Excursion: Geology of the Alps
3.-16. August 2009

Guide book

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Guides: Hugo Ortner (Innsbruck); Thorsten Nagel (Bonn); Andreas Wetzel (Basel)
Students and postdocs from Utah, Alberta, Salzburg and München

Panoramic view from Les Druges (St. Marcel/Aosta) to the north (Photo: B. Lammerer)

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Prepared by Bernd Lammerer

Program

Aug. 3 Monday:

14^{00}: Meeting at Munich, Luisenstrasse 37, Geology and Tectonics: Lecture room 222/ 1st floor
Workshop on the Geology of the Alps, Introduction to the Excursion
Accommodation in private rooms of participants from Munich or Youth Hostel Munich

Aug. 4 Tuesday:

Munich – Garmisch – Fernpass – Imst (150 km).

1- By cable car to the Muttekopf (Late Cretaceous sediments)

(Guide: Hugo Ortner).
Accommodation: Muttekopf Hütte (1.934 m); Phone: +43 (0) 664 - 123 69 28; Tel.: +43 (0) 54 14 - 8 64 56; Fax: +43 (0) 54 14 - 8 70 47; E-Mail: info@muttekopf.at costs: € 34 - 40 incl. Food.

Aug. 5 Wednesday:

Whole days walk in a high alpine area around the Muttekopf (2774 m):

1- Late Cretaceous "Gosau" sediments and geodynamic evolution of the Austroalpine nappes

(Guide: Hugo Ortner)
Accommodation: Muttekopf Hütte

Aug. 6 Thursday:

Imst – Landeck – Arlberg Pass – Sargans – Tannenbodenalp – Bad Ragaz – Vättis – Chur - Thusis (270 km).

1- Tannenbodenalp: panoramic view over the Helvetic nappes of the Churfirsten
2- Vättis: tectonic window of the External Aarmassiv

(Guide: Petra Veselá + Bernd Lammerer)
Accommodation: Camping Viamala, Thusis Telefon +41-(0)81-651-2472 Fax +41-(0)81-651-2472 info@camping-thusis.ch WWW www.camping-thusis.ch Contact: Pascale Zimmermann & Hugo Grieder

Aug. 7. Friday:

Thusis - Flims – Thusis (60 km)

1- Flims Landslide (largest landslide of the Alps)
2- Cassonsgrat (2700 m, by cable car): Glarus Thrust (geologic UNESCO world heritage site)

(Guide: Thorsten Nagel + Bernd Lammerer)
Accommodation: Camping Via Mala, Thusis (total 60 km)

Aug 8 Saturday:

Thusis - Passo San Bernardino (2065 m) – Mesocco – Roveredo – Bellinzona (120 km)

1- Via Mala gorge Bündnerschiefer of the Valais ocean and Erosion of the Hinterrhein
2- San Bernardino Pass : Adula nappe, Misox zone, Simano nappe

(Guide: Thorsten Nagel)
Accommodation: Camping TCS "Bosco di Molinazzo"; Roman Leonardi; Via San Gottardo 131; CH-6500 Bellinzona; Phone: ++41-91-829 11 18; Fax: ++41-91- 829 23 55; e-mail: camping.bellinzona@tcs.ch

Aug. 9 Sunday:

Bellinzona – Valbella – Bellinzona (70 km)

1- Ascent to Passo Trescolmen (2153 m): orthogneisses, eclogites, garnet-schists of the Adula nappe.

Whole day walk in high Alpine area.
(Guide: Thorsten Nagel)
Accommodation: Camping TCS "Bosco di Molinazzo"; Bellinzona;

Aug. 10 Monday:

Bellinzona _ Lugano – Mendrisio – Morbio Superiore arzo - Bellinzona (140 km)

1- Breggia gorge: Jurassic and Cretaceous sediments and synsedimentary tectonics

2- Arzo quarries: Early Jurassic extensional tectonics)

3- Meride Museum of Monte San Giorgio fossils (UNESCO world heritage site)

(Guide: Bernd Lammerer)

Accommodation: Camping TCS "Bosco di Molinazzo"; Bellinzona

Aug. 11 Tuesday:

Bellinzona – Val Verzasca - Centovalli – Sta. Maria Maggiore – Finero - Domodossola – Val Aosta (240 km)

1- Val Verzasca migmatites

2- Ivrea Zone near Finero

3- Via romana and Gallery between Donnas and Bard

4- If enough time: ascent from St. Marcel to Seissogne – Les Druges, (view to Mt. Blanc – Mt. Rosa – Matterhorn) and ascent to the old copper mine of Servette and the manganese mine of Praborna

(Guide: Bernd Lammerer)

Accommodation: Camping Aosta International Touring; Fraz. Arensod, 10 - 11010 Sarre (Aosta) - Italy
+39 0165 257061 campingtouring@libero.it

Aug. 12 Wednesday:

Aosta – Valtournanche - Cervinia – Aosta (110 km)

1- Ascent from Valtournanche to the Lago di Cignana (HP and UHP rocks of the Penninic ocean)

Whole day walk in high Alpine area

(Guide: Bernd Lammerer)

Accommodation: Camping Aosta International Touring

Aug. 13 Tursday:

Aosta – Tunnel Grand St. Bernard - Martigny (75 km)

1- Salvan Dorenaz basin: Permo – Carboniferous basin sediments, sedimentology of clastic sediments (alluvial fans, playa lakes, braided river systems)

(Guide: Andreas Wetzel and Petra Veselá)

Accomodation: Camping Martigny "Les Neuvilles" Rue du Levant 68, 1920 Martigny
Tel.: 027 722 45 44 Fax: 027 772 35 44

Aug 14 Friday:

US group: **Martigny – Zurich Kloten Airport or Munich (550 km)**

German group: Martigny – Trient - Lac D`Emosson - Martigny (70 km)

1- Aiguille Rouge basement, Triassic transgression, dinosaur footprints in Triassic quartzites

Guide: Andreas Wetzel and Petra Veselá

Accomodation: Camping Martigny "Les Neuvilles"

Aug. 15. Saturday:

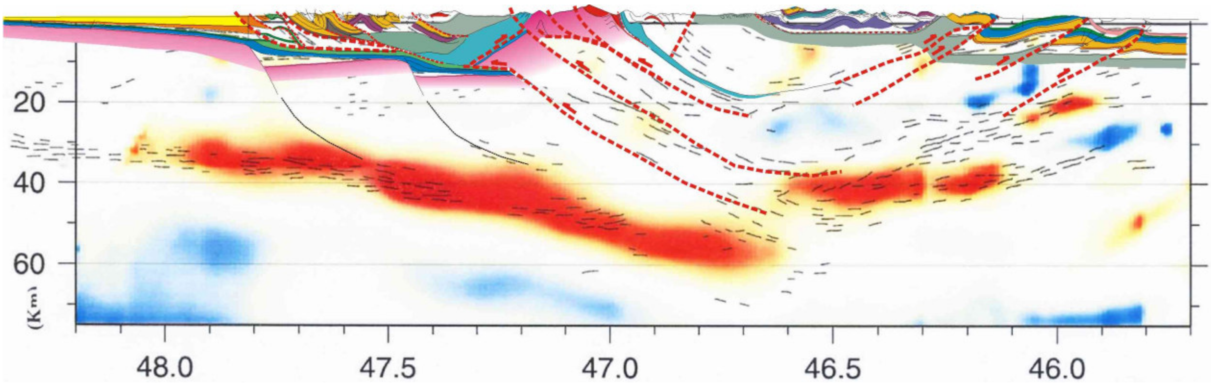
Martigny – Visp – Bürchen
geology of the Wallis area
Accomodation: tent or Chalet Friedrich

Aug. 16. Sunday:
Back to Munich (550 km)

Introduction

Considerable progress in our understanding of the Alps evolved from national research programmes during the 1980s and early 1990s conducted by the Swiss NFP20, the Italian CROP and the French ECORS programmes (Roure et al., 1990) in the Western Alps. These were partly integrated into the pan-European initiative of the N–S European Geotraverse (Blundell et al., 1992). The results gave rise to the idea of "indenter" -tectonics to describe the complex interactions of the European and the Adriatic– African continental plates during their collision, which started approximately 50 Ma ago, after the closure and subduction of the Penninic ocean beneath the Adriatic plate (Pfiffner et al., 1997; Pfiffner, 1992; Blundell et al., 1992).

In the Eastern Alps the TRANSALP project was conducted between 1998 and 2001 by partner institutions from Germany, Austria and Italy (TRANSALP Working Group, 2001, 2002). One of the main advantages of this profile is its length of about 340 km and the use of consistent acquisition and processing parameters, enabling application of modern imaging techniques. The design of the TRANSALP experiment was highly complex. Vibroseis near-vertical seismic profiling formed the core of the field data acquisition, complemented by explosive near-vertical seismic profiling, cross-line recording for three-dimensional control, active-source tomography or wide-angle recording of Vibroseis and explosive sources by a stationary array for velocity control and passive tomography by another 9- to 11-month stationary array for seismological/lithospheric studies.



Geological setting

The Eastern Alps are composed of a thin-skinned orogenic wedge, mostly of Adriatic origin to the north, a thick-skinned wedge to the south and an uplifted part of European basement and cover together with oceanic rocks in the center, the Tauern window. Its uplift re-deformed the Alpine edifice in Neogene time and led to lateral eastward extrusion of blocks between conjugate strike-slip faults in the nappe stack (Ratschbacher et al., 1991; Frisch et al., 1998) and to ductile stretching in the Tauern Window (Lammerer and Weger, 1998).

The excursion route crosses the flexural foreland basin of the Molasse and a narrow folded part, which was overthrust by its own Mesozoic substrata units (Helvetic nappes), and rootless oceanic sediments of the Rhenodanubian Flysch nappes. To the south, the Austroalpine nappes are represented in the northern part by the Northern Calcareous Alps (NCA), which were sheared off from their basement and folded already in Cretaceous time and contain the synorogenic Gosau sediments, which will be visited at day 1 and 2 at the Muttekopf.

Two subunits are exposed, the thin Allgäu nappe along the northern margin and the thick Lechtal nappe, which comprises most of the section north of the Inn valley (e.g. Linzer et al., 1995; Mandl, 2000; Auer and Eisbacher, 2003). The weakly metamorphic Paleozoic basement to the NCA stretches between the Inn valley and the Tauern Mountains, consisting of phyllites, quartzphyllites and minor volcanic and intrusive rocks. The Tauern window consists of imbricated and tightly folded Hercynian granitic sills and Paleozoic and Precambrian paragneisses and amphibolites of European origin in an overall east–west elongated dome structure with a steep southern limb (Lammerer and Weger, 1998; references therein). Steeply dipping foliation also occurs in the Austroalpine basement gneisses of the Adriatic plate further south until the prominent right-lateral Periadriatic fault, the presumed equivalent to the Insubric line in the Western Alps.

From here to the south, the Dolomite Mountains are crossed, mostly within or close to its quartzphyllite basement, which crops out in its southernmost position at the Agordo–Valsugana thrust. Further to the south, several minor thrusts give rise to the Tertiary Belluno basin and flexures, before the Mesozoic rocks plunge under the clastic sediments of the Venetian plain (Castellarin and Cantelli, 2000).

Geological background of the western Alps ***From:***
Stampfli, G. M., Borel, G. D., Marchant, R. & Mosar, J. 2002. *Western Alps geological constraints on western Tethyan reconstructions. In:Rosenbaum, G. and Lister, G. S. 2002. Reconstruction of the evolution of the Alpine-Himalayan Orogen. Journal of the Virtual Explorer, 7, 75 - 104.*

please refer to the web site of the tectonics group of the University of Lausanne!
http://www.unil.ch/igp/page23599_en.html

Introduction

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 and Stampfli, 1992; Manatschal, 1995; Froitzheim and Manatschal 1996; Bill et al., 1997), following the opening of the Central Atlantic ocean (Steiner et al., 1998). 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.). The NeoTethys ocean, whose closure was responsible for the formation of the Tethysides orogen, actually does not directly interfere with Alpine geology s.str., and the Alpine Tethys should be regarded more as an extension of the central Atlantic ocean in the Tethyan realm rather than a branch of the large and older NeoTethys ocean (Stampfli, 2000). In that sense, the onset of the Alpine cycle could be placed in the Carnian, a period corresponding to the final closure of PaleoTethys in the Mediterranean and Middle- East regions (Stampfli et al., 2001a; 2002a) and to the onset of rifting in the central Atlantic-Alpine domain (Favre and Stampfli, 1992).

The overall tectonic Evolution

We can regard the western Alps as issued from an accretionary prism related to the closure of the Alpine Tethys where different geological objects, corresponding to different stages of accretion, can be recognised:

- oceanic accretionary prism of the Piemont ocean (the western Alps portion of the Alpine Tethys), including crustal elements from the former toe of the southern passive margin (lower Austroalpine elements),

- accreted material of the Briançonnais terrain derived from the Iberic plate,

- accreted material of the Valais domain, representing the toe of the European (Helvetic s.l.) passive margin

- accreted material of the former European continental margin and rim basin,(Helvetic s.str. domain).

In time, one passes from the oceanic accretionary prism to the formation of the orogenic wedge proper that we place after the detachment or delamination of the subducting slab in the Early Oligocene (e.g. Stampfli and Marchant, 1995). The resulting heat flux allowed some more units to be detached from the European continental slabs and resulted in large scale subduction of continental material (Marchant and 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 accompanied and followed by the Paleogene flexure event of the lower European plate (Paleocene to Miocene) in front of the tectonic wedge.

Structural units seen on a present day Alpine crosssection, as in figures 1 and 2, represent remnants of palaeogeographic units whose main bodies were subducted (more than 90%). Each tectonic package is formed of obducted or underplated material with a geodynamic signature, which allows to replace them in their former palaeogeographic domain (i.e. rim basin, rift shoulder, passive margin, oceanic sequences, inversion basin, flexural bulge). In doing so, geodynamic markers (rift shoulders erosion, synrift deposits, denuded mantle) can be used to reconstruct the diverging phases, whereas palaeo-structures, ages of flysch deposits, of exotic terrain deposits and of metamorphism, as well as structural indicators, are used to reconstruct converging processes.

The structural framework of the western Alps

The present structural framework of the western Alps is shown in figures 1 and 2 (modified from Marchant, 1993; Marchant and Stampfli, 1997a; Marchant and Stampfli, 1997b), and are based on a structural model developed mainly by the Lausanne school (e.g Escher et al., 1997; Escher et al., 1988; Steck et al., 1997; Steck et al., 1989). This structural model separates the Alpine belt into different structural domains shown in different patterns on the figures. They are from north to south:

The Jura Mountains
This is the most external domain, incorporated in the Alpine chain last (from 11 to 3 Ma). It is characterised by typical thin skin tectonic with a décollement located in Triassic evaporitic deposits. This décollement horizon runs under the Molasse basin and is rooted in the external massifs overthrust (recent developments on this domain are found in Sommaruga, 1999).

The Molasse basin
Formerly representing a flexural foreland basin, it became a piggyback basin during the Jura overthrusting. It comprises a sedimentary sequence starting with a Mesozoic cover similar to the Jura, followed by a major hiatus from Late Cretaceous to Eocene or Oligocene due to the fore bulge uplift. The Molasse deposits started in Late Oligocene and lasted until Late Miocene and are made of shallow water marine or continental deposits. It is separated in a southern deformed unit (Subalpine molasse) and a less deformed northern domain, which covered a large part of the Jura before its folding. (Burkhard and Sommaruga, 1998).

The Préalpes
This is a composite terrain consisting of elements from the European margin at the base (Ultrahelvetic, Niesen nappes), an important sedimentary sequences derived from the Briançonnais domain s.l. forming the bulk of the Préalpes (Médianes Plastiques and Rigides, Brèche nappes) and the Nappes Supérieures (Gurnigel, Simme, Gets) formerly pertaining to the oceanic accretionary prism of the Alpine Tethys. The Préalpes are therefore quite exotic and represent the exported suture of the orogen thrusted over the external crystalline massifs. For an extended bibliographic data base on the Prealps see: http://www-sst.unil.ch/research/prealps/index.htm

The external massifs
The Aiguilles Rouges crystalline massif and its 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 unloading to the south 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
This classical domain of the Swiss Alps (Masson et al., 1980) consists of the Morcles (Doldenhorn), Ardon, Diablerets and Wildhorn nappes. The first two where 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 rifted Helvetic margin. Major uplift took place in Late Cretaceous and Paleocene time, related to the flexure of the European margin and 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 16). In France the Helvetic domain is called Dauphinois.

The Ultrahelvetic nappes
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
These highly metamorphic, mainly crustal nappes represent the former rift shoulder and syn-rift domains 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 now to former syn-rift deposits (Spring et al., 1992). Some Ultrahelvetic nappes and the Niesen nappe were certainly partially deposited on this lower Penninic nappes domain. The Niesen nappe contains Jurassic syn-rift deposits at its base and then inversion related Late Cretaceous clastics.

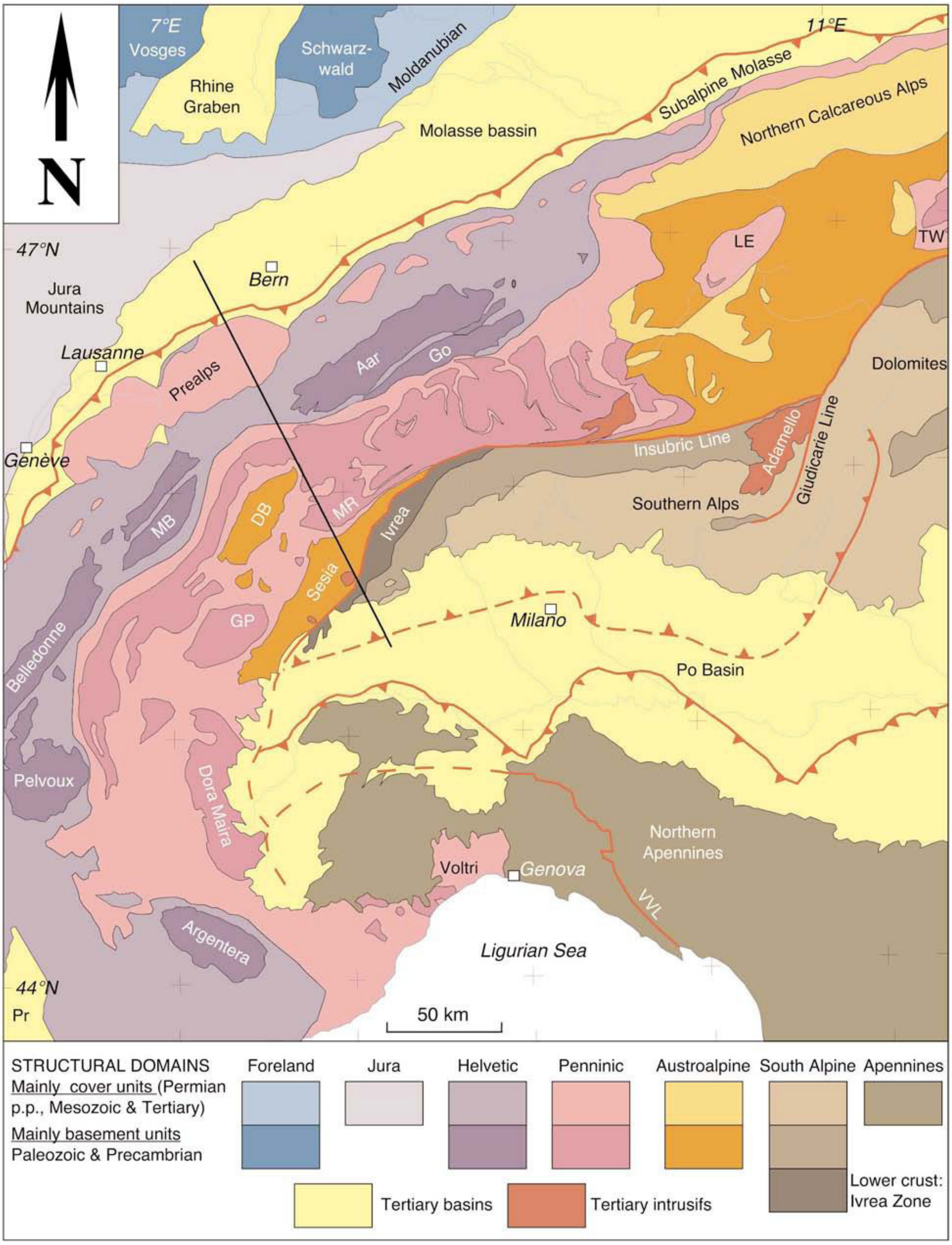


Figure 1. Tectonic map of the Western Alps (modified from Berthelsen, 1992b). Ao = Adamello intrusions; DB = Dent Blanche nappe; Go = Gotthard massif; GP = Grand Paradis massif; LE = Lower Engadine window; MB = Mont Blanc massif; MR = Monte Rosa nappe; Pr = Provence basin; TW = Tauern window; VVL = Villalvernia-Varzi-Levanto line.
Stampfli et al. 2002 Journal of the Virtual Explorer

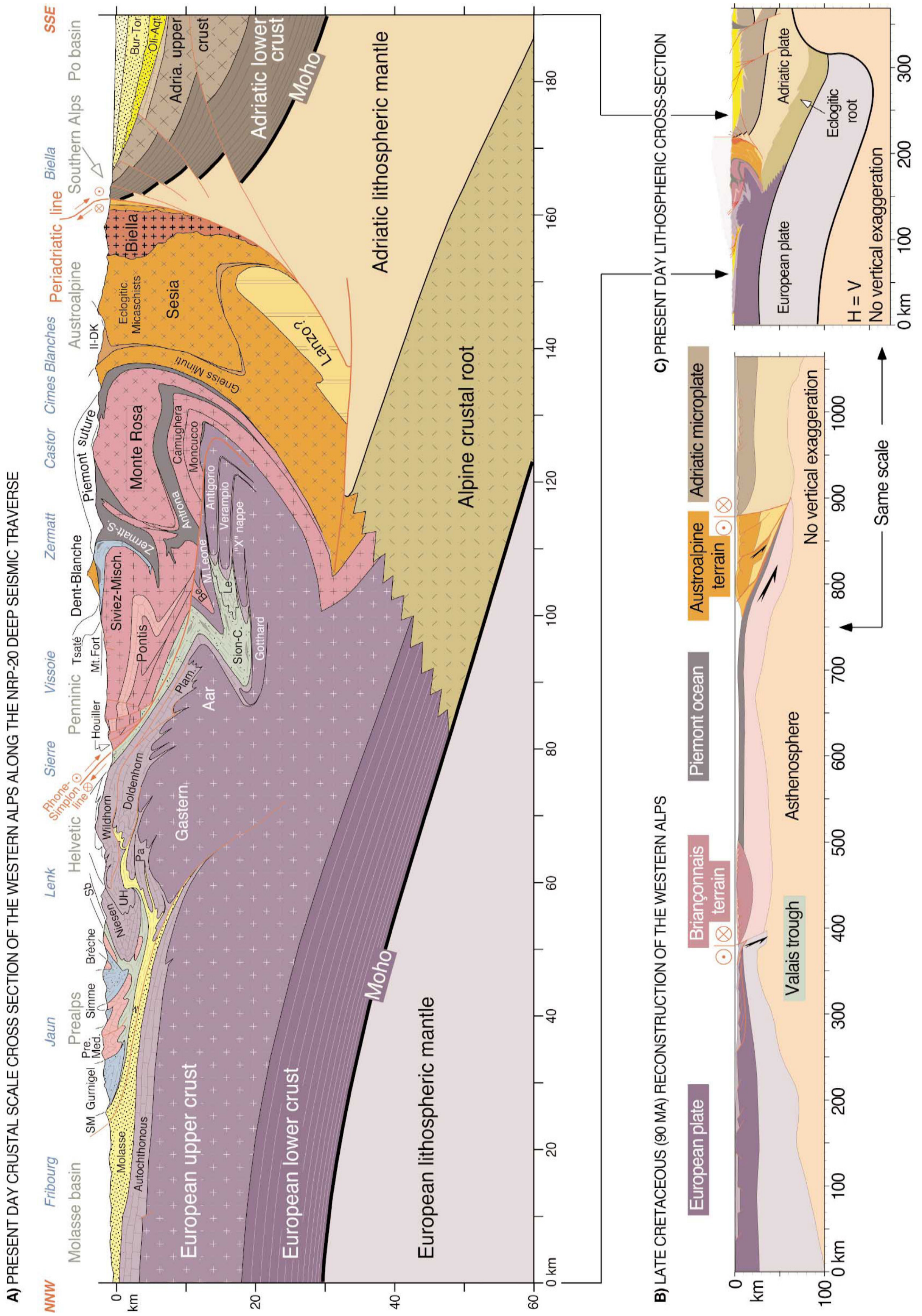


Figure 2. Cross-section of the western Alps on a western Switzerland transect, and simplified palinspastic model, (modified from Marchant, 1993; Marchant and Stampfli, 1997a). Location shown on figure 2 (thin black line).
Tethyan reconstructions Journal of the Virtual Explorer

The Valais suture zone

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 now as separate 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 and 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 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 Piemont 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.

The middle Penninic nappes

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 Médianes Plastiques were deposited, the Siviez-Mischabel basement nappes on which the Préalpes Médianes 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 Médianes 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 crosssection 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 Piemont ocean (the western Alps part of the Alpine Tethys ocean, its Italian part being referred to as Ligurian, its Austrian equivalent being often referred to as Penninic ocean)

The upper Penninic nappes

This domain represents the suture zone of the Alpine Tethys (Piemont 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, Briançonnais or even Helvetic e.g. Froitzheim, 2001), 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

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 Piemont suture zone (? parts of the Internal Massifs). The rest of this domain (the Dent Blanche klippe and Sesia zone) formed the former backstop of the Tsaté accretionary prism and was partly overthrusted by the Adriatic micro-continent (South Alpine units). Subsequently this overthrust was largely deformed by the 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 Western Alps orogen. This back stop was strongly affected by two events: the back folding of the internal units since Oligocene and the emplacement of the peri-Adriatic plutonic rocks derived from the detachmentdelamination of the Tethyan slab more or less at the same time.

The geodynamic framework of the western Alps

The plate tectonics of the western Tethyan regions The reconstructions shown in figure 4 to 9 are based on a tight pre break-up Permian fit (fig 3) as well as magnetic anomalies from the Central Atlantic. They have been constructed in a continuous effort to apply plate tectonic concept to palinspatic model of the western Tethys, moving away from pure continental drift model not constrained by plate limits (Stampfli and Borel, 2002). 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 (Stampfli et al., 2001c). This opening would be concomitant with the opening of the NeoTethys and the northward 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(Stampfli et al., 2002a), accompanied by the opening of back-arc basins (fig. 4, Meliata, Maliac and Pindos back-arc basins).

The Apulia Adria problem

A review of palaeomagnetic 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 Triassic (e.g. Channell, 1992; 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 (de Jonge et al., 1994; Wortel and Spakman, 1992) 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. 3 & 4), this leads 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 an African promontory from Middle Triassic to recent times and represents the eastern most Cimmerian element detached from Gondwana in Mediterranean – Neotethys, 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 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 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., 2001b; Vavassis et al., 2000). The northward subduction of a remnant PaleoTethys induced the opening of back-arc type basins already in Carboniferous-Permian times in a general context of cordillera collapse following the accretion of the Variscan terrains in Early Carboniferous time (Ziegler and Stampfli, 2001). 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 and 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 Western Alps.

Accelarating 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 in Late Triassic (Vavassis, 2001; Stampfli et al., 2002a). 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 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, Bulgaria, Yougoslavia, fig. 6; e.g. Baumgartner, 1985; De Bono, 1998).

The opening and closing of the Meliata-Maliac domain affected the whole Alpine history. First it created a generalised thermal subsidence starting in the Late Permian responsible for the large scale Triassic transgression on the Alpine regions. Then, its closure in the Late Jurassic (Kozur, 1991) affected the whole Austroalpine domain (the Austrian deformation phase) 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). The southward subduction of the Meliata-Küre system is certainly responsible for the (local) southward subduction of the European slab, a unique feature in the whole Alpine s.l. system where generally the southern continents are subducted northward (Africa, Arabia, India).

The Alpine Tethys rifting

Late Triassic ages obtained from shear zones in the Southern Alps (Hunziker et al., 1992; Schmid et al., 1987) 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 affected by transtension and emplacement of mafic material during the Permo-Triassic became subsiding rimbasins of the Jurassic Alpine Tethys (fig. 4). This is the case for the Médianes

(Subbriançonnais) and Helvetic rim basins. Paleozoic faults were also reactivated on a large scale 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, most likely using older structures. 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. Subsidence patterns of the marginal areas of the Alpine Tethys (Borel, 1995; Loup, 1992) together with stratigraphic and sedimentological records allow to place the onset of rifting in the Sinemurian (fig. 10). 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 to Bajocian age. Gabbros have been dated in several areas of the Alpine Tethys, the oldest dating 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). It should be emphasized here that up to 100 to 150 km of the nascent ocean were made of denuded continental mantle, based on comparison with the Galicia margin (e.g. Manatschal and Bernoulli, 1998; Müntener and Hermann, 2001), and, in view of the slow spreading of the ocean (Rampone and Piccardo, 2000), clear seafloor spreading was delayed in regard of the onset of thermal subsidence which is older than Bajocian, the Bajocian sequences being generally transgressive on the rift shoulders.

Rim Basins

Thermal subsidence of the Alpine Tethys margins in the 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 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 rimbasins 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 pelagic sequences. Therefore a general phase of sediment starvation characterises the distal part of the Alpine Tethys margins and the ocean itself. However, small carbonate platforms developed for a short time (Bajocian - Oxfordian) around the rift shoulders, e.g. the Briançonnais platform (Septfontaine, 1983) and in some south Helvetic units (Lempicka-Münch, 1996). They were linked to larger platforms along the rift shoulder as found in Corsica and Sardinia, and were finally drowned in the Late Jurassic; the south Helvetic platform 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 margin.

The "Valais ocean" question, and the Alps Pyrenees connection

From much, sometimes contradicting data concerning the Iberian drift (Stampfli, 1993; Stampfli et al., 1998) it can be determined that the opening of the Pyrenean rift system started in Late Jurassic times when Iberia was detached from New-Foundland. The minimal opening would be 200 km following the data from Malod and Mauffret (1990) or 350 km using the data of Sibuet and Colette (1991). Discrepancies stem from the former situation of Iberia with regard to New Foundland. The data of Srivastava et al. (1990) and of Srivastava and Verhoef (1992) allow to narrow down these differences and their proposal was applied for our model where a tight Permian fit for Pangea is used. 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. 6 & 7) and, therefore, certainly opened at the same time. The oldest magnetic anomaly in that part of the Atlantic is M0 (Hauterivian to earliest Aptian depending on the time scale…); however the sea-floor spreading is inevitably slightly older than the first clear magnetic anomaly.

In the Pyrenean region, the thermal event linked to the rising asthenosphere has been placed between 115 Ma and 80 Ma by Montigny (1986), it is regarded as Late Albian (≈ 100 Ma) by Peybernes (in: Debrand-Passard and Courbouleix, 1984). The Bay of Biscay opened later, together with the Aquitaine basin and the reactivation of the Iberian rift (Vergés and Garcia-Senz, 2001), during the rotation of Iberia, after the Valanginian (Moreau et al., 1992) and after the Atlantic opening between Iberia and America, therefore, in Late Aptian-Albian times; spreading there stopped in the Campanian (A33 anomaly, see figure 9). Thermal subsidence is active in the Late Cretaceous for the Pyrenees and the Gulf of Biscay region (fig. 10) where the Cenomanian is largely transgressive on the former rift shoulders (Peybernès, 1976; Peybernès and Souquet, 1984; Simo, 1986).

The major change of sedimentation in the Briançonnais domain, located on the Iberian plate, is found at the top of the Calcaires Plaquetés (Python-Dupasquier, 1990) and can be dated as Barremian. It corresponds to a general drowning and starvation phase of the area at that time. There is also a major sedimentary gap in the Albian that we relate to the thermal uplift of the Briançonnais. The Helvetic-Dauphinois margin is also affected by the Pyrenean rifting phase, clearly expressed on the subsidence curves for the Helvetic and Swiss plateau domain (fig. 10).

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 crustal attenuation however reached denudation of continental mantle (the Lherzolites of Lherz).

The "Valais trough", as recognised in the western Alps in the present days, is actually the remnant of trapped Piemont sea-floor and of the toe of the Helvetic margin (fig. 6 to 9). The "Valais ocean" (as defined in Stampfli, 1993) 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 because its suture was located exactly where the Algero-Provençal ocean reopened in Oligo-Miocene times (Roca, 2001). The southern margin of the Pyrenean ocean was the Briançonnais peninsula (fig. 11), its northern margin was the Corbières-Provençal domain from the Pyrenees to the Maures-Estérel massifs. The Pyrenean margin of the Briançonnais outcrops in the Galibier region of the French Alps (Toury, 1985), well known for its Late Jurassic Brèche du Télégraphe (Kilian, 1891). Recent investigation there (Luzieux and Ferrari, 2002), showed us the
development of a pull-apart type basins rapidly deepening under the CCD in Late Jurassic times and located to the north of the classic Briançonnais domain. This area developed facies different from the Swiss sub- Briançonnais during the Cretaceous, and is regarded as the most external sub-Briançonnais elements known so far. Its conjugate northern Provençal margin area is characterised by important erosion during the Oxfordian and the development of Albian basins deepening southward towards the ocean, followed by inversion related basins accumulating thousands of meters of upper Cretaceous clastic deposits in a northward migrating fore-deep type basin (Debrand-Passard and Courbouleix, 1984). This shows a southward closure of the Pyrenean ocean on a Provençal transect, whereas a northward subduction took place on a Pyrenean transect (Vergés and Garcia-Senz, 2001), also recorded in Sardinia (Barca and Costamagna, 1997; Barca and Costamagna, 2000), but again a southward subduction took place westward in the Biscay ocean.

