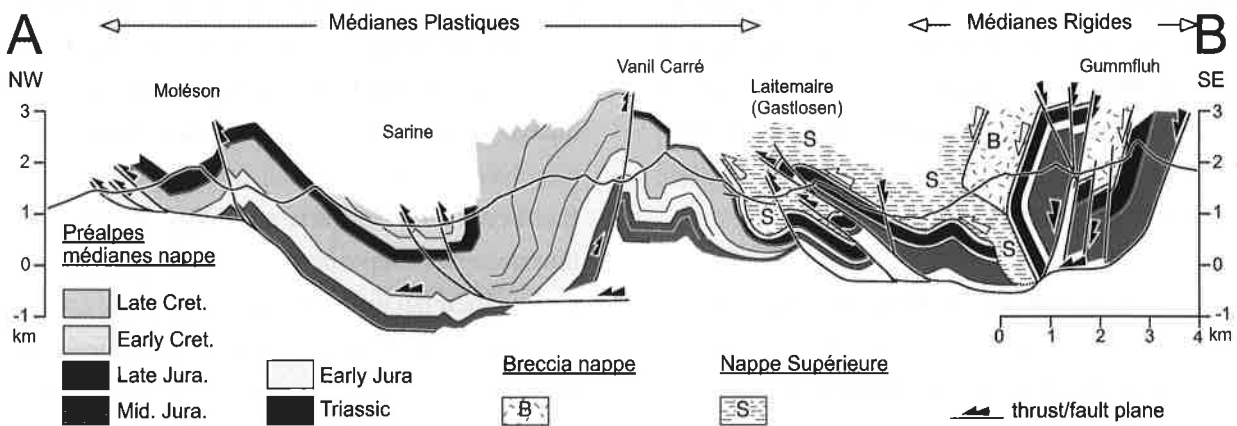
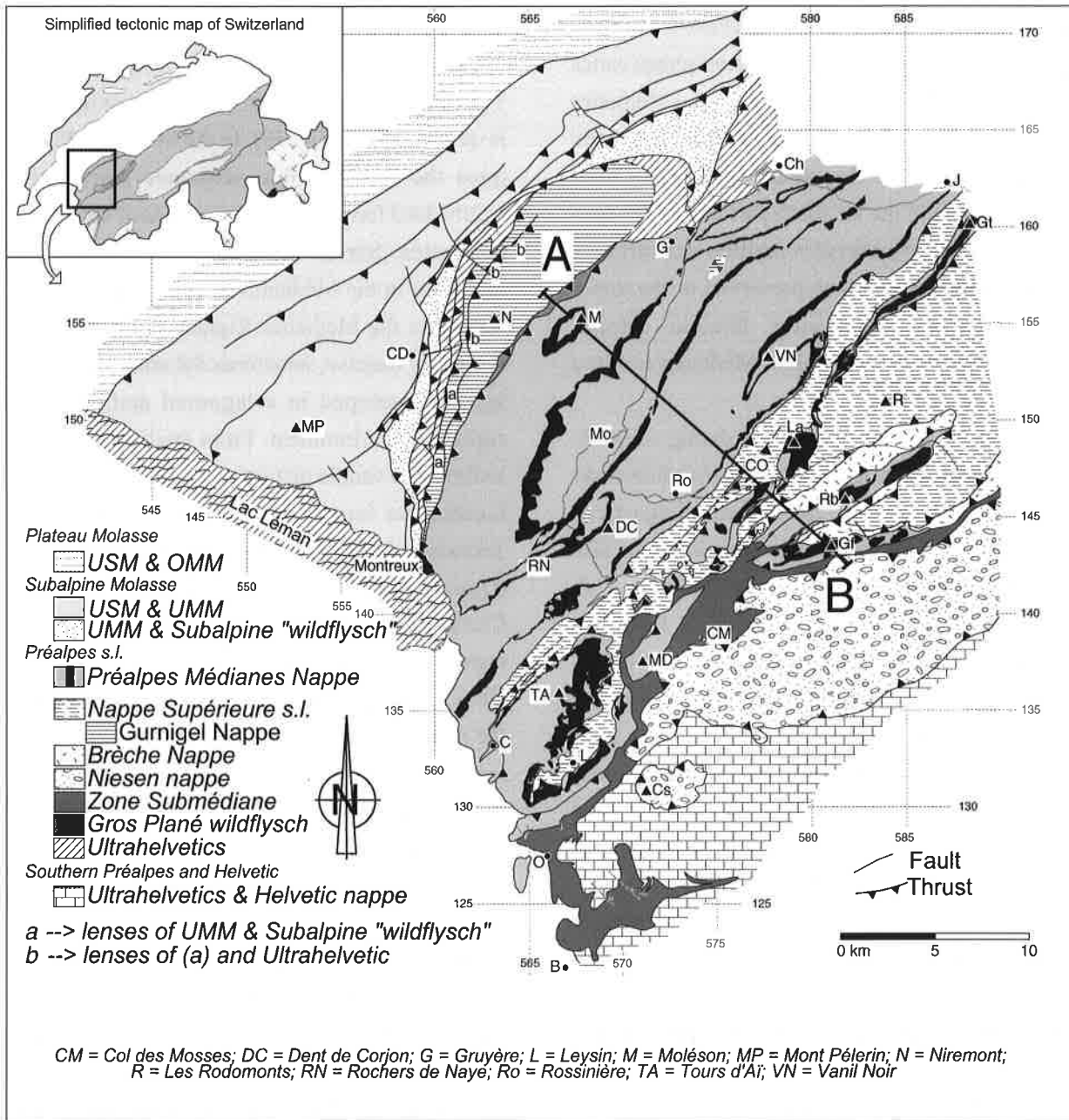
The trailing part of the nappe, the Médiannes Rigides, is formed by one major, in some places one or two minor, imbricated thrust slices that dip to the N/NW. These slices form fault-bend like folds that are cut by a large backthrust in their ramp portion. The imbricates dip gently to the north in the Simmental area of the eastern Préalpes Médiannes and are steeply dipping in the western region near Châteaux-d'Oex (fig. 36, 37).

Two major transport directions, N-S and NW-SE, towards the Alpine foreland have been determined for the Préalpes Médiannes. At the western termination late movements to the west in the complex frontal imbricate of the Médiannes Plastiques can be demonstrated (Mosar 1994; Mosar et al. 1996).

3 - TRANSPORTED METAMORPHISM AND INTERNAL DEFORMATION IN THE PRÉALPES MÉDIANNES

Numerous studies in recent years (see Mosar 1988; Mosar & Suppe 1988; Zahner & Mosar 1993; Jaboyedoff & Thélin 1995) showed that the

Fig. 37 - Tectonic map of the western Préalpes Romandes and their neighbouring structural units (the color version is in annex, map 5). Although the Nappe Supérieure of the Préalpes is made of four distinct tectonic units, only the Gurnigel nappe has been shown with a different pattern. UMM = Lower Marine water Molasse; USM = Lower Fresh water Molasse; OMM = Upper Marine water Molasse; OSM = Upper Fresh water Molasse. The cross section of the western Préalpes Médiannes Plastiques between Moléson - Sarine - Vanil Carré shows the deep Sarine syncline and the high structural relief of the Tsavas-Millelets anticline/Vanil Noir syncline/Dorena anticline.



Préalpes Médiannes underwent a very low-grade metamorphism ranging from diagenetic to epizonal (300-400°C) conditions. This metamorphism varies from diagenesis in the north (*i.e.* in the Médiannes Plastiques) to epizone in the south (*i.e.* in the Médiannes Rigides). This zonation also exists from top to bottom in the southern part of the Préalpes nappe stack. The epizonal conditions contrast with the diagenetic conditions preserved in the upper part of the underlying Niesen nappe. Metamorphism in the Préalpes Médiannes is thus a transported feature.

Rb/Sr, K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ dating of white micas from Triassic and middle Jurassic limestone yields ages of ~60 to 80 Ma in the trailing part of the Médiannes Rigides in the western Préalpes Médiannes (the Gummfluh imbricate: Masson et al. 1980; Huon et al. 1988; Cosca et al. 1992; the Amselgrat imbricate: de Coulon 1990, unpublished). These ages are interpreted by the authors as strong evidence for a Cretaceous tectonic and /or thermal event that was probably related to the initial stages of closure of the Alpine Tethys. We regard it as related to the translation of the Briançonnais peninsula and opening of the Pyrenean ocean and/or the closure of the Valais trough (see general introduction).

4-SEDIMENTOLOGY AND PALEO-TECTONICS

The Préalpes Médiannes are formed by limestones, dolomites, marls and shales ranging from Triassic to Tertiary in age (Badoux & Mercanton 1962; Plancherel 1979; Baud & Septfontaine 1980; Trümpy 1980; Baud et al. 1989; Plancherel 1990; Borel 1995; Mosar et al. 1996; Borel 1997; Borel & Mosar 2000)

To the N-NW, in the Médiannes Plastiques, a large basin is marked by an important thickness of sediments. To the south this subsiding domain turns into a ramp - associated with a continuously active

structural high controlled by a normal fault, that gives way to a platform and lagoonal environment in the Médiannes Rigides.

The sedimentary record in the two major domains is quite different (fig. 38). In the Médiannes Plastiques the stratigraphic sequence starts with interbedded formations of upper Triassic shales and dolomites. Series of Middle Triassic age are unknown in the Médiannes Plastiques but are dominant in the Médiannes Rigides. There they are formed by massive, sometimes dolomitic, limestone layers developed in a lagoonal and inter- to supratidal environment. From studies of Liassic sediments a simple picture arises, showing similar facies over large areas. Following the upper Triassic Dolomies Blondes which were uniformly deposited, the subsequent Liassic to Early Cretaceous paleogeographies and depocentres were mostly controlled by the passive growth of three normal paleofaults induced by the rifting of the Alpine Tethys (fig 35,36). The maximum extension of the Préalpes Médiannes depositional realm was reached at the end of the main rifting event (Borel 1997). Entrochal calcarenites, oolites, sandy as well lumachellic limestones sometimes containing cherts were deposited during Early to Middle Liassic. Locally (at the southern edge of the Médiannes Plastiques basin) during this period polyphase mega-breccias, karsts and fissure fillings developed. South of the structural high an ephemeral subsiding basin developed during the Early Liassic. Limestones of the Rhetian can be as thick as 200m. Sediments during the Early Liassic are interbedded limestones and marls and quartz-rich fine-grained limestones. These beds may have several hundred meters of thickness. Interbedded black marls and black argillaceous limestone were deposited in an anoxic environment during Toarcian times. An accelerated subsidence during the Late Liassic results in the appearance of deepwater argillaceous *Cancellophycus* limestone facies in the Médiannes Plastiques basin

Stratigraphy, Formations & Bathymetry in the Préalpes médianes

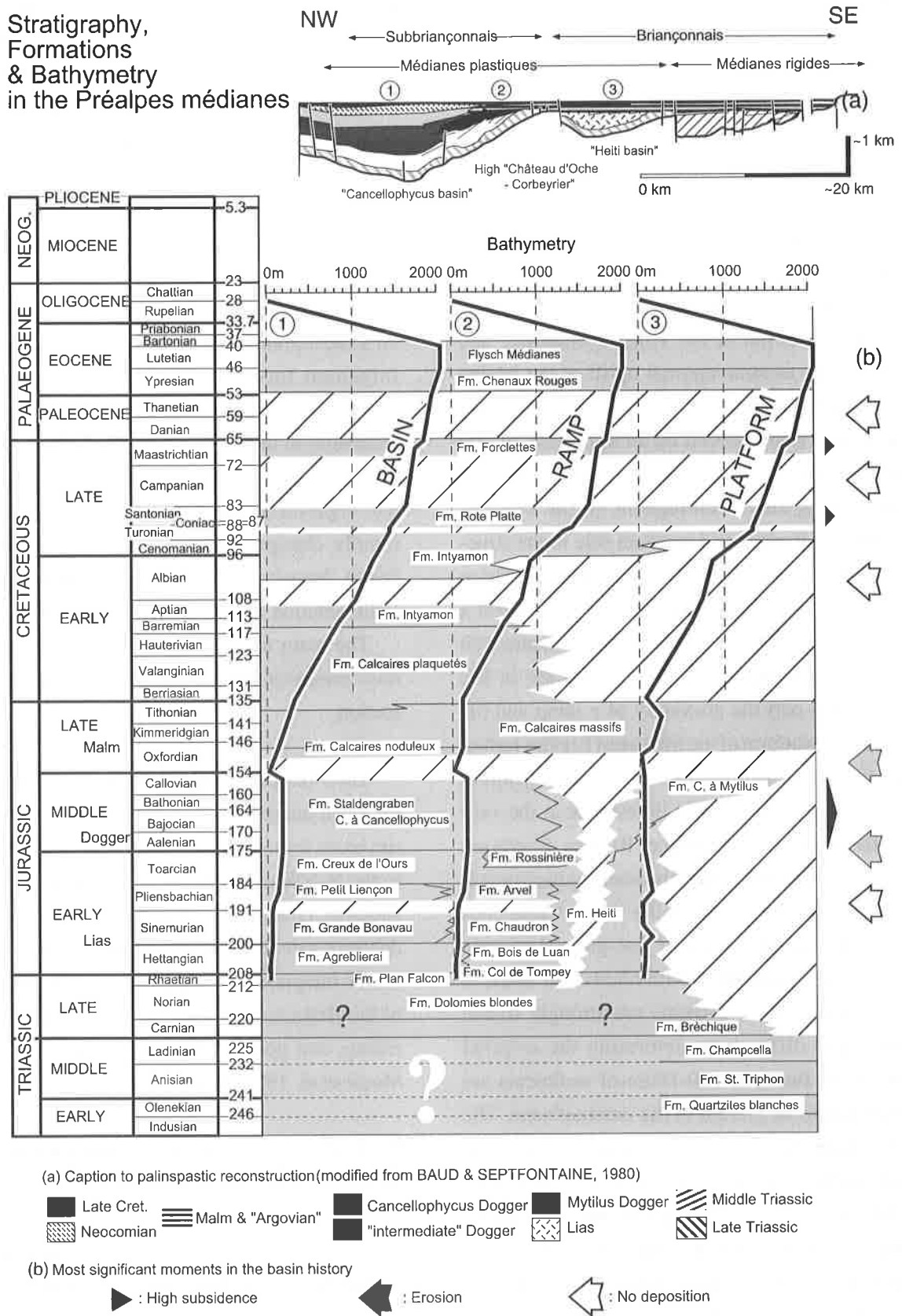


Fig. 38 - Stratigraphic table showing the times of deposition and sedimentary gaps (non-deposition and/or erosion) in the Préalpes Médianes Plastiques and Rigides, with the different formation for the Préalpes Médianes sedimentation domains. The simplified palinspastic reconstruction (Baud & Septfontaine 1980) shows the main paleogeographic regions of the Préalpes Médianes.

(**stop 6C**). This deposition continued into the Middle Jurassic (Dogger) with sporadic appearances of oolitic or spatic and sandy limestone layers. Slumps are frequent in a ramp setting towards the north.

To the south, in the Médiannes Rigides, a lagoonal facies developed with its marly and argillaceous limestone containing *Mytilus* and coal layers (**stop 6B**). This facies is transgressive on Triassic limestone in this meridional area, and paleokarsts are also common in the Triassic limestone and witness the tectonic/thermal uplift of the Alpine Tethys rift shoulder in Liassic time. The Late Jurassic is characterized by an important 20-300 meter thick competent limestone horizon outcropping almost throughout all the Préalpes Médiannes. It plays an important role in the structural development of folds and thrusts. These sediments are less argillaceous but still reflect a constant basin infill in the Médiannes Plastiques with resediments (**stop 6D**). They indicate in the meridional part the proximity of a ramp and the carbonate platform of the Médiannes Rigides further south. These sediments rest in stratigraphic unconformity on Triassic limestones in the very southern part of the Médiannes Rigides. Cherts are more frequent and bedding is more regular towards the top. The transition to the lower Cretaceous (Neocomian) succession of fine-grained micritic limestones interbedded with cherts and marls is progressive. The sediments are strongly folded making it difficult to determine the original thickness. Between 100-150m of sediments are estimated to be present in the external zone. The succession thins towards the center of the basin and pinches out completely to the south-east of Châteaux d'Oex. This pinching out can be explained by deposition along a slope close to an ancient coastline, as documented by calcarenites with large neritic bioclasts (echinoderms, corals, bivalves) and condensed red limestone.

An important change occurred during the late

Early Cretaceous as evidenced by the appearance of argillaceous, anoxic sediments (**stop 6E**). A different facies again appears during Late Cretaceous with interbedded limestone and gray, red and green wackestone or calcareous mudstone. These sediments are transgressive and terminate an erosional gap outlined by a mineralized hardground or an irregular erosional surface. The Couches Rouges Group of Late Cretaceous - Early Tertiary age rests with stratigraphic unconformity on a substratum of three different formations: the Intyamon formation to the N, the Calcaires Plaquetés formation and the Calcaires Massifs formation to the S. This unconformity is marked by a stratigraphic gap reflecting either erosion or non deposition. The Couches Rouges formation rapidly changes upward to schistose sand-rich flysch deposits of Tertiary age, thus ending the sedimentation history in the Préalpes Médiannes.

The main trend of the basin is E-W and the most prominent facies changes occur in a N-S direction.

The Préalpes Médiannes sedimentation realm has been interpreted in recent years in terms of a rim basin located in the Sub-briançonnais domain to the N-NW of the Briançonnais upper plate rift shoulder (see general introduction, fig.11). The different sedimentological and paleotectonic events can be integrated into the plate-tectonic evolution of the Briançonnais micro-continent (pre-rifting, rifting, and post-rifting tectono-thermal events; Mosar et al. 1996; Borel 1997).

5 - FOLD AND THRUST GEOMETRY IN THE W MÉDIANNES PLASTIQUES

From the external NW border towards the SE in direction of the Médiannes Rigides several distinct structural sub-units can be distinguished (fig.36, 37): [a] the external imbricates, divided into three different thrust slices forming the most fron-

tal tectonic structures of the Médiannes Plastiques; [b] the Gros Mology - Molard - Les Avants anticlinorium that forms the transition from the external imbricates to the “main body” of the Médiannes Plastiques; [c] The broad and deep Sarine or Intyamon syncline; [d] The Tsavas - Millets anticline that runs on the NW flanks of the mountains dominated by the Vanil Noir summit; [e] The Vanil Noir - Dent de Corjon - Rochers de Naye syncline forms a high mountain ridge that runs SE of the Sarine valley and extends till Lake Geneva; [f] The Dorena and Tinière anticlines are most prominent features in the Médiannes Plastiques bordering to the N the [g] Châteaux d’Oex syncline and further S the intermediate Médiannes (Gastlosen - Tours d’Aï) which further S is followed by the Médiannes Rigides.

Thrusts in the Préalpes Médiannes are common. They are generally buried under Quaternary sediments and can only be studied in a few outcrops. Thrusts of different importance and structural significance can be distinguished. The main thrust surface is formed by the basal décollement of the Préalpes Médiannes. It can be seen in the external imbricates, elsewhere it is buried below a folded hangingwall. From cross section construction, we see that this surface appears curved (folded?). It is along this décollement that the Préalpes have been detached from their basement and partly carried over and beyond the Helvetic units for a distance exceeding 100 km. The present-day depth to this basal décollement and its topography is a key element for reconstructing balanced/viable sections of the Préalpes Médiannes (see fig.36).

From recent studies this basal detachment is known to be located near sealevel, and exhibit a rather flat geometry in the eastern part of the Préalpes romandes near Thun (Vollmayr 1992; Pfiffner & Heitzman 1997; Erard 1999). This corroborates results from Mosar (1991) based on a balanced

cross section.

Interpretation of seismic data from the Lake Geneva – Rhône valley area of the Préalpes also suggest a rather shallow (sea level) present-day basal décollement surface in the frontal sector of the Préalpes (Steck et al. 1997; Sommaruga 1999). Unpublished data from seismic studies in the Cornettes de Bise in the Swiss Chablais (survey PSBR 8801-8805, recorded 1983 Prakla-Seismos, operator BEB Erdgas & Erdöl) and borehole data (from the E end of Lake Geneva - BEB Erdgas & Erdöl) indicate the same depth to the main detachment (seismic lines and borehole data are deposited at the Musée géologique de Lausanne). This depth to the basal décollement is in agreement with reconstructions in the Préalpes Romandes North of the Rhône river valley (Mosar 1994; Mosar 1997).

-Other thrust surfaces show less displacement and are associated with small imbricates and folds. Some may be overturned, steeply dipping towards the N due to stacking of several thrust slices. They are all splays of the main décollement. In turn smaller thrusts (reverse faults) branch off from these surfaces near their termination, in fold cores, to form pop-up like structures. Besides these foreland-vergent thrusts a series of hinterland-vergent or backthrusts, have been recognized and described (Revertera 1926). Those thrusts exhibit a top-to-the S-SE movement opposed to the general transport direction towards the NW onto the Molasse basin. They may originate on décollement and thrust ramps, as in the Médiannes Rigides (Niederhorn E Préalpes; Gumfluh-Rübli W Préalpes south of Châteaux d’Oex) or may result from inverted former synsedimentary faults (Tsavas anticline). In other places, they may be the result of triangle zones developing in response to space accommodation problems in fold cores (Dorena anticline or Rocher de la Motte in Chablais, fig.36).

An important and prominent structural feature

in the Préalpes Médiannes are vertical to sub-vertical faults or fault zones with horizontal displacement (Plancherel 1976; Plancherel 1979). They are mostly oriented N10 with a sinistral displacement and N110 with a dextral displacement. These zones separate domains with different orientations of fold axis directions, as well as areas where individual folds terminate and form en échelon relay structures. These faults have been interpreted as strike-slip faults rooting in the basement and affecting thus both the overlying and underlying structural units. Movement along these strike-slip fault systems would then be responsible for the fold-and-thrust development in the Préalpes Médiannes Plastiques. A different explanation, favored here, associates these faults with individual thrust and fold structures of the Médiannes Plastiques only (Mosar & Borel 1992). They are considered to be tear faults kinematically linked to the fold-associated thrust planes. In other places they form lateral ramps. Thus these faults can be considered transfer zones between two adjacent tectonic features. An example is formed by the western termination of the Dent de Broc syncline. By means of a tear fault, that runs in the Sarine river valley, it is juxtaposed to the W with the complex anticlinal structures of the Gruyère area. In the Montreux area the external imbricate thrusts trend NNE-SSW and show a strong sinistral displacement component. They can be interpreted as lateral ramps. Further south this zone turns into a larger sinistral tear fault zone affecting the whole of the western termination of the Médiannes Plastiques. Large, vertical and conjugate strike-slip faults are recognized in the Swiss portion of the Chablais Préalpes Médiannes. They offset thrust surfaces and folds and thus clearly postdate their development. This type of late faults is, however, not easily observed in the Préalpes romandes.

The Médiannes Plastiques are characterized by large-scale first order folds (fig. 36, 37). But

numerous second and third order folds can be observed in the middle Jurassic or the lower and upper Cretaceous multilayer lithologies. These folds are too numerous and too small to be represented in our cross sections. The large scale folds can be traced through large parts of the studied area and, when possible, fold axial surface traces have been mapped. It was thus possible to show that for many of these folds one has rather large domains with uniform dip separated by relatively small fold hinge zones.

6 -TRANSITIONS FROM THE PRÉALPES MÉDIANNES TO NEIGHBORING STRUCTURAL UNITS.

The transition from the Préalpes Médiannes nappe to the underlying and overlying structural units (fig. 37, -39) is marked by *mélange* units which are often termed *wildflysch* in the Alpine literature (Homewood & Caron 1982) (see also field trip 3). The contacts are characterized by special types of units and rocks such as *mélanges* and *rauhwackes* (*cornieules*) associated with anhydrites. For the Préalpes Médiannes two groups of *mélanges* can be distinguished: the supra-Préalpes Médiannes *mélanges* mostly related with the emplacement of the nappe Supérieure and the *mélanges* infra-Préalpes Médiannes (they are mostly supra-Helvetic) that are linked to the Préalpes Médiannes emplacement.

6.1 THE MÉLANGE ZONES

The Préalpes Médiannes rest on top of the Helvetic nappes in the S and the Subalpine Molasse and flysch in the N. All along this contact we observe the "Ultrahelvetic" and the "Zone Submédiane" units which have been interpreted in recent publications as *mélanges* similar to those formed in accretionary prisms. They were subsequently strongly overprinted by Alpine tectonics, especially during nappe emplacement

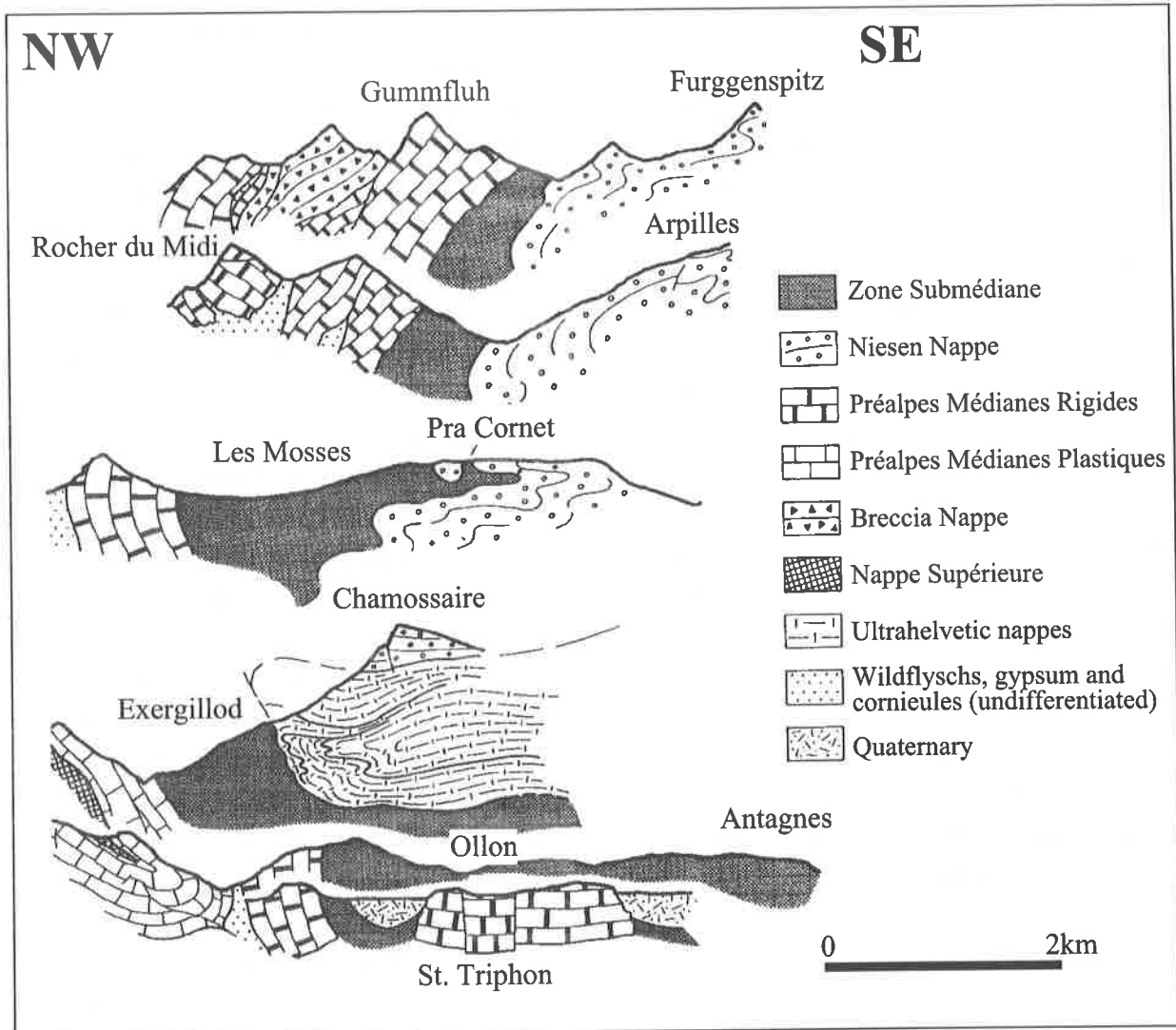


Fig.39 - Tectonic sections of the Zone Submédiane and its relation with the Préalpes Médiannes Rigides and the Niesen nappe. (Modified from Weidmann et al. 1976).

(Jeanbourquin 1994).

•The Zone Submédiane (Weidmann et al. 1976) is found below the basal décollement of the meridional part of the Préalpes Médiannes. It forms the contact with the underlying Niesen flysch nappe in the central and eastern Préalpes romandes and the Ultrahelvetic units in the western part of the Préalpes Médiannes along the Rhône valley. This zone is a mélange of blocks of various lithologies. Its origin has tentatively been located in the Valais trough domain, north of the Préalpes Médiannes depositional realm (Weidmann et al. 1976; Stampfli 1993; Jeanbourquin 1994). Very large outcrops of anhydrites and gypsum deposits are present along

the Rhône valley between Ollon and Bex. The Bex anhydrites, formerly attributed to the Ultrahelvetic units, can now be considered part of the Zone Submédiane. Gypsum-filled pressure shadows on large brecciated dolomite clasts (cm to m scale) and two fold axial directions developed in the intensely folded anhydrite bands, are evidences for important multiphase deformation.

The deformation, as well as the mixing of the mélange has largely been explained by thrusting along the Préalpes Médiannes basal décollement. Thus, the Zone Submédiane mélange is primarily tectonic in origin.

•The “Flysch with Couches Rouges lenses“

(Chaotic complex in fig. 10 of the introduction) is found on top of the Préalpes Médiannes as well as associated with the overlying Breccia nappe and/or Nappe Supérieure. These complexes can form rather continuous horizons and primarily contain characteristic slivers of hemipelagic Couches Rouges Formation.

•The “Ultrahelvetic” can be subdivided into upper and lower Ultrahelvetic mélanges. The lower ones are always associated with a specific Helvetic nappe, whereas the upper ones do not show this systematic correspondence and their kinematic development appears to be closely related to the evolution of the Préalpes nappes stack (Jeanbourquin & Goy-Eggenberger 1991; Jeanbourquin 1992; Jeanbourquin et al. 1992; Jeanbourquin 1994). They are formed by a succession of units and slivers (or nappes) in each of which virtual stratigraphic sequences have been defined (Badoux 1963; Badoux 1965; Lempicka-Münch & Masson 1993; Lempicka-Münch 1996). Their paleogeographic home is located to the south of the Helvetic domain and in the external Valais realm. They are located between the Préalpes nappes and the Helvetic nappes in the meridional Préalpes (see also **stop 3C**). The Ultrahelvetic mélange is also found beneath and in front of the Préalpes Médiannes nappe.

6.2 THE NIESEN NAPPE

During this field trip (Ft6) we shall also visit outcrops of the Niesen nappe. The Niesen nappe is formed mainly by large scale folds, which in the Col des Mosses area form a large N-NW verging recumbent anticline (Ackermann 1986; fig. 37, 39). Folds at all scales are common and show strong variations in fold axes orientation. Rock cleavage is omnipresent but has not been investigated yet. At the base of the nappe a small imbricated structure with slivers of rocks predating the flysch deposits is developed. This structure is overthrust by the “main body” of the Niesen nappe.

This tectonic unit comprises rocks of Triassic to Cretaceous age (with basement slivers at its sole) with north Pennine and Ultrahelvetic affinities, as well as an important series of middle Jurassic syn-rift deposits of the Helvetic margin (see introduction). These ante-flysch sediments are found in slivers only at the base of the nappe and no continuous section is known. The syn-rift deposits have been recently studied by Ringgenberg & Tomassi (2000). They consist of the *Couches à Posidonies de la Grande Eau*.

This Dogger formation belongs to the front of the Niesen nappe. Its geodynamic context has been related to the Alpine Tethys rifting. This formation was deposited in a deep-sea fan environment (see **stop 6F**). Its base is of Aalenian age and reworked microfossils give a Bathonian age to the top of the formation. If we compare the *Couches à Posidonies de la Grande Eau* with the contemporaneous deposits of the Brèche inférieure (Brèche nappe, see **Ft3 stop 3D**), it is evident that these two formations are connected to the same event (the rifting of the Piedmont part of the Alpine Tethys ocean) and were laterally continuous in Middle Jurassic times and located at the foot of the rift shoulder. Their present position on the same cross-section in the Alps implies a duplication of the Jurassic south European margin through the Cretaceous opening of the Pyrenean rift system.

The Niesen-Flysch, a Pyrenean flysch in the Alps

The flysch sequence rests above an unconformity on the Mesozoic sequences and comprises 400-1300m of terrigenous clastic sediments: mainly conglomerates with basement clasts, sandstones and shales. The age of the sediments is Maastrichtian to Lutetian (from calcareous plancton and foraminifera). Several different depositional units can be distinguished from bottom to top (fig. 40): the Frutigen flysch; the Niesenkulm flysch; the Seron flysch and Chesselbach flysch (Tertiary), all of which are deposited in a slope

environment. The Tauben flysch, a separate structural unit, shows similarities with the Seron and the Frutigen flyschs. Paleocurrent directions (mainly from the South) and sedimentary petrography suggest a strong tectonic control of the source area and of the morphology of the southern flank of the basin during deposition.

The basal unconformity (e.g. erosional surface between the Langy member and the Cretaceous Niesen flysch) is due to a pre-Maastrichtian tectonic event (Lugeon 1938; Homewood 1974). This unconformity can be observed in the bend of the road at the junction between Les Mosses road and the Diablerets road (**stop 6F**). A pre-Maastrichtian tectonic event is quite plausible looking at the Turono-Senonian geodynamic situation (fig. 5) where the closure of the Pyrenean ocean/ Valais trough (Stampfli et al. 1998)

generated the first movement of inversion in the French external Alps (Devoluy region, Huyghes & Mugnier 1995) and major clastic input in Provence (La Ciotat conglomerates, (Philip et al. 1987)). Therefore these first Pyrenean movements could also have affected the Helvetic margin before the main Campanian-Maastrichtian inversion, bringing some parts of the pre-Flysch basement to erosion. The created relief was named Cordillère Tarine by Barbier (1948) it represents the inverted parts of the Helvetic margin, source of sediments supply for the upper Cretaceous-lower Tertiary flyschs. A second inversion event is represented by the middle Eocene Meilleret flysch (Homewood 1974)(outcropping in the Col de la Croix region) which most likely cannibalised the Niesen flysch deposits.

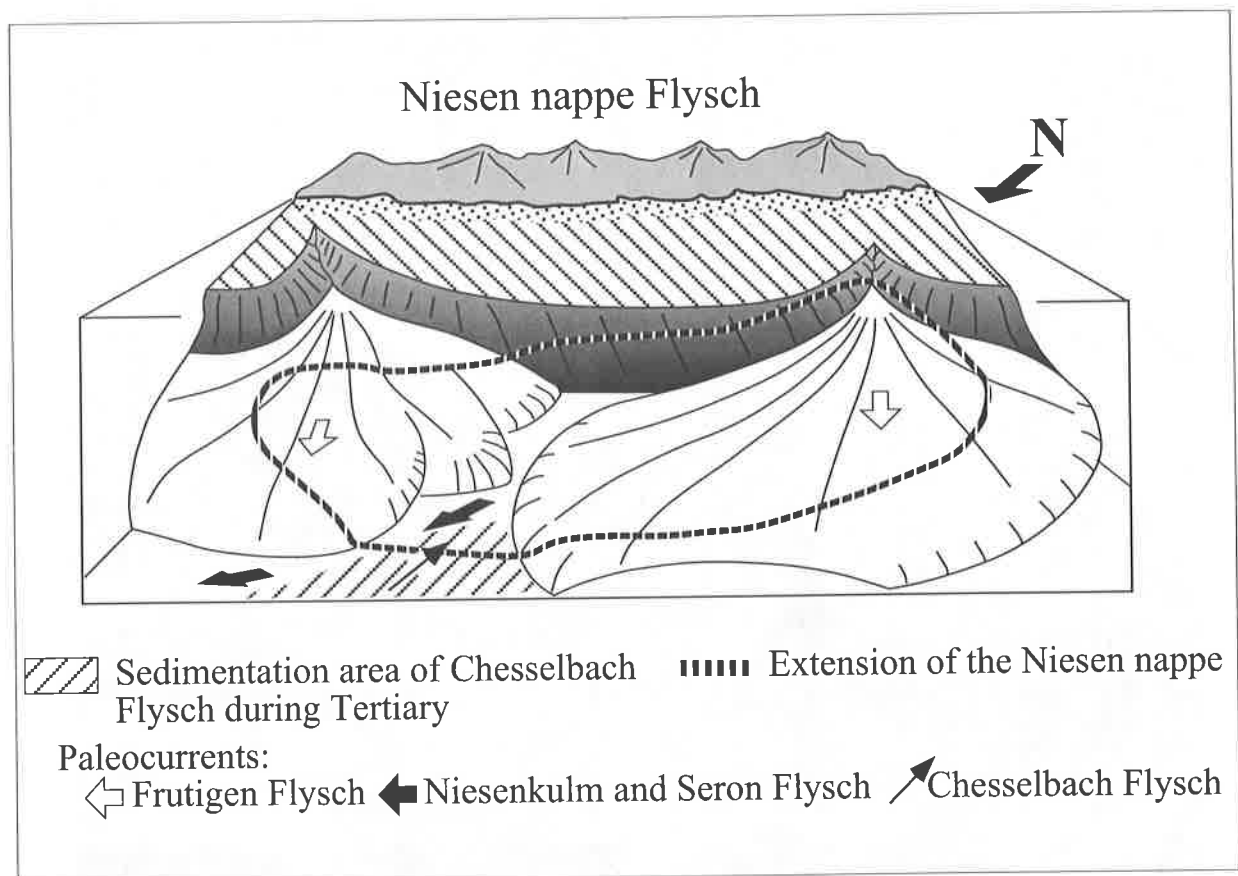
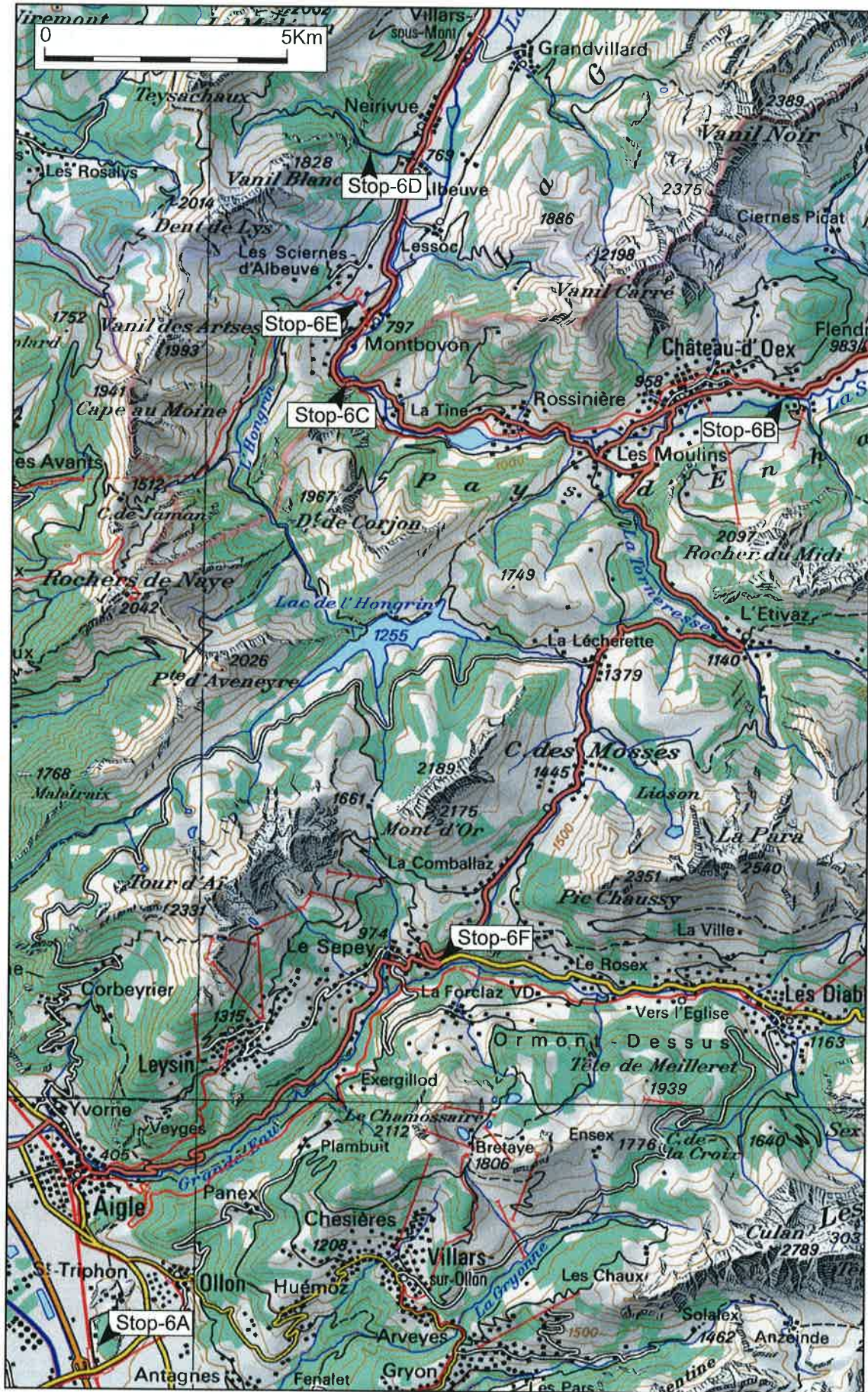