This Pyrenean orogen can be followed from the present Pyrenees eastward up to southern France (Provence), and continues in the Alps in the form of a large scale uplift 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. These flyschs clearly predate the Alpine collisional event in the Helvetic domain, characterised by the deposition of collisional flysch not before the earliest Oligocene.

Along the European margin it is interesting to note the similitude of facies between the Rheno-Danubian flysch 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; fig. 12). 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 opening north Atlantic ocean during the Cretaceous and Paleogene.

Exotic terrains and margin duplication
In the northern margin

As discussed above, the separation of the Iberic plate from North-America in the Late Jurassic (the onset of spreading is dated as M0, Early Aptian - 121Ma; fig. 8) implies a separation of the Iberic plate from Europe too. This separation is brought about by the Pyrenean pullapart rift system. The Briançonnais domain 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) in 1979. The most internal south-Helvetic (Ultrahelvetic) domains, the lower Penninic Simplo- Ticinese nappes and Valais zone (ie.Sion-Courmayeur zone, Sub-Médiane zone 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 was also sometimes referred as the north Penninic ocean or basin.

This displacement induced a duplication of the Piemont northern margin in the present day Alpine orogen from the French Alps up to the Engadine window, but is not considered here to have any extension further eastward.
The eastward escape of the Briançonnais was induced by the combined opening of the Atlantic and Bay of Biscay oceans and the large scale rotation of the Iberic plate together with Africa, as deduced from the magnetic anomalies from the Atlantic. This implies also a partial closure of the Valais trough already during the Late Cretaceous. The combined Africa-Iberia rotation was closely followed by a general shortening between Africa- Iberia and Europe, culminating into the Pyrenean orogen. But it was only during the Alpine collision that the Briançonnais domain became an exotic terrain, obducted in the Alpine accretionary prism (fig.12).

The Alboran plate (Rif, internal Betic, Kabylies, Peloritan, Corso-Sardinia and Calabria microplates, (Wildi, 1983) formed the southern margin of the Iberic plate (fig. 3 to 9). 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 starts in Cenomanian time, grading into major olistostromic deposits in Maastrichtian (Gübeli, 1982; Thurow, 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 under the Iberic plate started only 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. 12). 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 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 Ligurian basin, fig. 12) 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 oceans as a new generation of back-arc, accompanied by the drifting of the Calabrian micro-plate to its present position (fig. 13). In the southern margin The Early Cretaceous rotation of Iberia-Africa, 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-Vardar 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 and Wagreich, 1999) and references therin). 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. 7 to 9).
The subduction of the Piemont part of the Alpine Tethys is marked by HP/LT metamorphism of elements pertaining to the toe of the Austroalpine margin (Sesia massifs, Former back-stop of the prism) and to the accretionary prism s.str. (Tsaté nappe). Most of the eo-Alpine ages (Hunziker et al., 1992) 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 (Midle 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 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 and Häring, 1981). Before to become an active margin this western portion of the lower Austoalpine margin suffered a westward displacement in regards to Adria (Frank, 1987; Froitzheim et al., 1994; Trümpy, 1992) related to the closure of the Vardar/Hallstatt-Meliata oceanic realms and the important rotation of Africa at that time to which Adria was still attached. 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).

Subduction and obduction processes
Subduction and obduction processes affecting the oceanic sequences of the Piemont ocean are shown on figures 14 and 15. Spreading rates were obtained from the central Atlantic magnetic anomalies which allow to define the opening of the Alpine Tethys at least until Early Cretaceous. 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. 16) creating a general emersion of the Helvetic-Dauphinois and Subalpine domains since Paleocene time (Lihou, 1996, Allen et al., 1991, Burkhard and 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: (Ackerman, 1986) and Meilleret flysch (Homewood, 1974) (fig. 16). 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 (fig. 14) 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 (Dall'Agnolo, 2000; Steffen et al., 1993), which includes elements from the

Briançonnais margin (Couches Rouges, Breccia nappe). On the Briançonnais domain the deposition of the Médianes 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. 17), this is confirmed by metamorphic ages from the Briançonnais basement starting around 38 Ma (Markley et al., 1995) 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). 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 overthrusted on the Valais trough and a Late Eocene mélange is found at its base in the Valais (Pierre Avoi uni, Bagnoud et al. 1998).Along the same suture MORB are found in a few places in the Valais 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. The matrix of mélanges found directly under the Briançonnais are only dated in the Pierre Avoi unit in Valais, one cannot exclude that older mélanges (Late Cretaceous) are present, mainly in the Bündnerschiefer domain and related to the Cretaceous translation of the Briançonnais terrain. Elements from the Valais trough and the distal Helvetic margin were then accreted and are represented by the "Valaisan trilogy" (Aroley, Marmontains, Saint Christophe: Burri, 1958) made of poorly dated deep-sea sediments, outcroping nowadays in the Sion-Courmayeur zone.

Incorporation in the prism of elements pertaining to the Helvetic margin (fig. 12) is well constrained by the ages of the south Helvetic flysch (e.g. Jeanbourquin, 1994; Jeanbourquin and Burri, 1991) 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. Charollais et al., 1993; Kindler, 1988). Most of the attenuated crust from the Helvetic margin was subducted (Burkhard and Sommaruga, 1998).

General metamorphism of the south Helvetic domain spreads from 30 to 20 Ma (cooling ages)(Steck and Hunziker, 1994) and mylonites from the Helvetic nappes have been dated between 32 and 13 Ma (Kirshner et al., 1996). 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 Alpine Tethys slab detachment or delamination around 35-33 Ma (Wortel and Spakman, 1992; Stampfli and Marchant, 1995). This delamination corresponds to the emplacement of the peri- Adriatic 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 50 to 150 km of European plate (Marchant and Stampfli, 1997b).

Vergence of subduction

As we have seen above, the southward subduction of the Alpine Tethys ocean is related to the history of the Meliata-Maliac and Vardar domain. The change of vergence of the Alpine Tethys subduction westward is directly linked to the connection between the Alps and the Apennines (fig. 12). 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 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 palaeogeographic 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. In this framework where large scale constraints are taken into account, ophiolites of the Apenninic prism are regarded as mainly derived from the former Alpine prism which collided with the Iberic plate in Corsica and Calabria in Eocene times (fig. 12). These oceanic elements were re-mobilised when the remnant Alpine Tethys oceanic domain (Ligurian basin) south of the Iberic plate started to subduct northward.

Conclusions

We wanted to emphasize here the necessity to incorporate detailed field studies in large scale reconstructions in order to better constrain them. On the other hand, large scale reconstructions provide a unified and simple continental margins framework in which local information can be organized. In doing so, it can be shown in the Alps, that the very complex structural pattern does not correspond to a complex palaeogeography. If complexity there is, it is due to an interaction between different plates, which somehow have a common junction in the Alpine area. These European, African, Iberic, Alboran and Adriatic plates can be easily recognised through their sedimentary series and subsidence behaviour, and their interaction during collision processes through foreland basin evolution. We sum up here the major characteristics of the western

Tethyan area: First of all, the evolution of the eastern part of the orogen (Austria-Carpathes) has little to do with its western counter part (western Alps), they only go through a common fate only since the Late Cretaceous through a final and unified southward subduction of the Alpine Tethys and the Oligo-Miocene collision with Europe. Before that, the eastern Alps and Carpathian area are related to the closure of the Meliata-Vardar domain. Similarly, the western Alps and Apenninic area have little in common, they locally share bits and pieces of the Alpine Tethys ophiolitic sequences, but the comparison stops

at that. The Apenninic accretionary prism is issued from a northward subduction of a remnant Alpine Tethys sea-floor under Spain since Oligocene times. In Late Miocene, the slab roll-back eventually reached another oceanic remnant in the Ionian sea region consisting of the western most part of the NeoTethys ocean.
The fate of the Pyrenean domain is clearly related to a common wandering of Africa and Iberia between M0 and the Middle Eocene. This too often forgotten fact implies that the former northern margin of the Alpine Tethys had to be cut by the Pyrenean rift system, and a direct connection established between the two oceanic systems – Pyrenees-Biscay and Alpine Tethys. In order to avoid major Late Cretaceous collision or opening between Europe and Iberia, the plate limit had to be located along a grand circle, therefore the Iberian northern plate limit can be established with some confidence, as well as the former position of the Briançonnais domain. The ensuing duplication of the European margin can clearly be derived from thorough field investigation in the western Alps.

We hope to have provided here basic hard fact type information, which, somehow, should not be ignored by anybody who would be plate mover. But this is certainly not the end of the story, the Adria-Apulia separation/union being still far from a widely accepted solution, and we welcome any new information on the subject.

References (Selection)

Berthelsen, A. 1992b. Tectonic evolution of Europe – from Precambrian to Variscan Europe. In: D. Blundell, R. Freeman and S. Mueller (Eds.), A continent revealed - the European Geotraverse. University Press Cambridge, 153-163.

Bertotti, G., Picotti, V., Bernoulli, D. and Castellarin, A. 1993. From rifting to drifting: tectonic evolution of the South- Alpine upper-crust from the Triassic to the Early Cretaceous. Sedimentary Geology. 86, 53-76.

Bill, M., Bussy, F., Cosca, M., Masson, H. and Hunziker, J.C. 1997. High precision U-Pb and 40Ar/39Ar dating of an Alpine ophiolite (Gets nappe, French Alps). Eclogae geologicae Helvetiae. 90, 43-54.

Burkhard, M. and Sommaruga, A. 1998. Evolution of the Swiss Molasse basin: structural relations with the Alps and the Jura belt. In: A. Mascle, A. Puiddefabregas, H.P. Luterbacher and M. Fernandez (Eds.), Cenozoic Basins of western Europe. Special Publications of the Geological Society of London 134, 279-298.

Caron, C., Homewood, P. and van Stuijvenberg, J. 1980b. Flysch and Molasse of Western and Central Switzerland. In: R. Trümpy (Ed.), Geology of Switzerland, a guide-book, Part B, Geological Excursions. Wepf. and Co. Basel, 274-278.

Caron, C., Homewood, P. and Wildi, W. 1989. The original Swiss Flysch: A reappraisal of the type deposits in the Swiss Prealps. Earth-Science Reviews. 26, 1-45.

de Jonge, M.R., Wortel, M.J.R. and Spakman, W. 1994. Regional scale tectonic evolution and the seismic velocity structure of the lithosphere and upper mantle: the Mediterranean region. J. Geophys. Res. 99(B6), 12091-12108.

Dürr, S.B., Ring, U. and Frisch, W. 1993. Geochemistry and geodynamic significance of North Penninic ophiolites from the central Alps. Schweizerische Mineralogische und Petrographische Mitteilungen. 73, 407-419.

Escher, A., Hunziker, J., Masson, H., Sartori, M. and Steck, A. 1997. Geologic framework and structural evolution of the western Swiss-Italian Alps. In: O.A. Pfiffner, P. Lehner, P.Z. Heitzman, S. Mueller and A. Steck (Eds.), Deep structure of the Swiss Alps - Results from NRP 20. Birkhäuser AG. Basel, 205-222.

Escher, A., Masson, H. and Steck, A. 1988. Coupes géologiques des Alpes occidentales suisses. Mémoires de Géologie (Lausanne). 2, 11 p.

Faupl, P. and Wagreich, M. 1999. Late Jurasic to Eocene Palaeogeography and geodynamic evolution of the Eastern Alps. Mitteilungen der Österreichischen Geologischen Gesellschaft. 92, 79-94.

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

Froitzheim, N. 2001. Origin of the Monte Rosa nappe in the Pennine Alps – a new working hypothesis). Geol. Soc. America Bull. 113(5), 604-614.

Froitzheim, N. and Manatschal, G. 1996. Kinematics of Jurassic rifting, mantle exhumation, and passive-margin formation in the Austroalpine and Penninic nappes (Eastern Switzerland). Geological Society of America Bulletin. 108(9), 1120-1133.

Froitzheim, N., Schmid, S. and Conti, P. 1994. Repeated change from crustal shortening to orogen-parallel extension in the Autroalpine units of Graubüunden. Eclogae geologicae Helvetiae. 87(2), 559-612.

Froitzheim, N., Schmid, S. and Frey, M. 1996. Mesozoic paleogeography and the timing of eclogite-facies metamorphism in the Alps: a working hypothesis. Eclogae geologicae Helvetiae. 89(1), 81-110.

Gebauer, D. 1999. Alpine geochronology of the western Alps: new constraintsfor a complex geodynamic evolution. Schweizerische Mineralogische und Petrographische Mitteilungen. 79, 191-202.

Haas, J., Kovacs, S., Krystyn, L. and Lein, R. 1995. Significance of Late Permian-Triassic facies zones in terrane reconstructions in the Alpine-North Pannonian domain. Tectonophysics. 242, 19-40.

Hunziker, J.C., Desmons, J. and Hurford, A.J. 1992. Thirty-two years of geochronological work in the Central and Western Alps: a review on seven maps. Mémoires de Géologie (Lausanne). 13, 59 p.

Jeanbourquin, P. 1994. Early deformation of Ultrahelvetic mélanges in the Helvetic nappes (Western Swiss Alps). J. Struct. Geol. 16(10), 1367-1383.

Jeanbourquin, P. and Burri, M. 1991. Les métasédiments du Pennique inférieur dans la région de Brigue-Simplon. Lithostratigraphie, structure et contexte géodynamique dans le bassin valaisan. Eclogae geologicae Helvetiae. 84(2), 463-481.

Kilian, W. 1891. Note sur l'histoire et la structure géologique des chaînes alpines de la Maurienne, du Briançonnais et des régions adjacentes. Bulletin de la Société géologique de France. 3(XIX), 571-661.

Kozur, H. 1991. The evolution of the Meliata-Hallstatt ocean and its significance for the early evolution of the Eastern Alps and western Carpathians. In: J.E.T. Channell, E.L. Winterer and L.F. Jansa (Eds.), Paleogeography and paleoceanography of Tethys. Palaeogeogr. Palaeoclimatol. Palaeoecol. Elsevier 87, 109-135.

Lihou, J.C. 1996. Stratigraphy and sedimentology of the Sardona unit, Glarus Alps: Upper Crteaceous/middle Eocene deep marine flysch sediments from the Ultrahelvetic realm. Eclogae geologicae Helvetiae. 89(2), 721-752.

Loup, B. 1992. Mesozoic subsidence and stretching models of the lithosphere in Switzerland (Jura, Swiss Plateau and Helvetic realm). Eclogae geologicae Helvetiae. 85(3), 541- 572.

Malod, J.A. and Mauffret, A. 1990. Iberian plate motion during the Mesozoic. Tectonophysics. 184, 261-278.

Manatschal, G. 1995. Jurassic rifting and formation of a passive continental margin (Platta and Err nappes, Eastern Switzerland): geometry, kinematics and geochemistry of fault rocks and comparison with the Galicia margin, Diss. ETH Nr. 11188, Zürich

Marchant, R. 1993. The Underground of the Western Alps. Mémoires de Géologie (Lausanne). 15, 137 p.

Marchant, R.H. and Stampfli, G.M. 1997a. Crustal and lithospheric structures of the Western Alps: geodynamic significance. In: O.A. Pfiffner, P. Lehner, P.Z. Heitzman, S. Mueller and A. Steck (Eds.), Deep structure of the Swiss Alps - Results from NRP 20. Birkhaüser AG. Basel, 326-337.

Marchant, R.H. and Stampfli, G.M. 1997b. Subduction of continental crust in the Western Alps. Tectonophysics. 269(3-4), 217-235.

Markley, M., Teyssier, C. and Cosca, M. 1995. Deformation vs cooling ages: the application of the 40Ar/39Ar method to synkynematic white micas, G. S. A. Annual meeting, New-Orleans.

Masson, H., Herb, R. and Steck, A. 1980. Helvetic Alps of Western Switzerland. In: R. Trümpy (Ed.), Geology of Switzerland, part B. Wepf & Co. Basel, 109-135.

Müntener, O. and Hermann, J. 2001. The role of lower crust and continental upper mantle during formation of non volcanic passive margins: evidence from the Alps. In: R.C.L. Wilson, R.B. Whitmarsh, B. Taylor and N. Froitzheim (Eds.), Nonvolcanic rifting of continental margins: a comparison of evidence from land and sea. Geological Soc. Spec. Publ. 187, 267-288.

Schmid, S., Zingg, A. and Handy, M. 1987. The kinematics of movements along the Insubric Line and the emplacement of the Ivrea Zone. Tectonophysics. 135, 47-66.

Schmid, S.M., Rück, P. and Schreurs, G. 1990. The significance of the Schams nappe for the reconstruction of the paleotectonic and orogenic evolution of the Penninic zone along the NFP-20 East traverse (Grisons, eastern Switzerland). In: F. Roure, P. Heitzmann and R. Polino (Eds.), Deep structures of the Alps. Mém. Soc. géol. France 156; Mém. Soc. géol. Suisse, 1; Vol. spec. Soc. It.,1, 263-288.

Sommaruga, A. 1999. Décollement tectonics in the Jura foreland fold and thrust belt. Marine & Petrol. geol. 16, 111-134.

Spring, L., Reymond, B., Masson, H. and Steck, A. 1992. La nappe du Lebendun entre Alte Kaserne et le Val Cairasca (massif du Simplon); nouvelles observations et interprétations. Eclogae geologicae Helvetiae. 85(1), 85-104.

Stampfli, G.M. 1993. Le Briançonnais, terrain exotique dans les Alpes? Eclogae geologicae Helvetiae. 86(1), 1-45.

Stampfli, G.M. 1996. The Intra-Alpine terrain: a Paleotethyan remnant in the Alpine Variscides. Eclogae geol. Helv. 89(1), 13-42.

Stampfli, G.M. (Ed.), 2001. Geology of the western Swiss Alps, a guide book. Mémoires de Géologie (Lausanne) 36, 195 p.

Stampfli, G.M., Borel, G., Cavazza, W., Mosar, J. and Ziegler, P.A. (Eds.) 2001a. The paleotectonic atlas of the Peritethyan domain. CD ROM; European Geophysical Society p.

Stampfli, G.M. and Borel, G.D. 2002. A plate tectonic model for the Paleozoic and Mesozoic constrained by dynamic plate boundaries and restored synthetic oceanic isochrons. Earth and Planetary Science Letters, in press.

Stampfli, G.M., Marcoux, J. and Baud, A. 1991. Tethyan margins in space and time. In: J.E.T. Channell, E.L. Winterer and L.F. Jansa (Eds.), Paleogeography and paleoceanography of Tethys. Palaeogeography, Palaeoclimatology, Palaeoecology. Palaeogeography, Palaeoclimatology, Palaeoecology 87, 373-410.

Stampfli, G.M., Mosar, J., Favre, P., Pillevuit, A. and Vannay, J.- C. 2001c. Permo-Mesozoic evolution of the western Tethyan realm: the Neotethys/East- Mediterranean connection. In: P.A. Ziegler, W. Cavazza, A.H.F. Robertson and S. Crasquin-Soleau (Eds.), PeriTethys memoir 6: Peritethyan rift/wrench basins and passive margins, IGCP 369. Mém. Museum Nat. Hist. Nat Paris 186, 51-108.

Stampfli, G.M. et al. 1998. Subduction and obduction processes in the western Alps. In: A. Vauchez and R. Meissner (Eds.), Continents and their mantle roots. Tectonophysics 296(1-2), 159-204.

Steck, A., Epard, J.-L., Escher, A., Marchant, R. and Masson, H. 1997. Geologic interpretation of the seismic profiles through western Switzerland: Rawil (W1), Val d'Anniviers (W2), Mattertal (W3), Zmutt-Zermatt-Findelen (W4) and Val de Bagnes (W5). In: O.A. Pfiffner, P. Lehner, P.Z. Heitzman, S. Mueller and A. Steck (Eds.), Deep structure of the Swiss Alps - Results from NRP 20. Birkhaüser AG. Basel, 123-138.

Steck, A. et al. 1989. Coupe tectonique horizontale des Alpes centrales. Mémoires de Géologie (Lausanne). 5, 8.

Steck, A. and Hunziker, J. 1994. The Tertiary structural and thermal evolution of the Central Alps - compressional and extensional structures in an orogenic belt. Tectonophysics. 238, 229-254.

Trümpy, R. 1992. Ostalpen und Westalpen-Verbindendes und Trennendes. Jb. Geol. B.-A. 135(4), 875-882.

Venturini, G. 1995. Geology, geochemistry and geochronology of the inner central Sesia Zone (Western Alps - Italy). Mémoires de Géologie (Lausanne). 25, 183.

von Raumer, J., Stampfli, G.M., Borel, G.D. and Bussy, F. 2002. The organisation of pre-Variscan basement areas at the north- Gondwanan margin. International Journal of Earth Sciences. 91, in press.

Winterer, E.L. and Bosellini, A. 1981. Subsidence and sedimentation on Jurassic passive continental margin, southern Alps, Italy. American Association of Petroleum Geologists Bulletin. 65(3), 394-417.

Wortel, M.J.R. and Spakman, W. 1992. Structure and dynamics of subducted lithosphere in the Mediterranean region. Proceedings of the koninklijke nederlandse Akademie van Wetenschappen. Series B: Palaeontology, Geology, Physics, Chemistry, Anthropology. 95(3), 325-347.

Ziegler, P.A. and Stampfli, G.M. 2001. Late Paleozoic Early Mesozoic plate boundary reorganisation: collapse of the Variscan orogen and opening of Neotethys. In: R. Cassinis (Ed.), the continental Permian of the Southern Alps and Sardinia (Italy) regional reports and general correlations. Annali Museo Civico Science Naturali, Brescia Brescia 25, 17-34.

Zingg, A., Handy, M.R., Hunziker, J.C. and Schmid, S.M. 1990. Tectonometamorphic history of the Ivrea zone and its relationships to the crustal evolution of the Southern Alps. Tectonophysics. 182, 169-192.

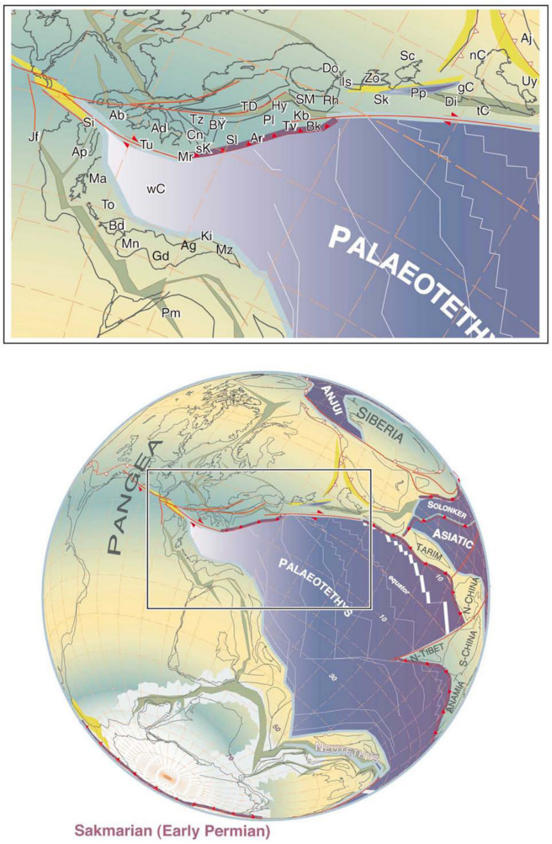
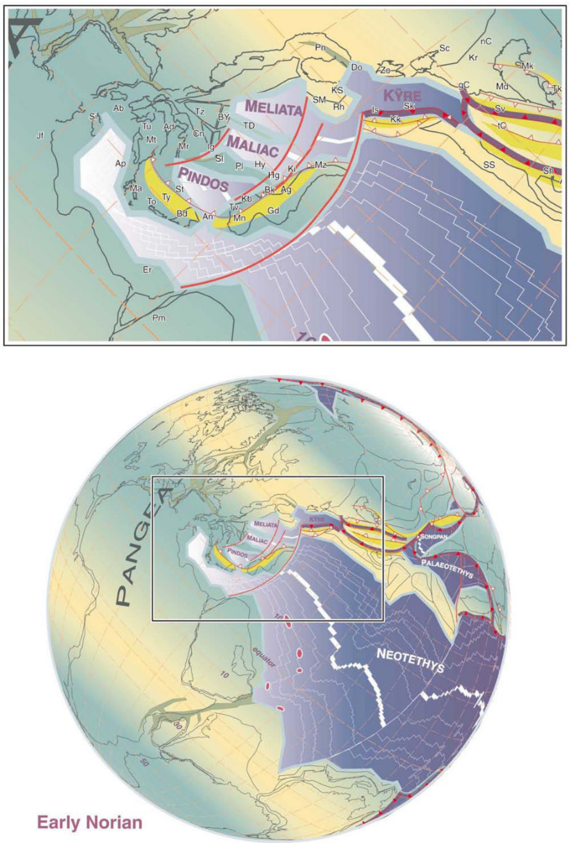


Figure 3.



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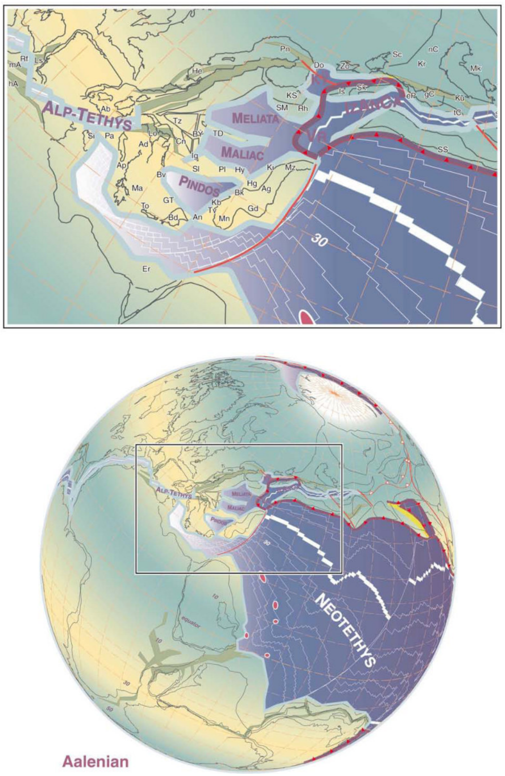


Figure 5

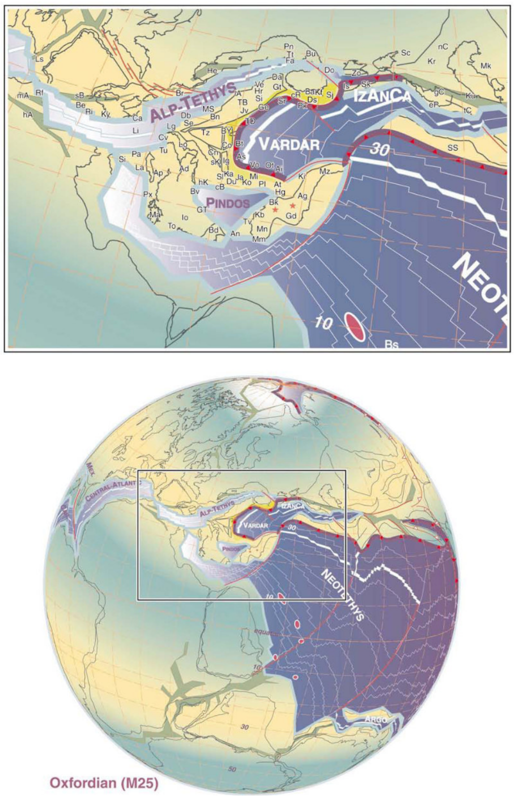
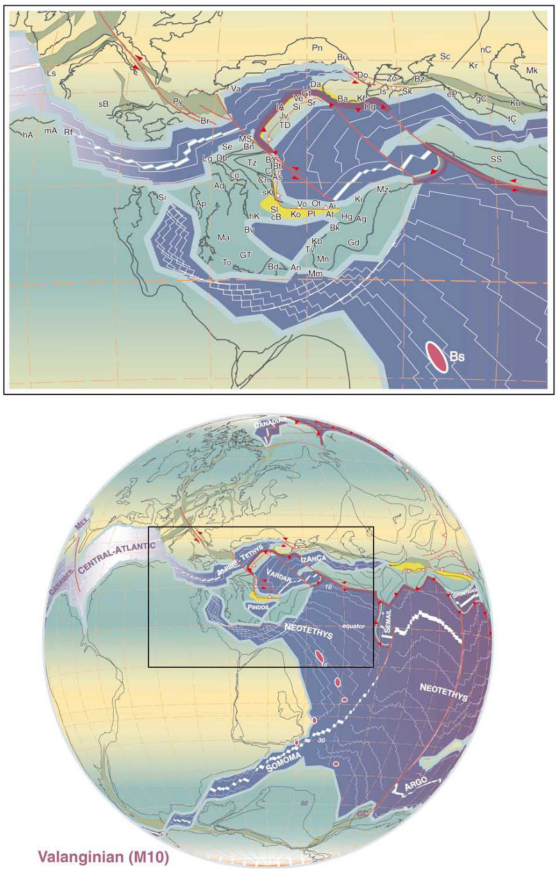


Figure 6



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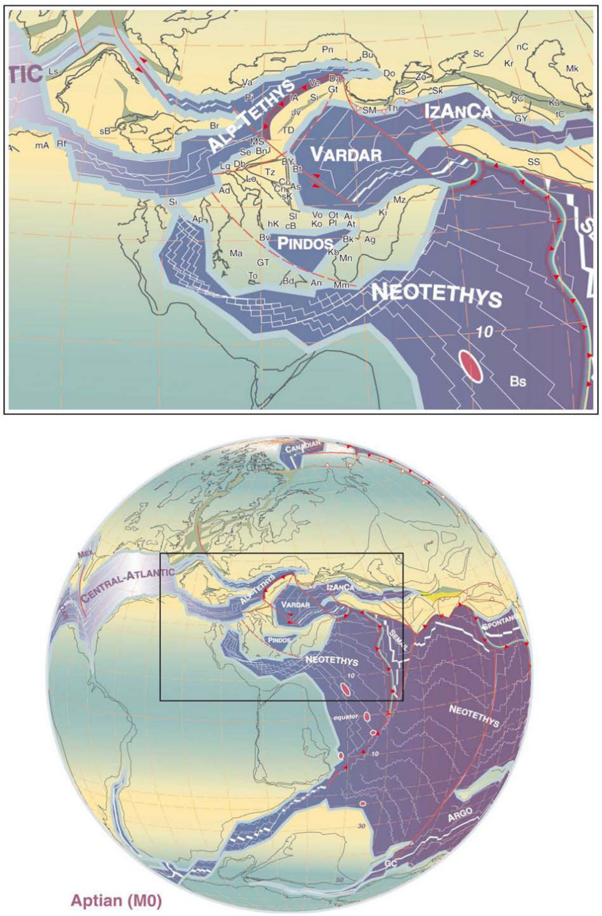


Figure 8

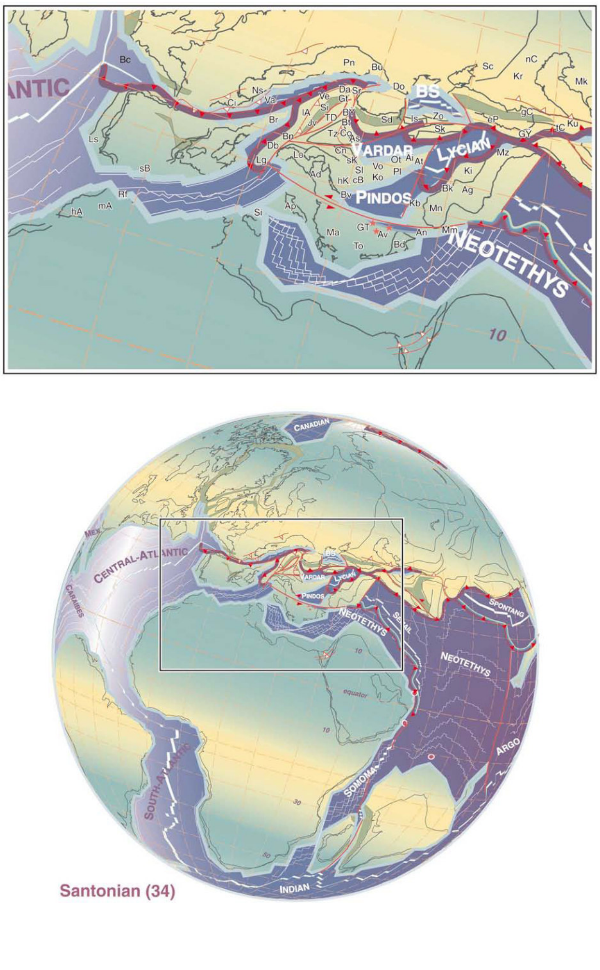


Figure 9

Figures 3 to 9.
Early Permian to Late Cretaceous reconstructions of the western Tethyan realm, (modified from Stampfli et al., 2001a; Stampfli and Borel, 2002), rift zone are shown in greenish color, foreland in orange color. Legend: AA, Austro-Alpin; Ab, Alboran; Ad, Adria s.str.; Ae, Abadeh; Ag, Aladag; Ah, Agh-Darband; Ai, Argolis ophiolites; Aj, Ajat; Al, Alborz; An, Antalya; Ap, Apulia s.str.; AP, Aspromonte, Peloritani; Ar, Arna accretionary complex; As, Apusenisouth, ophiolites; At, Attika; AT, Alpine Tethys; Av, Arvi ; Ba, Balkanides, externat ; Bc, Biscay, Gascogne; Bd, Beydaghlari; Be, Betic;
Bh, Bihar; Bi, Ba'id; Bk, Bolkardag; Bn, Bernina; Br, Briançonnais; Bs, Bisitoun seamount; Bt, Bator-Szarvasko ophiolites; Bu, Bucovinian; Bü, Bükk; Bv, Budva; By, Beyshehir; Bz, Beykoz basin; Ca, Calabride; cA, central Afganistan; cB, central Bosnia; Cc, Caucasus s.l.; cI, central Iberia; Ci, Ciotat flysch; Co, Codru; Cn, Carnic-Julian; CP, Calabria- Peloritani; cR, circum-Rhodope; Cv, Canavese; Da, Dacides; Db, Dent Blanche; DD, Dniepr-Donetz rift; Dg, Denizgören ophiolite (IP suture);
DH, Dinarides-Hellenides; Do, Dobrogea; Dr, Drina-Ivanjica; Ds, Drimos ophiolites; Du, Durmitor; Dz, Dzirula; eP, east Pontides; Er, Eratosthen; Fa, Fatric; Fr, Farah basin; gC, great Caucasus; Gd, Geydag; Ge, Gemeric; GS, Gory-Sovie; GT, Gavrovo-Tripolitza; Gt, Getic; Gü, Gümüshane; hA, high-Atlas; Ha, Hadim; He, Helvetic rim basin; Hg, Huglu; hK, high karst; Hr, Hronicum; Hy, Hydra; IA, Izmir-Ankara ocean; Ig, Igal trough; Io, Ionian; iP, intra-Pontides; Is, Istanbul; Ja, Jadar

; Jf, Jeffara rift; Jo, Jolfa; Jv, Juvavic; Ka, Kalnic; Kb, Karaburun; Kd, Kopet-Dagh ; Kf, Kotel flysch; Ki, Kirshehir; Kk, Karakaya forearc; Ko, Korab; KS, Kotel-Stranja rift; KT, Karakum-Turan; Ku, Kura; Kü, Küre ocean; Ky, Kabylies; La, Lagonegro; Lg, Longobucco; lA, lower Austroalpine; Lg, Ligerian ; Li, Ligurian; Lo, Lombardian; Ls, Lusitanian; Lu, Lut ; Ly, Lycian ophiolitic complex; mA, middle Atlas; Ma, Mani; Mc, Maliac rift/ocean; Md, Mozdak; MD, Moldanubian; Me, Meliata rift/ocean; Mf, Misfah seamount; Mi, Mirdita autochton; Mk, Mangyshlak rift; Mm, Mamonia accretionary complex; MM, Meguma- Meseta ; Mn, Menderes; Mo, Moesia; Mr, Mrzlevodice fore-arc; Ms, Meseta; MS, Margna-Sella; Mt, Monte Amiata fore-arc; Mz, Munzur dag; nC, north Caspian; NCA, North Calcareous Alps; Ni, Nilüfer seamount; Nr, Neyriz seamount; Ns, Niesen flysch; Ot, Othrys-Evia ophiolites; Pa, Panormides; Pd, Pindos rift/ocean; Pe, Penninic; Pi, Piemontais; Pk, Paikon intra oceanic arc; Pl, Pelagonian ; Pm, Palmyra rift; Pn, Pienniny rift; Px, Paxi; Py, Pyrenean rift; Qa, Qamar; Qi, Qilian; Qn, Qinling north; Qs, Qinling south; Rf, Rif, external; Rh, Rhodope; Ri, Rif, internal; Ru, Rustaq seamount; Rw, Ruwaydah seamount; sA, south Alpine; sB, sub-Betic rim basin; Sc, Scythian platform; sC, south Caspian basin; Sd, Srednogorie rift-arc; Se, Sesia (western Austroalpine); Sh, Shemshak basin; Si, Sicanian; Sj, Strandja; Sk, Sakarya; sK, south-Karawanken fore arc; Sl, Slavonia; Sm, Silicicum; SM, Serbo-Macedonian; Sr, Severin ophiolites; SS, Sanandaj-Sirjan ; sT, south-Tibet; St, Sitia; Su, Sumeini; Sv, Svanetia rift; Ta, Taurus, s.l.; Tb, Tabas; TB, Tirolic-Bavaric; tC, Transcaucasus; TD, Trans-Danubian; To, Talea Ori; Tr, Turan; Tt, Tatric; Tu, Tuscan ; Tv, Tavas+Tavas seamount; Ty Tyros fore arc; Tz, Tizia; uJ, upper Juvavic; Va, Valais trough; Ve, Veporic; Vo, Vourinos (Pindos) ophiolites; Vr, Vardar ocean; wC, western- Crete (Phyl-Qrtz) accr. cpl; WC, West-Carpathian; Ya, Yazd

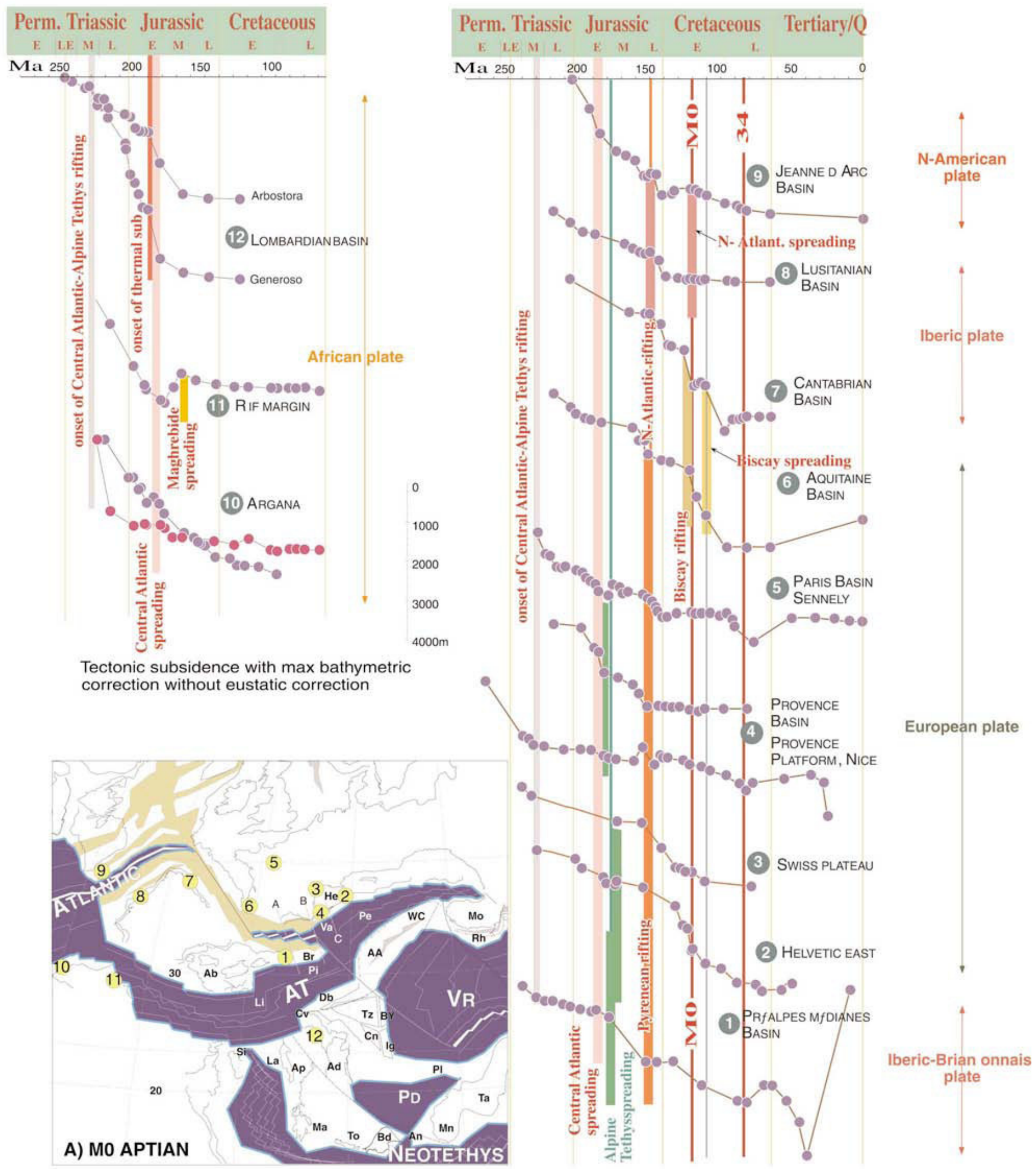


Figure 10. Synthetic subsidence curves for selected circum-Iberia and Atlantic domains (modified from Borel, 1997 and Stampfli and Borel, 2000) and location map. The subsidence program used to derived the curves is from R. Schegg., Time scale fromOdin (1994)

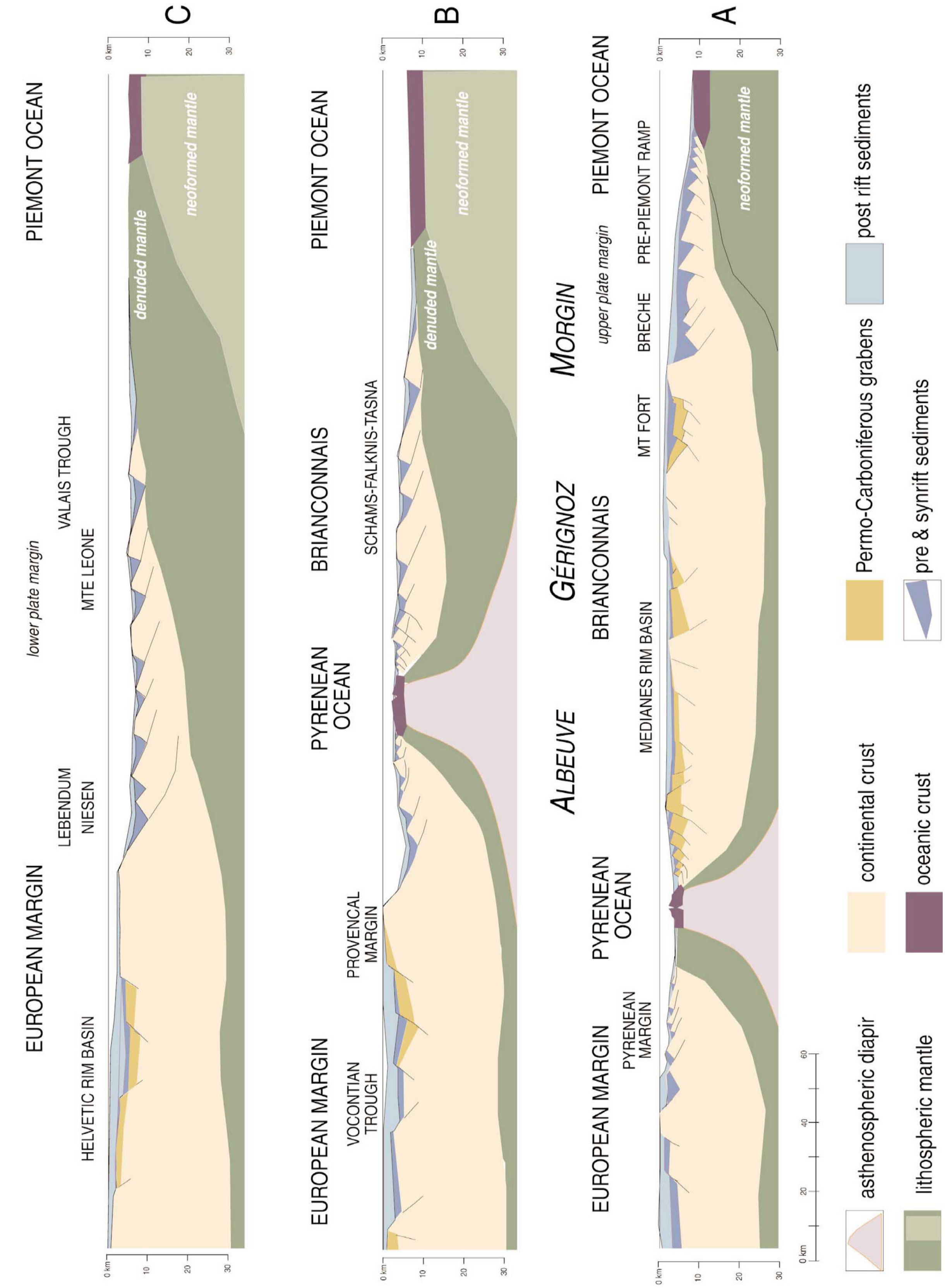


Figure 11. Early Cretaceous cross-sections of the European margin and position of the Pyrenean ocean. Location of cross-section in figure 10

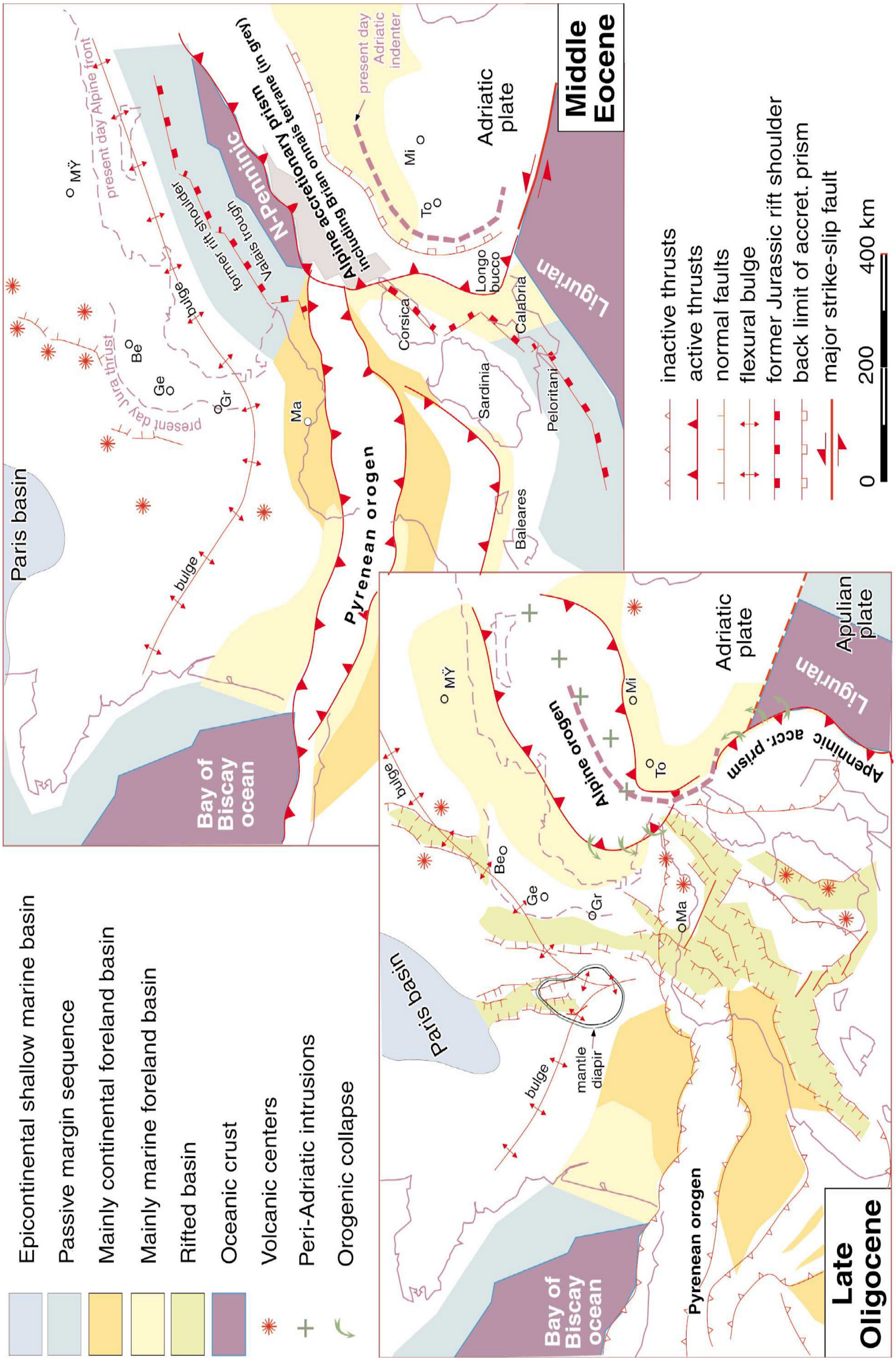


Figure 12. Middle Eocene and Late Oligocene maps of the Alpine Pyrenean area (modified from Stampfli et al., 1998)

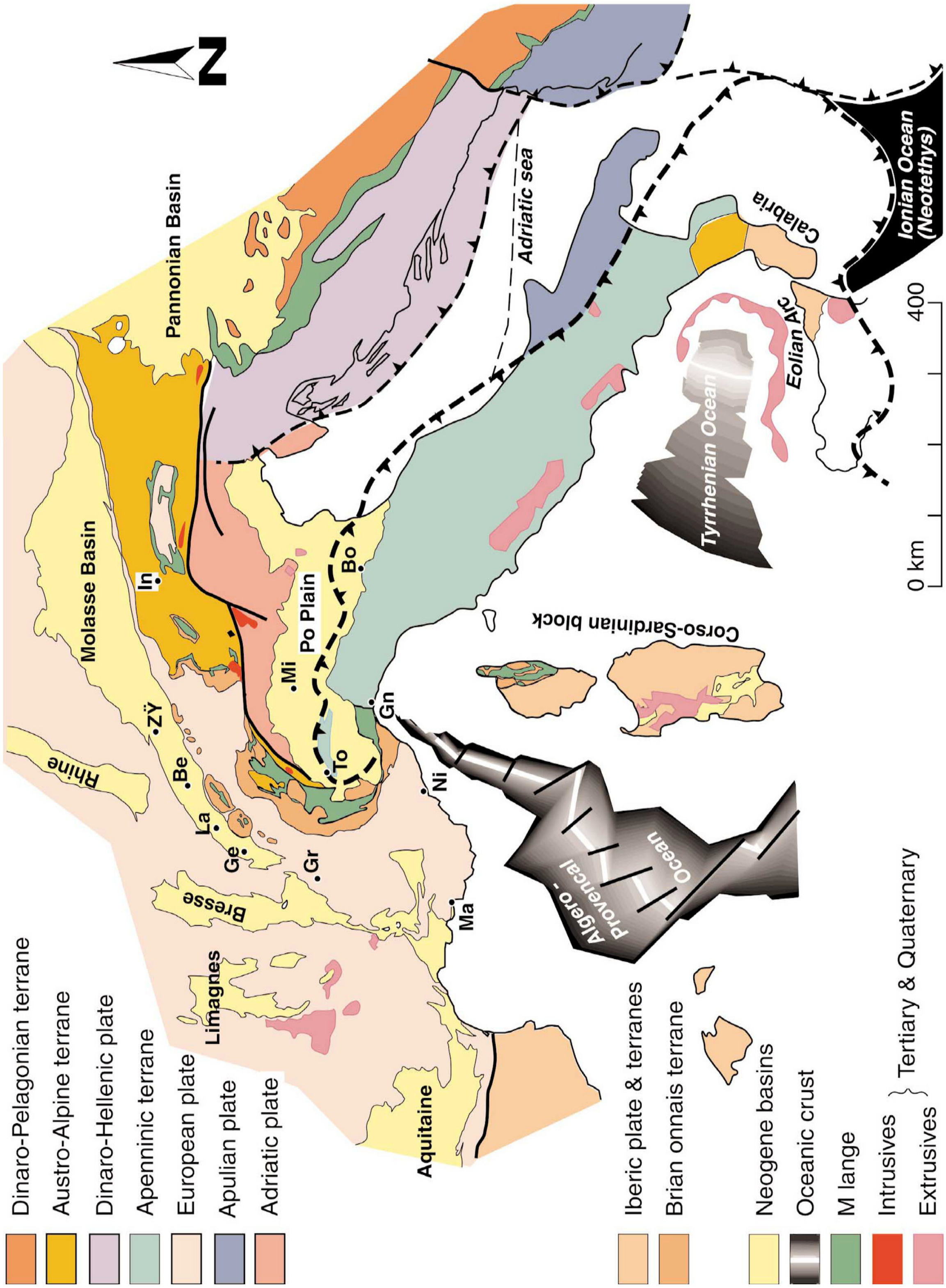


Figure 13. Simplified present-day tectonic map of the Alpine-Mediterranean area

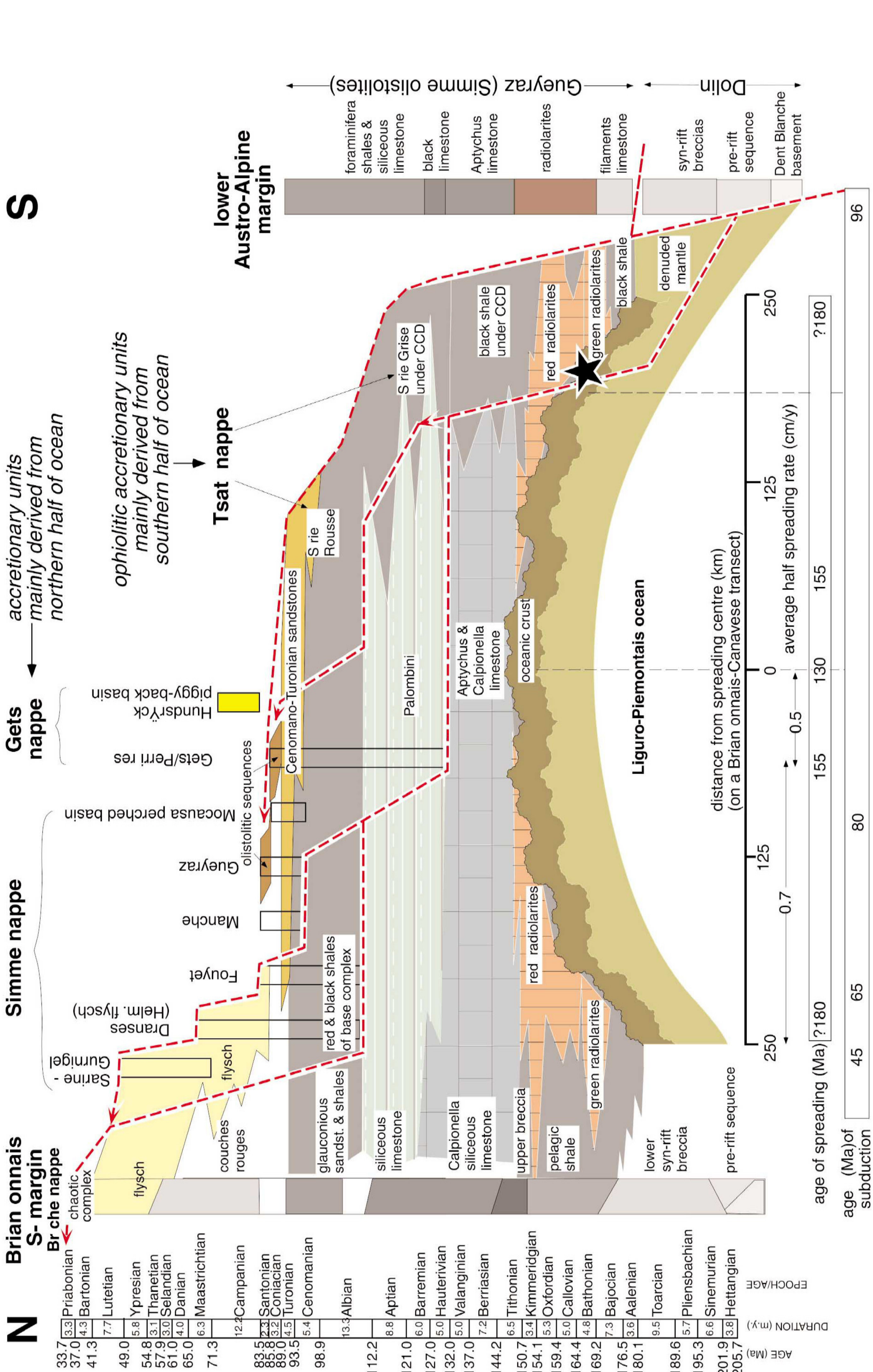


Figure 14. Stratigraphy and origin of the oceanic series and accretionary sequences found in the western Swiss Alps (modified from Stampfli et al., 1998)

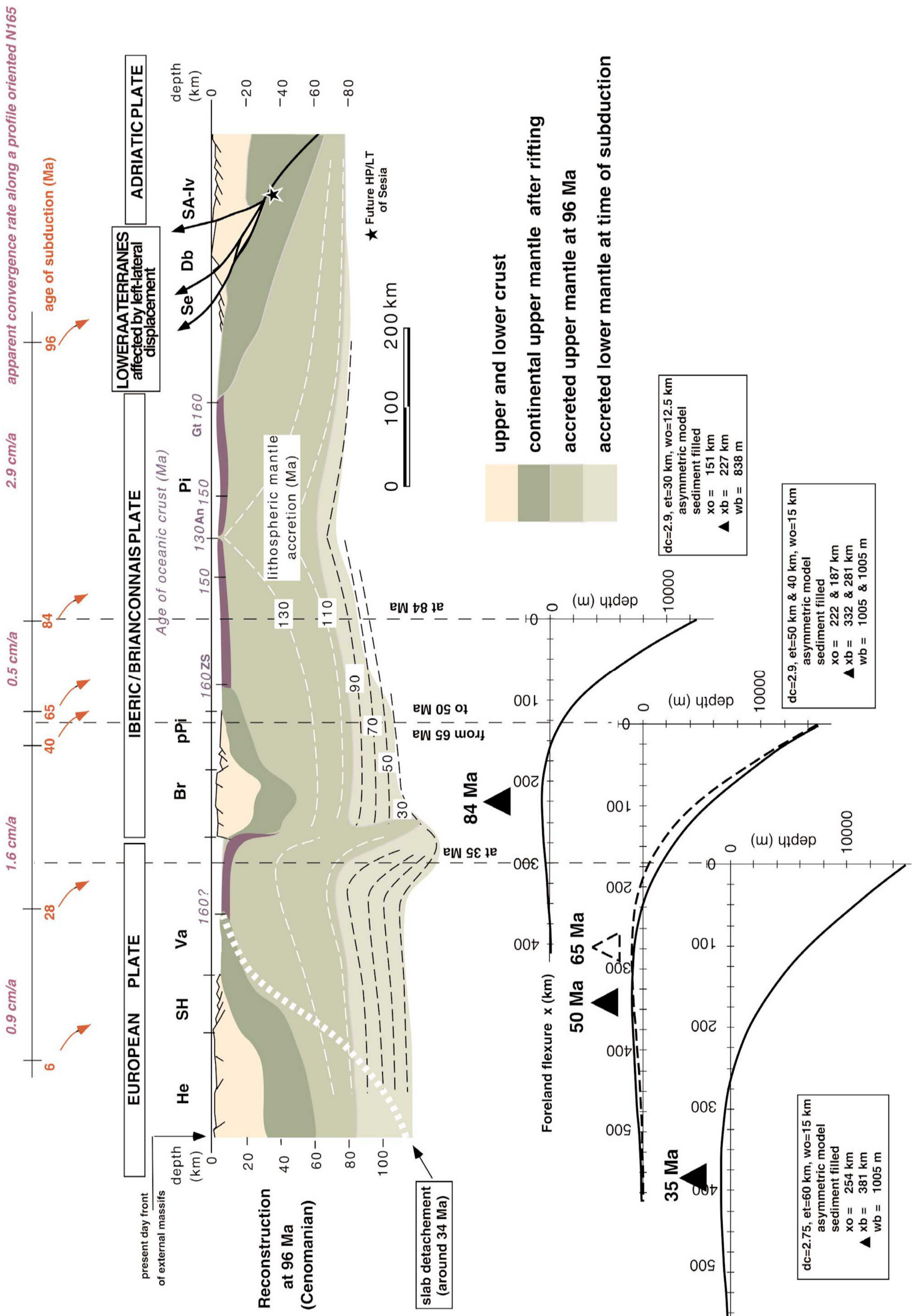


Figure 15. Evolution of the flexural bulge during the closing of the Alpine Tethys (modified from Stampfli et al., 1998). The flexural programme used to construct the curves is from M. Burkhard (Neuchâtel)

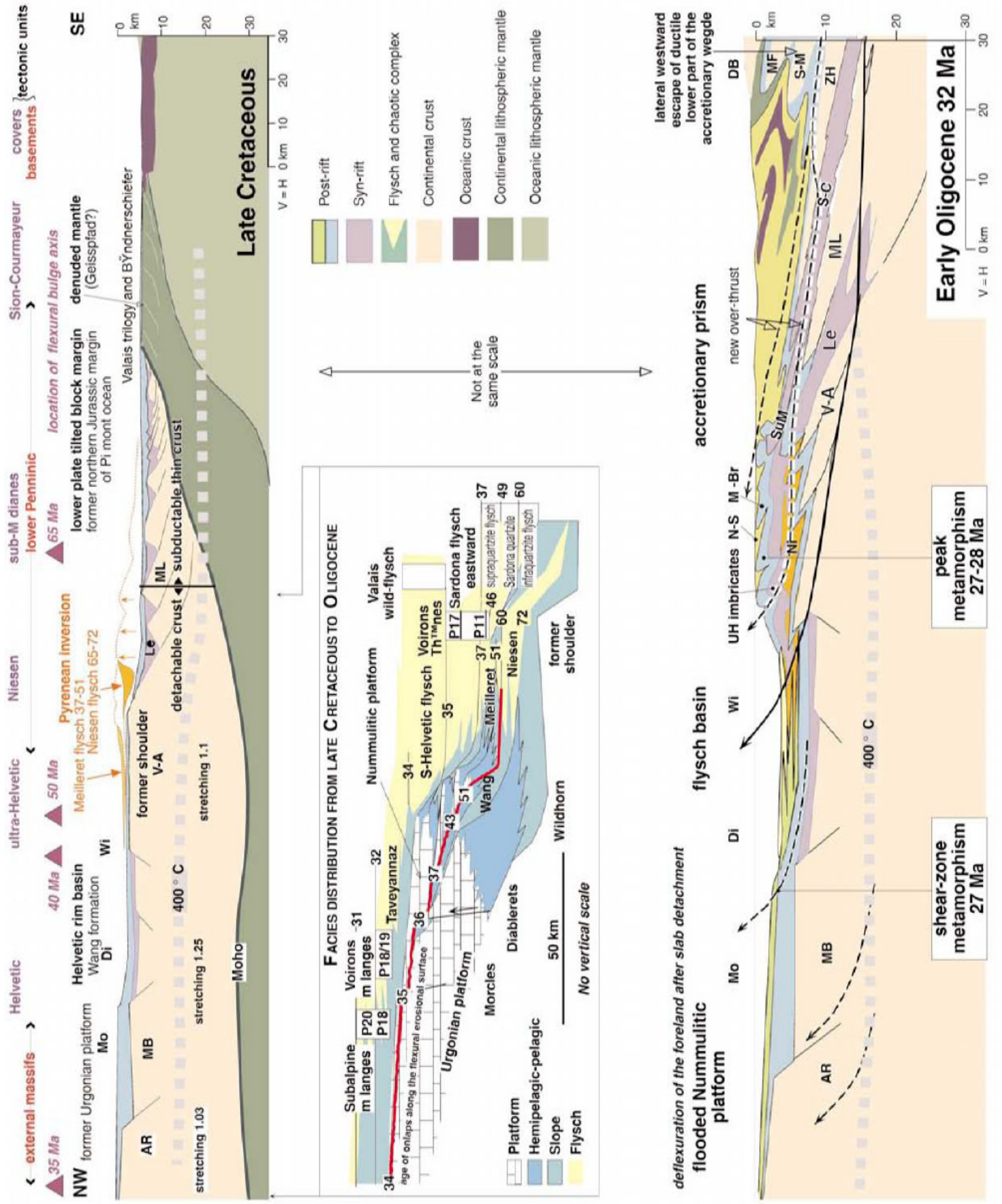


Figure 16. Reconstruction of the Helvetic margin in Maastrichtian and Early Oligocene (modified from Stampfli et al., 1998). See text for discussion and references about the timing of events. Stretching factor from Loup (1992). AR, Aiguilles rouges massif; Di, Diablerets nappe domain; DB, Dent Blanche 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 napes; Wi, Wildhorn nappe domain; ZH-Po, zone Houillère, Pontis nappes