Fig. 40 - Simplified sketch of the Niesen flysch depositional environments as reconstructed from sedimentary analysis (modified from Ackermann 1986)



II: EXCURSION OUTCROPS

Stops are located on the topographic map on the opposite page, a structural map corresponding to the same area is found in annex (MAP 5).

1 - THE MÉDIANES DOMAIN

STOP 6A - SAINT TRIPHON (NEAR AIGLE), RHÔNE VALLEY: PRÉALPES MÉDIANES RIGIDES, TRIASSIC PLATFORM SEDIMENTS; ALPINE AND PRÉALPES TECTONICS.

Location: Andonces quarry, close to St Triphon railway station (560410/120580)

Tectonics : The Saint Triphon tectonic klippe forms an overturned portion of the Préalpes Médiannes Rigides (fig. 39). Its present day position in the Rhône valley, disconnected from the rest of the Préalpes is still subject to debate. The Saint Triphon locality is also well known for its paleokarst and paleotectonic features (Baud & Masson 1975): exposed are several karst cavities filled with clasts and other deposits showing inverted graded bedding (fig. ft6-A). Conjugate fault sets and stylolithes corroborate an extensional Late Liassic-Dogger event in this Briançonnais sedimentation domain associated with the rifting of the Alpine Tethys.

Sedimentology : - Middle Triassic (Baud & Masson 1975; Baud 1987): The carbonate series of Saint-Triphon are made of 600-700m of littoral, peri-littoral and lagoonal sediments, often strongly bioturbated with few bivalves. The series comprises two formations: the Saint Triphon Formation (250m) of Anisian age and the Champcella Formation of Ladinian age formed by banded dark limestones and dolomites. In the visited Andonces quarry the upper portion of the Saint Triphon Formation is exposed. The calcirudites, calcarenites and calcilitites are deposited in a barrier and peri-littoral environment. Graded bedding, oblique

laminations, crossed bedding indicate the development of ripple marks, sandy bars, dunes etc.

W and S of Aigle we can see the panorama of the Chablais Préalpes with their different nappes and the structure of the Préalpes Médiannes, as well as their Molasse and Ultrahelvetic substratum and the Helvetic nappes of the Dents du Midi (see field trip 3, stop 3C).

STOP 6B - GERIGNOZ, E OF CHÂTEAUX D'OEX: PRÉALPES MÉDIANES PLASTIQUES - RIGIDES, MIDDLE JURASSIC, PLATFORM DEPOSITS.

Location: from Châteaux d'Oex we take the road in the direction of Saanen, after the locality of Les Granges we go down a steep concrete road leading to Gérignoz through a tunnel and a bridge on the Sarine river (570860/140690). The outcrop is located just before the tunnel which is dug out in the Calcaires Massifs (fig. ft6-B).

Tectonics : The Châteaux d'Oex synclinorium is a complex zone located to the north of the Tour d'Aï - Gastlosen structural unit (fig. 37). This large synclinorium forms a topographic low that is filled with rocks belonging to the Nappe Supérieure. This tectonic unit is thrust on top of the Médiannes Plastiques and both are folded together. Between Châteaux d'Oex and Lake Hongrin (Col des Mosses area) several outcrops of Cretaceous sediments belonging to the Médiannes Plastiques indicate the existence of folded and thrust structures at shallow depth beneath the surface. In the area above the Rhône valley two synclines can be clearly differentiated. This area corresponds to a former paleogeographic high bordered by sedimentary basins to the N and the S (Baud & Septfontaine 1980; Septfontaine 1983). The inversion of this paleotectonically complex region has led to an even more complex Alpine structure (Septfontaine 1995).

To the north of the syncline we observe the Dorena anticline (fig. 37), located in the Vanil Noir relief (the large mountain, looking north from the parking lot before to go down to **stop B**). Its trend is NE-SW and it forms an important and continuous structure in the southern Médiannes Plastiques. The anticline has a steep northern limb and a flat to gently southeast dipping central part. The moderately to steeply dipping meridional limb forms a dip slope that extends for almost 25km in a very straight manner.

At the bottom of, or along this dip slope, we find the major thrust contact with the overlying Nappe Supérieure (Châteaux d'Oex syncline). Several distinct segments can be recognized: from NE to SW we have a periclinal fold termination dipping to the NE, followed by a sub-horizontal central fold segment and an en echelon periclinal termination to the SW (transition to the Tinière anticline). The NE periclinal termination of the Dorena anticline is cut by a series of vertical faults perpendicular to the fold axes. They mainly show

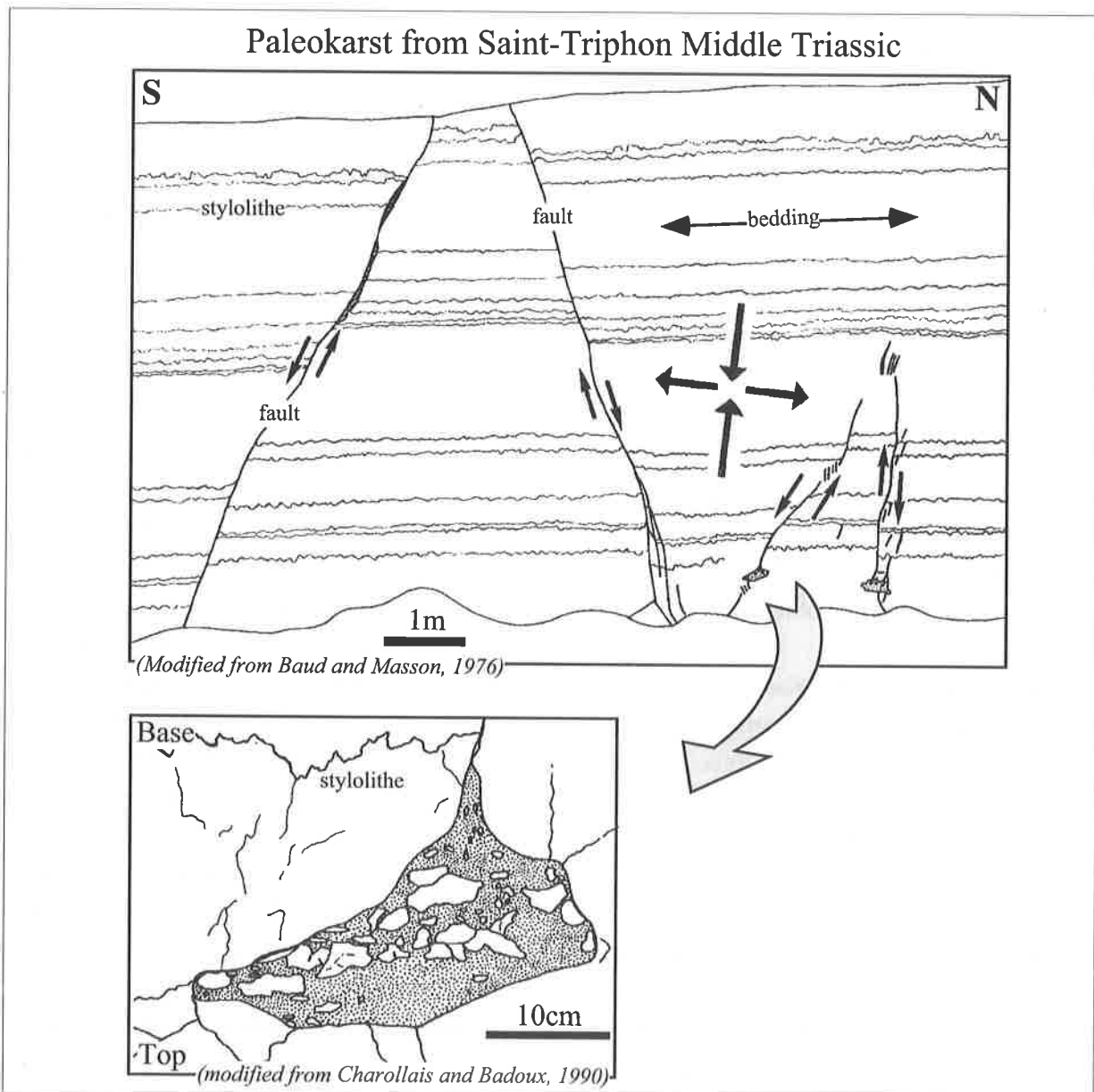


Fig. 16-A - Saint-Triphon karst and paleotectonic features: conjugate faults and stylolites indicate the general compression axes. Inverted grading in the karst filling indicates that the series are upside down.

normal offset. Bedding orientation is rotated from a rather flat position to steeper SE dips due to the normal movement of these faults. In its western portion the Dorena anticline is cut in its core by a thrust with a top-to-the-NW displacement.

South of the Châteaux d'Oex synclinorium we enter the realm of the Gastlosen-Laitemaire-Tours d'Aï structure. Developed as large fold in the Tours d'Aï area to the west, this tectonic structure changes to a SE dipping imbricate in the east (Laitemaire - Gérignoz area; fig.37). Another important syncline of the Nappe Supérieure forms the transition to the southern most portion of the Préalpes Médiannes: the Médiannes Rigides imbricates (Plancherel 1979; Mosar et al. 1996).

The trailing part of the nappe, the Médiannes Rigides, is a N/NW-dipping thrust sheet, although locally there are one or two minor imbricate thrusts (fig. 37). The imbricates are steeply dipping in the Rübli - Gummfluh area (the large mountain seen from the present stop when looking south). The major thrust sheets form fault-bend like folds that are cut by a large backthrust through the layers in the hanging-wall of their ramps. Because the fault-bend fold structure has been inclined to the NW

(its present day position), the backthrust, as well as associated splays, suffered a rotation thus appearing as normal faults with the meridional compartment downthrown by 200-400m. The backthrusts cuts not only through the Médiannes Rigides but also through the overlying Breccia nappe and Nappe Supérieure. Thus these two nappes have been emplaced on the Médiannes Rigides before the fault-bend fold and the associated backthrust developed (Mosar 1991). In the Mont d'Or and the Grande Eau, the Médiannes Rigides are reduced to slices consisting of series of Triassic age only (**Stop F**). Moreover, these slices (embedded in the Zone Submédiane) are overturned and dipping to the S-SE. A similar slice in the Rhône valley, the Saint Triphon klippe (**Stop A**), is in an upside down position and completely detached from the rest of the Préalpes Médiannes.

Sedimentology :

- Late Triassic (Baud et al. 1989): Intense and chaotic sedimentation processes formed limestone breccias with Triassic dolomite clasts. These probably reflect an episode of intense erosion during Late Liassic to Early Dogger times (collapse of

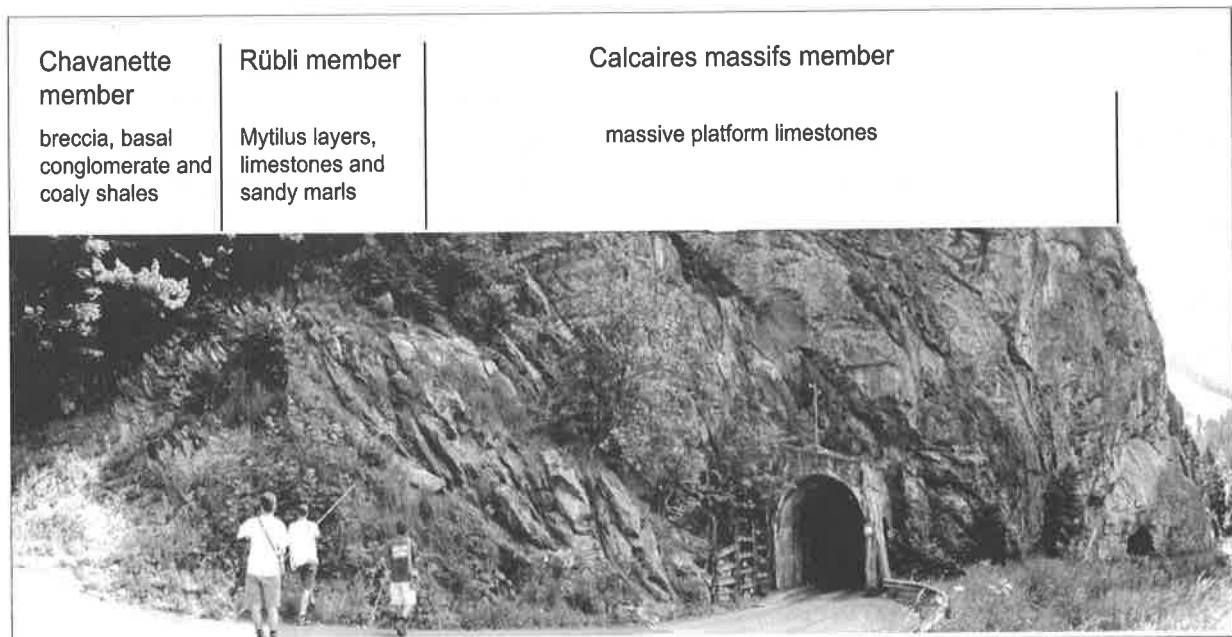


Fig. f16-B - Outcrop of the Couches à Mytilus (Mytilus layers) and Calcaires Massifs at Gérignoz just before the bridge on the Sarine River.

steep scarps).

- Early Jurassic (Lias): absent, non-deposition and/or erosion.

- Middle Jurassic (Dogger; Couches à Mytilus formation, Septfontaine 1983; Baud et al. 1989) : During the Middle Jurassic, the meridional part of the Préalpes Médiannes is represented by transgressive facies over middle Triassic sediments.

- Chavanette Member (Late Bajocian ?): coaly marls alternating with beds of yellow breccia. The reworked elements consist of dolomite and gray chert. The depositional environment is one of wooded swampy lowlands with locally coarse elements of continental origin (rivers).

- Rübli Member (Bajocian-Bathonian): at the top of the formation we find dark, massive lagoonal limestones interfingering with sandy deposits. They rest on top of more marine limestones. Below interlayered bedded limestones (mudstone) and fossiliferous (Mytilus) calc-schists can be seen. These deposits correspond to a tidal flat depositional environment with cyclic mud sedimentation. There were periodic emersions with vegetation cover.

- Calcaires Massifs (Bathonian to Early Cretaceous): these massive limestones are made of high energy platform deposits (grainstone to wackestone) and lagoonal mudstone.

Interpretation: we are dealing here with a relatively condensed section of shallow water deposits, transgressive southward on the Briançonnais rift shoulder. This Bajocian transgression is a witness of the thermal subsidence of the rift shoulder since that time, concomitant with acceleration of sea floor spreading in the Alpine Tethys. These condensed deposits contrast with basinal deposits found northward.

STOP 6C - PONT DE LA TINE (SARINE VALLEY): PRÉALPES MÉDIANNES PLASTIQUES, MIDDLE JURASSIC, BASINAL CANCELLOPHYCUS DEPOSITS

Location: The outcrop is located along the main road from Bulle to Château-d'Oex, just after the Sarine canyon a few kilometres upstream from Montbovon (560950/140700)

Tectonics : The outcrop is located in the subvertical eastern limb of the Sarine valley syncline (see description in **Stop D**) and forms the transition to the the Tsavas - Millets anticline and the Vanil Noir - Rochers de Naye syncline (fig. 37). It has a very important vertical to slightly overturned NW limb, as shown from the cleavage-bedding relationship (Pont de la Tine). The meridional limb (some 600m of Middle Jurassic limestones) is inclined to the SE and merges rather quickly into the southerly adjacent Vanil-Noir - Dent de Corjon - Rochers de Naye syncline.

The Tsavas-Millets anticline has a slightly box-shaped fold core with two axial surfaces that can be mapped in the field. The fold core is cut by a steeply north dipping thrust with top to the south displacement. The northwestern limb thickness is in excess of 2000m in the central Sarine valley section (1800m of middle Jurassic limestones) and thins along strike towards the NE as well as to the SW. Synsedimentary unconformities indicating north-westward opening deep sea fans emphasize the local overturning. Thus, bedding of the upper Jurassic sequence was dipping gently to the NW which is corroborated by N-NW directed deposition sense of re-sediments (see **Stop D**).

The Vanil-Noir - Dent de Corjon - Rochers de Naye synclines forms a large though discontinuous structure. The three names stand for three distinct portions (from NE to SW) of a larger structure separating the Tsavas-Millets anticline and the Dorena-Tinière anticline (**Stop B**). The Vanil Noir-Vanil Carré segment is the longest which is followed S of the Sarine river by the Dent de Corjon portion (S of road from Pont de la Tine to Château d'Oex). Overall these synclines have a tight (Rochers de Naye) to open (Dent de Corjon) box fold geometry, with a flat bottom where many second

order small scale folds developed in lower Cretaceous horizons. All three portions of the Vanil-Noir - Dent de Corjon - Rochers de Naye syncline form in their central section topographically elevated features. This type of inverted relief is characteristic of this part of the Médiannes Plastiques.