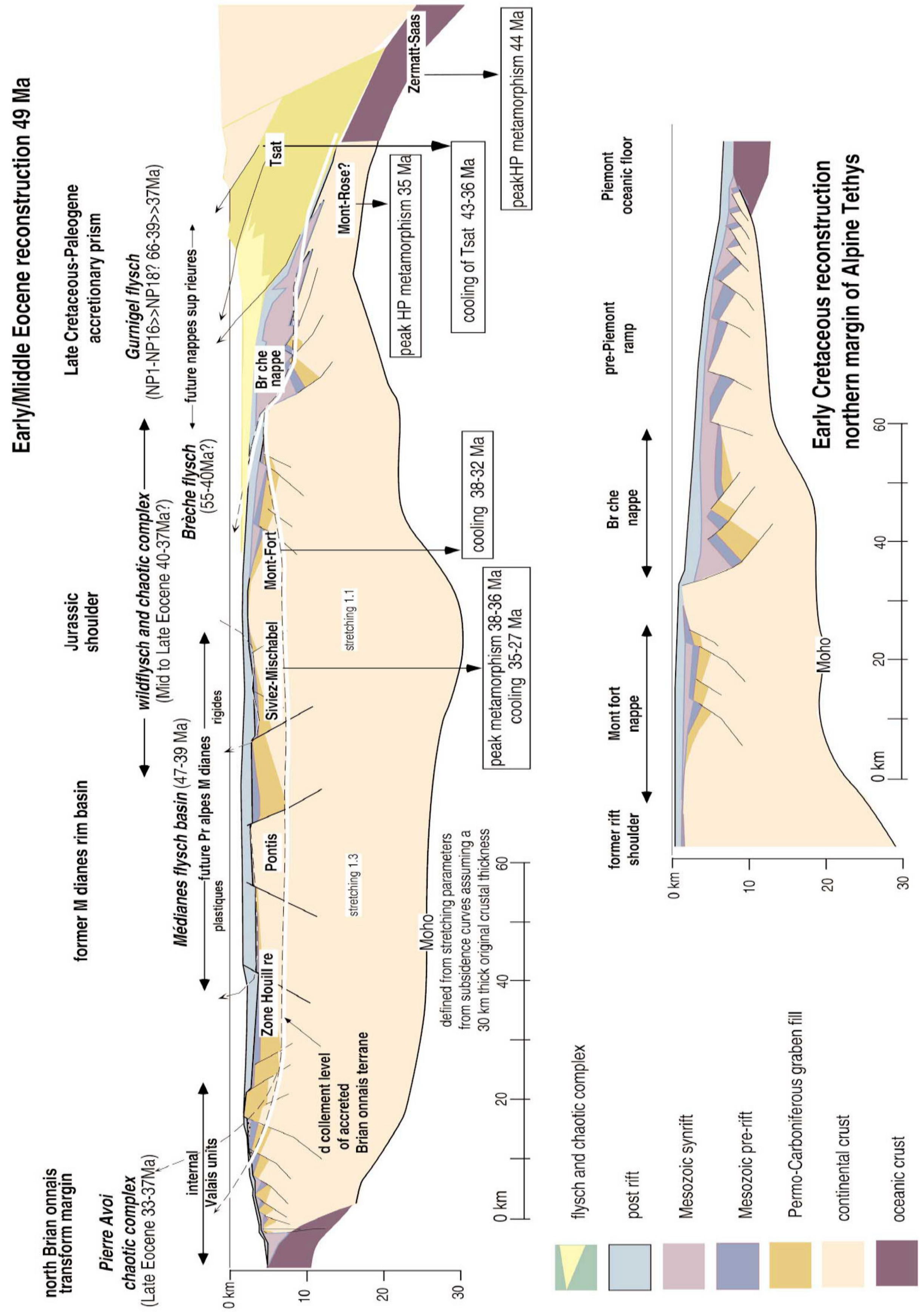
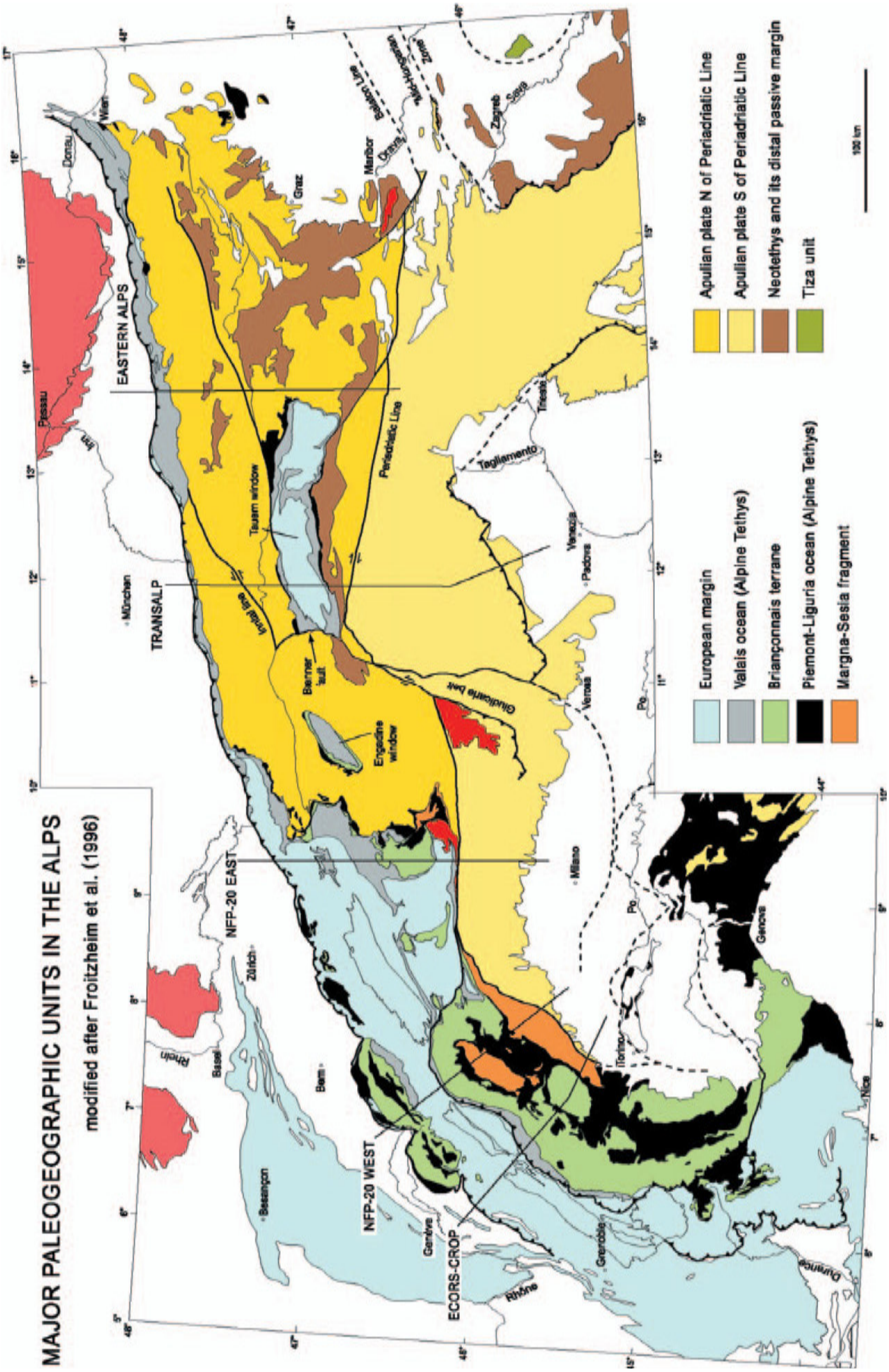


Figure 17 Reconstruction of the Briançonnais domain in Early Cretaceous and early Middle Eocene, modified from (Stampfli et al., 1998). See text for discussion and references about the timing of events. Stretching factor from Marchant and Stampfli (1997b)



Tectonic map of the Alps
http://pages.unibas.ch/earth/tecto/

Regional tectonics: from the Rhine graben to the Po plain, a summary of the tectonic evolution of the Alps and their forelands.

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Please refer to the tectonics groups of University of Basel

http://pages.unibas.ch/earth/tecto/

(html version by R. Bousquet & P. Dèzes)

1. Introduction

Firstly, this contribution gives a short overview of the overall architecture of the Western and Central Alps and their forelands (Po-plain and northern foreland) on the basis of three recent geophysical-geological transects, the locations of which are given in Figure 1. Secondly, the evolution of the Alpine system is discussed in time slices, starting with Cretaceous orogeny and ending with some evidence for very recent movements in the area of the Rhine graben. Some aspects of neotectonics and earthquake hazard are addressed as well, but only as far as they are directly related to tectonic movements which occurred during the geological past.

2. The major tectonic units of the Alps

Figure 1 presents an extremely simplified sketch map of the Alps which primarily highlights the transition from Central to Western Alps that will be discussed in profile view along three major transects.

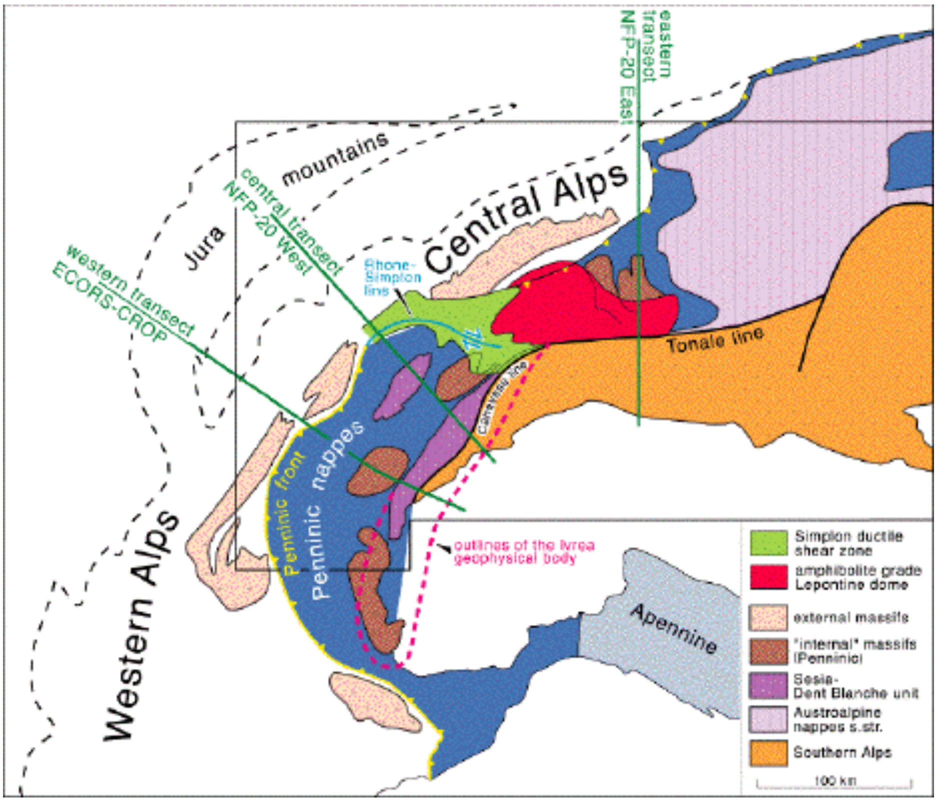


Fig. 1: Sketch map of the Alps, indicating locations of the three geophysical-geological transects depicted in Figure 3. From Schmid and Kissling (2000).

The Insubric line marks the northern and western boundary of the **Southern Alps**. The Southern Alps are characterized by a dominantly south-verging fold- and thrust belt whose southern tip is stratigraphically sealed by the Messinian (7 Ma) unconformity below the Po plain. At the base of this very young (Miocene) foreland prism we find the Adriatic middle and lower crust, including the Adriatic mantle, from which 10 to 15 km thick slices, consisting of basement and its Mesozoic sedimentary cover have been detached. This style of deformation points to the availability of a potential decollement horizon within the granitic upper crust within a depth interval of 10-15 km, corresponding to temperatures in the range of 250° to 375°C (assuming 25°C/km). Very probably, this depth corresponds to the brittle-plastic transition within granitic crust. This transition is due to the onset of crystal plasticity in quartz (at about 270°C according to Van Daalen et al., 1999, and/or reaction-enhanced ductility due to break-down reactions of feldspar at about 250°C according to Fitz Gerald and Stünitz, 1993). Before this post-collisional Miocene shortening, hence during Paleogene plate convergence and collision, the lithosphere of the present-day Southern Alps (the Adriatic plate) formed the upper plate, under which the Penninic Valais and Piemont-Liguria oceans, an intervening microcontinent (Briançonnais) and finally the European continental margin were subducted. Figure 1 also depicts the outlines of the **Ivrea geophysical body**, which represents the western edge of the Adriatic plate. **The Ivrea Zone**, a belt of South Alpine lower crustal rocks, is the surface expression of the Ivrea geophysical body. Because this lower crust has been exhumed to moderate depth, corresponding to less than 300°C already during Mesozoic rifting, it represents a particularly rigid part of the South Alpine basement at the WNW front of the Adriatic indenter.

Most of the about 100km Oligo-Miocene dextral strike slip along the E-W-striking eastern branch of the Insubric line (the Tonale line) has been taken up by dextral strike slip movements along the **Simplon ductile shear zone** and the **Rhone-Simplon line** (Schmid and Kissling, 2000). Hence, from Oligocene to probably recent times, the Western Alps are kinematically part of the WNW-moving **Adriatic indenter**, causing WNW-directed thrusting along the **Penninic front** of the Western Alps and within the Dauphinois foreland. The Rhone-Simplon line continues to act as a major discontinuity both in terms of seismic activity and the character of the stress regime up to the present day. During the latest stages of orogeny, this WNW directed indenting by the Adriatic plate possibly migrated further into the foreland, now also affecting the western Molasse basin and causing arcuate folding in the Jura mountains (Burkhard, 1990a).

The Central Alps are characterized by ongoing N-S-shortening during the Oligo-Miocene, i.e. coeval with WNW-ESE shortening in the Western Alps. These diverging transport directions necessitate an orogen-parallel extension, the effect of which is best documented by the **Simplon normal fault** and the exhumed amphibolite grade **Lepontine dome** in its footwall. Oligo-Miocene exhumation of the Lepontine dome is the combined effect of orogen-parallel extension, backthrusting along the Insubric line and fast erosion.

The units north of the Insubric line consist of the **Austroalpine nappes s.str.**, outcropping in eastern Switzerland only and extending into Austria. These units, although of similar paleogeographic provenance as the Southern Alps, consist of completely rootless flakes of basement and cover which have been detached (or delaminated) from their lithosphere already during Cretaceous orogeny. These nappes have been stacked towards the WNW and their former (Cretaceous) tectonic front runs almost perpendicular across the present-day Alps in Eastern Switzerland (Grisons). The **Sesia-Dent Blanche unit** of the Western Alps underwent an alpine tectono-metamorphic history which is different from that of the Austroalpine nappes s.str. and the Southern Alps (subducted near the Cretaceous-Tertiary boundary). However, its pre-alpine basement exhibits close similarities to that of the Southern Alps.

The **Penninic units** are of extremely heterogeneous paleogeographic provenance (remnants of oceanic lithosphere, a continental fragment referred to as Briançonnais, as well as basement of the European margin). Deformation is penetrative and polyphase, most of the Penninic units are overprinted by metamorphism (except for the Préalpes Romandes which have been detached and transported towards the northern foreland during the Eocene).

The **Helvetic nappes** have been detached from their former crystalline basement which must be looked for in the lowermost Penninic nappes. The units still attached to the European lithosphere consist of the **external massifs and their cover**, slightly detached from the lower crust during the Miocene, when deformation started to migrate into the foreland, eventually displacing the western **Molasse basin** and the **Jura mountains** by up to 30km from the Serravallian (12 Ma) onwards. The southern **Rhine graben** represents an Eo- Oligocene continental rift, kinematically linked to the Bresse graben situated west of the Jura mountains and ultimately to the opening of the Western Mediterranean basin (but not to the Alps). The geometry of Oligocene extensional faulting exerts a profound influence on Miocene to recent movements in the Jura mountains and their northern margin in the southern Rhine graben.

3. The major paleogeographic units of the Alps

Figure 2 depicts the following major paleogeographic units of the Alps, many of them being only preserved as extremely thin slivers, detached from the subducting lithosphere and accreted as slices (so-called nappes) to the upper plate (Austroalpine and South Alpine units).

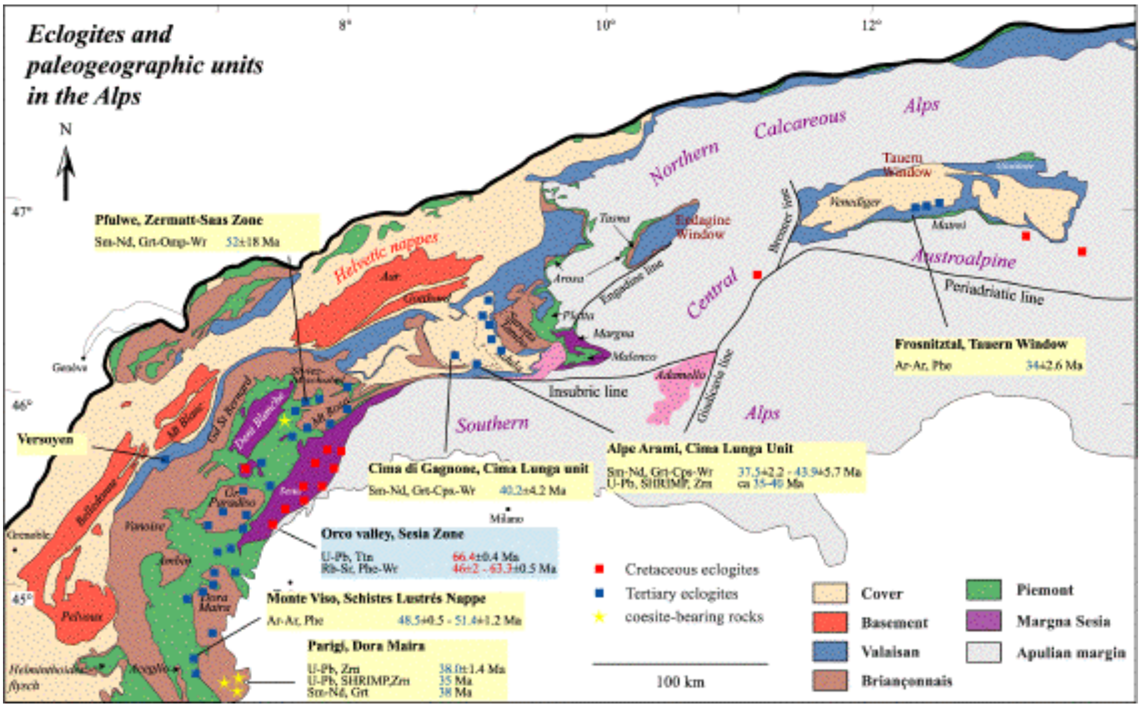


Fig. 2: Paleogeographic map of the Alps, indicating the present-day position of the major paleogeographic units of the Alps. Note that this is not a "tectonic" map in the traditional sense. After [Froitzheim et al. (1996)](#).

European margin: External massifs and their cover (extending northward underneath the Molasse basin) and Helvetic cover nappes, whose crystalline substratum has to be looked for within the deepest part of the Lepontine dome (lowermost "Penninic" nappes). Note that the European-derived basement can be traced southward almost to the Insubric line. This drastically illustrates very substantial exhumation of formerly subducted and newly accreted European lithosphere during the formation of the Lepontine dome. Its later exhumation is due to a combined effect of retro-flow (backfolding and backthrusting along the Insubric line) and unroofing by orogen-parallel extension and erosion during the post-collisional stages of orogeny. Note that some of these units also underwent Tertiary eclogitization.

Valais ocean: The remnants of this ocean predominantly consist of Cretaceous Bündnerschiefer, grading into Tertiary flysch and at least partly deposited onto oceanic lithosphere. Eclogitic mafics are preserved in the Versoyen of the Western Alps while blueschists and carpholite bearing rocks (low temperature — high pressure mineral) are preserved in the Engadine window. The Valais ocean opened near the Jurassic-Cretaceous boundary: The remnants of this ocean presently define a northern Alpine suture zone which closed during the Late Eocene.

Briançonnais microcontinent: This microcontinent was attached to Iberia and formed the northern passive continental margin of the Jurassic Piemont-Liguria ocean, before it broke off the European margin in conjunction with the opening of the Valais ocean. The Mesozoic cover of the Briançonnais microcontinent largely consists of platform sediments with frequent stratigraphic gaps ("mid-Penninic swell").
Its basement is preserved in the form of the Tambo-Suretta, Maggia and the Bernhard-M.Rosa nappes in the eastern, central and western Penninic realm respectively. Detached cover nappes form a substantial part of the Préalpes Romandes.

Piemont-Liguria basin: Oceanic lithosphere formed during the middle Jurassic to Early Cretaceous and is characterized by a classic alpine ophiolite suite. Sea floor spreading was followed by the deposition of radiolarites and aptycha limestones. During the Cretaceous the deposition of trench deposits (Avers Bündnerschiefer of Eastern Switzerland and schistes lustrées of Western Switzerland) indicates that the southern, i.e. the Apulian, margin of this basin had been converted into an active margin. In eastern Switzerland the Piemont-Liguria units (Arosa and Platta unit) were involved in top-WNW Cretaceous orogeny. However, the Piemont-Liguria ocean did not completely close before the onset of Tertiary orogeny.

Margna-Sesia fragment: A small fragment of the Apulian margin, that was rifted off Apulia during the opening of the Piemont-Liguria ocean, was later incorporated into the accretionary wedge along the active northern margin of Apulia.

Apulian margin: North of the Insubric line this southern margin is only preserved in the form of rootless basement and cover slices (Austroalpine nappes s.str.). South of the Insubric line it corresponds to the Southern Alps and their lithospheric substratum, the Adriatic plate (part of the larger "Apulian" plate). A third oceanic domain, the Meliata-Hallstatt ocean, formed during the Triassic. It is only preserved in eastern Austria and merely plays a role for understanding Cretatceous (Eoalpine) orogeny.

4. Three major Alpine transects and their deep structure

The major features common to all three transects, schematically sketched in Figure 3, are
(1) ESE to south directed subduction of the European lithosphere,
(2) a gap between European and Adriatic Moho,
and (3) the presence of wedge-shaped bodies of lower crust, largely decoupled from the piling up and refolding of thin flakes of upper crustal material (the Alpine nappes).
However, there are very substantial differences in the geometry and kinematic evolution of the eastern transect (Figure 3c) as compared to the western and central transects (Figure 3a and Figure 3b, respectively).

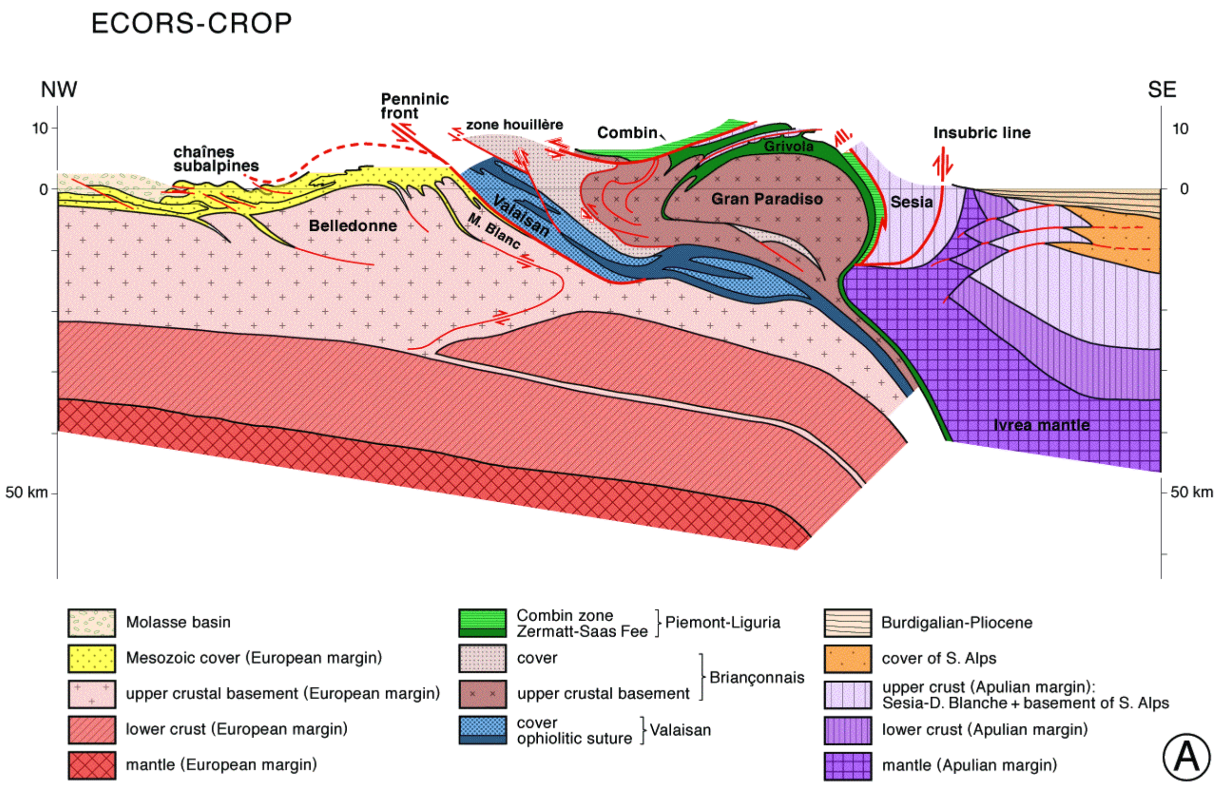


Fig. 3a

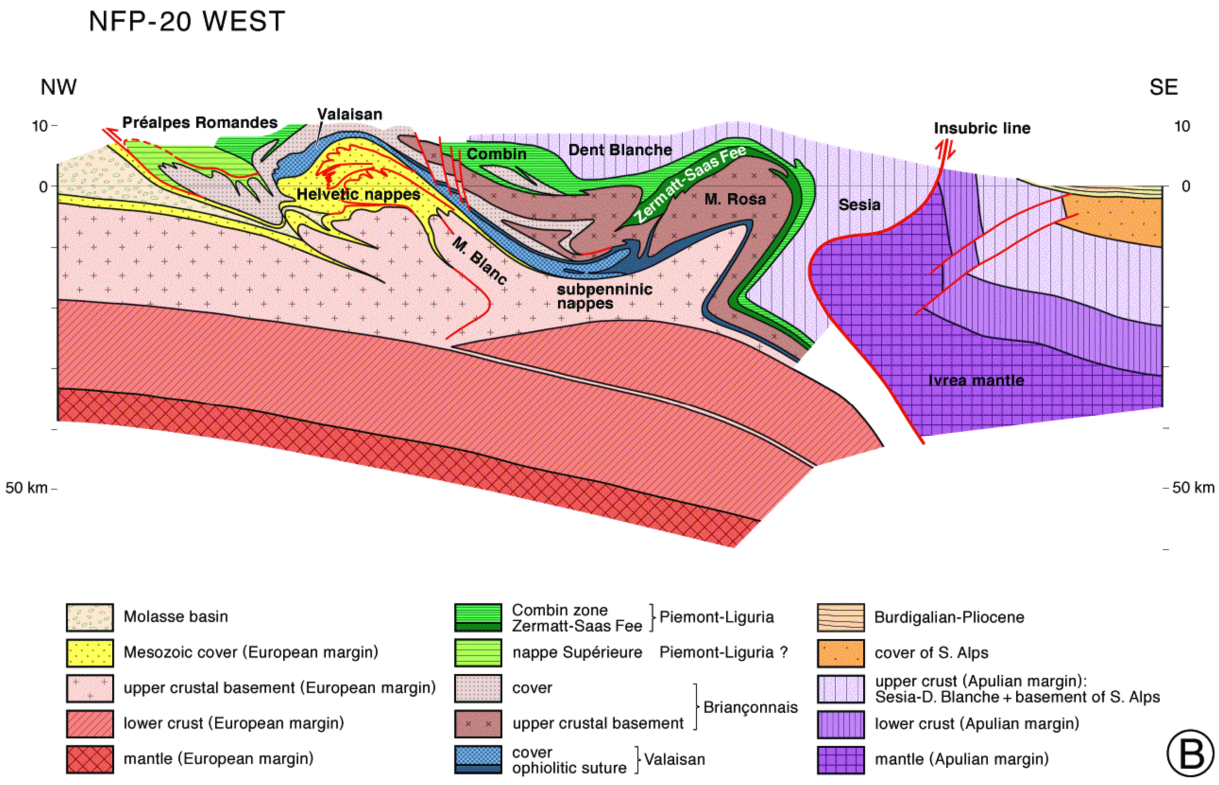


Fig. 3b

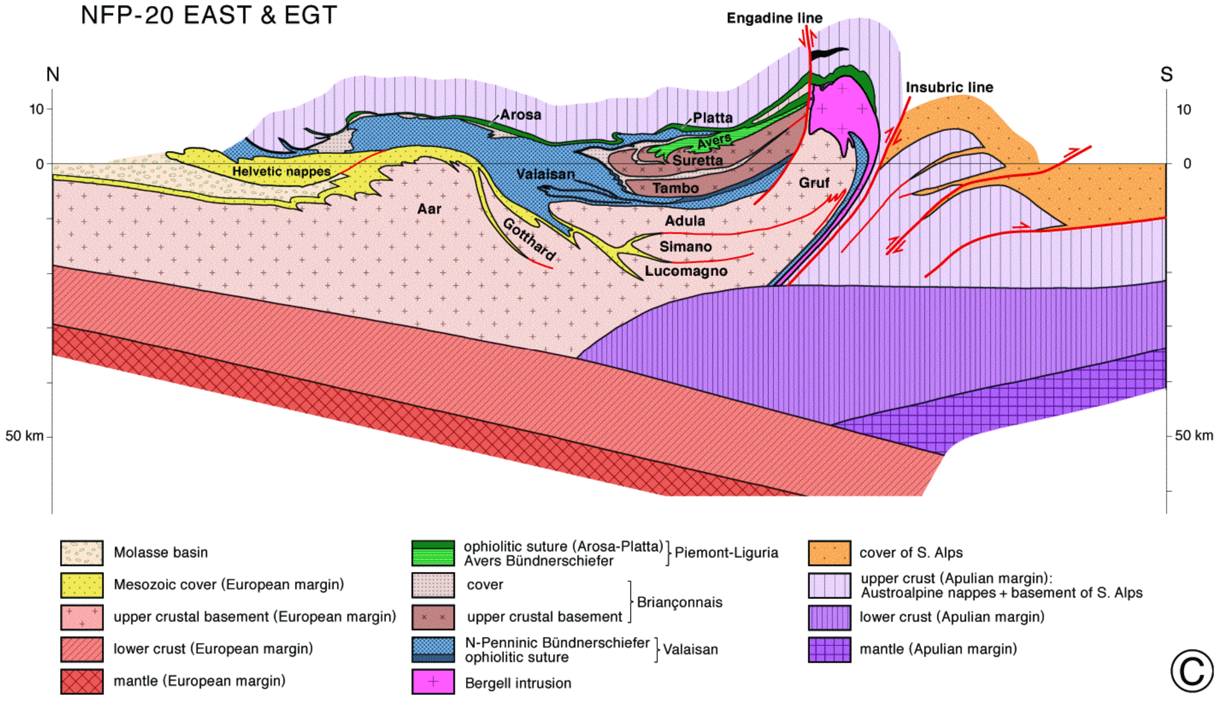


Fig. 3c

Fig. 3: Three schematic geophysical-geological cross sections through the western and central Alps (profile traces indicated in Figure1). Superimposed circles mark weel-locatable earthquake foci for the 1980-1995 time period, projected into the sections from within a 30 km couloir.

The following observations can be made:

1. In the eastern (NRP-20 East) transect (Figure 3c) the Adriatic Moho descends northward and toward its contact with the lower (European) crust), while in the central and western transects (Figure 3a and Figure 3b) this same Adriatic Moho rises toward the surface when approaching the contact zone with the European lithosphere. This contrast finds its expression also in the surface geology. In the eastern transect the southern Alps form an impressive south vergent foreland fold and thrust belt ("retro-wedge") riding above the Adriatic lower crust, while this same Adriatic lower crust is exposed in the Ivrea zone, situated at the SE end of the central and western transects. The Ivrea zone and Ivrea geophysical body wedge out eastward and do not extend into the area covered by the eastern transect.

2. In the eastern (NRP-20 East) transect a wedge of Adriatic lower crust is found above European lower crust and below European upper crust at its northern tip. This slice of Adriatic lower crust was wedged into the European lithosphere during the Miocene, splitting apart along the boundary between the upper and lower crust. For the western (ECORS-CROP) and central (NRP-20 West) transects a somewhat similar but geometrically different process of wedging is inferred. In these latter two cases, however, the lower crustal wedge is interpreted to be derived from the European lithosphere. Hence the wedges of lower crust seen in the eastern (Figure 3c) and the western and central transects (Figure 3a and Figure 3b), respectively, are of different origin and, thus, cannot be laterally connected. The observation discussed above, which implies that the Adriatic lower crust descends below the Penninic nappe stack in the eastern profile while it rises to the surface in the western transect, independently supports the conclusion that the two lower crustal wedges are not laterally connected.

3. In the area immediately north of the Insubric line, the eastern transect (NRP-20 East) exhibits substantial back thrusting and backfolding of all the Penninic nappes, including the Valais suture zone. This was associated with exhumation of the amphibolite-grade Lepontine dome and the deep-seated Bergell. In the western transect (ECORS-CROP), however, back thrusting does not affect the Valais suture zone and appears to be restricted to the units above this suture (within the Briançonnais upper crust). Note that Barrovian-type amphibolite-grade rocks have not been exhumed to the surface in the western transect, where the Insubric line only exhibits minor vertical offset.

4. The orogenic lid of the Austroalpine nappes, under which Penninic and Helvetic nappes were accreted in the eastern transect, is absent in the western and central transects.

Well-located earthquake foci have been orthogonally projected over a maximum distance of 30km onto the transects of Figure 3. Comparing (Figure 3a, Figure 3b and Figure 3c) significant differences are evident in the depth of seismogenic regions. The maximum depth of earthquakes is situated near the Moho in the northern and southern forelands along the eastern traverse (Figure 3c). Beneath the Penninic units, i.e., within the Lepontine metamorphic dome, they are restricted to the thickened upper crust. Also note that the Adriatic lower crustal wedge is quiescent. Coincidence of the lower limit of seismicity with predicted isotherms based on thermal modelling (Okaya et al., 1996) suggests that the 500°C isotherm controls the cataclastic-plastic transition. Quiescence within the Adriatic lower crustal wedge further suggests that stress transmission between the European and Adriatic lithosphere is largely restricted to upper crustal levels in case of the central Alps.

In contrast, the earthquake distribution along the western and, to a lesser degree, along the central transects exhibits a wide, east dipping corridor of foci affecting the entire transect, including the allochthonous European lower crust (Figure 3a and Figure 3b). Thus, mechanical coupling and stress transmission between the Adriatic microplate and the European lithosphere occur along a deep-reaching seismogenic zone. This indicates a contrasting (with respect to the eastern transect) present-day thermal regime, primarily caused by the following substantial differences in the kinematic evolution. Firstly, oblique convergence and collision in the western Alps before 35 Myr ago must have led to a significantly smaller volume of accreted radiogenic upper crustal rocks, as compared to the central Alps, the latter being characterized by head-on convergence and collision. Secondly, double verging displacements of the central and western Alps after 35 Myr ago allowed for orogen-parallel extension in the central Alps (Lepontine dome), associated with updoming of the isotherms. The Penninic realms of the central and western Alps differ significantly not only in deeper crustal architecture but also in the thickness of the seismogenic zone.

Some of the earthquake foci (down to a depth of about 10-15km) depicted in Figure 3 are known to be associated with normal faulting within the axial zone of the Alps, while strike-slip and/or thrusting mode prevails in the northern and southern forelands of the Alps (Sue et al., 1999; Maurer et al., 1997). The cause of this normal faulting within the central parts of the Alps is yet unclear (gravitational collapse and/or buoyant rise of the lithospheric root). However, compression in both forelands suggests ongoing compressional coupling between Adriatic and European plates, although no focal solutions are available yet for the deep (>15km) foci depicted in Figure 3.

In summary, the discussion on the three transects reveals major differences between Central Alps (Figure 3c) and Western Alps (Figures 3a and 3b). As shown in Figure 1, the limit between these two different segments of the Alpine chain coincides with the Rhone-Simplon line which continues to be seismically active and which separates different present-day stress domains (Maurer et al., 1997).

5. Inferences concerning the rheological behaviour from a tectonicians point of view

The maximum depth of the seismogenic zone (assumed to coincide with the maximum depth of cataclastic, i.e. friction-controlled and dilatant deformation mechanisms) is a widely disputed topic. Below some partly speculative inferences will be made, based on field observations and deductions from the geometry of the present-day deep structure of the Alps. Observations made by structural geologists focussing on the study of deformation microstructures indicate the onset of crystal plasticity at vastly different temperatures for different minerals under natural strain rates:

Anhydrite (decollement horizon in the context of Jura-folding) may deform by crystal plasticity above about 70°C (Jordan, 1994). **Calcite** exhibits significant non-cataclastic deformations above about 180°C (Burkhard, 1990b). **Quartz** starts to deform by crystal plasticity above some 270°C (van Daalen et al., 1999). **Feldspar** does not start to deform by crystal plasticity below some 450-500°C (Tullis, 1983), despite the predictions based on the application of experimentally derived flow laws, which indicate crystal plasticity from about 380°C onwards. However, break-down reactions in feldspar may promote reaction-enhanced ductility at lower temperature (Fitz Gerald and Stünitz, 1993), provided that water is available. Minerals such as **hornblende** and **pyroxenes** are definitely more flow resistant than feldspar and **Olivine** does not start to flow by crystal plasticity below 700°C (Handy and Zingg, 1991).

At first sight these data suggest a fairly shallow base for the seismogenic zone in the quartz-rich upper crust. However, elevated pore pressures are able to displace the brittle-plastic transition to greater depth (i.e. higher temperatures). On the other hand, deformation of more mafic lower crustal rocks is predicted to be controlled by cataclastic deformation at temperatures lower than some 450-500°C (i.e. down to Moho-depth for an undisturbed geotherm within the foreland), assuming that their strength is controlled by feldspar in the absence of significant amounts of quartz. Hence, contrary to the predictions based on the extrapolation of experimentally determined flow laws, lower crustal rocks may be very flow resistant and may deform by cataclastic mechanisms. Hence, it is not surprising to find deep foci within lower crustal rocks as found in Figure 3 in case of the Alpine forelands. However, lower crustal rocks may become weak within overthickened crustal roots of mountain belts an/or if heat flow is elevated.

The geometry of the deep structure along the transects given in Figure 3 independently suggests that **lower crustal rocks are flow resistant, in contrast to a common belief amongst earth scientists that the lower crust is generally "weak"**. Lower crustal wedging demands the lower crust to remain little deformed (or undeformed) and calls for decollement horizons at the top, as well as at the base of the lower crust. Since quartz starts to deform by crystal plasticity already above 270°C, the upper crust may easily detach from the lower crust. Given the high strength contrast between feldspar and olivine, detachment at the base of the lower crust may also occur, provided that temperatures at around 450° to 500°C (onset of crystal plasticity in feldspar) are reached within the lowermost continental crust.

It is interesting to note that the Adriatic lower crustal wedge in the NRP-20 East profile (Figure 3c) is presently aseismic, in contrast to the European lower crustal wedges in Figures 3a and 3b. This points to differences in the thermal regime between the Central and Western Alps. In case of the NRP-20 East profile the lower limit of seismicity roughly coincides wirh the 500°C isotherm as predicted by thermal modelling along this transect (Okaya et al., 1996). This independently supports the inference that the base of the seismogenic zone in the lower crust does indeed coincide with the cataclastic-plastic transition for feldspar near the 500°C isotherm.

6. Evolution of the Alpine system and its forelands in time slices

This discussion largely focuses on the evolution along the eastern (NFP-20 East) transect of Figure 3c where timing is best constrained. Figure 4 gives a timetable of orogeny, while Figure 5 depicts cross sections along this eastern transect for different time slices.

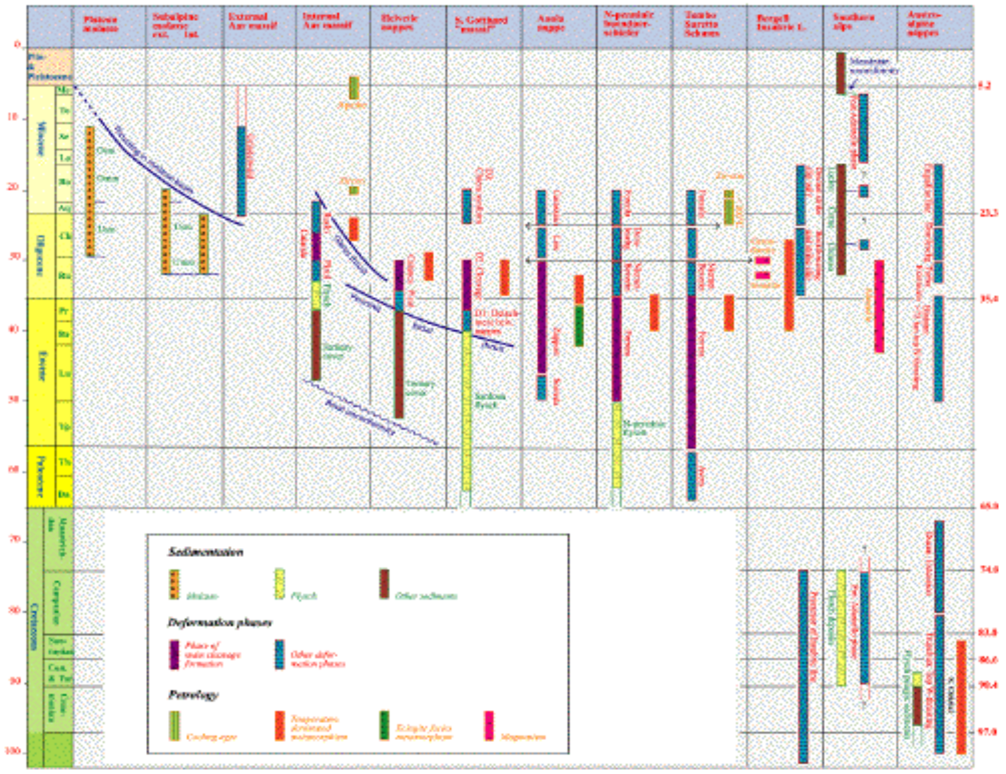


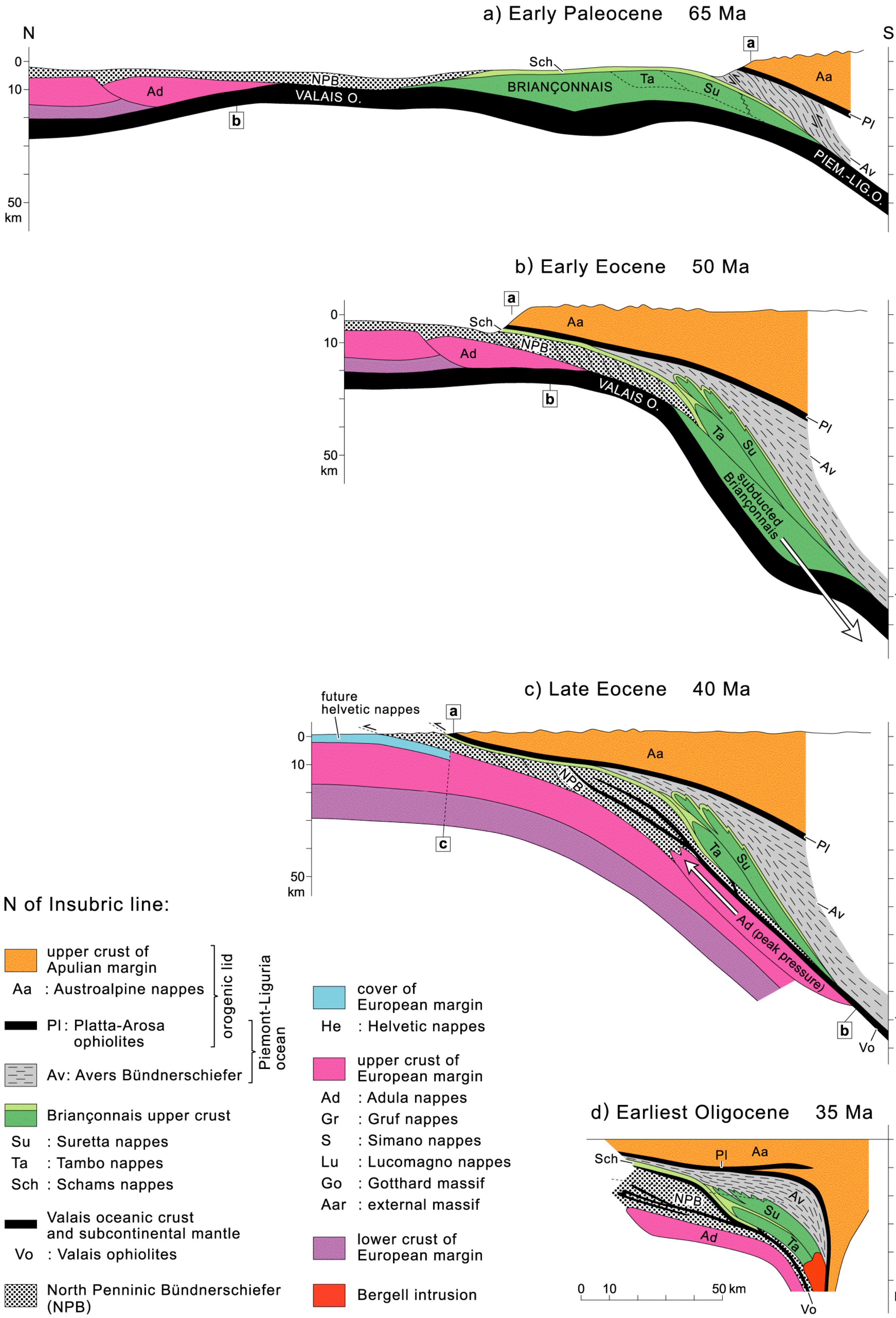
Fig. 4: Correlation table showing an attempt to date deformation phases and metamorphism along the eastern transect (NPF-20 East). For an extensive discussion of timing constraints see Schmid et al. (1997), for a brief overview, see Schmid et al. (1996). From Schmid et al. (1996).

6a. Cretaceous orogeny

Cretaceous (or Eo-alpine) orogeny in the Eastern Alps is regarded as independent and unrelated to Tertiary orogeny because of its different kinematic scenario (top WNW, hence almost orogen-parallel thrusting) and because it is separated from Tertiary convergence by an extensional event during the Late Creatceous ("Ducan extension" in Figure 4). Apart from the Austroalpine nappes, it only affects the Piemont-Liguria units of Eastern Switzerland (Arosa-Platta) while the rest of the Penninic units remains largely unaffected by this orogeny which did not propagate further to the west beyond Eastern Switzerland, nor down-section into the Briançonnais units.

The attribution of a pre-Adamello phase in the Southern Alps (main deformation of Miocene age) to Cretaceous orogenic activity is uncertain, but a precursor of the Insubric line must have been active (separation between the detached crustal flakes of the Austroalpine nappe system from the Adriatic lithosphere which remained intact). However, the southern margin of the Piemont-Liguria margin represented an active margin, as documented by the accretionary wedge of the schistes lustrées and by the eclogitization of the Sesia unit at around the Cretaceous-Tertiary boundary.

During the various stages of Tertiary orogeny, the pre-structured Austroalpine nappe system, together with the Arosa-Platta ophiolites, formed a rigid upper plate (referred to as "orogenic lid" in Figure 5), of which the Southern Alps (not depicted in Figure 5a to 5c but present at the southern margin of these figures) formed part.



a) Early Paleocene 65 Ma
N
S
0
10
50
km
Sch
Ta
Su
Aa
NPB
Ad
VALAIS O.
BRIANÇONNAIS
Pl
Av
PIEM.-LIG.O.
a
b
b) Early Eocene 50 Ma
subducted Briançonnais
100
c) Late Eocene 40 Ma
future helvetic nappes
Ad (peak pressure)
Vo
c
d) Earliest Oligocene 35 Ma
50 km
N of Insubric line:
upper crust of Apulian margin
Aa : Austroalpine nappes
Pl : Platta-Arosa ophiolites
orogenic lid
Av: Avers Bündnerschiefer
Piemont-Liguria ocean
Briançonnais upper crust
Su : Suretta nappes
Ta : Tambo nappes
Sch : Schams nappes
Valais oceanic crust and subcontinental mantle
Vo : Valais ophiolites
North Penninic Bündnerschiefer (NPB)
cover of European margin
He : Helvetic nappes
upper crust of European margin
Ad : Adula nappes
Gr : Gruf nappes
S : Simano nappes
Lu : Lucomagno nappes
Go : Gotthard massif
Aar : external massif
lower crust of European margin
Bergell intrusion

Fig. 5a-d

6b. Early Tertiary convergence and subduction (65-50 Ma)

During the Paleocene the Briançonnais terrane enters the subduction zone, thereby closing the last remnants of the Piemont-Liguria ocean in eastern Switzerland, the youngest sedimentary cover of which now forms an accretinary wedge consisting of the Avers Bündnerschiefer (Figure 5a). Very probably, this southern ocean remained open for a longer period of time in the Western Alps. After some 200 km of N-S-convergence (1.3 cm per year) the distal margin of Europe (future Adula nappe) enters the subduction zone at around **50 Ma**, now closing the Valais ocean (Figure 5b). Penetrative deformation during this time interval is largely restricted to the southernmost Penninic units, i.e. the Briançonnais terrane (Tambo-, Suretta- and Schams nappes, see Figure 4) and the Avers Bündnerschiefer of the Piemont-Liguria ocean (Figure 5b).

6c. Tertiary collision (50-35 Ma)

During the middle and late Eocene (i.e. between Figure 5b and 5d) some additional 200 km N-S plate convergence (corresponding to 1.5 cm per year) were taken up by the incorporation of the Valais ocean and the distal European margin into a growing accretionary wedge below the orogenic lid formed by the Austroalpine nappes. Figure 4 illustrates the migration of deformation and metamorphic events towards the northern foreland, reaching the area of the future Helvetic nappes by the end of the Eocene.

Note that a total of some 400 km N-S convergence across the Central Alps asks for substantial sinistral strike slip movement across the future Western Alps (see discussion in Schmid and Kissling, 2000). Hence, the western Alps formed under a sinistrally transpressive scenario during Early Tertiary convergence and collision, with W-directed movements post-dating Tertiary collision (see post-collisional stage 1).

Since the Alpine nappes in Figure 5 exclusively consist of thin slices of upper crustal basement and/or its cover, detached from their lower crustal and mantle substratum, all European (and Valaisan) lower crust (including parts of the upper crust) must have been subducted together with the mantle lithosphere (Figure 5c). Hence, N-vergent nappe stacking during this collisional stage took place within an accretionary wedge which starts to grow as more non-subductable upper crustal granitic material of the European margin enters the subduction zone. Radiogenic heat production within this granitic basement, perhaps in combination with slab break-off (depicted in Figure 5e) leads to a change in the thermal regime and to Barrovian-type (Lepontine) metamorphism.

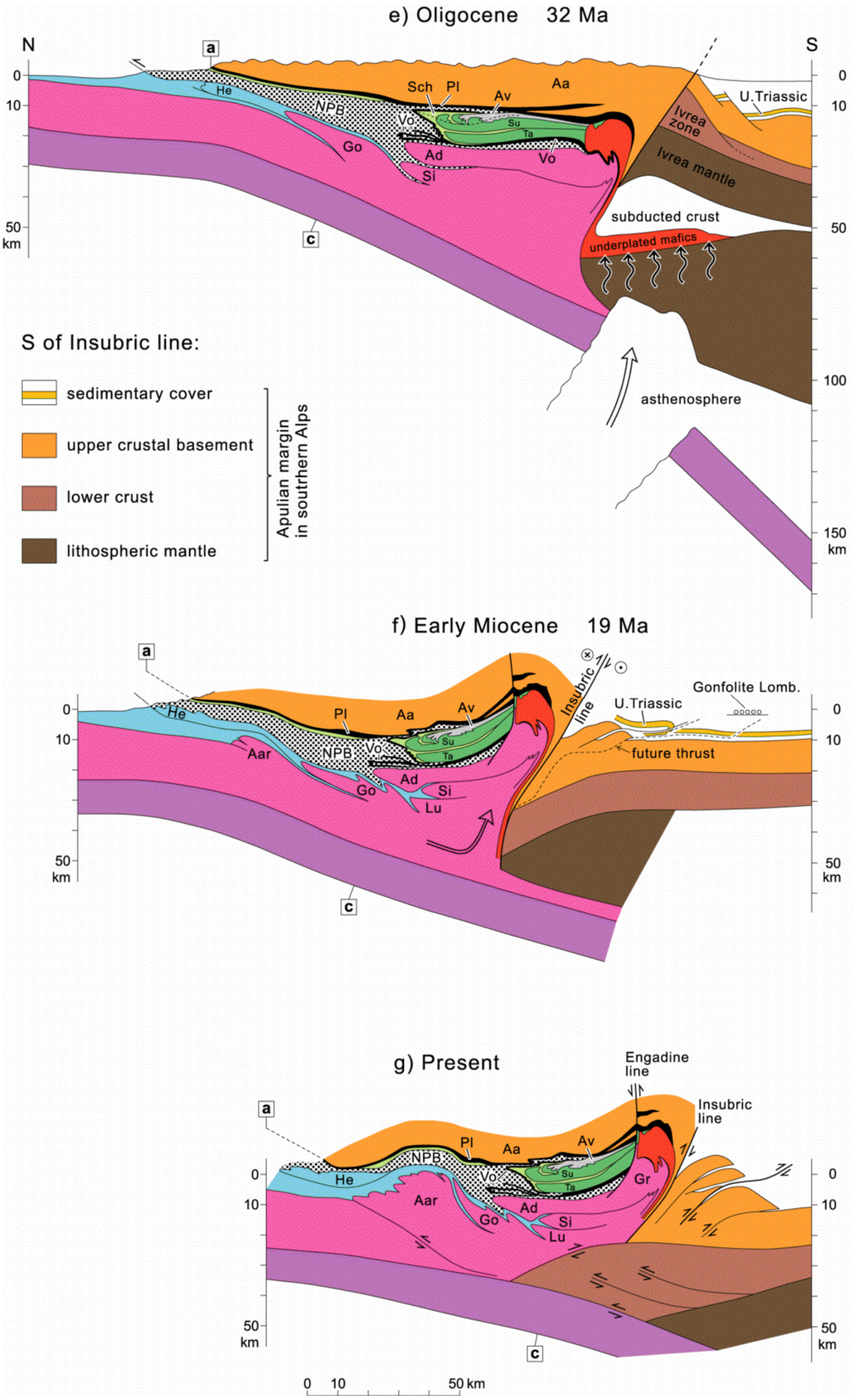


Fig. 5: Scaled and area balanced sketches of the kinematic evolution of the eastern Central Alps from early Tertiary convergence (a-b) to collision (c) and post-collisional shortening (d-g). From Schmid et al. (1996)

6d. Post-collisional stage 1 (35-20 Ma)

Further growth of the accretionary wedge leads to a situation whereby part of the material entering the subduction zone has to be retro-thrusted above the steeply N-dipping Insubric line towards the surface (Figure 5e and 5f). A "singularity point" (Beaumont et al., 1994) develops within the lower part of the upper crust, separating the subducting part of the European crust from that part of the wedge which is back-thrusted and -sheared in order to be exhumed by erosion (this singularity point is near the bent arrow depicted in Figure 5f).

As can be seen from Figure 4, forward thrusting in the Helvetic nappes (i.e. the Glarus thrust) is contemporaneous with retro- or backward thrusting along the Insubric line. Hence the orogen now becomes a bivergent one with a southern and northern foreland. Interestingly, this transition into bivergent thrusting coincides with increased rates of erosion due to the pop-up of the central Alps between fore- and retro-thrusts. This provokes the transition from flysch to molasse type sedimentation in the northern foreland.

N-S directed plate convergence during this first post-collisional episode amounting to about 60 km, slowed down to some estimated 0.45 cm per year. In map view, this time interval coincides with the WNW-directed movement of the Adriatic plate, now decoupled from the central Alps via dextral strike slip movement along the Tonale line (some 100 km). Kinematically, the western Alps are now part of the WNW-moving Adriatic plate and are separated from the Central Alps along the Simplon ductile shear zone and later on by the Rhone-Simplon line (see Figure 1).

Note that continental rifting in the Rhine and Bresse grabens falls into this same time interval. However, this rifting is kinematically unrelated to shortening across the Alpine system which remains in compression throughout.

6e. Post-collisional stage 2 (20-7? Ma)

Further crustal overthickening within the central part of the Alpine orogen by bivergent (retro- and pro-wedge) thrusting eventually led to a rather rapid propagation of the deformation front from the Insubric line towards the Po plain (Southern Alps), as well as towards the northern foreland (thrusting at the base of the Aar massif and within the southern Molasse basin) at around **20 Ma** ago. This is depicted in Figure 5e while the timing constraints are given in Figure 4. Regarding the Southern Alps, deformation stopped at around **7 Ma** ago (Messinian unconformity).

In the northern foreland, however, the situation is more complex. During the late Serravallian (**12 Ma**), deformation suddenly stepped further into the foreland, now also incorporating the western part of the Molasse basin and the Jura mountains into the orogenic wedge (Burkhard and Sommaruga, 1998). Whilst decollement along Triassic evaporites is recognized by most authors as being responsible for this forward stepping of the deformation front onto the northernmost Jura mountains up to the southern Rhine and the Bresse grabens two questions remain open:

(1) Did thin-skinned deformation stop at around **7 Ma** in the Jura mountains, i.e. contemporaneously with foreland deformation in the Southern Alps?

(2) How exactly did the arc of the Jura mountains form: clockwise rotation of the western part of the Molasse basin and the northern Alps as proposed by Laubscher (1961), or W to NW-directed indentation of the western part of the Central Alps as proposed by Burkhard (1990a)

In regard to the first question we will argue below that present-day deformation is thick-skinned, hence it is likely that Jura-folding was a short-lived event (**12-7 Ma**). Regarding the second question we favor an indentation model since there is evidence for counterclockwise rather then clockwise rotation of the Adriatic plate during the Miocene (Schmid and Kissling 2000).

Assuming that relatively fast plate convergence across the Alpine system of Switzerland stopped at around **7 Ma**, the 60 km plate convergence over the duration of this second post-collisional episode amounts to about 0.5 cm per year. Hence plate convergence remains practically unchanged between **35 and 7 Ma**. It will be interesting to compare this figure of 0.5 cm per year to present-day shortening estimates across the Alpine system.

6f) Recent movements in the Upper Rhine graben area

Figure 6 depicts some recent results from work in progress in the framework of EUCOR-UGENT (M. Giamboni, unpublished) concerning the area of the Upper Rhine graben in the Sundgau area west of Basel. The Sundgau Schotter have been deposited during a very short time interval from **3.2 to 2.6 Ma** according to [Petit et al. (1996)](). Presently they outcrop within a 20 km wide corridor between the Vosges and the frontal Jura mountains (see outlines of the base of the Sundgauschotter indicated in Figure 6). The base of these Sundgauschotter forms an excellent reference horizon for inferring relative vertical movements during the last **3 Ma** or so (their basal part needs not to have been deposited **3.2 Ma** ago everywhere, but certainly before **2.6 Ma** ago) provided that this basal contact may be assumed to be near-horizontal at the time of deposition. The fact that these gravels were deposited in a braided river environment indicates that their basis may be assumed to be nearly planar, with some very minor slope from E to W which cannot substantially modify the picture emerging from Figure 6 which is the following:

The contour map of the base of the Sundgauschotter (Figure 6) suggests substantial relative vertical uplift of the southernmost part of the depositional corridor in respect to the northernmost occurrences (in the order of 250m). Moreover, two very pronounced en-echelon anticlines, gently folding the base of the Sundgauschotter are inferred north of the Vendlincourt fold in the Rechésy area, these gentle folds being directly observable within Upper Jurassic limestones and Oligocene deposits below the base Sundgauschotter. Note also that the base Sundgauschotter is affected by at least part of the folding to be observed in the Ferrette fold.

The geometry of these folds, particularly in the area immediately east of Montbéliard, suggests thick-skinned reactivation of basement faults formed during Oligocene rifting. Hence, we propose a thick-skinned origin for approximately NNW-SSE-directed ongoing shortening, as indicated by the northernmost (post **3 Ma**) folds in the Basel area affecting the Sundgauschotter and very probably going on at present. This suggests that thin-skinned Jura folding may indeed have stopped some **7 Ma** ago. Such a postulate is compatible with (1) the present-day stress field in the Jura mountains, as determined by in situ stress measurements (Becker 1999) which indicates that the Jura belt is no longer an active thin-skinned fold and thrust belt and (2) by the inference that the historical Basel earthquake reactivated a deep-seated basement fault ([Meyer et al., 1994]()), and (3) the ocurrence of intra-crustal earthquakes within the Molasse basin (Figure 3c).

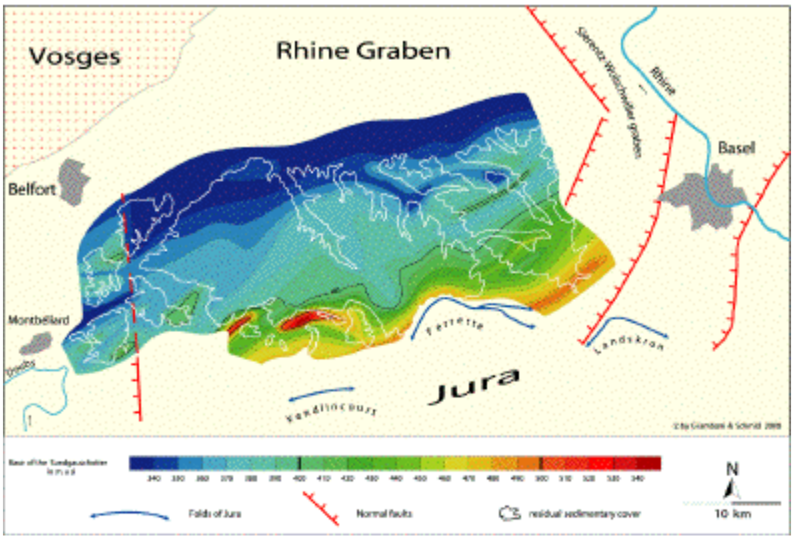


Fig. 6: Contour map of the base pf the Pliocene (3.2-2.6 Ma) Sundgau gravel deposits W of Basel. Assuming an originally planar base of these braided river deposits, the contours indicate folding with an amplitude of up to 200m. This indicates substencial tectonic movements during the last 3 Ma, post-dating thin-skinned Jura-tectonics. From [Giamboni and Schmid](), work in progress.

References

Burkhard, M. (1990a). Aspects of large-scale Miocene deformation in the most external part of the Swiss Alps (Subalpine Molasse to Jura fold belt). Eclogae geol. Helv., 83, 585-602.

Burkhard, M. and Sommaruga, A., 1998. Evolution of the western Swiss Molasse basin: structural relations with the Alps and the Jura belt. In: Mascle, A. et al. (eds), Cenozoic Foreland Basins of Western Europe. Geol. Soc. Special Publ., 134, 279-298.

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

Handy, M.R. and Zingg, A., 1991. The tectonic and rheological evolution of an attenuated cross section of the continental crust: Ivrea crustal section, southern Alps, nortwestern Italy and southern Switzerland. Geol. Soc. America Bull., 103, 236-253.

Maurer, H., Burkhard, M., Deichmann, N. and Green A., 1997. Active tectonism in the central Alps: Contrasting stress regimes north and south of the Rone valley. Terra Nova, 9, 91-94.

Okaya, N., Freeman, R., Kissling, E. and Mueller, S., 1996. A lithospheric cross section through the Swiss Alps, Part 1, Thermokinematic modeling of the Neoalpine orogeny. Geophys. J. Int., 125, 505-518.

Petit, Ch., Campy, M., Chaline, J. and Bonvalot, J., 1996. Major hydrographic changes in Alpine foreland during the Pliocene-Pleistocene. Boreas, 25, 131-143.

Schmid, S.M., Pfiffner, O.A., Froitzheim, N., Schönborn, G., and Kissling, E., 1996. Geophysical-geological transect and tectonic evolution of the Swiss-Italian Alps. Tectonics, 15, 1036-1064.

Schmid, S.M., Pfiffner, O.A. and Schreurs, G., 1997. Rifting and collision in the Penninic zone of eastern Switzerland. In: Pfiffner O.A. et al. (eds.), Deep Structure of the Alps, Results from NFP 20, 160-185.

Schmid, S.M. and Kissling, E., 2000. The arc of the western Alps in the light of geophysical data on deep crustal structure. Tectonics, 19, 62-85.

Schönborn, G., 1992. Alpine tectonics and kinematic models of the central Southern Alps. Memorie di Scienze Geologiche, XLIV, 229-293.

Part II
Excursion itinerary

Aug. 4 Tuesday:
Munich – Garmisch – Fernpass – Imst (150 km).
By cable car to the Muttekopf (Late Cretaceous sediments)
(Guide: Hugo Ortner).

We cross the alpine Border at Eschenlohe, pass Garmisch Partenkirchen and the landslide of the Eibsee, which was dated by Jerz & Poschinger 1995 to 3500 y b.p. (300 - 400 Mio. m^3)

Stop 1 Fernpass Landslide: 4150 y b.p. (U/Th on Aragonite); Albert Penck reckognized it as a landslide, ist volume is around 1 km^3 (3. largest landslide of the Alps)

Christoph Prager, Karl Krainer, Veronika Seidl & Werner Chwatal: SPATIAL FEATURES OF HOLOCENE STURZSTROM-DEPOSITS INFERRED FROM SUBSURFACE INVESTIGATIONS (FERNPASS ROCKSLIDE, TYROL, AUSTRIA). *Geo.Alp,* Vol. 3, S. 147–166, 2006

One of the largest mass movements in the Alps, the catastrophic Fernpass rockslide in the Northern Calcareous Alps (Tyrol, Austria), shows long run-out distances up to at least 15.5 km in length. Detailed field studies show evidence that preferentially in medial to distal accumulation areas the Sturzstrom kinematics could have been favoured by water-saturation of its low permeable fine-grained substrate.

2. Location and Geology: The Fernpass is situated in the western part of the Northern Calcareous Alps, approx. 45 km westnorthwest of Innsbruck (Tyrol, Austria) and enables an important North-South-passage between the Tyrolean Inn valley in the South and Bavaria (Germany) in the North. Its apex (1332 m a.s.l.) and the valley floors to both sides are covered by at least 16.5 km^2 wide spread rockslide deposits, attributing to one of the largest mass movements in the Alps.

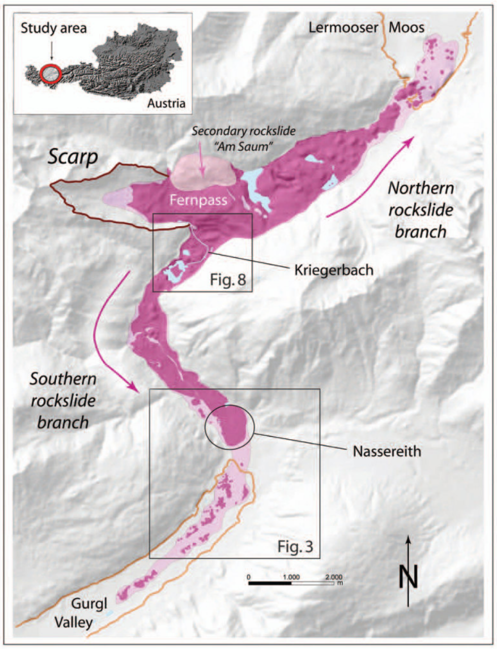
2.1. Scarp area: The source area of the Fernpass rockslide is located within the southernmost Lechtal nappe, a polyphase and heteroaxial folded and faulted major thrust unit of the western Northern Calcareous Alps (Eisbacher & Brandner, 1995). Here the calcareous rockslide debris originated from an exceptional deeply incised, wedge-shaped niche with a present maximum elevation of 2231 m a.s.l., indicating a failure volume of about 1 km^3. The scarp is made up by several hundred metre thick alternations of thinbedded platy dolomites, limestones and marls belonging to the bituminous Seefeld Fm (Norian, Upper Triassic). These incompetent and low permeable rocks represent an intraplatform-basin succession within the upper Hauptdolomit Group (Norian), one of the main rock units in the Northern Calcareous Alps (Brandner & Poleschinski, 1986, Donofrio et al., 2003).
At Fernpass lithological parameters and bedding conditions, but above all, complex intersection of brittle fault systems and fracture zones control the formation of preferred sliding planes and the block size distribution. Crucial slope vulnerabilities result from polyphase faulting along three dominant fault systems, i) E-W-trending normal and reverse faults, ii) NE-trending sinistral and iii) NW-trending dextral faults. Cataclasis induced from brittle faulting along the prominent NE-trending Loisach fault system (Eisbacher & Brandner, 1995) and along a NW-trending fault at Nassereith enabled fluvio-glacial erosion and valley- deepening (Fig. 1). This caused stress redistribution of the valley-slopes and uncovered favourable orientated sliding planes permitting subsequent slope instabilities.