Sedimentology : - Middle Jurassic (Dogger, limestone and clay; Calcaréo-argileuse Formation, Septfontaine 1983): outcropping at this location is the Staldengraben Formation (Bathonian-Callovian) formed by argillaceous Zoophycus limestone (fig. 38) over a distance of 200m. This marly limestone is strongly schistose, bioturbated and pyrite is omnipresent. Ammonites and belemnites can be observed. In the upper part (Callovian) one can find sandy resediments interlayered with turbiditic structures. Flute-, groove- and load structures together with borrows and feeding traces can readily be seen at the base of the beds. These sediments have been deposited into the so-called "Cancellophycus domain".

-Late Jurassic (Malm, Calcaires Massifs Formation, (Heinz 1985; Heinz & Isenschmid 1988; Baud et al. 1989): The base of the Calcaires Massifs Formation is accessible downstream of the bridge over the Sarine River. The contact with the limestone and clay formation of the Dogger is not visible. The base of the Malm series typically consists of interlayered nodular and resedimented limestones (**stop D**).

STOP 6D - GORGES DE L'ÉVI, W SARINE VALLEY: PRÉALPES MÉDIANES PLASTIQUES, LATE JURASSIC - CRETACEOUS, BASINAL DEPOSITS.

Location: On the road from Bulle to Chateau-d'Oex, one takes a small road at the end of the Neirivue village on the right, leading to a gorge cut in the Jurassic limestones just west of Albeuve (560970/150220).

Tectonics : This outcrop is located in the western limb of the Sarine valley syncline (fig. 37). This syncline runs from the Dent de Jaman east of Montreux (Lake Lemman) along the lower Hongrin river towards the NE. It widens to a width of almost 5km in the Sarine valley. From Albeuve northeast, its orientation changes to a more E-W direction and the structure runs towards the ENE past the Dent du Chamois towards the Jogne river. By its dimensions the Sarine or Intyamou syncline forms the most important structure N-NE of the Rochers de Naye syncline. Its topographic culmination is around 2000m for the Jurassic-Cretaceous limit. The center of this structure reaches a depth of about -1km. The western limb of the syncline is dipping with 30-40° towards the SE and brings the Late Jurassic-Cretaceous limit from an elevation of 1800-2000m in the Vanils ridge down below the Sarine valley (750m). Drill holes and studies of the thickness of Quaternary sediments (Pugin 1988) show that the top of the Cretaceous-Tertiary sediments is at an altitude of about 400m in the center of the synform, between Montbovon and Lessoc. The Jurassic-Cretaceous boundary can thus be put at 200-300m, assuming a 100-200m thick sediment pile for the whole Cretaceous-Tertiary sediment pile. Inside the syncline two topographic relief culminating around 840m are formed by a series of tight folds in the Late Cretaceous layers. The eastern limb of the Sarine syncline is sub-vertical. The Late Jurassic-Cretaceous limit outcrops again beyond the adjacent Tsavas-Millelets anticline at an altitude of 2200m in the Vanil Noir syncline.

The NE continuation of the Sarine syncline extends beyond the Dent du Chamois, east of the Sarine river. The syncline clearly splits into two distinct smaller synclines separated by a thrust-associated anticline (Chenevart 1945; Chatton 1974). Beginning E of the Sarine river the synclinal bottom (Cretaceous-Jurassic boundary) raises to an altitude of about 1200m south of the Dent de

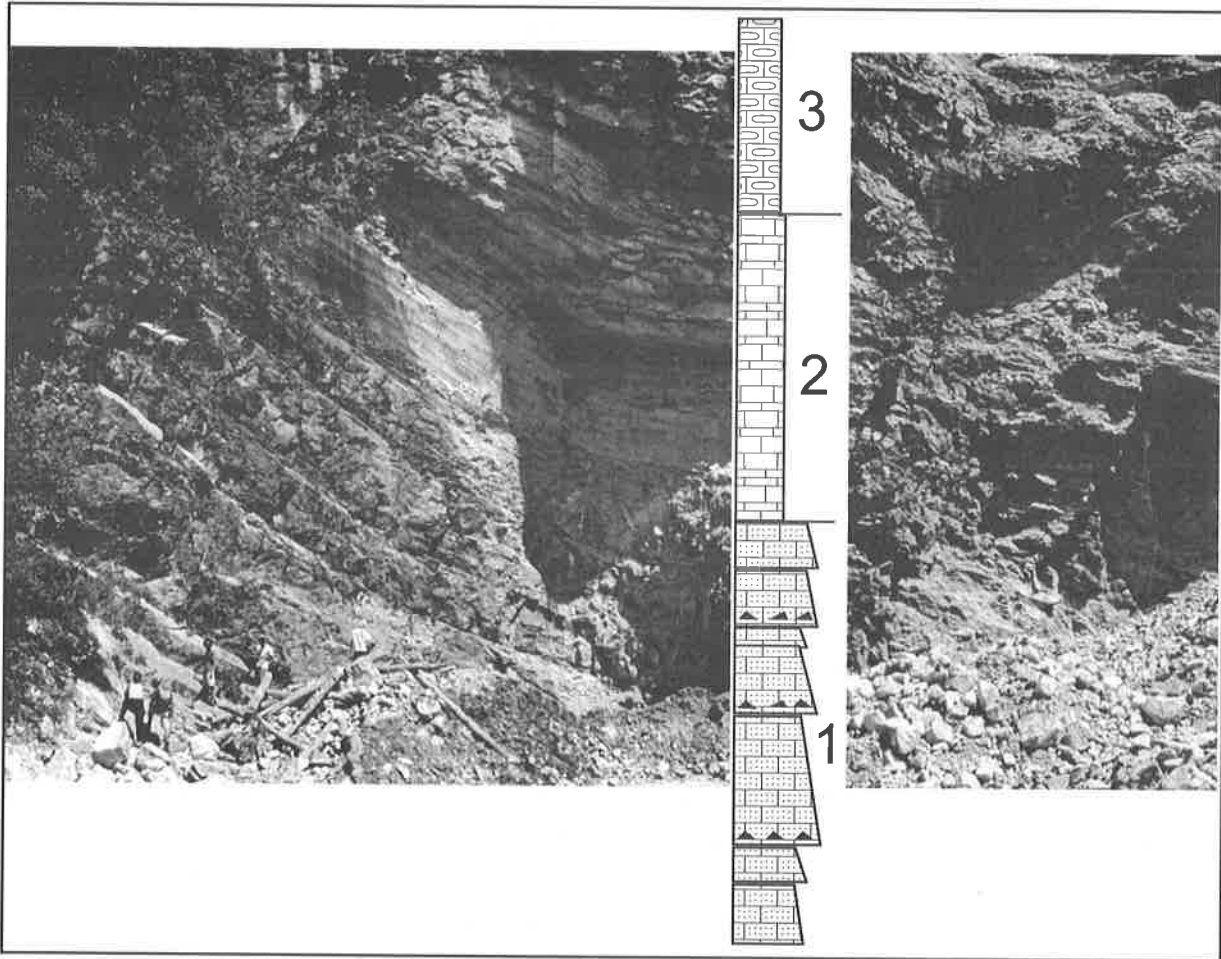


Fig. 16-D - Outcrop of the Calcaires massif in the quarry in the Gorges de l'Évi: 1, resediments (late Kimmeridgian - early Tithonian); 2, Albeuve Beds: green and beige bio-micritic nodular limestone. 3, Calpionnellidae Limestone with cherts (Tithonian)

Chamois. To the SW also the syncline splits up into two minor 30-40° NE dipping synclines. The Jurassic-Cretaceous boundary plunges from a height of about 2000m at the Dent de Hautaudon towards the Sarine valley center, where fold axis have shallow to sub-horizontal plunges.

In the northern part of the syncline, a small klippen of Nappe Supérieure can be found. Thus this large structure forms a deep synform bordered on each side by steeply dipping panels. The center and the lower part of the eastern limb are cross cut by two thrusts that uplift the southeastern part. These thrusts die out toward the Dent de Jaman where the main syncline splits into two minor synclines.

Sedimentology :

- Late Jurassic (Malm, Calcaires Massifs For-

mation, Heinz 1985, Heinz & Isenschmid 1988): The Calcaires Massifs formation is exposed in a quarry that is exploited for aggregate. West of Albeuve the upper part of these limestone represents the distal facies of a carbonate platform ramp of the external Médianes Plastiques during the Late Jurassic. From top to bottom one can observe:

— Calpionnellidae Limestone (Tithonian): fine, well stratified, micritic limestone beds with numerous bedded and nodular cherts. Several meters below the top of the formation a small centimeter thick calcarenite layer can be observed. Beds are thinner towards the top of the formation and “dikes” of chert can be seen injected along fault planes. Load structures are common and can be seen at the base of the beds.

— Albeuve Beds: green and beige bio-micritic

nodular limestone. The nodular aspect is underlined by green anastomosing clayey seams highlighting cm thick limestone nodules. Centimeter thick green and gray clay deposits can be recognized along undulating limestone bed surfaces.

— resediments (Late Kimmeridgian - Early Tithonian): they are formed by thick limestone layers. One can observe a fine graded bedding. The intraclasts are formed by carbonate debris resulting from the erosion of the Briançonnais platform further to the south. These sediments are allochthonous in the Préalpes médianes plastiques basin.

— nodular limestone: represent the autochthonous sediments in the basin and are formed by interbedded marls and fine-grained nodular limestone.

- **Early Cretaceous** (Neocomian, Calcaires Plaquetés Formation, (Boller 1963). The outcrop consists of small cliffs along the road leading from the parking to the massive limestone quarry. The succession consists of grey fine-grained micritic limestone interbedded with dark cherts and marls. Dark spots within the limestones result from bioturbation. Few calcarenite are present. Chert layers are absent at the top of the formation. The age of the succession is Berriasian to Barremian, based on a rich pelagic fauna of ammonites, belemnites, calpionellidae, planctonic foraminifers and radiolarians. The sediments are strongly folded making their thickness calculation difficult. In the external zone, between 100-150m of sediments are estimated to be present. The succession thins towards the centre of the basin and pinches out completely to the south-east in the Châteaux d'Oex syncline. This pinching out can be explained by deposition along a slope close to an ancient coastline as documented by calcarenites with large neritic bioclasts (echinoderms, corals, bivalves) and condensed red limestone.

STOP 6E - HONGRIN RIVER, SW OF MONTBOVON (SARINE VALLEY): PRÉALPES MÉDIANES PLASTIQUES, CRETACEOUS SEQUENCE.

Location: On the road from Bulle to Chateau-d'Oex, one takes a small road just before the junction between the Sarine river and the Hongrin river, on the right between Albeuve and Montbovon, leading to the Comba d'Avau (560960/140900), then the pedestrian track going down to the Hongrin River. The first outcrops (middle Cretaceous) are on both side of the river just up-stream from the bridge:

- the Aptian siliceous limestone with radiolarians are on the left side of the river,
- the Albian black marls on the right hand side.
- Turonian planctonic grey limestone are found a few hundred meters away along the track as one reaches a small tunnel going under the railway, here is the top of the Intyamou formation.
- characteristic red coloured outcrops of the Couches Rouges with the hard-ground described below, are found as one continues along the pedestrian track toward Montbovon. After passing a few farm houses, one reaches an asphalted road again and a few hundred meters away, after joining a second road red limestone outcrops are visible (560920/140830), and can be followed along the path going down to Montbovon.

Tectonics : This series of outcrops is located in the central part of the Sarine valley syncline (see description in **Stop D**).

Sedimentology :

- **Middle Cretaceous** (Intyamou formation, fig. 38, Python-Dupasquier 1990):

Comba d'Avau Member (Aptian-Early Albian) located in the Gruyère syncline, this series attains thicknesses up to 35m. It is always in conformity with the underlying Calcaires Plaquetés Forma-

tion. The formation outcropping in the Hongrin river bed at Comba d'Avau, consists of marl and black, bioturbated marly limestone containing pyrite, interbedded with laminated marly limestone (biomicrites). Interfingering limestone and siliceous calcarenites with flat or mamelonated bedding surfaces occur towards the top of the series.

The Chällihorn member (Late Albian-Turonian) is found in the cliff dominating the Hongrin River to the south and along the track towards the railway tunnel. It consists of micritic limestone and marlstone of grey colour rich in planctonic fauna, some darker marly interval and siliceous levels are also present.

Pelagites and hemipelagites of Aptian to middle

Turonian age were deposited on a ramp inclined towards the north. Sedimentary structures suggest a depositional environment from storm dominated deeper areas (in the north) through ramp deposits to sediments locally cut by mud slides (to the south).

- **Early Tertiary - Late Cretaceous** (Couches Rouges Group, Guillaume 1986):

The Couches Rouges Group rests with stratigraphic disconformity on a substratum of three different formations (fig. 38): the Intyamou Formation to the N, the Calcaires Plaquetés Formation and the Calcaires Massifs Formation to the S. This unconformity is marked by a stratigraphic gap reflecting either erosion or non deposition. At their

Pra du Pont Hardground Profile

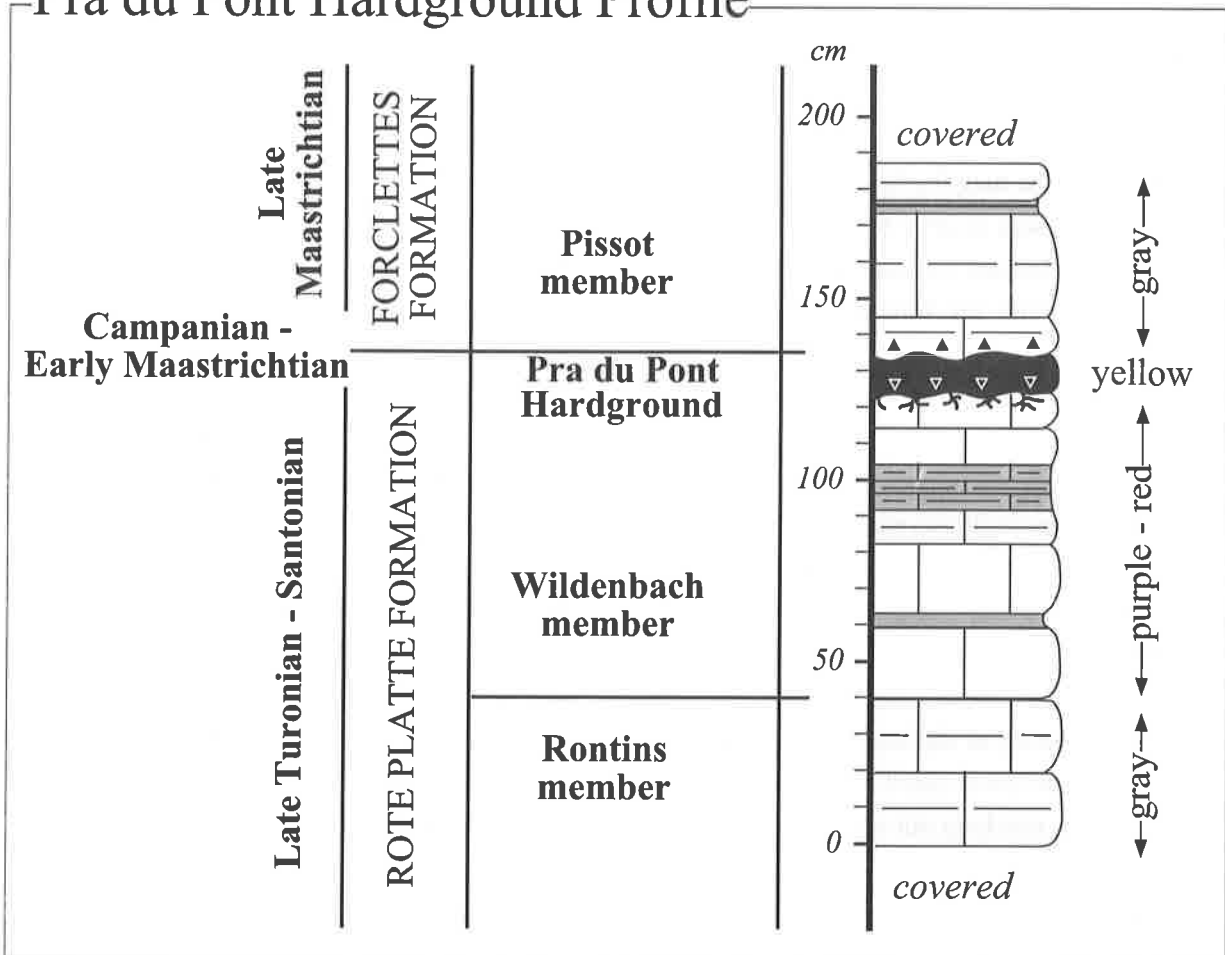


Fig. f16-E - Description of a stratigraphic sequence as observed in the Couches Rouges in the Sarine valley or at Pra du Pont in the Hongrin river valley (modified from Guillaume 1986).

top the Couches Rouges Formation rapidly passes to flysch deposits closing the sedimentation history in the Préalpes Médiannes. The series observed in the Hongrin valley are formed by:

- Rote Platte Formation (Late Turonian-Santonian): interbedded limestone and grey, red and green wackestone or calcareous mudstone. These sediments are transgressive and terminate an erosional gap outlined by a mineralized hardground or an irregular erosional surface.

- Pra du Pont hardground (fig. ft6-E) (Campanian - Early Maastrichtian): mineralized and silicified hardground with irregular basis penetrating the underlying limestone. Contains glauconite grains.

- Forcettes Formation (Late Maastrichtian to Early Eocene): gray packstones and grainstones, rich in microfauna. The fine bedding (5cm) shows evidence of sedimentary structures, laminations, ripples and graded bedding. Carbonates are more abundant than argillites; locally the limestones are interbedded with shales. Some siliceous beds are rich in radiolarians.

2 - THE NIESEN DOMAIN

STOP 6-F COL DES MOSSES: ZONE SUBMÉDIANNE, NIESEN NAPPE.

Location: On the road from Aigle to the Col des Mosses at the intersection with the road leading to Les Diablerets, ESE of Le Sepey. The top of the section with the contact between the Niesen flysch and the massive Jurassic sandstones is exposed in a quarry just at the cross-road (the series is overturned!) (571300/134400), the rest of the section is along the road, the base of the section is located close to the crossing to the old road and the bridge on the Les Diablerets road (571625/134550).