2.2. Accumulation area: Seismic investigations near the Fernpass apex indicate a substantially steepened and undercut slope toe, where the top of the bedrock units is situated in about 700–800 m a.s.l. (Prager et al., in preparation). The overlying 500–600 m thick soft rock units are assumed to originate mainly from the Fernpass rockslide and decrease significantly in thickness laterally. Morphologically the medial to distal accumulation areas are characterised by largescale transversal debris ridges and trenches, exemplary formed Toma-hills and associated funnel- to basin-formed depressions, some of them filled with kettle-like lakes. According to the morphological definition by Abele (1974:119), the well-known Toma are "isolated, cone- to pyramidal- or roofshaped elevations, predominately made up by rockslide debris and characterised by more or less planar hill slopes with constant inclination". Formerly these typical hummocky characteristics of several large rockslides in the Alps, among them also the Fernpass slide, were believed to represent late-glacial deadice scenery (Abele, 1964, 1969). In contrast, Abele (1991a, 1997) favoured that the intensively structured rockslide scenery may result from pull-apart mechanisms, generating Horst- and Graben-like transversal debris ridges and depressions, during the rapid rockslides motion on water-saturated substrates. But the internal rockslide kinematics that generated the typical cone-shaped Toma, featuring sub-circular basal planes and occurring often as isolated individuals, is not established till now.

However, at Fernpass neither these unsmoothed structures nor the rough scarp shows any signs of glacial overprints and indicates a post-glacial genesis of the hummocky rockslide scenery. This assumption was backed up by the cross-check of three independent and remarkably well coinciding dating methods. Rockslide-dammed torrent deposits, situated close to the scarp-front, yielded a C-14 minimum age of at least 3380-3080 cal. BP, whereby an age of between 3300 and 4600 cal. yrs BP (Mid- Holocene) is assumed for the base of this backwater sequence. This coincides well with two cosmogenic radionuclide Cl-36 exposure ages of large-scale sliding planes at the scarp, which are 3600 ± 900 and 4800 ± 1100 yrs old. In addition, small-scale successions of the rockslide deposits are lithified by previously not mentioned carbonate cements. These have been dated by the Th-230/U-234 disequilibrium method and yielded a minimum age of about 4150 ± 100 yrs for the accumulation of the southward- deflected rockslide deposits. All age data indicate a failure event in the middle Holocene at about 4100–4200 yrs BP (Prager et al., in review). The internal structure of the Holocene Fernpass rockslide is characterised by chaotic deposits featuring varying block-size separation and fragmentation.

Upper parts of the proximal depositional facies contain large angular blocks up to a few metres in diameter and occasionally even slabs of 100´s metres in side length. Due to dynamic disintegration and abrasion of the surging debris, medial to distal areas are built up by subangular to even edge-rounded components of different size (centimetre to several metres) mixed with abundant fine interstitial material. Sieve analyses of 9 gravelly samples yielded approx. 5–30 weight-percent clay to silt-sized matrix. The basal sliding plane of the Fernpass rockslide, supposed to show fine attritionbreccias, is not exposed. Drilling data show that basal deposits of the adjacent Tschirgant-rockslide are made up by silt- to clay-sized, calcareous deposits with a low hydraulic permeability of about 4.0 - 5.0 x 10-9 m/s (Hartleitner, 1993).

2.3. Travel path of the Fernpass rockslide Due to the oblique impact on its opposite slope, the failing rock masses were proximally piled up to a few hundred metre thick succession and subsequently split into two channelled but diametrically opposed Sturzstrom branches (Fig. 1). The northern Sturzstrom, containing the majority of the debris volume, shows a comparable low deflection angle and surged at least 10.8 km towards northeast on to the aggradation plain of the Lermooser Moos at approx. 970 m a.s.l. This accumulation path is kinematically coherent and thus no further aim of this study.



In contrast, the trajectory flow of the southern Sturzstrom is characterised by unusually high deflection angles. First, the eastward sliding debris was deflected from the proximal accumulation area about 140° to southwest and subsequent, due to channelling in the narrow valley, about perpendicular to southeast. Then, after a travel distance of approx. 11 km from the topmost scar, this Sturzstrom entered at Nassereith an unconfined alluvial plain. But instead of continuing its further run-out path straight on, the debris curiously turned about 90° to southwest and flew another 4 km down the Gurgl valley to its lowermost accumulation point at 790 m a.s.l., covering a total run-out distance of at least 15.5 km. Based on this, both rockslide branches show extremely low angles for their overall slopes, i.e. the ratios of drop height versus run-out length along the channel line (referred to as "Fahrböschung", Heim 1932). This may be used as a geometrical criterion to describe landslides mobility and at Fernpass measures about 6.7° for the northern respectively 5.3° for the highly and curiously deflected southern branch.

Nearby in the Inn valley, there are several other large landslides:
The *Tschirgant landslide* deposits are the complex product of multiple failures, which can be dated to between ca. 3750 and ca. 3050 BC. It occupies a large area of 13 km² but its volume is only 240 Million m³.

Köfels landslide (Ötztal, Tyrol, Austria). About 8,700 years ago, some 3 cubic kilometers rock slided down the western side of the Ötztal (Heuberger 1994). The rubble blocked the *Ötztaler Ache*, which later formed a small canyon (Maurach). Friction caused the Gneiss at the base of the landslide to melt, forming a glassy rock known as ***frictionite***.

Stop 2 and the rest of this and the following day: Imst – Alpjoch: Meeting with Dr. Hugo Ortner (Innsbruck)

Please refer to the website of Hugo Ortner (Innsbruck):
http://www.uibk.ac.at/geologie/staff/the_hugo/muttekopf.html

Aug. 5 Wednesday:

Whole days walk in a high alpine area around the Muttekopf (2774 m):
Late Cretaceous "Gosau" sediments and geodynamic evolution of the Austroalpine nappes

Muttekopf Gosau beds (Late Cretaceous):

In the Eastern Alps compression during orogeny in the Upper Cretaceous caused crustal thickening, isostatic uplift and gravitational adjustment of the unstable orogenic wedge. This process triggered extensional basin formation on the back of the orogen (Gosau Basins). The basin fill of the Muttekopf Gosau Basin is arranged in megacycles, the first one comprising alluvial fan sediments and "Inoceramus marls" of the Lower Gosau Complex (FAUPL et al. 1987) of Santonian age. Three other cycles follow (Upper Gosau Complex, Campanian to Maastrichtian), consisting of turbiditic fining upward sequences, that are indicative for extensional tectonics during basin formation, as subsidence events prevent formation of autocyclic coarsening upward sequences and therefore prograding of the turbidite system. Deposition of the 1st and 2nd Megacycle occured below the CCD (Carbonate Compensation Depth). The carbonate rich 3rd Megacycle was deposited probably below the CCD after a period of palaeogeographic reorganisation (uplift?) in the source area. (after Ortner 1994).

The Upper Cretaceous sediments of the Muttekopf area are preserved in a large WSW-trending syncline in the Hauptdolomite of the Inntal Nappe (Fig. 15B). During Tertiary (post-Gosauian) compression, the Larsennscholle, a equivalent to the Krabachjoch Nappe, was thrust onto the Gosauian deposits. According to Ampferer (1930), the northern margin of the Larsennscholle was part of the Gosau basin. The Gosau transgresses onto the Middle to Upper Hauptdolomite.

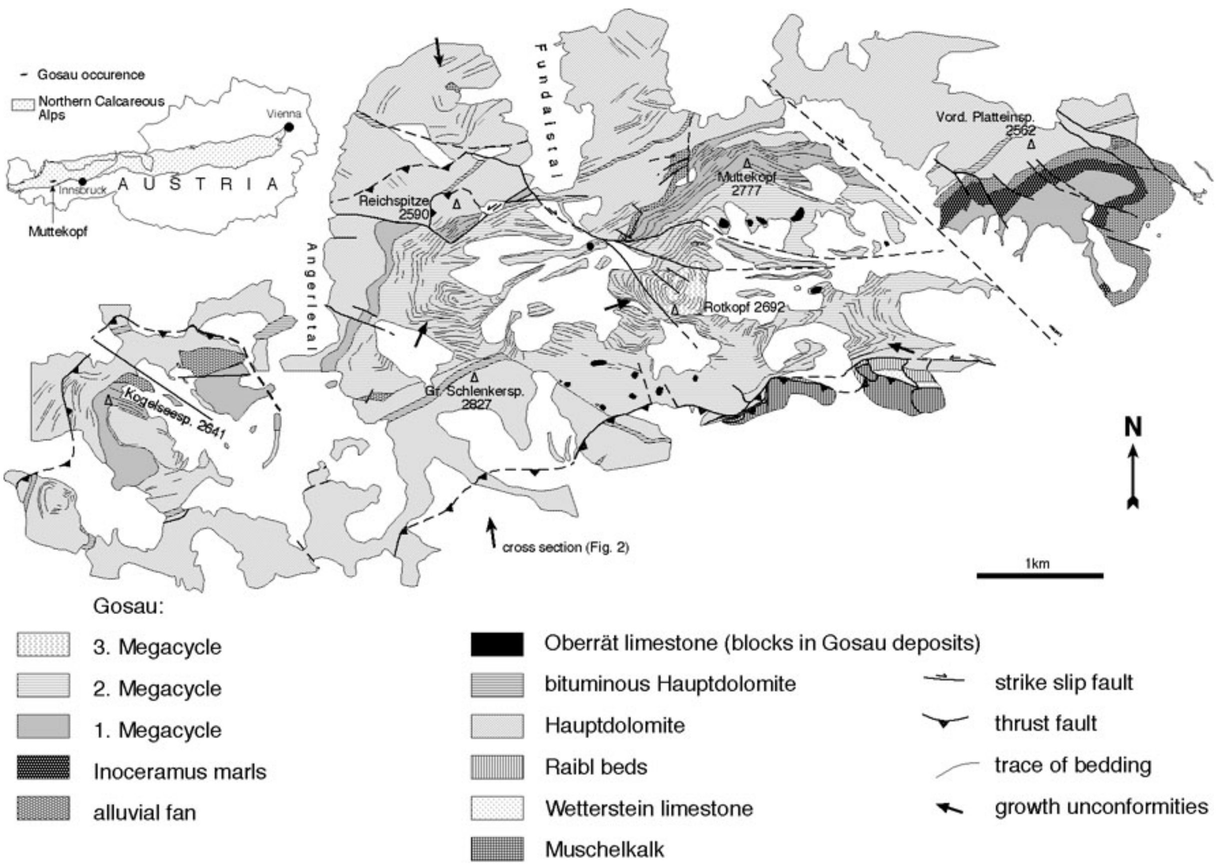


Fig. 1 Geological sketch of the Muttekopf area. Inset: position of the Muttekopf area in the Northern Calcareous Alps.

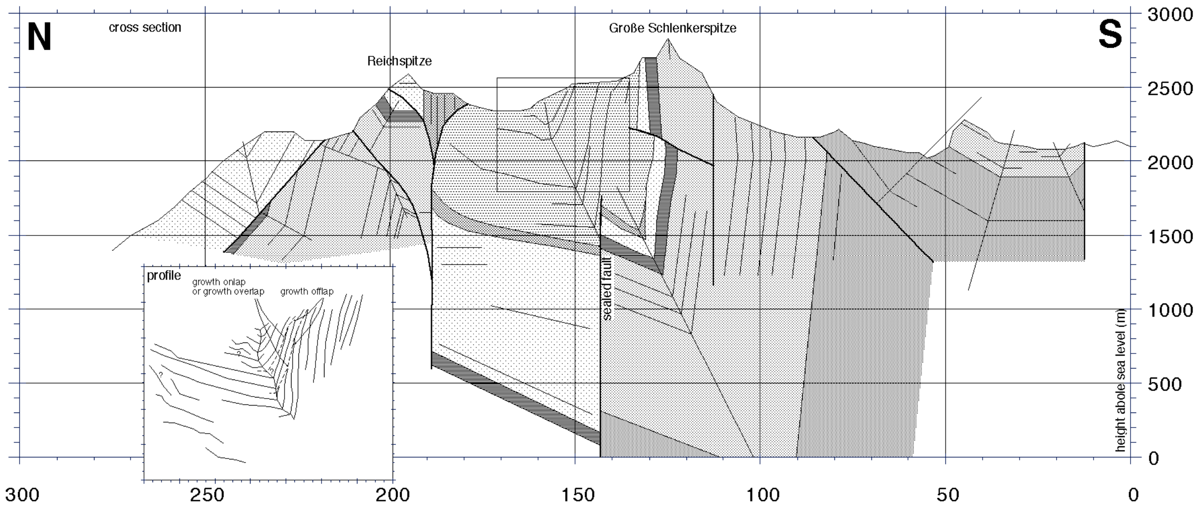


Figure 2: *Cross section across the investigated area. Inset: profile of the central part of the syncline showing a combined onlap-offlap pattern.*

Stratigraphy

Lower Gosau Subgroup :

Sedimentation in the Muttekopf area began before the Coniacian - Santonian boundary with deposition of thin braided river deposits followed by a thick alluvial fan (up to 300m thick) that is restricted to the easternmost part of the Gosau outcrops (Plattein). The alluvial fan is followed by a horizon of conglomerates with perfectly rounded clasts, which is interpreted to represent a transgressive lag at the boundary to the overlying Inoceramus marls (Haas 1991). Similar conglomerates are widespread at the base of the Gosau deposits and seem to be a remnant of a relative sea level rise. The silt- to sandstones of the Inoceramus marls contain a variety of marine fossils (including Inoceramus undulatoplicatus), that were used to date Coniacian - Santonian boundary (Ampferer 1912, Leiss 1988, 1990).

Upper Gosau Subgroup

The Inoceramus marls are overlain by marls, sandstones, breccias and conglomerates of the Upper Gosau Subgroup with a sharp contact, but no unconformity. These turbiditic deep water deposits are organized in three fining upward Megacycles, that reach a maximum thickness of about 300m. Each Megacycle is composed of following elements:

1 Megabreccias: A thick Megabreccia at the base of the cycle in the proximal part of the Megacycle

2 Thick Bedded Turbidite Association: A sediment stack, that consists of thick (up to 40cm) sandstone beds with the full or incomplete Bouma sequence in alternation with very thick turbiditic marls (up to 15m). In many cases the sandstone beds develop from conglomerate or breccia layers by grading.

3 Thin Bedded Turbidite Association: A sediment stack, which is formed by an alternation of very thin sandstone beds (up to 10cm, predominantly ca. 2cm) of the Bouma Tb or Tc interval with dark, sometimes laminated marls, that reach thicknesses up to 30cm. Thick breccia beds are intercalated with the alternating marls and sandstones, slumping is common.

The lithofacies associations grade into each other from proximal to distal. Megabreccias are restricted to the proximal parts of the Megacycles. Each of the megacycles has different heavy mineral and clast associations, and different sediment transport directions (Ortner 1990, 1992, 1994a,b; Fig. 15C).

The age of the deposits of the Upper Gosau Subgroup is poorly constrained. The turbiditic marls occasionally contain corroded nannoplancton and rare foraminifera. The onset of deep water sedimentation is dated by a microfauna found by Dietrich & Franz (1976) in the "lower part of the deep water deposits" as Upper Santonian. Two nannofossil specimens from the lower and upper part of the 1st Megacycle yielded an age of CC17 (Upper Santonian to Lowest Campanian) or younger. The base of the 2nd Megacycle has an age of Lower Campanian to Lower Maastrichtian (CC18) or younger, and the upper part of the 2nd Megacycle has an age of Upper Campanian to Lower Maastrichtian (CC21-23a; M. Wagreich pers. comm. 1993-1995). The 3rd Megacycle was dated by planctonic foraminifera to the Upper Maastrichtian to Danian (Oberhauser 1963) and by nannoplancton as Upper Maastrichtian (Lahodinsky 1987).

The 1st Megacycle

The 1st Megacycle unconformably overlies the Hauptdolomite in most areas. At the southern margin of the outcrops, the sediments onlap on to the Hauptdolomite. The first Megacycle has very coarse grained beds in the westernmost part of the Gosau outcrops (Plattigspitze) and becomes successively finer to the east (Fig. 18A). In the Muttekopf section the entire first megacycle consists of the Thin Bedded Turbidite Association. There, the fining upward trend is only visible in the thicknesses and clast sizes of the breccia beds. At the northern margin of the Gosau outcrops, a huge slab of Hauptdolomite (ca. 1 km long, 100m thick; Fig. 15B, Fig 16B) forms the base of the first megacycle.

The 2nd Megacycle

The 2nd Megacycle starts with a Megabreccia, that contains large blocks of Upper Rätian limestone (100m maximum diameter), Hauptdolomite and slabs of redeposited Gosau sediments of the 1st Megacycle. At the southern margin of the Gosau outcrops (Brunnkar), chaotic megabreccias directly overlie Hauptdolomit and reach a thickness of 100m. To the north, the same breccia horizon overlies the 1st Megacycle, and therefore seals a topography, possibly generated along a normal fault, that was filled by the 1st Megacycle (Fig. 18D). There, the megabreccia bed is ca. 20 m thick and normally graded. From the thickness change of the megabreccia bed, slump folds and ripple marks, a north dipping slope and sediment transport from the south to the north can be inferred. The main part of the 2nd Megacycle consists of theThick Bedded Turbidite Association. West of the Galtseitjoch, the megabreccia bed is replaced by an unconformity in the southern part of the basin (Schlenkerkar unconformity, Fig. 16B), that disappears to the north. The sediments of the 1st Megacycle were slightly steepened and eroded before sedimentation of the 2nd Megacycle.

The 3rd Megacycle

The third Megacycle crops out around the Rotkopf and deeply cuts into the sediments of the second Megacycle. This erosional unconformity at the base of the 3rd megacycle was called "Rotkopf" unconformity by Wopfner (1954). The amalgamated breccias and megabreccia beds onlap onto the 2nd Megacycle and thicken from ca. 10m in the east to 100m in the west. This indicates a sediment transport from east to west.

Interpretation of the Megabreccias and fining upward Megacyles

Megabreccias and fining upward cycles of the Upper Gosau Subgroup of the Muttekopf Gosau area are part of a large, approximately north dipping slope on the Austroalpine nappes into the Rhenodanubic Flysch basin. In the Muttekopf area, sedimentation was very coarse clastic, as several normal faults cut the slope. Synsedimentary movements caused large scale slumping and periodical collapse of parts of the slope. This process redeposited large slabs and boulders of the basement of the basin as well as Gosau sediments, predominantly by giant debris flows. During the tectonic events, topography was generated. The removal of the topography is reflected by the fining upward cycles of sedimentation. In analogy to similar deposits in rift basins of the Lower Austroalpine Ela Nappe in the Engadine Dolomites (Eberli 1987), extensional faulting at the southern margin of the Muttekopf Gosau basin is suspected. Small sealed normal faults are present in the Schlenkerkar (Fig. 16B). However, the geometry of the Schlenkerkar unconformity clearly demonstrates, that there was folding during Upper Cretaceous sedimentation. Measuring of the fold axis trends in the sealed Hauptdolomite anticline in the Parzinnkar (Fig. 16A) and below and above the Schlenkerkar unconformity (Fig. 16B), show that the direction of compression changed from top-to-NW to top-to-NNW. Top to NW movements were active before Gosau deposition until the end of sedimentation of the 1st Megacycle, top to NNW movements were active after Gosau deposition. The compression direction rotated about 20°.

Wagreich (1993a) discussed a model of tectonic erosion for subsidence of the Upper Gosau subgroup and Brandner & Ortner (1995) suggested a model of slab pull. Both models can account for synchronous thrusting and extension.

Hugo Ortner 1999:
Growing folds and sedimentation of the Gosau Group, Muttekopf, Northern Calcareous Alps

Synorogenic clastics were deposited on the growing orogenic wedge of the Northern Calcareous Alps from the Early Cretaceous to the Miocene. Sedimentation of the Gosau Group took place between the Turonian and the Late Eocene. Generally, the Gosau Group is subdivided into the Lower Gosau Subgroup, dominated by terrestrial to near-shore deposits, and the Upper Gosau Subgroup consisting of deep marine clastics. The Gosau deposits monitor the tectonic evolution in time during deposition.
The Gosau Group of the Muttekopf area is preserved in a large WSW-trending syncline (Muttekopf-Sinnesbrunn-syncline; Niederbacher 1982) in the Hauptdolomit of the Inntal Nappe, one of the large tectonic units of the western Northern Calcareous Alps. Three-dimensional analysis of the succession was facilitated by excellent outcrop conditions.
In the Muttekopf area the Lower Gosau Subgroup is only preserved in the easternmost part of the outcrops (Plattein area). In the western part, the thickness of these deposits is reduced and they were locally removed erosively before onset of

sedimentation of the Upper Gosau Subgroup. A NW-trending dextral fault separating the Plattein area from the more western outcrops is interpreted to have been active between the Coniacian und Santonian, leading also to vertical movement, that prevented erosion in the Plattein area. More to the west, another dextral NW trending fault east of Schlenkerspitze (Fig. 1) with a lateral offset of about 1km is sealed by sediments of the Upper Gosau subgroup. In areas between dextral faults, folds with a wavelength of about 100 m with NE-trending fold axes are truncated and sealed by sediments of the Upper Gosau subgroup.

The Upper Gosau Subgroup in the Muttekopf area consists of a succession of marls, sandstones and breccias that can be subdivided in three fining-upward megacycles (Ortner 1994). The first and the second megacycle were deposited from Late Santonian to Early Maastrichtian times. The first megacycle is restricted to the northern limb of the syncline and thins towards the south and west. Most of the sediments in the Muttekopf area are part of the second megacycle.

The profile across the syncline in the second megacycle shows several progressive unconformities (cf. Riba 1976) in the southern, steep limb of the fold, that become conformities in the northern flat limb of the syncline. Growth synclines are characterized by zones of growth offlap-onlap (Ford et al. 1997), as seen in the profile (Fig. 2, Detail). The wedging of the first megacycle towards the south could therefore be interpreted as growth onlap. Both the axis of the syncline in the Gosau deposits and in the anticline in the south in the underlying Hauptdolomite trends WNW.

Two conclusions can be drawn from these observations: 1) Structures of an older deformational event are sealed by Gosau sediments. During this event, dextral faulting along NW-trending faults was accompanied by minor folding with NE-trending axes.

2) The Upper Gosau Subgroup of the Muttekopf area was deposited during active shortening, the area of deposition has to be interpreted as piggy back basin. The asymmetry of the cross sections is not due to hidden normal faults (as suggested by Ortner 1994), but is the effect of deposition onto growing folds and is in accordance with widespread large scale slumping, deposition of megabreccias and with cannibalisation of large amounts of Gosau sediments.

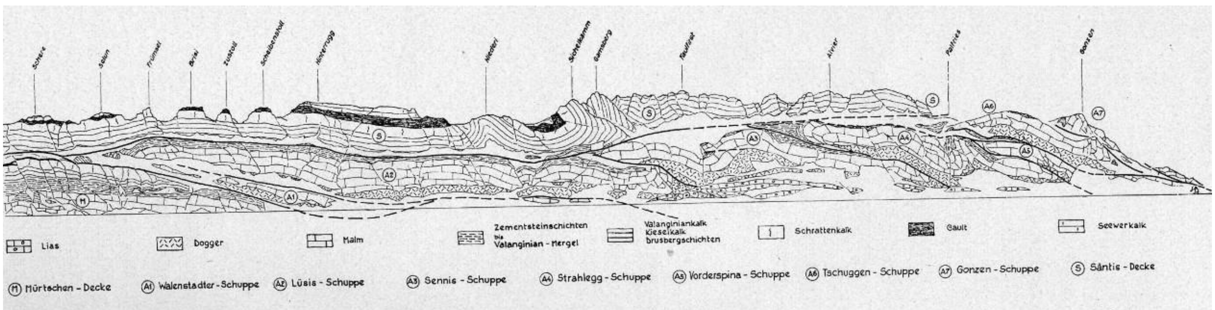
Aug. 6 Thursday:

Imst – Landeck – Arlberg Pass – Sargans – Tannenbodenalp – Bad Ragaz – Vättis – Chur - Thusis (270 km).
Tannenbodenalp: panoramic view over the Helvetic Mürtschen- Axen- and Säntis nappes of the Churfirsten
Vättis: tectonic window of the External Aarmassiv

(Guide: Petra Veselá + Bernd Lammerer)

Accommodation: Camping Viamala, Thusis Telefon +41-(0)81-651-2472 Fax +41-(0)81-651-2472 E-Mail: info@camping-thusis.ch WWW www.camping-thusis.ch Contact person: Pascale Zimmermann and Hugo Grieder

Stop 1: From Tannenbodenalpe panoramic overview over the Churfirsten and its nappe systems:



Figur 7. *Aufriss der östlichen Churfirsten und der Alviergruppe* (von S. Dollfus umgezeichnet nach R. Helbling, 1938)

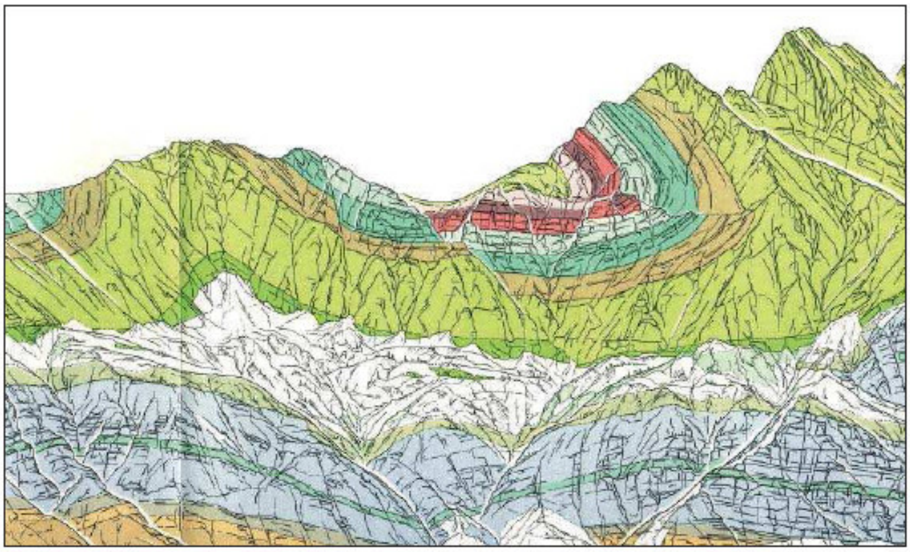
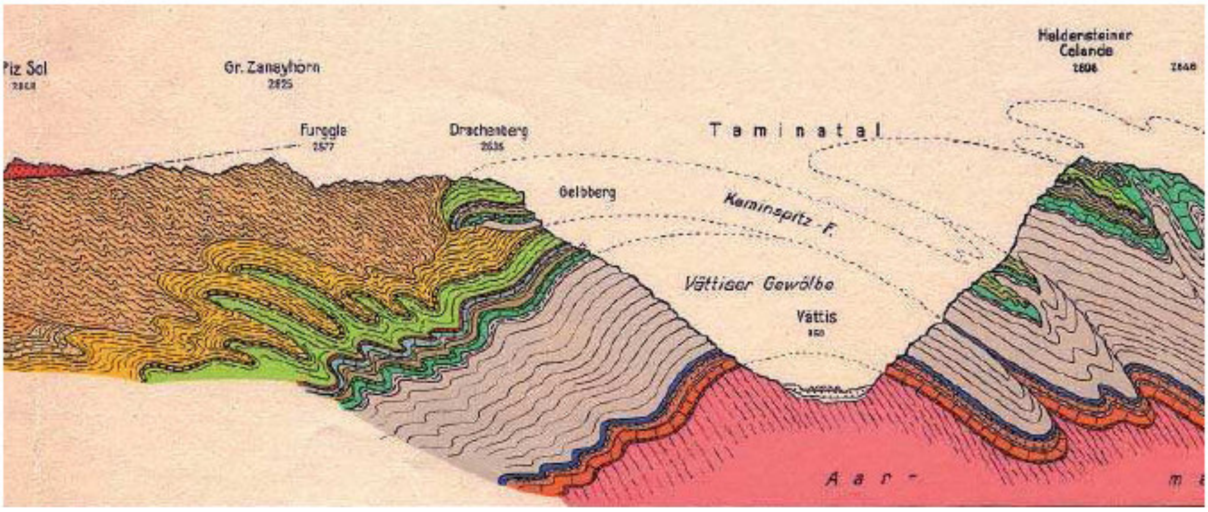


Abbildung 23: Geologische Interpretation der Sichelkammfalte: rostorange = Bommerstein-Formation mit Molser-Member, braun = Reischiben-Formation, blauviolett = Quinten-Formation, olivgrün = Zementstein-Formation, weisslichgrün = Palfries-Formation, grün = Betlis-Formation, senfgelb = Kieselkalk-Formation, orange = Drusberg-Formation, dunkel- und hellgrünblau = Unterer und Oberer Schrattenkalk, dunkel- und hellbordeauxrot = Garschella-Formation, hellgrün = Seewer-Formation. Nach HELBLING (1938).

Stop 2: Vättis window

In 2008, the mountainous region on the west side of Vättis, the Tectonic Arena Sardona, has been declared a world heritage site by the UNESCO. The unique stone formations in this area are readily recognizable for every observer. Not only do they build a spectaculair landscape, but their existence was material to the development of modern geological theories. Near Vättis is the "Vättner Fenster", a geological window, where the rock layers making up the Alps from the Aaremassif onwards are visible.



Vättis window after Oberholzer 1933

Aug. 7. Friday:

Thusis - Flims – Thusis (60 km)

Flims Landslide (largest landslide of the Alps)
Cassonsgrat (2700 m, by cable car): Glarus Thrust (geologic UNESCO world heritage site)

(Guide: Thorsten Nagel + Bernd Lammerer)

Accommodation: Camping Via Mala, Thusis (total 60 km)

Stop 1: Vorderrhein gorge and Flims landslide

Vorderrhein gorge: 15 billion cubic meters of rocks and mud slid 8300 y b.p. into the valley and dammed up the Rhine. Over the years though, the Rhine ate its way through the soft limestone, creating the canyon-like valley of the Vorderrhein Gorge. The flanks are well-forested and contrast the grey-silver of the stone well. In the valley, the Rhine meanders though the canyon, creating small peninsulas, a singular landscape not to be found anywhere else in Switzerland.

S. Ivy-Ochs, A.v. Poschinger, H.-A. Synal and M. Maisch 2008: Surface exposure dating of the Flims landslide, Graubünden, Switzerland .- **Geomorphology** Volume 103, Issue 1, 1 January 2009, Pages 104-112

Abstract: Sixteen boulder and bedrock surfaces related to the Flims landslide (volume 8–12 km^3) were dated with ^{36}Cl and ^{10}Be. Exposure ages range from 4900 ± 250 yr to 15,440 ± 1480 yr, including corrections made due to snow cover and karst erosion. Ages of 11,410 ± 590 yr and 13,340 ± 1090 yr were obtained on bedrock surfaces outside of the landslide zone. These are minima for deglaciation of Segnes valley and provide constraints on possible amounts of inherited nuclides. Based on seven boulder ages, which range from 8200 ± 260 to 9520 ± 990 yr, a mean of 8900 ± 700 years is calculated for the Flims landslide. We exclude three outliers (one significantly older and two significantly younger than the others) and the ages from the Cassons bedrock site from the mean calculation. The latter is excluded as shielding due to snow and vegetation cover is difficult to constrain there. The oldest boulder on the landslide yielded an age (15440 ± 1480 yr) more than 5000 years older than any other boulder age, suggesting that it included the surface of the pre-slide bedrock. The exposure ages are consistent with a single failure event. The erratics and patches of till lying on the landslide debris must have been carried piggy-back on top of the landslide rather than having been deposited by the late Pleistocene Vorderrhein Glacier. The Flims rockslide is about the same age as the Köfels (9800 yr) and Kandertal (9600 yr) rockslides. All three occurred during the marked transition to warmer, wetter conditions during the early Holocene.

Nicolas Pollet, Roger Cojean, Réjean Couture, Jean-Luc Schneider, Alexander L. Strom, Claire Voirin, and Patrick Wassmer : A slab-on-slab model for the Flims rockslide (Swiss Alps).- Can. Geotech. J. **42**(2): 587–600 (2005) | doi:10.1139/t04-122 | © 2005 NRC Canada

Abstract: The Flims rockslide is the largest landslide in the Alps, with an estimated volume of 12 km^3. It resulted from a prehistoric high-speed movement of a large limestone mass. Several main factors influenced the mobility of the Flims rockslide: (*i*) the steep slopes of the Rhine River valley that blocked the spreading of the rock debris out of the limits of Rabiusa and Carreratobel tributary valleys; (*ii*) the resisting forces taking place at the base of the rockslide by friction and substratum obstacles; and (*iii*) the rock mass evolving to a granular state, as observed in the deposits, in which coherence of the original rock massif has been preserved. We expect that most of the energy was consumed by impacting on the opposite slope, forcing the rock mass to stop. Lateral parts and some portions of debris, which entered valleys of the right tributaries of the Rhine River, created tongues by rock avalanche motion, indicating transport velocity. These rock masses eroded the valley fill to create a large mixed mass at the toe of the rockslide deposits. Thus, the Flims rock slope movement can be classified as a rockslide to its middle section and as rock avalanches at its lateral margins. A slab-on-slab model is proposed to characterize transformation of the rock mass during transport, with different stages of motion. Beginning as a rockslide, a delaminating process took place to produce a multislab shearing motion. Shearing and fracturing create dilatancy of the sliding rock debris, with spreading constrained by topographic effects. Dynamic disintegration processes explain the production of fine particles and are at the origin of the granular state of the deposits. Lateral sections of the debris mass continued to flow in the absence of topographic constraints.

Stop 2: Flims – Cable car to Cassonsgrat: Glarus Thrust

http://dic.academic.ru/dic.nsf/enwiki/10564527 The **Glarus thrust** (German: "Glarner Überschiebung") is a major thrust fault in the Alps of eastern Switzerland. Along the thrust the Helvetic nappes were thrusted more than 100 km to the north over the external Aarmassif and Infrahelvetic complex. The thrust forms the contact between older (Helvetic) Permo-Triassic rock layers of the Verrucano group and younger (external) Jurassic and Cretaceous limestones and Paleogene flysch and molasse. The Glarus thrust crops out over a relatively large area in the cantons Glarus, St. Gallen and Graubünden, due to its horizontal orientation and the high local relief. Famous outcrops include those at Lochsite near Glarus (the town) and in a mountain cliff called "Tschingelhörner" between Elm and Flims (in the same cliff is a natural hole called the Martinsloch).

World heritage: Thrust faults of this kind are not uncommon in many mountain chains around the world, but the Glarus thrust is a well accessible example and has as such played an important role in the development of geological knowledge on mountain building. For this region the area in which the thrust is found was declared a geotope, a geologic UNESCO world heritage site, under the name **Swiss Tectonic Arena Sardona**. The area of this "tectonic arena" encompasses 32.850 hectare of mainly mountainous landscape in 19 communities between the Surselva, Linthtal and Walensee. In the arena are a number of peaks higher than 3000 meters, such as Surenstock (its Romansh name is "Piz Sardona", from which the name comes), Ringelspitz and Pizol. In 2006 the Swiss government made a first proposal to declare the region world heritage to the International Union for Conservation of Nature (IUCN). The IUCN then did not find the area to have an extraordinary or universal value and denied the proposal. The Swiss made a new, this time successful proposal in March 2008. The region was declared world heritage in July 2008, because "the area displays an exceptional example of mountain building through continental collision and features excellent geological sections through tectonic thrust."

History: The first naturalist to examine the Glarus thrust was Hans Conrad Escher von der Linth (1767-1823). Escher von der Linth discovered that, contradictionary to Steno's law of superposition, older rocks are on top of younger ones in certain outcrops in Glarus. His son Arnold Escher von der Linth (1807-1872), the first professor in geology at the ETH at Zürich, mapped the structure in more detail and concluded that it could be a huge thrust. At the time, most geologists believed in the theory of geosynclines, which states that mountains are formed by vertical movements within the Earth's crust. Escher von der Linth had therefore difficulty with explaining the size of the thrust fault. In 1848 he invited the British geologist Roderick Murchison, an international authority, to come and look at the structure. Murchison was familiar with larger thrust faults in Scotland and agreed with Eschers interpretation. However, Escher himself felt insecure about his idea and when he published his observations in 1866 he instead interpreted the Glarus thrust as two large overturned narrow anticlines. This hypothesis was rather absurd, as he admitted himself in private.

Eschers successor as professor at Zürich, Albert Heim (1849-1937), initially stuck to his predecessors interpretation of two anticlines. However, some geologists favoured the idea of a thrust. One of them was Marcel Alexandre Bertrand (1847-1907), who interpreted the structure as a thrust in 1884, after reading Heims observations. [*aut|Bertrand, M.; **1884**: "Rapports de structure des Alpes de Glaris et du bassinhouiller du Nord", Société Géologique de France Bulletin, 3rd series, v. 12, pp. 318–330.*] Bertrand was familiar with the Faille du Midi, a large thrust fault in the Belgian Ardennes. Meanwhile, British geologists began to recognize the nature of thrust faults in the Scottish Highlands. In 1883, Archibald Geikie accepted that the Highlands are a thrust system. [*aut|Geikie, A.; **1883**: "On the Supposed Pre-Cambrian Rocks of St. David's", Quarterly J. of the Geol. Society **39(1-4)**, p. 261-333. http://jgslegacy.lyellcollection.org/cgi/content/abstract/39/1-4/261]*

The Swiss geologists Hans Schardt and Maurice Lugeon then discovered in 1893 that in western Switzerland, Jurassic rock layers are on top of younger molasse too, and argued that the structure of the Alps is a large stack of nappes, large sheets of rock that had been thrusted on top of each other. [*aut|Schardt, H.; **1893**: "Sur l'origine des Préalpes romandes", Eclogae geologicae Helvetiae **4**, pp. 129–142.*] At the turn of the century, Heim was also convinced of the new theory. He and other Swiss geologists now started mapping the nappes of Switzerland in more detail. From that moment on, geologists began recognizing large thrusts in many mountain chains around the world. However, it was still not understood where the huge forces that moved the nappes came from. Only with the arrival of plate tectonic theory in the 1950's an explanation was found. In plate tectonics, the horizontal movement of tectonic plates over the Earth's soft asthenosphere causes horizontal forces within the crust. Presently, geologists believe most mountain chains are formed by convergent movements between tectonic plates.

Fig. … Glarus Thrust below Tschingelhörner and Ofen. At left end: Piz Grisch (2902 m). At right end: Martinsloch and Pass dil Segnas (Segnespass, 2627 m). The thrust itself is marked by a thin (20-50 cm) very light grey layer of Lochseiten limestone. Above the thrust: "Helvetic" Permian "Verrucano". Below the thrust: (i) "Subhelvetic" slivers of light grey Cretaceous limestones and slightly yellowish/brownish Eocene greensands and (ii) "Infrahelvetic" sardona Flysch unit (almost black Eocene shales, sandstones and quartzites, grey cretaceous Globotruncanen limestone and -marls. At far left below the thrust also black Tertiary shales from Blattengrat Flysch Unit. http://www.denbrok.ch/geology/eth/old/glarus/tschingel-123.html

Aug 8 Saturday:

Thusis - Passo San Bernardino (2065 m) – Mesocco – Roveredo – Bellinzona (120 km)

Via Mala gorge Bündnerschiefer of the Valais ocean and Erosion of the Hinterrhein
San Bernardino Pass : Adula nappe, Misox zone, Simano nappe

(Guide: Thorsten Nagel)

Accommodation: Camping TCS "Bosco di Molinazzo"; Roman Leonardi; Via San Gottardo 131; CH-6500 Bellinzona; Phone: ++41-91-829 11 18; Fax: ++41-91- 829 23 55; e-mail: camping.bellinzona@tcs.ch

Thusis Via Mala gorge
The Viamala, a gorge cut deep into Graubunden slate by the Hinterrhein river, is not only an impressive natural monument but also an outstanding testimony to bold road building by our forefathers. The history of traffic through the Viamala Gorge is marked by bold building projects. The Romans were the first to tackle such difficult topography. Two galleries cut in the rocks on the west bank show the remains of their pack-animal trails, still visible from the kiosk on the other side of the gorge. The name Viamala – bad route – dates back to the Middle Ages when the Roman path fell slowly but surely into ruin. It was not until the 15th century that the route was revived: in the «Viamala Letter» dated 1473, the municipalities of Thusis, Masein and Cazis committed themselves «…to hew, repair and set about the Reichsstrasse and the path between Thusis and Schams, the so-called «Fyamala». A well organized trade traffic then grew up, which helped the people along the transit routes of Splügen and San Bernardino become very prosperous.

Fig.: Scetch of the Via Mala gorge from J.W. Goethe

Aug. 9 Sunday:

Bellinzona – Valbella – Bellinzona (70 km)

Ascent to Passo Trescolmen (2153 m): orthogneisses, eclogites, garnet-schists of the Adula nappe. Whole day walk in high Alpine area.

(Guide: Thorsten Nagel)

Accommodation: Camping TCS "Bosco di Molinazzo"; Bellinzona;

Thorsten Nagel *(2008):* Tertiary subduction, collision and exhumation recorded in the Adula nappe, central Alps. Geological Society, London, Special Publications*; 2008; v. 298; p. 365-392;*

The Adula nappe in the Central Alps represents a lithospheric mélange assembled in a south-dipping subduction zone during the Tertiary orogenic cycle. It consists of several heterogeneous lobes which are stacked in a forward-dipping duplex geometry. Eclogites, garnet peridotites and garnet-white-mica schists record southward-increasing peak pressure conditions which culminate at 12–17 kbar/500–600 °C in the north and 30 kbar/800–850 °C in the south. Some studies infer even higher peak pressures for the garnet peridotite body of Alpe Arami. The present-day metamorphic field gradient for peak pressures exceeds the lithostatic pressure gradient. So far, only eclogites and garnet peridotites from the Cima Lunga complex in the south and the adjacent Southern Steep Belt have yielded Tertiary metamorphic ages for the peak-pressure stage. Some recent studies propose that the Adula nappe got assembled after the formation of high-pressure assemblages in eclogites and garnet peridotites and reject regional high-pressure conditions in Tertiary times. This scenario, however, is in conflict with the observed continuity of metamorphic field gradients and post-peak-pressure structures. Amphibolite facies conditions post-date formation of the Central Alpine nappe stack. In this paper, the associated field gradient is explained through southward-increasing temperatures during near-isothermal decompression.

The main mylonitic foliation in the Adula nappe post-dates peak-pressure conditions. It is associated with top-to-the-north shearing and southward-increasing amounts of decompression from eclogite facies to amphibolite facies conditions. Also, the present-day supra-lithostatic field gradient for peak pressures probably results from this deformation phase and is here related to substantial vertical flattening during northward shearing. All subsequent structures affect established nappe boundaries. Pervasive Oligocene deformation events in the Adula nappe are coeval with intense shearing along the so-called Insubric mylonites and occur during ongoing isothermal decompression to around 5 kbar. They are associated with orogen-oblique to orogen-parallel stretching of unspecified amount which may considerably contribute to the exhumation of the Lepontine dome already before the onset of the well-known Miocene extension.

Aug. 10 Monday:

Bellinzona _ Lugano – Mendrisio – Morbio Superiore arzo - Bellinzona (140 km)

Breggia gorge: Jurassic and Cretaceous sediments and synsedimentary tectonics
Arzo quarries: Early Jurassic extensional tectonics)
Meride Museum of Monte San Giorgio fossils (UNESCO world heritage site)

(Guide: Bernd Lammerer)

Accommodation: Camping TCS "Bosco di Molinazzo"; Bellinzona

Stop 1: Arzo

Southern Ticino, Switzerland : Geological archive of the evolution of the Mesozoic alpine Tethys Ocean.-
IAS Newsletter 194 October 2004: Super Sedimentological Exposures: Arzo & Breggia
Compiled by Helmut Weissert (Dept. of Earth Sciences, ETH-Z), http://www.iasnet.org/news/newsletter/194.pdf

Introduction
The localities Arzo and Breggia can be regarded as “classical” geological archives of the tectonic and oceanographic evolution of the alpine Tethys Ocean. In addition, Arzo is located at the foot of the 1,097 metres high Monte San Giorgio built up of a Middle to Late Triassic shallow-water carbonate succession. The Middle Triassic Grenzbitumenzone, outcropping along the Mte San Giorgio has become famous for its rich and unique fish and reptile fauna. Today Monte San Giorgio is a UNESCO World Heritage Site.

A short geological history:
About 280 million years ago, the supercontinent Pangea was cut by numerous transtensional graben structures, which can be recognized from the Ural Mountains through Europe into the Appalachians. These graben structures were filled with continental deposits during the Permian and the early Triassic. Clastic sediments of Permo-Triassic age form the oldest sedimentary rocks of the Monte San Giorgio mountain. The early extensional tectonics was accompanied by volcanic activity. In the Southern Ticino region, volcanic (andesites, rhyolites) and volcanoclastic rocks are outcropping near the city of Lugano. The Late Permian and Early Triassic red clastic sediments are overlain by middle and late Triassic dolomites, limestones and marlstones which were deposited in a shallow marginal sea of the opening alpine Tethys Ocean. The Monte San Salvatore along the Lago di Lugano is built up of a middle Triassic dolomite sequence which was formed in a current influenced shallow platform environment. The middle Triassic sediments outcropping today on the nearby Mte San Giorgo were deposited in an isolated shallow marine basin. Due to restricted water circulation, laminated claystones, marlstones and dolomites were sedimented under low oxygen and anoxic conditions (Fig. 1).

These sediments, known as “Grenzbitumenzone”, contain up to 40% organic carbon. This horizon contains a rich fauna of vertebrates with abundant reptiles and fish. A thick succession of dolomites and limestones was formed under Sabkha-type conditions during the Late Triassic (Dolomia Principale, Calcare di Zu). Starting with the Early Jurassic the alpine Tethys Ocean was affected by accelerated rifting. This resulted in the development of several listric normal faults. One of these faults, the Lugano fault, separated a rapidly subsiding basin Generoso Basin) to the east of the fault from a submarine high forming the footwall of this fault (Lugano high).

Today, the signature of this rifting tectonic activity along the Lugano fault is contained in Triassic-Liassic sedimentary rocks of the submarine high (Locality Arzo) and in the basin infill sediments of the evolving Generoso Basin (Locality Gole della Breggia).
Tectono-sedimentary breccias were formed along the margin of the evolving submarine high. These breccias are overlain by a condensed succession of Early Jurassic red marine limestones, recording a progressive deepening of the submarine high. The evolving Generoso Basin east of the submarine high, was filled by up to several thousands of meters of turbiditic siliceous limestones. A distinct change in basin infill history occurred during the late Liassic. The turbidite succession was replaced by pelagic sediments of middle Jurassic to middle Cretaceous age. Today this pelagic succession is preserved in the Breggia section. Red limestones rich in ammonites (Rosso ammonitico) formed during the Toarcian and are overlain by pelagic lamellibranch limestones, radiolarian cherts and limestones of middle to late Jurassic age and by white nannofossil limestones which are of early Cretaceous age. These pelagic sediments provide a unique and continuous record Jurassic-Cretaceous alpine Tethyan oceanography. The facies of the middle-Jurassic-Late Cretaceous pelagic sediments was controlled by the basin topography, by water depth (calcite compensation depth) and by physical and chemical oceanography.

The most prominent Early Cretaceous black shale episode known as Oceanic Anoxic event 1a (Livello Selli) is missing at the Bregga locality. Other black shales of Barremian and Albian-Cenomanian age document the peculiar conditions in Cretaceous oceans resulting in a widespread and often global deposition of sediments enriched in organic carbon.

The Late Cretaceous black shales are intercalated with red, green and white limestones, marlstones and claystones of the Scaglia Variegata, Scaglia Bianca and Scaglia Rossa. These sediments deposited during middle and Late Cretaceous reflect changing tectonic and oceanographic conditions in a slowly closing alpine Tethys Ocean. A succession of turbiditic sandstone was formed during the Late Cretaceous and it is interpreted as the first flysch succession in the western Southern Alps.

Locality 1: Arzo quarry

Theme: Fractured Late Triassic carbonate platform with polyphase breccias, so called “Neptunian Dykes” -
Explanation: Tectonic activity along the opening alpine Tethys Ocean was intensified in the Late Triassic and Early Jurassic. A record of the tectonic activity along a newly developing continental margin is preserved in several small quarries near the village Arzo. There, shallow-water carbonates (limestones and dolomites of the Norian Hauptdolomite Formation and the Rhetian successions) are overlain by red bioclastic limestones with crinoids and brachiopods (Broccatelli Formation) and by red ammonite- bearing limestones (Besazio Limestone. Early Jurassic). In places, the Triassic limestones and dolomites are cut by synsedimentary faults and graben structures of up to several meters in width anwhich are classified as Neptunian Dykes. The graben structures and fractures were filled with synsedimentary polyphase breccias.

Polyphase fragmentation of the Triassic platform carbonates resulted in breccias with multiple generations of clast and matrix formation. Theses breccias are known as Macchia Vecchia. On a larger, regional scale, the fragmentation of the evolving Mesozoic southern continental margin of the alpine Tethys resulted in listric fault structures separating submarine highs from continental margin basins

Locality 2: Parco delle Gole di Breggia

Theme: A Mesozoic paleoceanography of the Tethys Ocean.

The path through the Park begins by the side of a small lake and continues along the river to the old cement factory and along the river to Morbio Superiore.

Explanation: The section in the Gole della Breggia or Breggia Gorge is best studied starting at the upper end (locality Morbio superiore). A well-prepared path in the park allows one to follow the Liassic-Cretaceous section along the Breggia gorge walking upward through time.

Up to several thousands of meters of turbiditic limestones, which were deposited in the rapidly subsiding early Jurassic Generoso Basin, are today forming the Monte Generoso to the northeast of the gorge. Along the excursion path only the uppermost meters of this sections can be studied. Grey, siliceous limestones with rare evidence of turbidity current deposition are alternating with silty marls (*Moltrasio Formation*).

A dramatic change in the depositional environment occurred during the Late Liassic, when turbidite sedimentation was stopped and red pelagic carbonates of the *Rosso Ammonitico Formation* (Toarcian)
were accumulated. The Ammonitico Rosso limestones do not outcrop along the path. They can only be visited at the river. There, the Rosso Ammonitico succession shows a prominent cyclicity pattern. These cycles record oceanographic changes controlled by changes in orbital parameters (20ky cycles according to Weedon, 1989).
Rosso Ammonitico limestones are overlain by a succession of red *Limestones with pelagic bivalves* (Bathonian-Bajocian). One large, up to 10m thick slump (Foto 2) and several small slump deposits within these limestones document how pelagic sedimentation in the Generoso Basin was also controlled by local basin morphology. The slump deposits can be seen from the path crossing the valley high above the river. The red pelagic limestones are overlain by green and red *radiolarian cherts* of middle to Late Jurassic age. The radiolarites form the steepest part of the gorge. Radiolarites were deposited below the middle to Late Jurassic calcite compensation depth (CCD). A peculiar oceanography with highly productive surface waters favouring the bloom of radiolarians and good ventilation of deep water favoured the formation of the red radiolarian cherts. The radiolarian cherts are overlain by radiolarian limestones of Oxfordian-Tithonian age. The transition to radiolarian limestones (Oxfordian) marks a progressive deepening of the Late Jurassic CCD and a change in Tethyan Oceanography.
Most prominent is the following change in pelagic sedimentation at the transition from the Jurassic to the Cretaceous outcropping in the old quarry at the base of the gorge were deposited during the Berriasian to Barremian (*Maiolica Formation*). Several slump deposits are intercalated with the wellbedded limestones. In the uppermost part of the up to 140 m thick white limestone formation we can recognize black shale
intercalations. These black shales were formed under dysoxic or anoxic conditions during the Barremian. The top of the Maiolica Formation is marked by a hardground containing glauconite. The hardground was formed during Late Barremian and Ealy Aptian. Due to this stratigraphic hiatus spanning the Late Barremian and early Aptian, the black shale deposits formed during OAE1a (Livello Selli) are not found in the Breggia section. From Late Aptian to Cenomanian hemipelagic marls and limestones of the Scaglia group were deposited. A significant increase in clay content coincides with beginning of convergence tectonics in this part of the alpine Tethys Ocean. Flysch sedimentation started during the Late Cenomanian. These flysch deposits can be studied at the very end of the Breggia gorge section.

Literature
Arthur, M.A., Premoli Silva, I., 1982. Development of widespread organic-carbonrich strata in the Mediterranean Tethys. In *Nature and origin of Cretaceous carbonrich facies* (eds. S. Schlanger, M.B. Cita), p.7-54.

Bernoulli, D. 1964. Zur Geologie des Monte Generoso. Ein Beitrag zur Kenntnis der südalpinen Sedimente. Beitr. Geol Karte, Schweiz. 134pp.

Bernoulli, D., Caron, C., Homewood, P., Kälin, O., Van Stujivenberg, J., 1979. Evolution of continental margins in the Alps. Schweiz. Min. Petr. Mitt., 59, 165-170.

Hsü. K.J., 1976. Paleoceanography of the Mesozoic Alpine Tethys. GSA, Spec. Paper, 170, 44p.

Weissert, H., 1981. Depositional processes in an ancient pelagic environment: the lower Cretaceous Maiolica of the Southern Alps. Eclogae geol. Helv., 74, 339-352.

Wiedenmayer, F. 1963. Obere Trias bis mittlerer Lias zwischen Saltrio und Tremona (lombardische Alpen). Die Wechselbeziehungen zwischen Stratigraphie, Sedimentologie und syngenetischer Tektonik. Eclogae Geol. Helv. 56, 529-640.

Rudolf Stockar,2003. Parco delle Gole della Breggia, Guida Geologica, Museo cantonale di storia naturale, CH-6900 Lugano.

IAS Meeting of Sedimentology, 2001, Davos, Switzerland, Excursion guides (a few copies are still available from the author of this article) General Information

Tourist Web site: http://www.mendrisiotourism.ch
Web address UNESCO Site: h t t p : / / w h c . u n e s c o . o r g / pg.cfm?cid=31&id_site=1090

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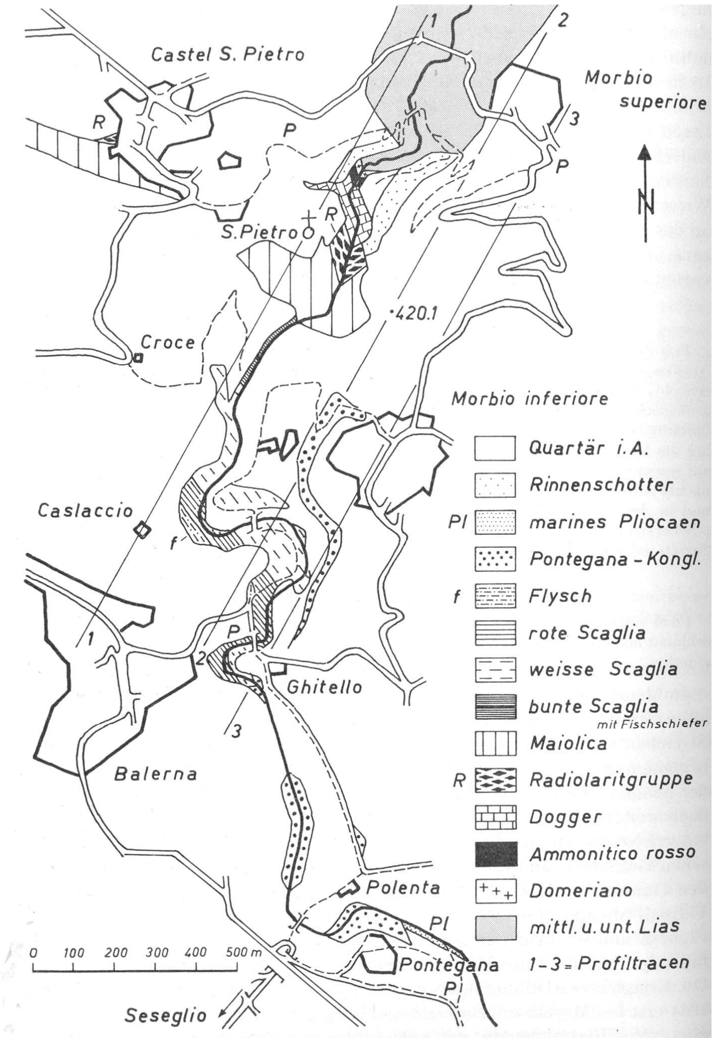
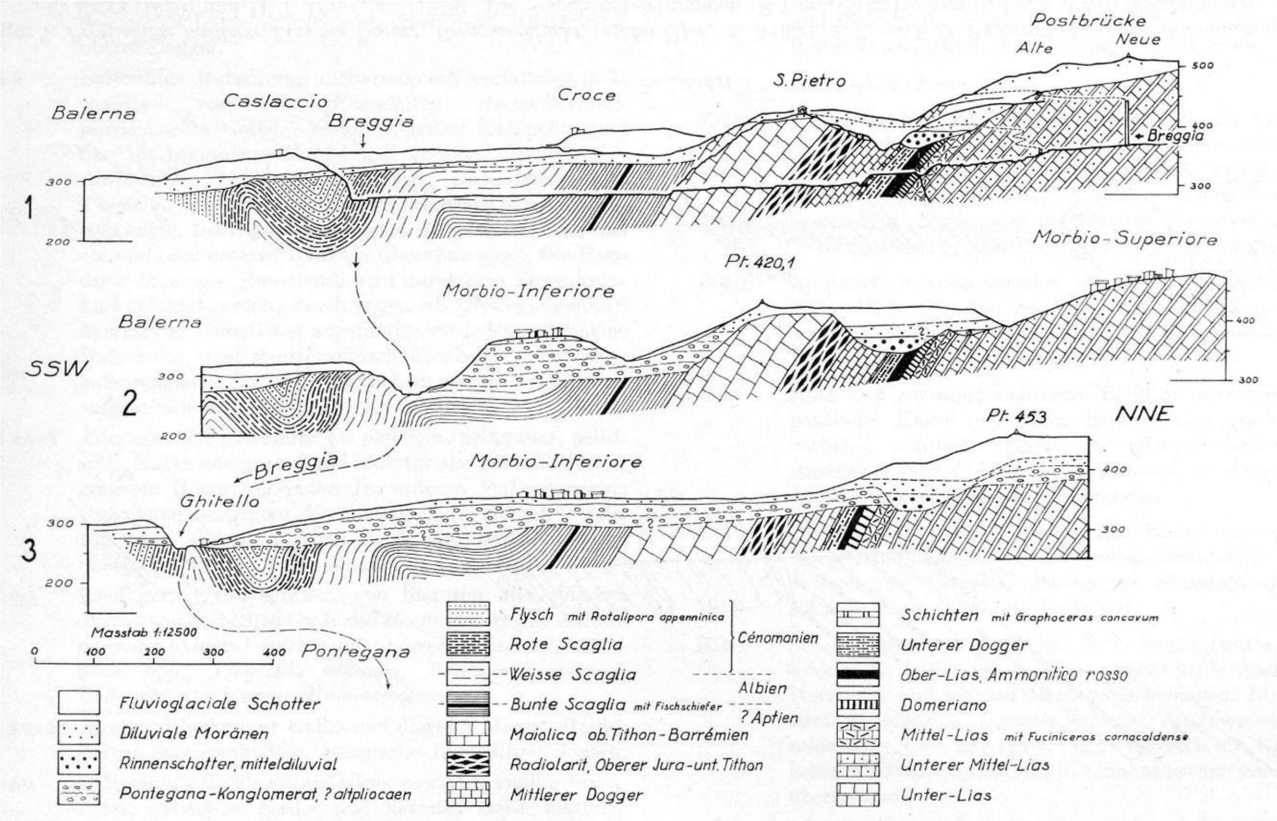
Stop 2: Breggia gorge:

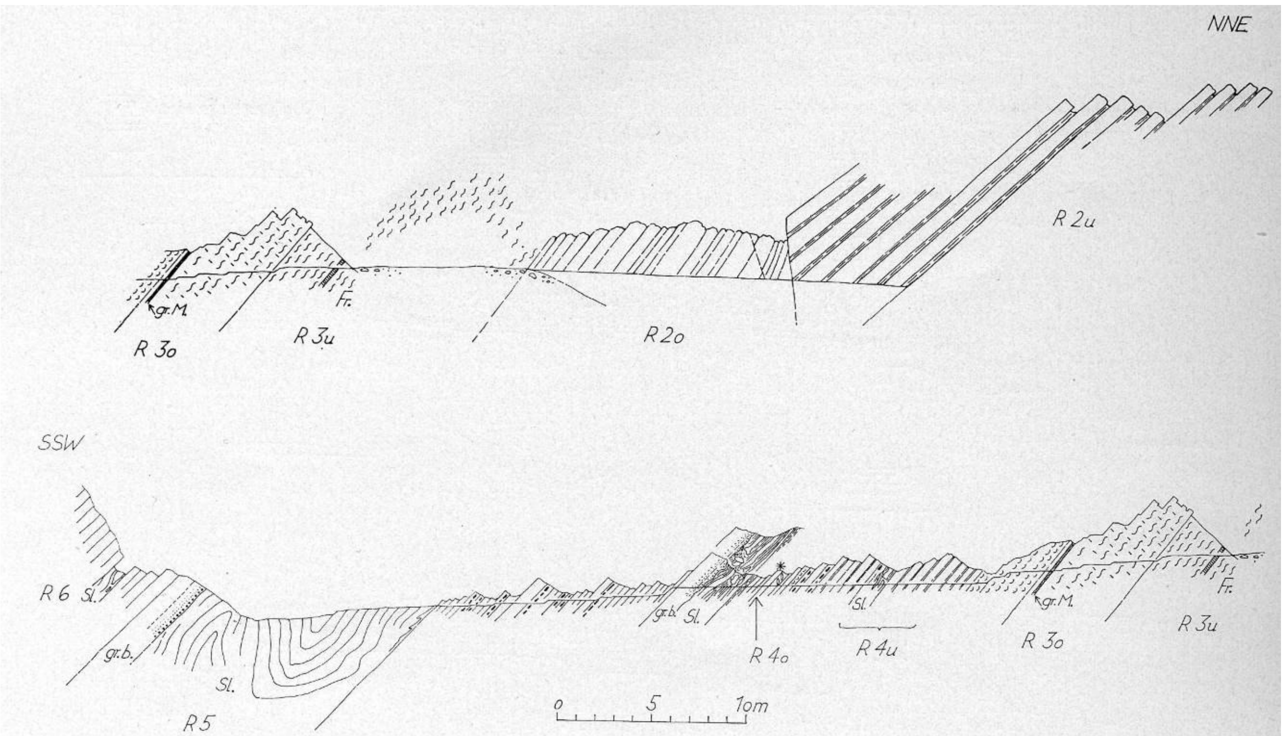
Numerous components, both natural and human, are present in the Park, but the particular geological contents of the Breggia Gorge make it one of the most important geo-topes in Switzerland and the first Swiss geo-park. Along this part of the river the natural section created by the excavation of the water has brought to light a geology profile that is almost continuous between the Jurassic and the Tertiary periods (more than 80 million years).
Numerous pieces of evidence of ancient seas are present in the rocks of the Gorge: layers extremely rich in fossils, remains of underwater landslides, witness to climatic changes in eras long before man appeared on Earth, and proof of volcanic eruptions.



Figur 5. *Geologische Kartenskizze des Gebietes der Breggiaschlucht* nach Aufnahmen von L. Vonderschmitt



Figur 6. *Geologische Profile 1:12500 durch das Gebiet der Breggiaschlucht* nach L. VONDERSCHMITT (1940)



Figur 7. Profilskizze mittlerer Lias bis Dogger, Breggiaschlucht, rechtes Ufer, ca. l :350, z.T. nach D. BERNOULLI (1964). Ammoniten nach C. BENZ (1920) und D. T. DONOVAN (1958). Die Schichtbezeichnungen (R l etc.) wurden von C. Renz (1920) übernommen.

R 6 Gut gebankte, graue, rote und grünliche, kieselige und mergelige Kalkpelite mit Radiolarien und Schälchen pelagischer Lamellibranchiaten. Hornstein. Mittlerer bis oberer Dogger.

R 5 Submarine Rutschung: disharmonisch verfalteter, z.T. chaotisch verformter (Fliessfalten, «Lamellierung», phacoidisches Gefüge) Komplex heller Kalkpelite und lila- bis braunroter Mergel und knolliger Mergelkalke. Ammoniten des oberen Toarcien (Dumortieria spp., Pleydellia spp.), des Aalenien {Tmetoceras scissum, Erycites fallax, Ludwigia murchisonae, Oraphoceras concavum etc.) und des unteren Bajocien (Sonninia spp.). Die Basis der submarinen Rutschung wird durch eine Bewegungsfuge gebildet, welche durch gegen SE (Bewegungssinn!) überkippte Fliessfalten augenfällig wird. Die submarine Rutschung wird stratigraphisch überlagert durch eine kalkarenitische Bank mit deutlichem graded bedding und aufgearbeiteten Pelitgeröllen an der Basis.

R 4 o-5 Alternanz gut gebankter bis plattiger, hellgrauer, pelitischer Kalke und grauer und lilaroter Mergel. z.T. dünne, graurote Homsteinbänder. Im unteren Teil submarine Rutschung hellgrauer Mergelkalke und Mergel, welche von einer kalkarenitischen Bank mit graded bedding überlagert wird.

B4o Knolliger, gelblichgrauer, von lilaroten Mergelflasern durchzogener, pelitischer Kalk, 25 cm. Zahlreiche Ammoniten des oberen Toarcien und unteren Aalenien: Dumortieria spp., Pleydetlia aalensis, Tme.tow.ras scissum, Ludwigia murchisonae, Hammatoceras spp.

R4u-4o Alternanz hellgrauer Kalke und lilaroter Mergel. Radiolarien, Spongiennadeln, pelagische Lamellibranchiaten.

R 4u Hellgraue, z.T. lilarot gefleckte, zuweilen knollig texturierte, politische Kalke und lilarote, meist blättrige Mergel. Zuoberst kleinere submarine Rutschung. Ammoniten des oberen Toarcien: Dumortieria spp.

R 4u-3o Hellgraue politische Kalke und lilarote Mergel. Rotbraun und weiss gefleckte, flaserig-knollig texturierte Mergel und Mergelkalke. Posidonien, Pleydellia aalensis. (Grenzposidonienschichten von G. RENZ), ob. Toarcien.

R 3 Ammonitico ROSSO.

R3o Gut gebankte, rote und rot-weiss gefleckte, knollige Mergelkalke mit Ammoniten des mittleren Toarcien: Paroniceras spp. Phymatoceras erbaense, Erycites spp., Hammatoceras spp. - gr.M. Grünliches Mergelband mit exotischem Block von permischem Quarzporphyr (L. VONDERSCHMITT, 1940; heute nicht mehr sichtbar).

R 3 u Rote und rot-weiss gefleckte, knollige Mergel und Mergelkalke mit Ammoniten des unteren Toarcien: Mercaticeras mercati, Hildocems spp., Phymatoceras tiroknse, etc. Fr. = Horizont mit Frechiella spp.

R 2o Rote und rot-weiss gefleckte, knollige oder brekziöse, pelitische Kalke und Mergelkalke. Graue, gelbe und rötliche, pelitische Kalke mit Mergelzwischenlagen. Amaltheus stokesi, Pleuroceras spinatum, Arieticeras spp., Meneghiniceras lariense etc. Domerien.

R 2 u Hellgraue, gelblich verwitternde Kalke mit siltigen Mergelzwischenlagen. Limonitisierte Ammoniten: Protogrammoceras bassanii, Fuciniceras cornacaldense. Ob. Pliensbachien s. str. (ob. Carixien)

R l (Vom Beginn des Profils ca. 20 m bachaufwärts aufgeschlossen): hellgraue, gelblich verwitternde Kalke mit Hornstein und siltigen Mergelzwischenlagen. Limonitisierte Ammoniten; üptonia jamesoni, AcantJiopleuroceras valdani etc. Unt. und mittl. Pliensbachien s. str. Mächtigkeit um 100 m, gegen unten in Lombardischen Kieselkalk übergehend.