Tectonics : The region of Le Sepey-La Forclaz

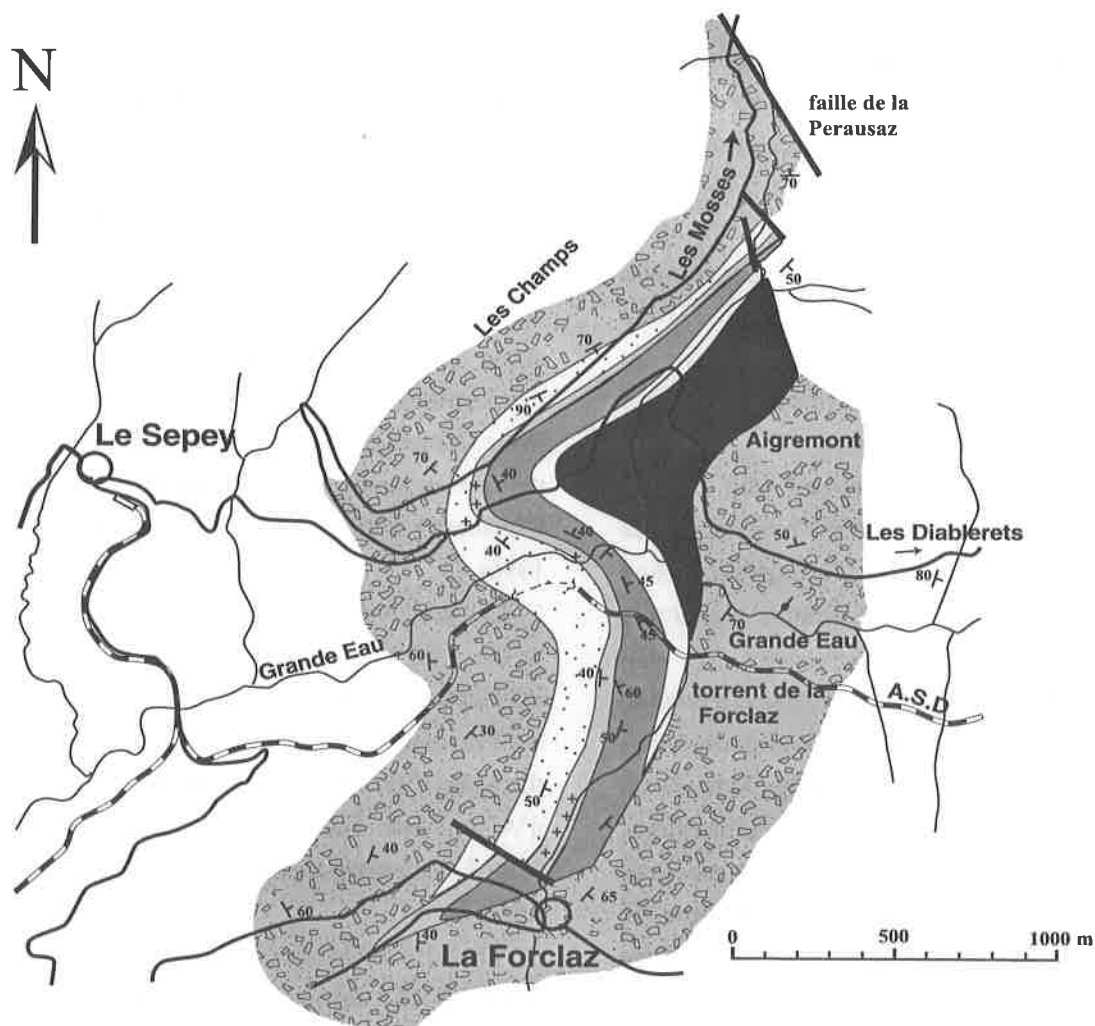
(VD) in the Prealps, contains in the core of the front of the Niesen nappe a Dogger detritic formation called the *Couches à Posidonies de la Grande Eau* (Lugeon 1938 and Badoux & Homewood 1978) in clear stratigraphic contact with the Cretaceous Niesen Flysch. The formation is part of the Murgaz tectonic slice in which some series show the Cretaceous Flysch overlapping directly on the Triassic or on the Malm (Mac Connel 1951). The pre-Flysch basement of the Niesen nappe also contains lower and upper Cretaceous series in some places (Ackermann 1986)

Located to the SE is the Niesen flysch nappe; to the NW we can see the Mont d'Or tectonic imbricate made of Triassic sediments belonging to the Préalpes Médiannes Rigides (fig. 37). Beyond the Mont d'Or structure, to the W, lies the Tours d'Aï anticline. The Tours d'Aï anticline connects to the Leysin syncline which is cored by the Nappe Supérieure.

The different Médiannes Rigides imbricates (fig. 36) are mostly surrounded by terrains belonging to the Zone Submédiane (fig. 37), except where covered by the Breccia nappe. Between the Gummfluh-Rübli imbricates and to the N and W of the Mont d'Or inverted imbricate, the Zone Submédiane (Lombard et al. 1974; Weidmann et al. 1976) is in contact with the Nappe Supérieure.

Sedimentology of the Jurassic syn-rift sequence: The studied formation has been divided into 6 members (« Schistes à miches », Grès et marnes de la Raverette, Conglomérats du Leyderry, « Grès de passage », Marnes et calcaires du torrent de La Forclaz, Grès de Langy) by (Badoux & Homewood 1978). Ringgenberg & Tomassi 2000 have renamed the formation and the members formally as follows (figure ft6-F2):

The *Grande Eau formation* is divided in 6 members:



- | | | | |
|--|---|-----|--------------------------------|
| | Flysch du Niesen (Crét. sup)
Niveau des "calcaires blancs" | | Conglomérat du Leyderry |
| | Grès de Langy | | Grès et marnes de la Raverette |
| | Marnes et calcaires du torrent de la Forclaz et Grès de passage | | Schistes à miches |
| | | +++ | Gisements à Posidonies |

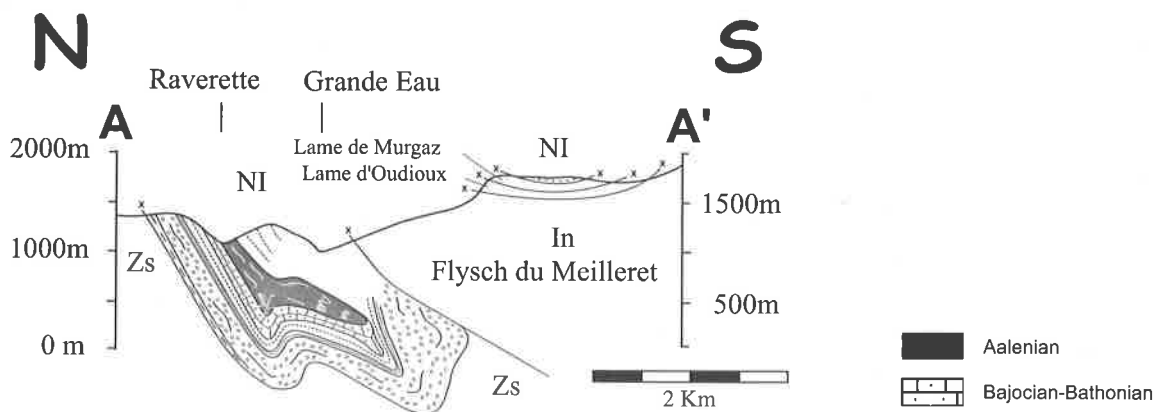


Figure ft6-F1 - Geological map and geological cross-section of the Niesen nappe in the La Forclaz region, modified from Badoux and Homewood 1978.

- schistes à miches member
- Raverette member
- Leyderry member
- transitional member
- Forclaz member
- Langy member

Schistes à miches member

This member is a characteristic facies of the Zone des Cols that can be found in many places.

It is a black marl full of micas containing calcareous, cherty and pyritic concretions that gave sometimes ammonites and Belemnites. Homewood and Badoux have found a specimen of *Leioceras* giving an Aalenian age to the formation.

At the top of the series, one can find a 2m chaotic level with a matrix of the same composition as the rest of the schiste à miches, but containing segmented beds of micaceous sandstones. It can be interpreted as a mudflow or

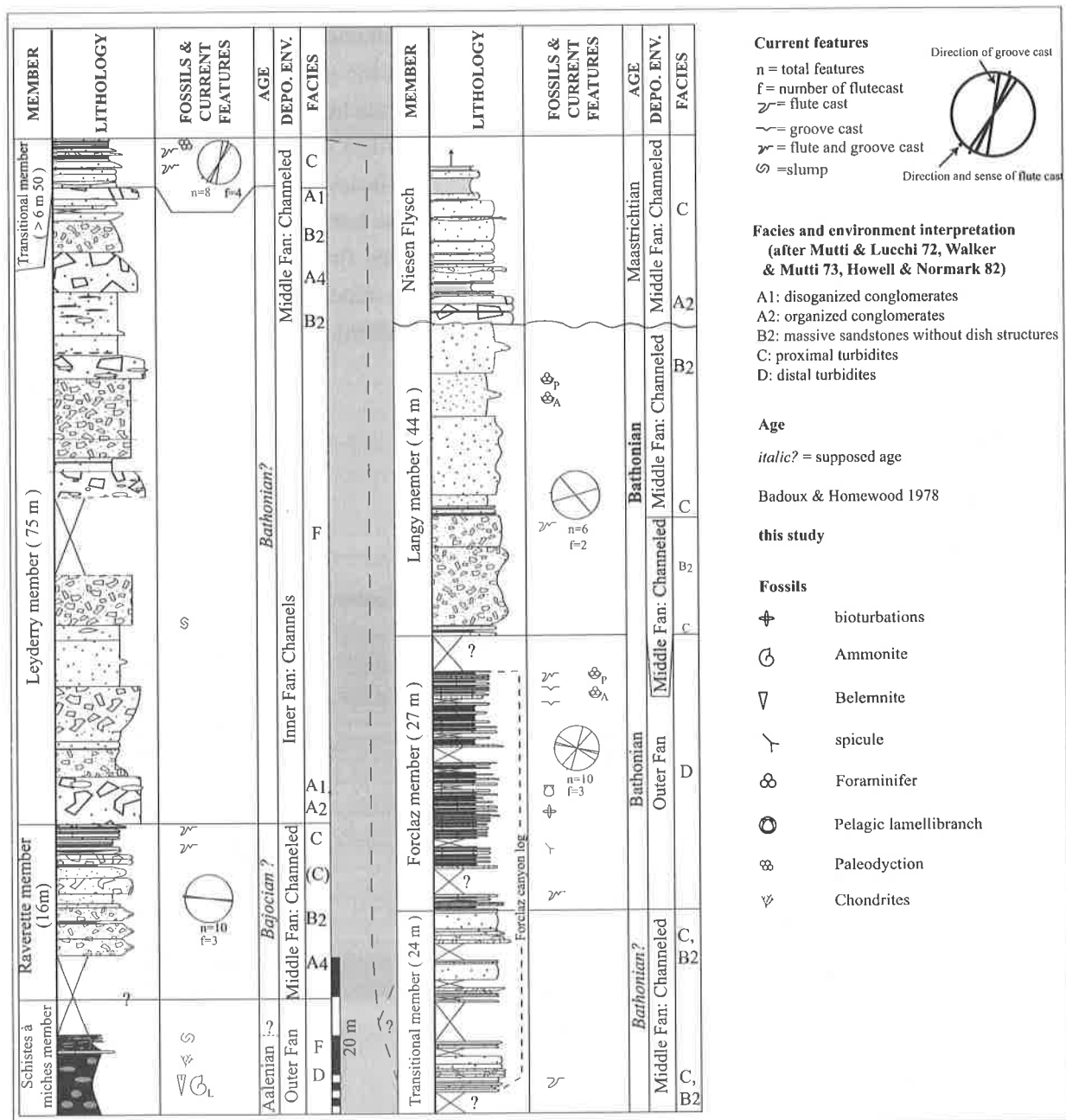


Fig. ft6-F2 - Synthetic lithological section of the Grande Eau formation, modified from Ringgenberg and Tomassi 2000.

a slump.

The transition to the Raverette member is not visible, but just before the slump, there is a calcareous and micaceous polygenic bed of sandstone with Chondrites that looks like the Raverette member. Therefore, the transition seems to be progressive and there is no reason to put a tectonic contact between those two members.

These shales have been deposited in an anoxic environment above the CCD. They can be interpreted as resulting from the dissolution of the Triassic and Liassic platforms in Toarcian time (Favre & Stampfli 1991). The top of the series is characterized by some turbiditic events and slumping of the upper beds, marking the onset of coarser detritic activity.

Raverette member

This member, which is 16m thick on the Sepey-Diablerets road and around 40m in the Grande Eau

canyon, is composed at its base by calcareous microconglomeratic metric massflows with millimetric to centimetric elements, separated by thin layers of micaceous marl and calcareous sandstones. Going up in the series, the size and granulometry of the beds become finer, Tab Bouma sequences and graded beddings occur. We often found mud balls forming bed-like levels limits.

In the matrix, the main elements are represented by quartz (very often polycrystalline), feldspar, black limestones with spicules and dolomite. The cement is dolomitic, probably of secondary origin.

The detritic event initiated at the top of the "schistes à miches" became more important with the deposition of deca-centimetric to metric massflow (facies A4 and B2) and turbidites (facies C). The transition from B2 to C indicates a thinning- and fining-upward sequence, which is typical of a middle fan (channeled suprafan from Walker & Mutti). This implies the installation of a

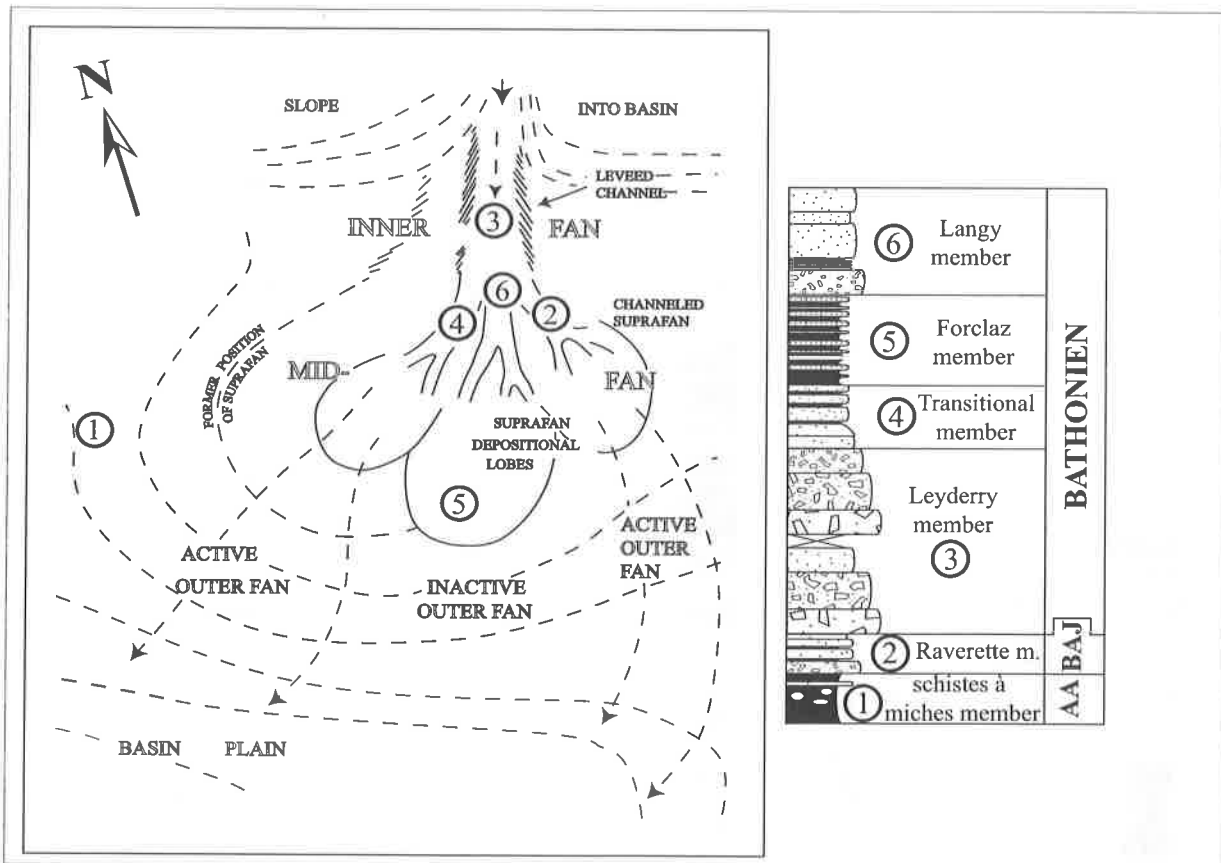


Fig. f16-F3 - Depositional environments for the middle Jurassic Grande Eau formation, modified from Ringgenberg and Tomassi 2000.

deep-sea fan.

Leyderry member

These polygenic conglomerates are 50m thick on the Sepey-Diablerets road and less than 25 m in the Grande Eau canyon. They have no internal structures (disorganized conglomerates) and elements are decimetric. The main elements are dolomite, gneiss, quartzite, black limestones, rhyodacite, dolerite, rhyolite, green sandstones, volcanoclastic rocks and green rocks of probable metamorphic origin. The calcareous gritty matrix looks like the Raverette member but we also find contemporaneous resedimented oolites.

In analogy with the Infra-Niesen series in which Lempicka-Münch (1996) described Bathonian oolites, the presence of oolites in this member could mean a Bathonian age. A Bajocian age can be proposed for the Raverette member in which no oolites are found.

This member comprises conglomerates of A1 facies, a chaotic slump of F facies that can be interpreted as the infill of an inner fan channel. At the top, thickness and granulometry decrease, we found graded bedded pebbly sandstones (facies A4). This progressive removal of the sediments source could be explained by a relative high-stand which could correspond with the main Bathonian high-stand. This trend of thinning and fining upward continued until the deposition of La Forclaz member, which is almost basinal.

Transitional member

This member forms the transition between the Leyderry and the Forclaz members. It is more than 6m thick on the Sepey-Diablerets road and 26m in La Forclaz canyon. It is formed by metric and decimetric calcareous turbidites, which consist of quartz, feldspar and many micritic clasts containing echinodermic fragments, spicules, foraminifers (*Lenticulina*, *Trocholina* and *Syphovalvulina*) and oolites with quartz and feldspar cores. As in

the Raverette member, the granulometry of the beds become finer, and their thickness decreases upward. Tabd Bouma sequences can be identified. The classical turbidites (facies C) of this member shows a lot of graded bedded and current features. We also have found *Paleodyction* traces, which characterize a deep environment (c. 2000m) and the member shows a thinning- and fining-upward sequence. All those elements allow the transitional member to be placed in the channeled suprafan.

Forclaz member

Good outcrops of this member are only found in La Forclaz stream (571600/133325). It is 30 to 50m thick and composed of turbiditic sandy limestones which alternate with marls. The proportion sand/marls is 1:1 and Tabc sequences are present. The detritic elements point to a more basinal environment than for the other members, e.g.: limestone with little quartz, feldspar and white mica but a lot of spicules, recrystallized echinoderms, *Lenticulina*, *Nodosoaria* and little or no resedimented platform fragments. Marls interbeds are sometimes full of *Posidonia bronni* or *bositra* and ichnotraces. Badoux & Homewood (1978) found *Protopenneroplis striata* (Weynschenk) and *Archeosepta platierensis* (Wernli) which indicates a Bathonian age. The transition to the Langy member is gradual, the bed thickness and the detritic fraction increase upward.

Here, distal turbidites alternate with hemipelagic marly intercalations (facies D). This implies a suprafan depositional lobe position. Actually, the important calcareous proportion, the low detritic fraction and the absence of platform resediments, clearly show that this member is more distal. This fits well with the Bathonian high-stand which could have completely drowned the source of influx. At the top, turbidites become more sandy and less calcareous before the transition to the Langy member which comprises a lot of platform resediments.

Langy member

The base and the middle part of this member consist of decimetric beds of coarse calcareous sandstones and argillaceous intercalations. The rest is made of metric to decametric grain flow of coarse calcareous sandstones and calcareous microconglomerates without intercalations. The elements are made of quartz, feldspar and contain up to 30% of limestone clasts, principally oolites with quartz and feldspar cores but also echinodermic fragments and foraminifera (*Protopenneroplis striata* and *Archeosepta platierensis*) which indicate a Bathonian age. In the microconglomerates, the detritic fragments are very diversified: gneiss, quartzite, dolomite, black limestones, volcanoclastic rocks and some rhyolites.

The base and the middle part are formed by classical turbidites (facies C). The rest are massive grain flow (facies B2) with no clay. This member could be placed in the channeled suprafan.

As shown in figure ft6-F2 the main current features are N-S and E-W, the sense is from NNE to SSW or from E to W. Regarding the general situation of the Helvetic margin, we assume that NNE-SSE is the main direction of influx and the E-W directions are due to the channeling dispersal.

The petrological study of the clasts allows the stratigraphy of the source to be reconstructed:

- Bathonian: carbonate platform
- Liassic: black limestones with spicules
- Triassic: dolomitic, calcareous and white quartzites
- Permo-Carboniferous : volcanic (rhyolites, rhyodacite, dolérite, welded tuff), green sandstones
- Hercynian basement : gneiss fragments

Interpretation

- The Grande Eau formation has been sedimented in a deep-sea fan environment, except

the schistes à miches member, characteristic of a large part of the south European platform.

- The erosion of layers from basement to Dogger show that the source of detrital influx was a submarine rise which has been uplifted for a long time before to be colonized by a Bathonian platform. Furthermore, the reworked elements of this platform (echinoderms fragments, oolites, foraminiferas) can be found in the Grande Eau formation and in the more external Infra-Niesen units (Lenk nappe, Lempicka 1996): the stratigraphy of the Lenk nappe (Blaue Schüpfe and Mulkerblatten series) shows 20-50m of fine argillaceous sandstones (Bajocian), 30-60m of sandy limestones (Bathonian) with 3m of microconglomerate at the top. Therefore, a submarine rise separated two distinct environments, a shallow one represented by the Infra-Niesen units with some detritic and organogenic influx and another one, much deeper, represented by the Grande Eau formation. Therefore, this submarine rise is considered as the separation between the South Helvetic rim basin domain and the Piemont rift domain, and interpreted as the Helvetic rift shoulder.

- As seen above, the Bathonian ends with microconglomerates, then an erosional surface takes place (probably submerged, H. Masson oral communication) followed by Oxfordian shales. The Callovian starvation is more likely due to the final drowning of the rift shoulder following rapid thermal subsidence, the Oxfordian shales corresponding then to the distal progradation of the Helvetic shelf. However starvation lasted for a longer period as the larger part of the shelf progradation was trapped in the Helvetic rim basin

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Annexes

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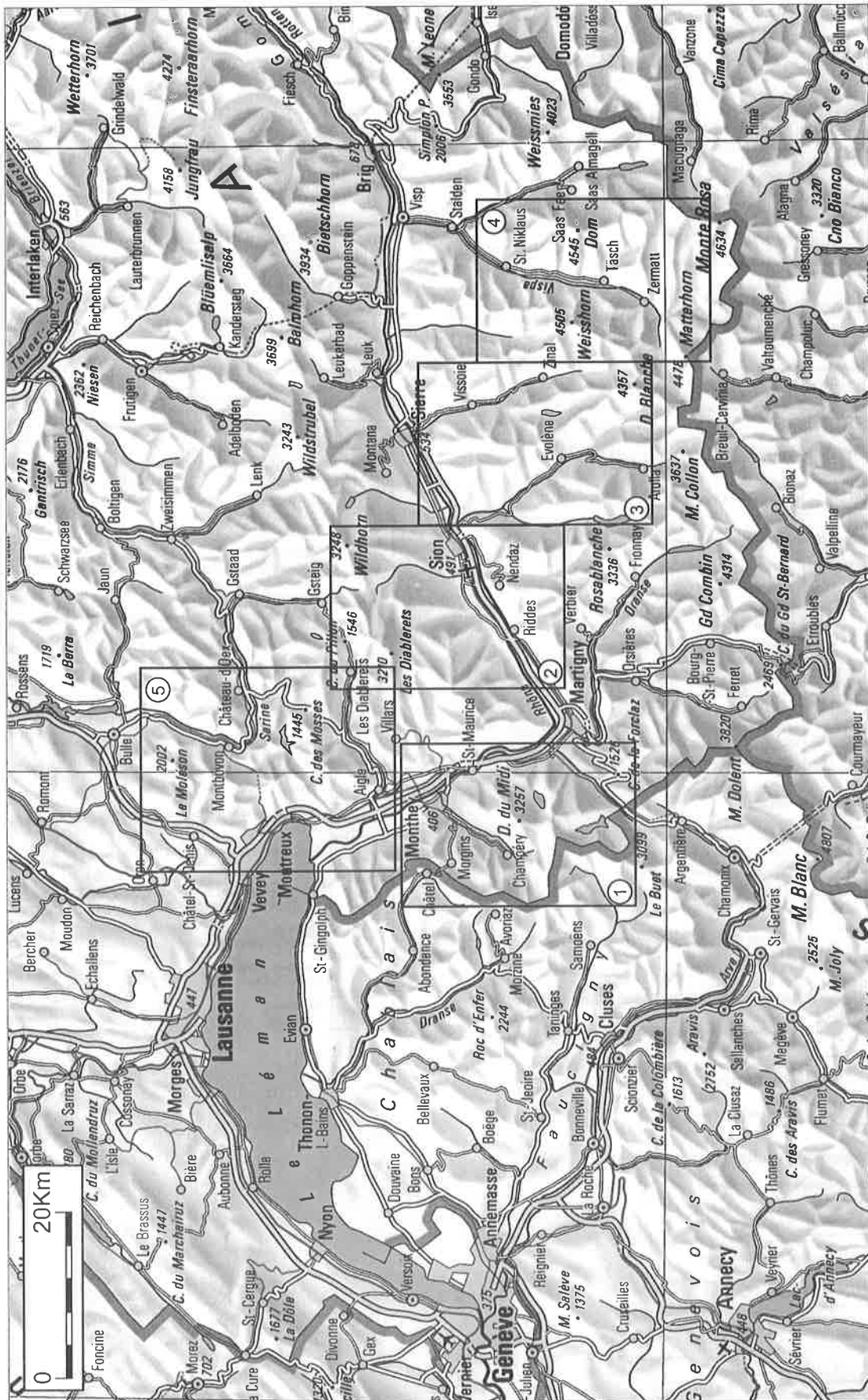
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Annex 1: Location of the structural maps in the pouch in the back cover



Annex 2: Legend of the tectonic maps

Quaternaire

1	Dépôts quaternaires non différenciés des grandes vallées
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Helvétique s.l.

Molasse subalpine

Molasse d'eau douce inférieure

3	Grès (Molasse grise) <i>Aquitainien</i>
4	Grès et conglomérats (Molasse rouge, poudingue du Mont Pèlerin, Molasse à charbon) <i>Chattien</i>

Molasse marine inférieure

5	Marnes et grès (grès de Vaulruz) <i>Oligocène inférieur</i>
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Flysch subalpin

6	Marnes et grès <i>Oligocène inférieur</i>
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Massif des Aiguilles Rouges

7	Marnes et grès (flysch parautochtone) <i>Oligocène inférieur</i>
8	Marnes et grès (flysch autochtone) <i>Oligocène inférieur</i>
9	Quartzites, dolomies, calcaires et marnes <i>Trias-Eocène</i>
10	Conglomérats (poudingue de Vallorcine), grès, pélites et rhyolites <i>Permo-Carbonifère</i>
11	Granite de Vallorcine <i>Carbonifère supérieur (307 ± 2 Ma, Bussy & Hernandez 1997)</i>
12	Gneiss et migmatites <i>Protérozoïque-Paléozoïque</i>

Nappe de Morcles

17	Calcaires, marnes et grès (y compris flysch) <i>Crétacé - Oligocène inférieur</i>
18	Calcaires et marnes <i>Dogger-Malm</i>
19	Calcaires et marnes <i>Lias</i>
20	Calcaires dolomitiques, dolomies et gypse <i>Trias</i>

Nappe d'Ardon - Massif du Mont Blanc interne

34	Calcaires, marnes et grès (y compris flysch) <i>Crétacé - Oligocène inférieur</i>
35	Calcaires, dolomies, schistes marneux et pélitiques <i>Trias-Jurassique</i>
39	Gneiss <i>Protérozoïque-Paléozoïque</i>

Nappe du Wildhorn (sud du Rhône)

Unité (nappe) des Diablerets

50	Calcaires et marnes <i>Jurassique</i>
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Unité (nappe) du Mont Gond

52	Calcaires, marnes et grès <i>Jurassique</i>
----	--

Unité (nappe) du Sublage

53	Calcaires, marnes, dolomies et grès <i>Trias-Jurassique</i>
----	--

Nappe du Wildhorn (nord du Rhône)

Unité (nappe) des Diablerets

55	Calcaires, marnes et grès <i>Crétacé - Oligocène inférieur</i>
56	Calcaires et marnes <i>Dogger-Malm</i>

Unité (nappe) du Mont Gond - Buntstock

57	Calcaires, marnes et grès <i>Crétacé - Oligocène inférieur</i>
58	Calcaires et marnes <i>Dogger-Malm</i>

Unité (nappe) du Sublage

59	Calcaires, marnes et grès <i>Crétacé - Oligocène inférieur</i>
60	Calcaires et marnes <i>Dogger-Malm</i>
61	Calcaires et marnes (avec parfois quartzites, schistes pélitiques, dolomies et évaporites) <i>Lias (parfois Trias-Lias)</i>
62	Quartzites, schistes pélitiques, dolomies et évaporites <i>Trias</i>

Nappes ultrahelvétiques

71	Grès <i>Eocène</i>	Flysch de la Dent de Valère
73	Calcaires et marnes <i>Lias</i>	Nappe de Box
74	Dolomies, évaporites et pélites <i>Trias</i>	Nappe du Laubhorn
75	Dolomies, évaporites, calcaires et marnes <i>Trias-Dogger</i>	Nappe de Höchst
76 H	Grès, conglomérats et pélites (flysch) <i>Eocène</i>	Nappe d'Arveyes
78	Argilites, marnes et calcaires (y compris flysch) <i>Dogger-Malm et Eocène</i>	Nappe du Sex Mort
77	Calcaires, marnes et grès (y compris flysch) <i>Dogger-Eocène</i>	Nappe d'Anzeinde
78	Calcaires et marnes <i>Malm-Crétacé</i>	Nappe de la Plaine Morte
79	Wildflysch <i>Oligocène</i>	

Pennique inférieur

Unités infra-Niesen

81	Wildflysch <i>Eocène</i>	Zone du Luchberg
82	Argillites, grès, marnes et calcaires <i>Lias-Malm</i>	Zone de Brand- Mülkerblatten
83	Conglomérats, grès et marnes (flysch) <i>Eocène</i>	Nappes du Moilleret et d'Ochsenweid

Zone de Sion-Courmayeur

110	Marbres, dolomies, argillites, grès, schistes et calcschistes <i>Trias-Crétacé</i>	Unité de Ferret
112	Marbres, calcschistes, schistes, quartzites et brèches (y compris flysch) <i>Trias-Crétacé</i>	Unité du Roignais-Versoyen
114	Schistes, grès, quartzites, marbres et conglomérats <i>Eocène-Oligocène</i>	Unité de la Pierre Aval

Nappe du Niesen

104	Calcaires et marnes <i>Trias-Jurassique</i>	Unité d'Oudioux-Chamosaire
105	Calcaires et marnes <i>Trias-Jurassique</i>	Unité de Murgaz
108	Conglomérats, grès, calcaires et marnes (flysch) <i>Maastrichtien</i>	Unité du Pic Chaussy
107	Conglomérats, grès, calcaires et marnes (flysch) <i>Maastrichtien</i>	Unité de Joblé-Arpilles- Brenleires-Arnenhorn
109	Conglomérats, grès, calcaires et marnes (y compris flysch) <i>Trias-Lutétien</i>	Unité de la Sarine-Kander

Pennique moyen

Zone Submédiane

108	Dolomies, évaporites, calcaires, marnes et grès (y compris flysch) <i>Trias-Eocène</i>
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Zone Houillère

Partie externe

115	Quartzites, dolomies, calcaires et évaporites <i>Trias</i>
117	Schistes graphiteux, grès, conglomérats et prasinites <i>Carbonifère supérieur - Permien</i>

Partie interne

117	Schistes graphiteux, grès, conglomérats et prasinites <i>Carbonifère supérieur - Permien</i>
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Nappe des Préalpes médianes plastiques

118	Calcaires, marnes et grès <i>Malm-Eocène</i>
119	Calcaires, marnes, dolomies et évaporites <i>Trias supérieur - Dogger</i>

Nappe du Flysch à lentilles de Couches rouges

120	Argillites, grès et lentilles calcaires <i>Eocène</i>
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Nappe des Pontis

121	Quartzites, calcaires et dolomies <i>Trias inférieur-moyen</i>
122	Grès, schistes et conglomérats <i>Carbonifère supérieur - Permien</i>

Nappe des Préalpes médianes rigides

129	Calcaires, dolomies, évaporites et grès <i>Trias moyen - Eocène moyen</i>
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Nappe de Siviez-Mischabel

130	Calcaires, dolomies et schistes marneux (série du Barrhorn) <i>Mésozoïque et Eocène</i>
131	Quartzites <i>Permo-Trias</i>
132	Gneiss granitiques de Randa et du Portjengret <i>Permien (269 ± 2 Ma, Bussy et al. 1996b)</i>
133	Grès et schistes <i>Permo-Carbonifère</i>
135	Micaschistes, gneiss, amphibolites et métagranites <i>Protérozoïque et Paléozoïque</i> (gneiss granitique de Thyon: 500 +3/-4 Ma, Bussy et al. 1996a)

Nappe de la Brèche

137	Brèches carbonatées, calcaires, dolomies et argillites <i>Trias supérieur - Eocène</i>
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Nappe du Mont Fort

138	Dolomies et marbres <i>Trias</i>
139	Quartzites et schistes <i>Permo-Trias</i>
140	Rhyolites <i>Permien</i>
141	Métagrauwackes, micaschistes et metabasites (basaltes, gabbros et tufs) <i>Paléozoïque</i>
142	Micaschistes et gneiss <i>Paléozoïque</i>

Nappe du Mont Rose

143	Marbres et flysch à blocs <i>Trias-Crétacé</i>
144	Zone mylonitique entre les zones du Mont Rose et de Macugnaga <i>Crétacé ou Tertiaire ?</i>
145	Gneiss granitiques <i>Permo-Carbonifère (310 ± 20 Ma, Frey et al. 1976)</i>
146	Gneiss, micaschistes et amphibolites <i>Protérozoïque-Paléozoïque</i>

Pennique supérieur

Zone de Zermatt – Saas Fée

150	Calcschistes et métabasites <i>Jurassique–Crétacé</i>
151	Métabasaltés (prasinites) et roches vertes non différenciées, localement entremêlées de calcschistes <i>Jurassique</i>
152	Métagabbros <i>Jurassique (gabbro de l'Allalin: 164 ± 2.7 Ma, Rubatto 1998)</i>
153	Serpentinites <i>Jurassique</i>

Nappe du Tsaté

157	Calcschistes, localement à intercalations de métabasites; 157a: Marbres siliceux <i>Jurassique–Crétacé</i>
157a	
158	Métabasites, localement entremêlées de calcschistes <i>Jurassique</i>
159	Métagabbros <i>Jurassique</i>
160	Serpentinites <i>Jurassique</i>

Nappe des Cimes Blanches

154	Arkoses, quartzites, dolomies, marbres et brèches <i>Permien–Crétacé</i>
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Nappe des Gëts

161	Grès, pélites, conglomérats et calcaires (flysch à lentilles d'ophiolites) <i>Crétacé (bloc ophiolitique: 166 ± 1 Ma, Bill et al. 1997)</i>
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Nappe du Frilihorn

155	Arkoses, quartzites, dolomies, marbres et brèches <i>Permien–Crétacé</i>
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Nappe des Dranses

163	Grès, calcaires et pélites (flysch à helminthoïdes) <i>Crétacé supérieur – Paléocène</i>
-----	---

Austroalpin

Système Dent Blanche – Mont Mary – Pillonet – Sesia – Seconde zone dioritico-kinzigitique

Séries du Cervin (Ce), du Mont Dolin (Do), de Rolsan (Ro) et du Pillonet (Pi)

166	Dolomies, marbres et brèches <i>Trias supérieur – Jurassique moyen</i>
-----	---









Série d'Arolla et zone de Sesia

+	166	+	Gneiss granitiques <i>Paléozoïque (gneiss de Cervinia: 289 ± 2 Ma, Bussy et al. 1998)</i>
	169		Métadiorites et métagabbros <i>Paléozoïque (métagabbro d'Anzasca: 288 ± 2/-4 Ma, F. Bussy et J. Hunziker, comm. orale)</i>
	170		Gneiss, métapélites, marbres et métabasites <i>Protérozoïque–Paléozoïque</i>

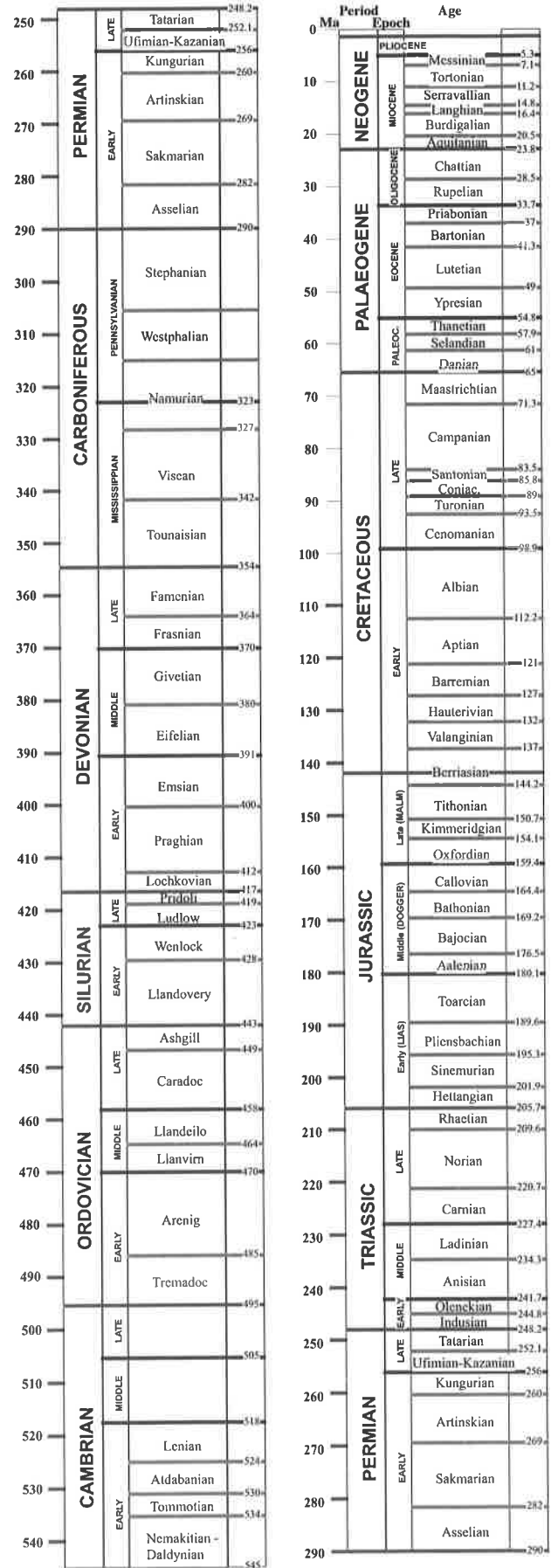
Série de Valpelline

174	Gneiss et micaschistes (kinzigites et strolonites), marbres et amphibolites <i>Protérozoïque–Paléozoïque</i>
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Éléments structuraux

	40	Direction et pendage de la stratification ou de la schistosité dominante
		Direction de stratification ou de schistosité dominante verticale
	10	Direction et plongement d'axe de pli des nappes de charriage
		Axe horizontal de pli des nappes de charriage
	34	Direction et plongement d'axe de pli de vergence S ou SE
		Contact tectonique entre unités, contact tectonique important au sein d'une unité
		Failles tardives (RS = Ligne Rhône–Simplon) <i>Oligocène–Quaternaire</i>
		Mylonites de Miéville (Mi) <i>Permo-Carbonifère</i>

Annex 3: Time table

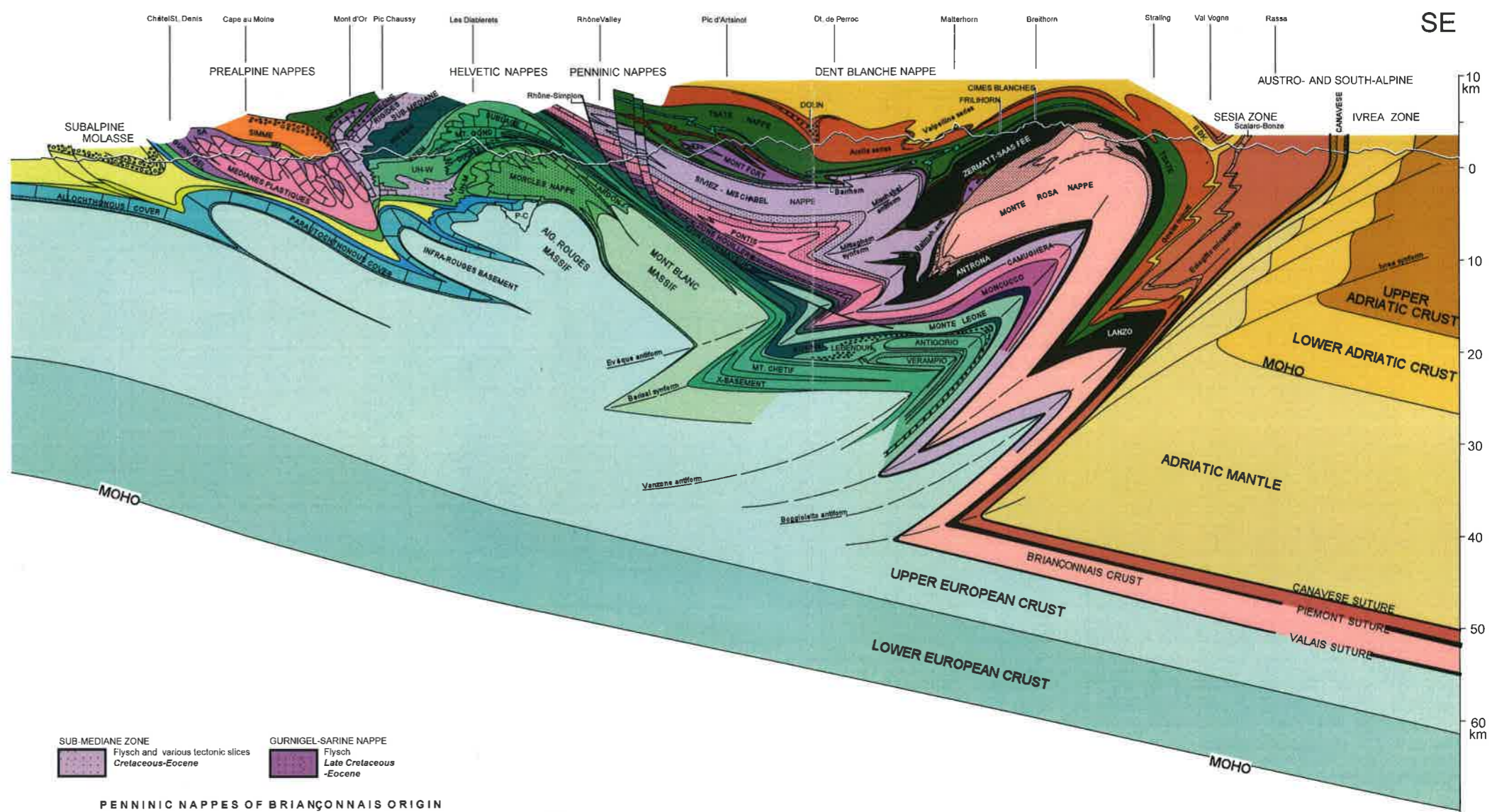
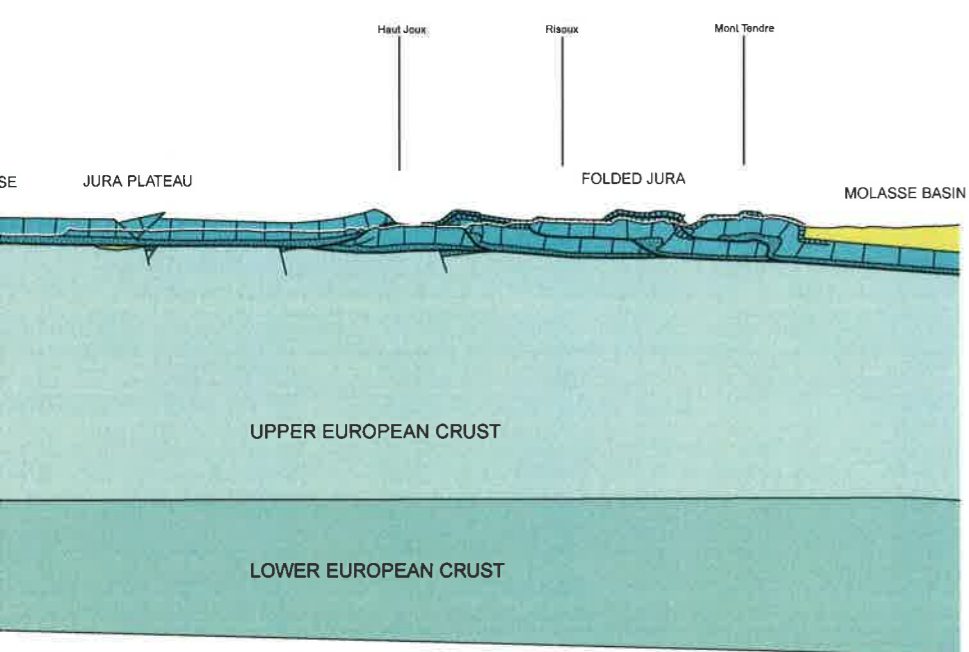


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LOWER PENNINIC NAPPES

LEBENDUN NAPPE Conglomerates, schists and gneisses <i>Paleozoic and Mesozoic</i>	MONTE LEONE NAPPE (+ Faldbach cover) Quartzites, dolomites, evaporites metagreywackes and gneisses <i>Paleozoic-Triassic</i>
INFRA-NIESEN ZONE Limestones and flysch <i>Jurassic-Eocene</i>	ANTIGORIO NAPPE Metagreywackes (Baceno) Schists and gneisses <i>Paleozoic</i>
TEGGIOLO ZONE (TE) Quartzites, dolomites, calcschists and breccias <i>Triassic-Jurassic</i>	VERAMPIO NAPPE Metagreywackes (Baceno) Schists and gneisses <i>Paleozoic</i>

ULTRAHELVETIC NAPPES (UH-W)
Quartzites, dolomites, evaporites, limestones, marls and flysch
Triassic-Eocene

ULTRAHELVETIC NAPPES (UH-M)
Quartzites, dolomites, evaporites, limestones, marls and flysch
Triassic-Eocene

HELVETIC NAPPES

WILDHORN NAPPES (DIABLERETS, MT.GOND AND SUBLAGE) Limestones, marls, shales, flysch <i>Triassic-Early Oligocene</i>	MONT CHETIF MASSIF AND X-BASEMENT Dolomites, shales, limestones <i>Triassic-Middle Jurassic</i>
Dolomites, limestones, marls and shales <i>Triassic-Jurassic</i>	Gneiss, amphibolites <i>Paleozoic and older</i>
ARDON NAPPE Limestones, marls, shales <i>Cretaceous-Early Oligocene</i>	MONT BLANC MASSIF (INTERNAL) Quartzites, dolomites, shales, limestones <i>Triassic-Late Jurassic</i>
MORCLES NAPPE Limestones, marls, shales, flysch <i>Late Jurassic-Early Oligocene</i>	Quartz porphyry, granulites and gneisses <i>Paleozoic and older</i>
Limestones, marls and shales <i>Early and Middle Jurassic</i>	MONT BLANC MASSIF (EXTERNAL) Quartzites, dolomites, shales, limestones <i>Triassic-Cretaceous</i>
Quartzites, evaporites, dolomites <i>Triassic</i>	Schists, granulites and gneisses <i>Paleozoic and older</i>

LOWER PENNINIC COVER NAPPES PARTLY ORIGINATED IN THE VALAIS ACCRETIONARY PRISM

SION-COURMAYEUR ZONE (FLYSCH MAINLY)
Calcschists, schists, quartzites, breccias and conglomerates, locally with basal slices of basalt, serpentinite and dolerite.
Late Cretaceous

ROSSWALD SERIES (FLYSCH MAINLY)
Calcschists, siliceous limestones, schists and conglomerates, locally with slices of rhyolites and meta basic rocks
Late Cretaceous

INFRA MONCUCCO OPHIOLITES
Serpentinites and metagabbros
Cretaceous ?

PENNINIC NAPPES OF BRIANÇONNAIS ORIGIN

BRECHE NAPPE Breccias, radiolarites, limestones and flysch <i>Late Triassic-Eocene</i>	MONT FORT NAPPE Quartzites, dolomites, marbles and breccias <i>Triassic-Liasic</i>	MONTE ROSA NAPPE Dolomites, marbles and meta flysch with slices and dykes of meta basalt <i>Triassic-Cretaceous</i>
MEDIANES RIGIDES NAPPE Dolomites, limestones, marls and flysch <i>Middle Triassic-Eocene</i>	SIVIEZ - MISCHABEL NAPPE In most parts: Quartzites, evaporites <i>Early Triassic</i> In Barhorn Zone with marbles and schists <i>Triassic-Eocene</i> Conglomerates, schists, metagranulites <i>Permo-Carboniferous</i> Gneisses and amphibolites <i>Paleozoic and older</i>	CAMUGHERA ZONE Meta granulites, gneisses <i>Paleozoic and older</i>
MEDIANES PLASTIQUES NAPPE Limestones, shales, marls and flysch <i>Lias-Eocene</i> Dolomites, evaporites <i>Late Triassic</i>	PONTIS NAPPE Quartzites, dolomites <i>Early and middle Triassic</i> Conglomerates, schists and metagranulites <i>Permo-Carboniferous</i> Gneisses, metagranulites and amphibolites <i>Paleozoic and older</i>	MONCUCCO ZONE Paragneisses, micaschists <i>Paleozoic and older</i>
ZONE HOUILLERE Quartzites and dolomites <i>Triassic</i> Metagreywackes, conglomerates and schists <i>Carboniferous-Triassic</i>		

OPHIOLITE NAPPES OF PIEMONTE ORIGIN AND ASSOCIATED COVER NAPPES

GETS NAPPE
Flysch with slices of ophiolites and gneiss
Cretaceous

DRANSES NAPPE (DRA)
Flysch
Late Cretaceous

SIMME NAPPE
Flysch and conglomerates
Middle Cretaceous

TSATE NAPPE (ACCRETIONARY PRISM WITH OPHIOLITES)
Serpentinites, metagabbros metabasalts and calcschists
Middle Jurassic - Cretaceous

FRILHORN NAPPE
Dolomites, marbles, quartzites breccias and conglomerates
Permian-Cretaceous

CIMES BLANCHES NAPPE
Dolomites, marbles, quartzites breccias and conglomerates
Permian-Cretaceous

ZERMATT-SAAS FEE AND ANTRONA OPHIOLITE ZONES
Serpentinites, metagabbros, eclogites metabasalts and calcschists
Middle Jurassic-Cretaceous

LANZO OPHIOLITE ZONE
Lherzolites, serpentinites, metagabbros eclogites, metabasalts and calcschists
Middle Jurassic-Cretaceous

DENT BLANCHE NAPPE AND SESIA ZONE

Valpelline and Znd Diorite-Kinzigle Zones
Schists and gneisses of amphibolite to granulite facies (lower crust)

Carving unit
Calcschists
Jurassic

Scalero-Bonze units
Conglomerates, quartzites, dolomites, calcschists
Permian - Jurassic

Doln cover zone
Dolomites, marbles and breccias
Late Triassic - Middle Jurassic

Arolla Series and Gneiss Minuti
Orthogneisses, metagabbros and schists
Late-Paleozoic

Eclogitic Micaschists Complex
Schists, gneisses and granulites
Late Paleozoic

AUSTRO AND SOUTH-ALPINE NAPPES

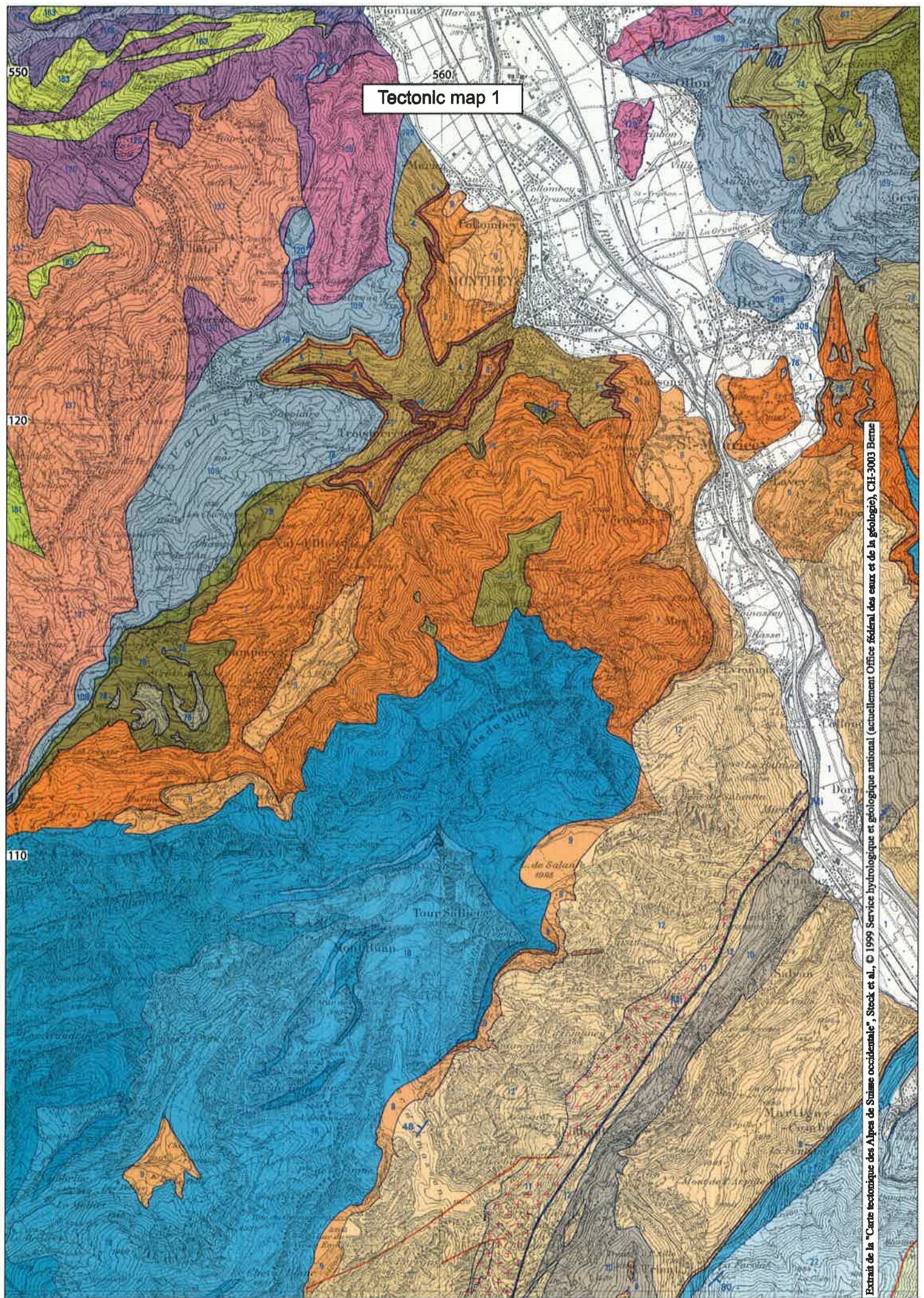
CANAVESE ZONE
Conglomerates and andesite dykes
Tertiary

IVREA ZONE
UPPER CONTINENTAL CRUST
Phyllites, micaschists, gneisses and granulites
Carboniferous and Permian

LOWER CONTINENTAL CRUST
High grade gneisses, metagabbros and serpentinites
Early Paleozoic and older

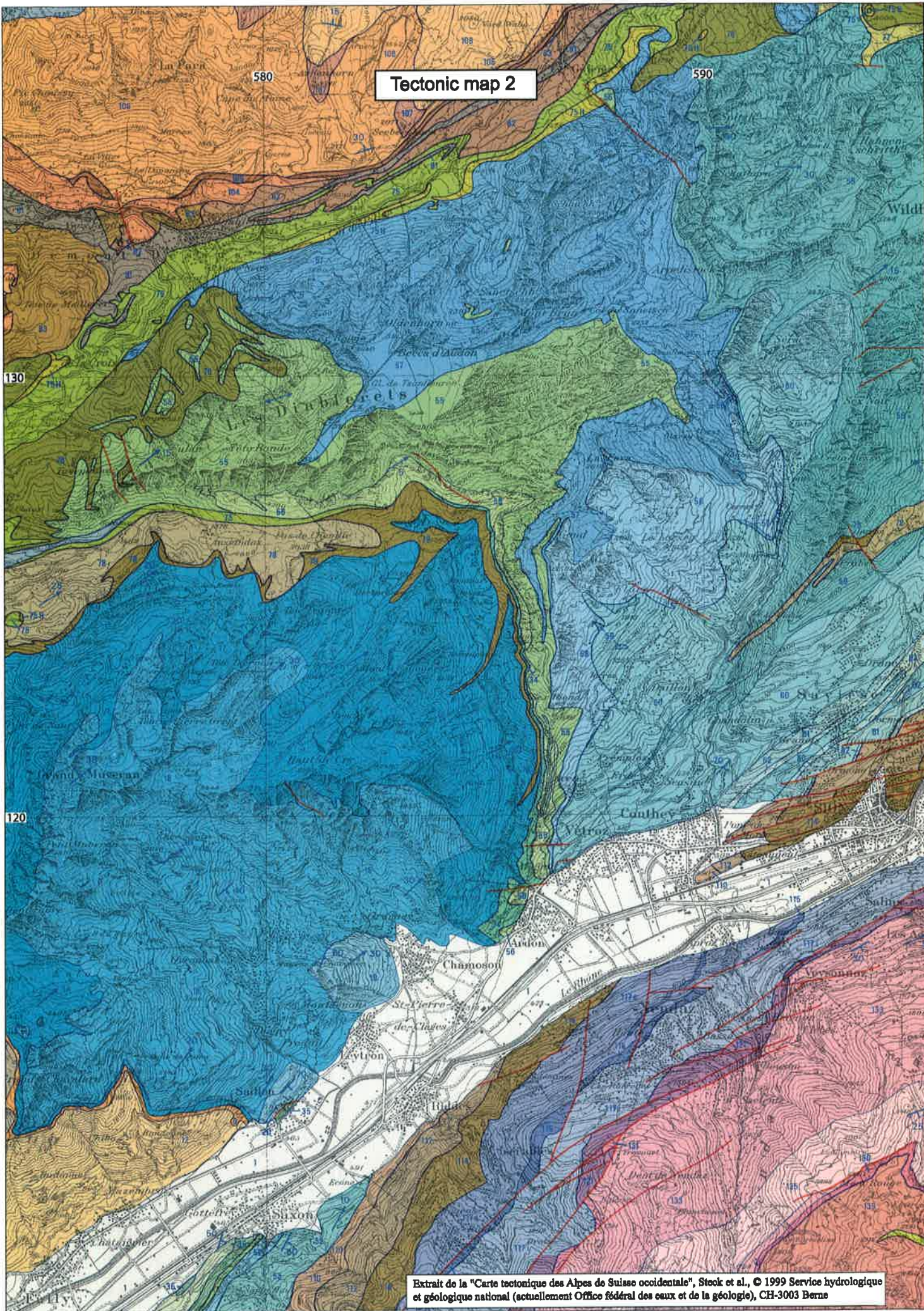
Constructed by :
A. Escher, H. Masson, A. Steck, J.-L. Epard, R. Marchant, M. Marthaler, M. Sartori and G. Venturini. 1994
Section sciences de la terre BFSH-2 CH-1015 LAUSANNE
Computer graphics by Marc Escher

Tectonic map 1

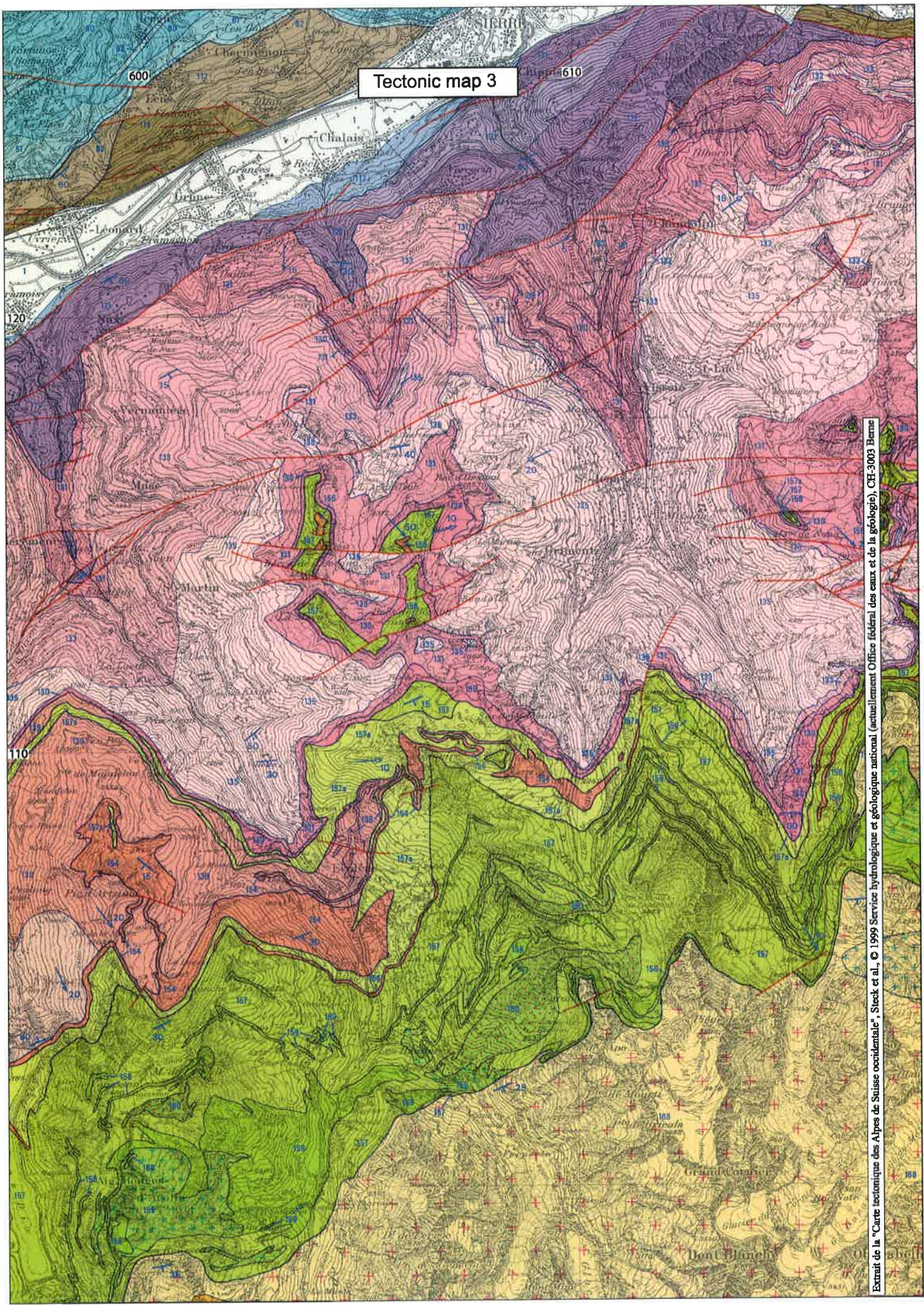


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Tectonic map 2

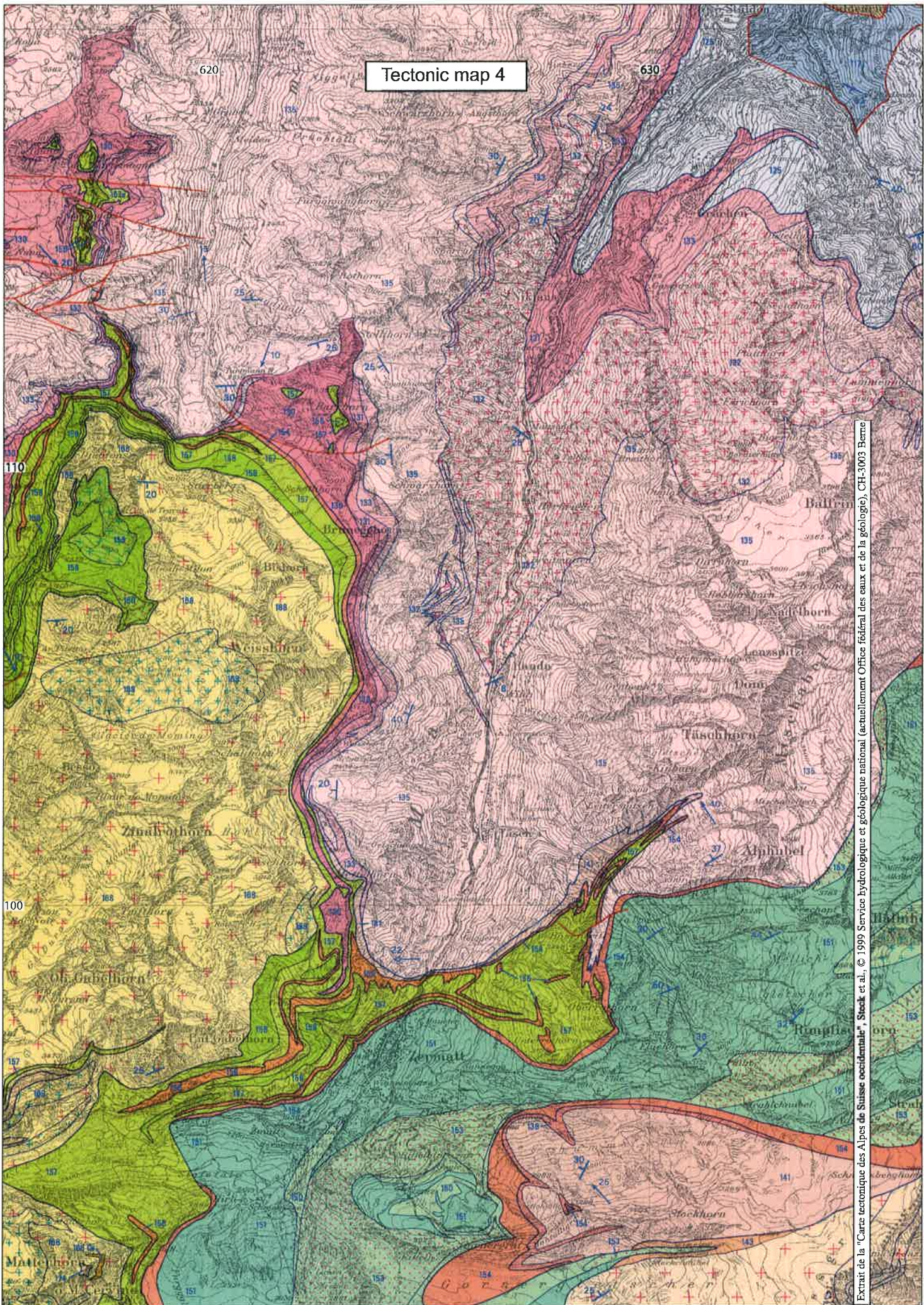


Tectonic map 3

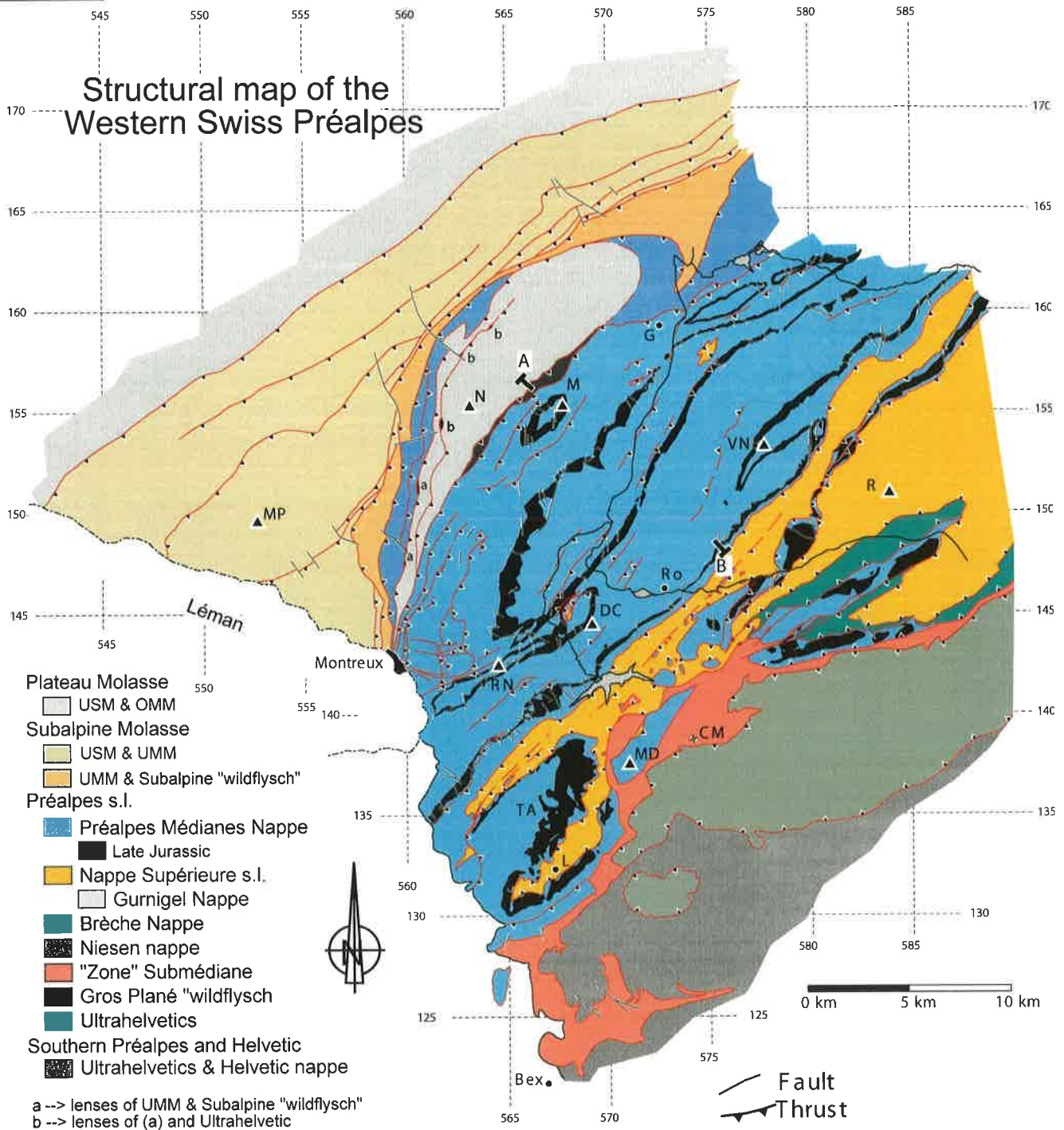


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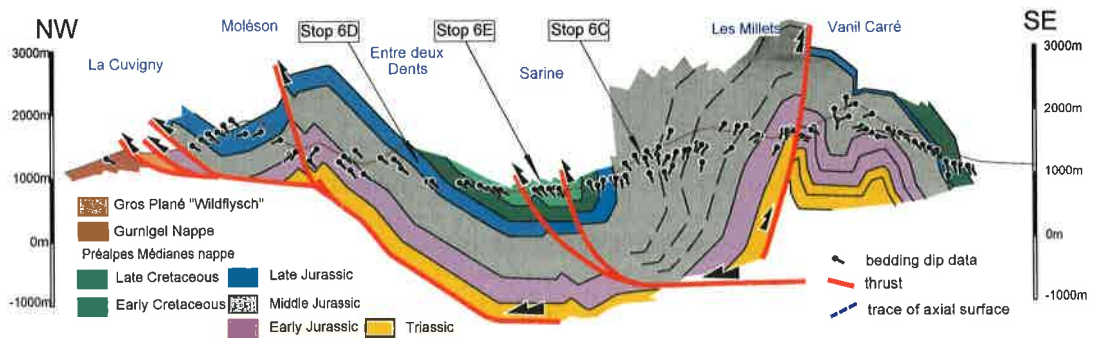
Tectonic map 4

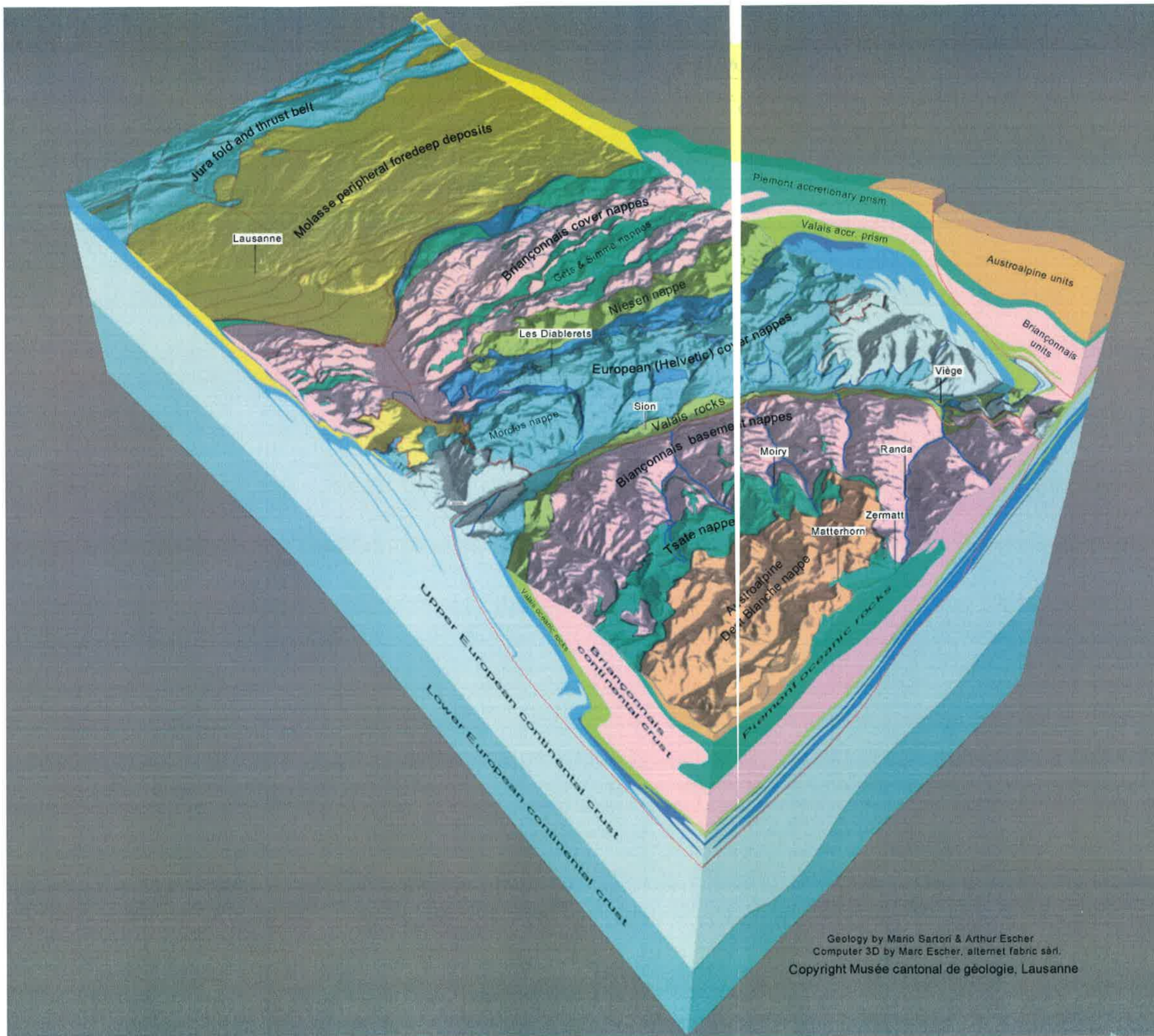


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CM = Col des Mosses; DC = Dent de Corjon; G = Gruyère; L = Leysin; M = Moléson; MP = Mont Pélerin; N = Niremont; R = Les Rodomonts; RN = Rochers de Naye; Ro = Rossinière; TA = Tours d'Aï; VN = Vanil Noir





Geology by Mario Sartori & Arthur Escher
 Computer 3D by Marc Escher, alternet fabric sàrl.
 Copyright Musée cantonal de géologie, Lausanne

Block diagram showing the main geologic units in the internal Western Swiss Alps and their relation with the actual topography (Escher & Sartori, Musée de Géologie, Lausanne, 2001).

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- No. 16 VANNAY J.-C. 1993. Géologie des chaînes du Haut-Himalaya et du Pir Panjal au Haut-Lahul (NW-Himalaya, Inde). Paléogéographie et tectonique. 148 pp., 44 text-figs., 6 pls.
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