Sl = Slumping (submarine Rutschung); gr.B. = graded bed.

Aug. 11 Tuesday:

Bellinzona – Val Verzasca - Locarno – Centovalli – Sta. Maria Maggiore – Finero - Domodossola – Val Aosta (240 km)

Val Verzasca migmatites
Ivrea Zone near Finero
Via romana and Gallery between Donnas and Bard,
If enough time: ascent from St. Marcel to Seissogne – Les Druges, (view to Mt. Blanc – Mt. Rosa – Matterhorn) and ascent to the old copper mine of Servette and the manganese mine of Praborna
(Guide: Bernd Lammerer)
Accommodation: Camping Aosta International Touring; Fraz. Arensod, 10 - 11010 Sarre (Aosta) - Italy
+39 0165 257061 campingtouring@libero.it

Stop 1: Val Verzasca:
Please refer to the link: http://titan.minpet.unibas.ch/aliens/smpm/SB85_25.pdf

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Tertiary migmatites in the Central Alps: Regional distribution, field relations, conditions of formation, and tectonic implications

Thomas Burri, Alfons Berger, and *Martin Engi*

Alpine migmatites are almost entirely confined to the Southern Steep Belt (SSB), the regional-scale, transpressional shear-zone at the southern margin of the Central Alps. Migmatites surfacing in more northerly parts of the Lepontine area are derived from pre-Alpine, probably Variscan or older periods of partial melting, connected to the intrusion of bodies of granitic to quartz-dioritic composition, as well as mafic and granitic dykes. Except for the Bergell and Novate intrusives, Tertiary igneous activity in the Central Alps is limited to *in-situ* migmatisation and to the intrusion of aplitic and pegmatitic dykes and smaller (<50 m) granitoid bodies. Two processes contributed to the origin of this migmatite belt: (1) In the course of regional Barrovian metamorphism, water-assisted partial melting of granitoid rocks was induced in a large part of the southern Lepontine area, commonly leading to a maximum of 10–25 vol% total leucosome. (2) In pelitic rocks of the southeastern Lepontine area, a smaller leucosome- fraction is found, which is essentially a result of muscovite dehydration melting. Pressure-temperature conditions of partial melting estimated for amphibole-bearing leucosomes are 0.6–0.8 GPa and 700 } 50 °C, indicating mid-crustal partial melting. Thermally retentive chronometers set fairly tight limits for this event at 25–30 Ma. The spatial relationship between migmatites and SSB, as well as the styles of variable deformation in the leucosomes, indicate that partial melting and deformation were coupled processes. Observations suggest that the focused deformation in the SSB led to episodic injection of hydrous fluids, which in turn triggered water-assisted partial melting and associated strain partitioning into the "weak" partially molten rocks. Processes of partial melting in the migmatite belt appear to be continuous in time and space with the regional thermal history that produced upper amphibolite facies metamorphism without partial melting in adjacent areas to the north, i.e. outside the SSB.

Geology of the Southern Steep Belt (SSB) and adjacent areas

The Swiss–Italian Central Alps form an Alpine Barrovian-type metamorphic belt (Niggli and Niggli, 1965; Niggli, 1970; Wenk, 1975; Frey and Ferreiro-Mählmann, 1999; Engi et al., 2004). Temperature generally increases from north to south (Todd and Engi, 1997; Frey et al., 1999) reflecting the south-vergent subduction of the European plate below the Apulian plate and the subsequent uplift of hot subducted lithosphere during continent-continent collision. In detail, P–T relations are complicated due to the relaxation of the isotherms that accompanies deformation and exhumation of the nappe stack (Engi et al., 2001; Burg and Gerya, 2005). In the internal part of the Central Alps, a moderately dipping nappe stack is exposed, which forms the metamorphic core of that part of the mountain belt, the so-called Lepontine Dome (Fig. 1). The Lepontine nappesconsist of basement units, dominated by Variscan granitoids that intruded the metamorphic basement. Several of these nappes also contain migmatites, which appear to be genetically related to these intrusives. The most important units in the study area are classic basement thrust sheets (the Monte Rosa, Maggia, Antigorio, Simano and Leventina nappes) or very heterogeneous tectonic mélange units (e.g. Adula, Cima Lunga and Orselina units) containing abundant eclogite relics. These latter units include what Trommsdorff (1990) termed an *Alpine lithospheric mélange* and have collectively been interpreted as TAC-fragments (Tectonic Accretion Channel),

accreated along the subduction plate boundary (Engi et al., 2001). The Lepontine Dome structure is bordered to the north and south by elongate steep belts. The east–west striking Southern Steep Belt (SSB) is the Alpine nappe stack (Milnes, 1974; Heitzmann, 1987). The entire nappe stack swings from a flatlying to moderately dipping orientation, into a subvertical and even overturned orientation. The SSB itself has classically been divided into different units, historically called "Zonen" or "Züge", i.e. elongate, discontinuous trails (e.g., Kern, 1947; Zawadynski, 1952; Knup, 1958). These map-scale subdivisions essentially reflect changes in the spectrum of rock types found in the mélange, thus emphasising local diversity. However, the different zones may have several characteristics in common, such as the occurrence of similar orthogneiss1 rock types, of migmatites with variably deformed leucosomes, or of eclogite relics. The SSB contains the most convincing evidence of Alpine anatexis and I intrusion. Pegmatitic and aplitic dykes, meter to decameter-sized granitic bodies, as well as *in-situ* migmatites, are widespread within this belt stretching E–W from the Bergell to Domodossola (Fig.1). Especially in the central part of this E–W belt, high-T mylonites are common. In general, the observed grain sizes in many rock types are exceptionally small, given the high metamorphic grade.
To the south, the belt is truncated by the E–W running Insubric Line (termed Tonale-Line east of Locarno, and Canavese-Line west of Locarno (e.g., Schmid et al., 1989; Steck and Hunziker, 1994; Schärer et al., 1996), a major ductile to brittle shear zone of the Central Alps. This shear zone separates the Southern Alps (Apulian plate), which were weakly metamorphosed and moderately deformed during the Alpine orogeny, from the high-grade metamorphic Central Alps (European plate) to the north. North of the Insubric Line, metamorphic grade reached amphibolitefacies conditions on a regional scale (Trommsdorff, 1966; Wenk, 1970; Niggli, 1970; Engi et al., 1995), but relics of eclogite and granulite facies occur inside the TAC-units (Engi et al., 2001).

Migmatites in the Central Alps

Migmatites surface in many parts of the Alps and, where they occur in units, which attained but low to moderate temperatures in the Alpine cycle, it is clear that these rocks are of pre-Alpine origin (Frey et al., 1999). Based on structural and petrologic arguments it is also well established, that within the Lepontine area, old basement units were overprinted by Alpine metamorphism (e.g. polycyclic metamorphism in the Suretta nappe, Nussbaum et al., 1998; basement in the Bodengo area: Hänny, 1972). Inside these basement units, migmatites of pre-Alpine age, apparently related to the intrusion of large granitoid bodies (Variscan intrusives), are widespread. In the southern part of the Lepontine Alps, temperatures over 650 °C were reached in the Tertiary, but the age of migmatisation has long been a matter of debate (e.g., Klemm 1906/1907; Grubenmann, 1910; Gutzwiller, 1912; Niggli, 1950; Wenk, 1970). The discussion of the relative age has in part revolved around the particularly abundant stromatic migmatites, where relations between deformation and partial melting are structurally complex. This debate inspired several detailed studies, which were carried out in the eastern part of the SSB (e.g., Blattner, 1965; Hänny, 1972) and in the central Valle Verzasca area (Sharma, 1969), located further north. Isotopic studies (Hänny et al., 1974) revealed older migmatites overprinted by local partial melting during Alpine metamorphism in the eastern SSB. In this same section of the belt, migmatisation is contemporaneous with magmatic activity related to the Bergell intrusion, the latter starting at 32 Ma (e.g., Berger et al., 1996). The deeper portions of the Bergell pluton (western end) preserve a protracted magmatic history with a final crystallisation at 28 Ma (Oberli et al., 2004). Monazite dating in the gneisses and migmatites yielded 29–26 Ma (Köppel et al., 1981; Hänny et al., 1974) and similar ages were obtained for discordant veins and dykes of the SSB (Schärer et al., 1996; Romer et al., 1996). Hence partial melting in the eastern and central part of the SSB most likely occurred in this time interval. This is further corroborated by isotopic ages obtained using other thermally retentive chronometers, which date the amphibolite facies metamorphism (Vance and O'Nions, 1992; Nagel, 2000), to which partial melting is related. The youngest in trusives observed in the Alps crosscut the latest ductile Alpine structures and occur as aplites and small intrusions (i.e. Novate granite), which have been dated at ~25 Ma (Romer et al., 1996; Schärer et al., 1996; Liati et al., 2000). A particularly young, Miocene age of 20 Ma (lower intercept) was obtained for a completely discordant dyke at Lavertezzo (Romer et al., 1996).

Stop 2 Centovalli area

West of Locarno the Insubric Line bends towards the southwest, whereas the SSB maintains its roughly E–W direction (Heitzmann, 1987; Schmid et al., 1989; Fig. 1). The opening gap between the SSB and the Insubric Line is taken up mostly by the Sesia-unit. In this western segment of the SSB, the Insubric Line did not correspond to the main boundary during subduction and back-thrusting, but rather transects this older element (Schärer et al., 1996). Due to major displacements along the younger ductile to brittle Insubric and Centovalli Lines, amphibolite-facies fabrics and assemblages in the SSB were locally severely overprinted by greenschist-facies brittle deformation.

Stop 3: Ivrea zone.

Geology of the Finero ultrabasic/basic complex (Ivrea Zone)
(Mainly after Kruhl & Voll, 1979a,b; Zingg, 1990; Handy et al, 1999)

The Ivrea Zone is situated at the internal border of the Alpine arc, south and east of the Insubric Line. A large gravity anomaly and inverted seismic velocities have lead to the model of an exposed crustal section comprising the following units (Fig.2.1):
(l) the Ivrea Zone with peridotites, mafic rocks and paragneisses in the amphibolite and granulite facies (lower and basal intermediate crust);
(2) the Strona-Ceneri and Val Colla Zones with granitic gneisses, paragneisses and micaschists in the amphibolite facies (intermediate crust);
(3) the Late Paleozoic to Tertiary sedimentary cover of the aforementioned units (upper crust). The crustal section (Fig.2.2) is not strictly continuous, since it is truncated by several, variously aged faults.

The oldest part of the crustal section is preserved in the medium-grade basement units, which are interpreted to be the overprinted remains of an Ordovician (440-480 Ma) magmatic arc or forearc complex. During Variscan subduction this arc was tectonically underplated by a Carboniferous accretion-subduction complex (320-355 Ma) containing metasediments and slivers of Rheic oceanic crust presently found in the Ivrea-Verbano Zone. During the late stages of Variscan convergence (290-320 Ma), lithospheric delamination triggered magmatic underplating and lead to polyphase deformation under amphibolite to granulite facies conditions. This was broadly coeval with extensional exhumation and erosion of the Variscan- overprinted Ordovician crust presently exposed in the Strona-Ceneri and Val Colla Zones.

Post-Variscan transtensional tectonics (270-290 Ma) were associated with renewed magmatic underplating, mylonitic shearing, and incipient exhumation of the lower crust in the Ivrea- Verbano Zone. This coincided with the formation of elongate basins filled with volcanoclastic sediments in the upper crust. Early Mesozoic, Tethyan rifting of the southern Alpine crust (180-230 Ma) reduced crustal thickness to 10 km or less. In the lower crust, most of this thinning was accommodated by granulite to retrograde greenschist facies mylonitic shearing. The lower crust was exhumed along a large, noncoaxial mylonitic shear zone that was linked to asymmetrical rift basins in the upper crust. The composite structure resulting from this complex evolution is probably typical of thinned, late Variscan continental crust on the passive margins of western Europe. Alpine faulting and folding (20-50 Ma) fragmented the crustal section. The originally deepest levels of the crustal section in the Ivrea-Verbano Zone as well as some segments of the basement-cover contact were steepened, whereas other parts of the crustal section, particularly the Strona-Ceneri Zone, underwent only minor to moderate Alpine rotation.

The Finero Complex: The Finero Complex (Fig. 2. l) is the largest ultramafic body in the Ivrea Zone. It is composed of meta-peridotite and metagabbro layers and - during late Variscan times — deformed under, first, granulite and then amphibolite and greenschist facies conditions. The body is bound by the Insubric Line against the Alpine orogen (i.e. those parts of the Alps which experienced deformation and metamorphism during Alpine times) in the northwest. The true nature of the body is still under discussion. Most probably it represents a layered intrusion of upper mantle melts into the lowermost part of the continental crust. However, the original geometry of the body is not clear. There are repetitions of peridotite layers which may be due to
(l) multiple intrusion,
(2) repeated differentiation processes or
(3) processes of self-organization in the magma chamber, or
(4) post-intrusion isoclinal folding during still high-temperature conditions. In addition, it has been suggested that the complex was affected by metasomatic alteration resulting in pervasive presence of amphibole and phlogopite in the peridotite layers (Zanetti et al., 1999).

J. H. KRUHL and G. VOLL 1979: EXCURSION-GUIDE TO THE WESTERN PART OF THE FINERO PERIDOTITE-METAGABBRO – COMPLEX OF THE IVREA ZONE (N. ITALY).- MEMORIE DI SCIENZE GEOLOGICHE Volume XXXIII, pp. 17 - 26, fig. l, Padova 1978 - 79

INTRODUCTION

The Ivrea Zone follows the S-margin of the Alps from W of the Dora Baltea valley to near Locarno. It consists of granulite- and amphibolite facies metasediments and intercalated metabasic and meta-ultrabasic bodies. These bodies are concentrated in the N. The Finero- Complex is the largest of them. This zone and especially the ultrabasic bodies have attracted attention for important questions: are they the uppermost part of the earth mantle, and are they connected to higher levels along profiles towards the S into the Series dei Laghi? Is their appearance just S of the Alpine margin connected causally to the formation of the Alps?

The following hypotheses of formation have been discussed for the Finero Complex:

- The Complex is part of a solid upper Mantle and ploughed up during either prealpine times or in connection with alpine orogeny.
- The Complex is a layered intrusion which formed either in the mantle below a continental or an oceanic crust. It was then metamorphosed and lifted up.
- The complex is a layered intrusion which formed in the deeper part of a continental crust. It was then metamorphosed and lifted up.

To us it seems likely that possibility l can be ruled out. The compositional layering in meta- gabbros and peridotites predates every folding and metamorphism. It suggests cumulate formation of a layered, multiple intrusion. Further indications for this mode of formation are: chromite and other spinel layers which often show vague indications of current bedding; preservation of small idiomorphic olivines included in large pyroxenes or small idiomorphic pyroxenes in large olivines: both are caught by adhesion in a melt; occasional preservation of cumulate- and intercumulate textures in thin section.

Age dates provided by HUNZIKER (1974) and HUNZIKER and ZINGG (1978) suggest that the whole Ivrea Zone and the Finero Complex with it reached a higher position long before and without causal connection to the alpine orogeny. If the Complex were a layered intrusion formed within a solid higher mantle it is hard to understand why such parts of the solid mantle have not been ploughed up as well. The very thick sequence of metapelites and their sometimes more complicated orogenic history in comparison with basic bodies in them do not allow this part to be underlain by oceanic crust. We, therefore, think it most likely that the Finero Complex formed as a layered multiple intrusion within the deepest part of a continental crust.

GRANULITE FACIES DEFORMATION AND METAMORPHISM

This deformation and metamorphism were the most penetrative. They occurred together with formation of kinzigites, stronalites and metabasites in the whole Ivrea Zone. Age dates of HUNZIKER (1974) yield «caledonian» times for this event. There are no traces of older deformations and metamorphisms at least in the meta-ultrabasic and -basic rocks. In the Finero Complex this deformation formed isoclinal cm- to m-folds. They are common in the metagabbros but rare in metaperidotites. In the former these folds are isoclinal and no general sense of overfolding can be recognized. All granulite facies folds have a pronounced axial plane cleavage expressed by flat sides of grains, (100) of hornblendes and pyroxenes and (001) of phlogopite, c of hornblendes and pyroxenes and features of olivine-orientation (KRUHL and VOLL, this volume) indicate the pronounced stretching lineation which may or may not be recognized from grainshapes megascopically. This stretching lineation and fold axes are parallel. They plunge 0-30° SW on the S-limb of later greenschist facies antiforms which fold this lineation to a NE plunge on the N-limbs. Compositional layering and isoclinal granulite facies folds show strong thinning and boudinage in the metagabbros and in pyroxene-hornblende layers of peridotites. Boudin axes are normal to the stretching lineation (str). Mineral parageneses produced during this metamorphism have already been described by LENSCH (1968,1971).

AMPHIBOLITE FACIES DEFORMATION AND METAMORPHISM

This deformation was separated from the granulite facies deformation by strong static annealing, still in granulite facies. The amphibolite facies deformation caused total deformation of the northern and southern metagabbros and of central parts of the inner metagabbro. Furthermore shearplanes - dm to 10m thick - formed within all peridotites. Olivine, clinopyroxene, hornblende and plagioclase recrystallize partially or totally. Metagabbro-pyroxenes are partially replaced by hornblende,garnets by hornblende + clinopyroxene or hornblende alone, in both cases together with plagioclase. Clinopyroxenes, hornblendes and plagioclases may recrystallize - garnet and orthopyroxene are never found recrystallized. They are obviously not stable. Especially the instability of orthopyroxene indicates the transition to amphibolite facies.

Open folds are produced with axes plunging 30 - 60°. NE on S-limbs of later greenschist facies anticlines. Axial planes dip 70° NNW. Looking NE axial planes and short limbs of the folds are rotated clockwise. Angles between limbs are 40 - 120°, hinge thickening is not pronounced, A second cleavage is produced as axial plane cleavage to these folds or alone. It is expressed as flat sides of newly formed grains, especially of recrystallization aggregates (plagioclase, hornblende) and by (100) of hornblendes. Stretching is parallel to fold axes. It deviates clearly from the strike of the granulite facies deformation. It is expressed as long axes of grains, recrystallization aggregates and c-orientation of hornblende. Again static annealing outlasts deformation. It causes formation of polygonal plagioclases in recrystallization aggregates and replacement coronas around garnet with cellular fabrics. Only locally deformation interferes with the formation of such coronas. Then stretching haloes are formed at the ends of garnets, filled with the replacement minerals.

Shearplanes in the peridotites formed during this act form an anastomosing network roughly parallel to the general layering. Within these shear zones olivine and phlogopite recrystallize drastically. Olivine shows superplastic flow. Hornblende too may recrystallize. Orthopyroxenes are bent and torn apart, spinels are fractured. The drifting apart of fragments and of recrystallization aggregates depicts the stretching direction within these shear zones. It coincides with the stretching direction of more uniformly deformed metagabbros and peridotites. Locally the shear planes of these blastomylonitic shear zones may be folded round str. Long and short limbs of such folds indicate clockwise rotation looking NE along the axes — just as for major amphibolite facies folds (B_2). We call such folds refolding s_2 of shear planes B'_2. It seems important to point out that in and outside of shear zones axes of folds and stretching direction always coincide. Static annealing is less pronounced in these shear zones and it may be that within them deformation lasted longer.

GREENSCHIST FACIES DEFORMATION AND METAMORPHISM

This deformation occurred after regional cooling. It caused formation of the large antiform recognized by LENSCH (1968, 1971) and a further one towards the N, Fold axes plunge 40 - 60° WSW, axial planes dip steeply NNW. Major folds are usually concentric, minor folds often complicated by wedging and disharmonic folding in fold hinges. Within the peridotites planes of compositional layering bent round these folds are covered with flexure gliding slickensides produced during formation of and normal to the axes of these folds. Thin layers of talc and serpentine, formed during this flexure gliding may cover these surfaces. This folding produced no penetrative reorientation of minerals and no cleavage related to these folds can be detected. Folds, cleavages and stretching lineations of first and second deformation are folded round these axes. Therefore these structures alter their attitudes from S- to N-limbs of the greenschist- facies folds by simple refolding. Shear zones of amphibolite facies deformation are also folded round these latest folds.

Cold faults are numerous and it is largely uncertain whether they belong to this act of cold folding or to the later one of folding of the Insubric Line. Because of its possible significance we mention one fault bordering the NW- part of the phlogopite peridotite from its W-end to 463.40/5105.76. E from there this fault could not be found. As a contrary there the contact is formed by unfaulted granulite facies rocks rich in hornblende and pyroxene (464.30/5106.43). At 463.40/5105.76 this fault is covered with serpentine and carries vertical slickensides, indicating downward movement of the N-part.
It is still uncertain whether the formation of these greenschist facies antiforms is of alpine or prealpine age. It is, however, certain that formation of the Insubric Line cutting these folds and folding of this Insubric plane is younger.

THE INSUBRIC LINE

The Insubric Line cuts the greenschist facies folds. N of the Insubric Line 4 late alpine acts of folding may be distinguished under high greenschist facies conditions. T increases northward. Directly N of the Insubric Line T had fallen to lower greenschist facies T when the 3rd act of these folding stages refolded the Insubric Line. Before the Insubric plane was folded it cut the greenschist facies antiforms. These movements occurred under high greenschist facies conditions. Folds folding the Insubric Plane are open. They produce (looking W) an S-shaped major fold. The short limb of this fold covers the W- continuation of the peridotite greenschist antiforms. Within this flat cover numerous open smallscale folds with equal limbs are produced. Axes of all these folds plunge 30° WSW, axial planes dip steeply NNW. A crenulation cleavage was developed under lower greenschist facies conditions. It dips steeply NNW. Quartz still recrystallizes but micas are reoriented only partially.

EXCURSION TO THE W-PART OF THE FINERO COMPLEX

The area can be reached either from Locarno - Malesco - S^{ta} Maria Maggiore - Malesco) - Finero.
Or from: Cannobio (Lago Maggiore) through the Cannobino valley – Finero.
Coordinates are given from sheet S.ta Maria Maggiore F.o 16 della Carta dTtalia, 1:25000.

Point l (road Malesco - Finero: 463.40/5107,65) - Gneiss zone of the Monte Rosa nappe-root, N of the Insubric Line. Prealpine basement in (prealpine) high amphibolite facies metamorphism. Biotite-plagioclase-K-feldspar-augengneiss. Discordant pegmatite and aplite dykes of prealpine age. Prealpine compositional layermg. Alpine flattening is superimposed, It results in only one alpine cleavage: strike: 65 dip: 75 NW. Prealpine quartz, biotite, plagioclase and K-feldspar are recrystallyzed. Recrystallized plagioclase is oligoclase (25-30 Mol% An). Recrystallization aggregates are strongly elongate in the direction of first alpine stretching: strike SW-NE, plunge: 30-50 SW. Discordant dykes are folded by this flattening with variable axes depending on the original attitude. Prealpine folds become more isoclinal. Alpine flattening to >1/5 of original thickness.

Point 2 (463.40/5107.32), at the same road S of last point - Original rock: prealpine basement with garnets, diameter up to 3 cm. This rock continues along the strike into Val Loana towards the SW. There it contains occasionally very much and coarse prealpine sillimanite. These features together with rutile needles in the quartz suggest proximity to the kinzigite series. Coarse plagioclases and heavy minerals may be preserved from the parent rocks, The rock is penetrated by a l. alpine cleavage which again carries a l. alpine str-lineation, plunging 15-30° SW. Garnets are broken and fragments drifted apart in this direction. Most parent minerals are completely recrystallized or changed. Within the fine grained matrix of quartz, oligoclase, biotite and light mica small garnets, staurolites (often radiating) have formed. Fragments of old garnets are rimmed by new alpine one. Further W kyanite and chloritoid have been found in the same rocks, also staurolite, chloritoid and andalusite of alpine age together. At point 2 an intercalation of 1 m garnet amphibolite may be seen. Its garnets are relics from prealpine times. They are deformed as the garnets in the schists. Plagioclase is recrystallized entirely, hornblende nearly entirely. The new, green hornblendes are oriented with (100) in $s_{1\ alp}$ and with c parallel $str_{1\ alp}$. Amphibolites of similar history but without garnets may be observed 200 m S from point 2 at the road side.

Point 3 (463,60/5106,96), at the highest point N of Piano di Sale - Prealpine gneisses,completely phyllonitized by alpine deformation and metamorphism, A penetrative $s_{1\ alp}$ cleavage shows a phyllitic habit. It is refolded by second alpine folds with axes plunging 60 - 90° NE. Short limbs are rotated clockwise (looking E). Parent minerals are completely recrystallized or changed, apart from single coarse plagioclases or muscovites. These rocks are dose to the border of greenschist/amphibolite fades. Tiny alpine garnets appear. A $str_{1\ alp}$ may be recognized parallel to $B_{2\ alp}$.

Point 4 (463.66/5106.92), just S of the Pass - At both sides of the road meta-subvolcanic dykes of meta-granite-porphyry may be seen. The dykes are 20 - 200 cm thick and intercalated irito schists of point 3. The dikes are intruded parallel to a prealpine layering and have suffered the same alpine deformation as the schists, i.e. a penetrative $s_{1\ alp}$ and refolding round $B_{2\ alp}$ (very steep axes parallel to $str_{1\ alp}$). $S_{2\ alp}$ has 1-2 cm-distances.

Phenocrysts of plagioclase still preserve a fine magmatic zoning and allanite-epidote crystals included from the melt. Apatite phenocrysts contain tiny slender zircones included from the melt. Biotite phenocrysts have secreted rutile needles which were then changed to sphene during alpine metamorphism. K-Feldspar phenocrysts are rare. Quartz phenocrysts are strongly flattened and extended parallel $str_{1\ alp}$ (more than 10 times) under complete recrystallization. Cracks in the feldspars are filled with adularia and sometimes epidote and prehnite. The matrix consists of fine newly formed quartz, feldspars and micas, sharply oriented during alpine deformation. Such dykes also occur just N of point 3. They are common in this zone and range in composition to trachyandesite and andesite.

Point 5 (463.84/5106.67) - Here one crosses the Insubric Line. Directiy N of it, 70 m NE of the road under trees, the last phyllonitic gneisses are found. They resemble those of point 3 but contain many porphyroclasts of plagioclase, muscovite and turmaline in a fine grained and very sharply deformed matrix without alpine garnets.

Directiy S of the Insubric Plane - 20 m E of the road under trees - granulite facies metagabbros are the first rocks on the S-side. They are rich in ortho- and clinopyroxenes, contain little hornblende and show but weak amphibolite facies deformation. They have, however, suffered extensive greenschist facies alteration (formation of chlorite and epidote from garnet, of pale amphiboles from pyroxenes) under static conditions. 20 m along the road towards the S lead to an intercalation of schists rich in oriented clinozoisite with much posttectonic chlorite, Serpentine and colourless amphibole.

50 m S of point 5 garnet-metagabbros are strongly deformed by amphibolite facies shearing (recrystallization of brown hornblende, clinopyroxene and plagioclase). The posttectonic greenschist alteration is still strong.

140 m from point 5 along the road the metagabbros are still strongly deformed and recrystallized by amphibolite facies deformation. Here the Greenschist-facies alteration is missing or weak. This situation persists along the road into the metagabbros. The amphibolite facies alteration of garnets has been syntectonic her causing coronas to form which are extended parallel to the strike of this deformation.

Point 6 to point 7 (464.07/5106.30 to 463.98/5106.24), near chapel N of Finero -
At point 6: border between the most northerly hornblende peridotite and the garnet metagabbro directiy S of it. The hornblende peridotite shows granulite facies deformation and very fine grained olivine recrystallized at grain boundaries of coarse olivines. Stretching of granulite facies age plunging gently SW can be recognized from long axes of olivines. The rocks are rich in green spinel surrounded by serpentine which also fills cracks often together with talc.
In the typical garnet metagabbro S of the border an amphibolite facies shearing starts about 4-5 m S of the border and becomes more penetrative away from it. The stretching lineation plunges 30° NE. 2-3 m of the metagabbro near the peridotite are banded, rich in plagioclase and poor in garnet. Within this zone sapphirine is concentrated (LENSCH, 1971 a). It occurs in the peridotite immediately adjacent to the metagabbro too. There it forms rims around green spinels. This peridotite is rich in orthopyroxene, hornblende and clinopyroxene. In the metagabbros sapphirine is concentred in layers. It forms large crystals up to 2 cm diameter which show mechanical twinning. Furthermore, as cellular intergrowths with plagioclase (An 80-90) and clinopyroxene, and as smaller isolated grains. These metagabbros are dominated by granulite facies deformation. Here too, however, late and static greenschist facies alterations are frequent.

Point 15 (464.10/5105.93) - This point is reached by descending the path from Finero to Provola, to the bridge across the Cannobino River. Concentric folds of greenschist facies stage may be seen in the phlogopite peridotite. Wave lengths are between 1 and 10 m. Fold axes strike: 55°, plunge: 75° SW. These folds fold the phlogopite (001) parallel orientation and a plagioclase dyke which follows this plane. It is deformed strongly. A dykelet branching off and cutting the layering shows less deformation. The dyke contains biotite.

Point 19 (464.87/5 105.95), at Ponte Provola
Not far N from this bridge the phlogopite peridotite shows good compositional layering with different phlogopite contents. It is sharply parallel oriented. A few m NW of the bridge the contact between phlogopite peridotite and inner garnet -metagabbro is well exposed. The metagabbro shows a very good compositional layering. 3 m downstream from the contact the metagabbro contains several isoclinal granulite facies folds with wave lengths and amplitudes within the m-range.

Point 20 (464.89/5105.92), downstream from point 19 - Here the contact between the inner metagabbro and the hornblende-peridotite is well exposed. At this contact a coarse hornblende-pyroxene-layer finishes the metagabbro. Within the peridotite S of it several dm-layers with green giant amphiboles are parallel to the compositional layering. They are boudinaged. Boudin axes are vertical or plunge steeply E.
Going SE through the hornblende peridotite one finds several Cr-spinel layers, some of which display a vague current bedding, while others show folds which may well be flattened magmatic slumping. Some layers rich in hornblende branch off the normal hornblende-rich layers and are folded during granulite facies flattening. The layers of normal attitude from which they branch off remain unfolded. They are, however, occasionally cut obliquely by the plane of flattening, i. e., the granulite facies cleavage. This arrangement of hornblende layers may be due to magmatic current bedding.

Point 21 (464.97/5105.80) - S-margin of the hornblende peridotite against the southern metagabbro = basischer Hauptzug. This margin is formed from several m of coarse hornblende-pyroxene rock. Directiy S of these rocks a zone of ultra-blastomylonites, several m wide, starts. Directly at the contact these contain calcite-rich layers. The calcites have still been affected by deformation but they contain porphyroclasts of the metagabbro minerals. S of the carbonate-rich layers the blastomylonites are black, extremely fine grained and resemble pseudotachylites. They are not pseudotachylites, however, but consist of very fine grained brown hornblendes and plagioclases resulting from recrystallization in amphibolite facies shear zones. Large porphyroclasts of brown hornblende and — in many layers — of pyroxenes and plagioclases are contained in this fine grained matrix. Within this zone, deformed in amphibolite facies, stretching plunges NE with 35°. After the amphibolite facies deformation and under static amphibolite facies conditions coronas replacing garnet formed round garnet porphyroclasts.

All these rocks were affected by later greenschist facies alterations without further deformation. These alterations cause chlorite, pale amphiboles, epidote and, partly, carbonate to form. This deformation wanes towards the S into the metagabbros but it does not disappear completely.

Aug. 12 Wednesday: Aosta – Valtournanche - Cervinia – Aosta (110 km)
2- Ascent from Valtournanche to the Lago di Cignana (HP and UHP rocks of the Penninic ocean) Whole day walk in high Alpine area.
(Guide: Bernd Lammerer)
Accommodation: Camping Aosta International Touring; Fraz. Arensod, 10 - 11010 Sarre (Aosta) - Italy
+39 0165 257061 campingtouring@libero.it

Val Tournanche / Lago di Cignana (HP and UHP rocks of the Penninic ocean)
(mainly after Van der Klauw et al., 1997; Reinecke, 1998; Reinecke et al., 1994)

Please refer to original article: ftp://ftp.hardrock.rub.de/lib/1997/van%20der%20Klauw%20etal%201997.pdf

The Val Toumanche (from Chaillon to Breuil; Fig.3.1) cuts the subhorizontal nappe pile of the inner arc of the Western Alps. In its lower parts Zermatt-Saas Fee-Ophiolite Zone (the oceanic crust and sedimentary cover of the former Penninic ocean) is exposed. It is overlain by the Combin Zone and the Dent Blanche nappe which represents a relic of the western margin of the Adriatic plate and builds the top of the Matterhorn.

Stop 1: UHP rocks in the vicinity of Lago di Cignana (Fig.4.2). UHP metamorphic oceanic crust of the Zermatt-Saas Zone (Piemontese Zone) at Lago di cignana, Valtournance, Italy) (from Reinecke, van der Klauw, Stöckhert, 1994)
UHP metamorphic eclogites and associated metasediments of the Piemontese-Ligurian ocean crop out immediately south and west of Lago di Cignana in the area extending ca. 1.2 km in N-S and E-W directions from the dam (1950-2250 m above sea level). To the east and northeast, the UHP rocks disappear under scree covering the steep walls of the Valtournanche valley. In the north/ west and southwest, the UHP rocks are tectonically overlain ("Combin-thrust" of Ballevre and Merle 1993) by greenschists and calcschists of the Tsate nappe (Combin zone) and by a huge serpentimte slice (southern slope of Mt. Pancherot). Relic Mg-poor almandine-grossular garnets and poikiloblastic epidote in the greenschists preserve inclusions of glaucophane (Wagner-Zweigel, 1993; Le Goff, 1986; Dal Piaz, 1976) which suggest derivation from garnet-epidote-glaucophanites indicative of a former low-grade, high-pressure Metamorphism.

Metamorphic rocks in structurally higher positions stacked upon the Tsate nappe compnse phengite-chlorite-epidote-albite-schists and -gneisses (Arolla series of the Austroalpine Dent Blanche nappe). In the studied area these rocks do not reveal any obvious high-pressure relics,

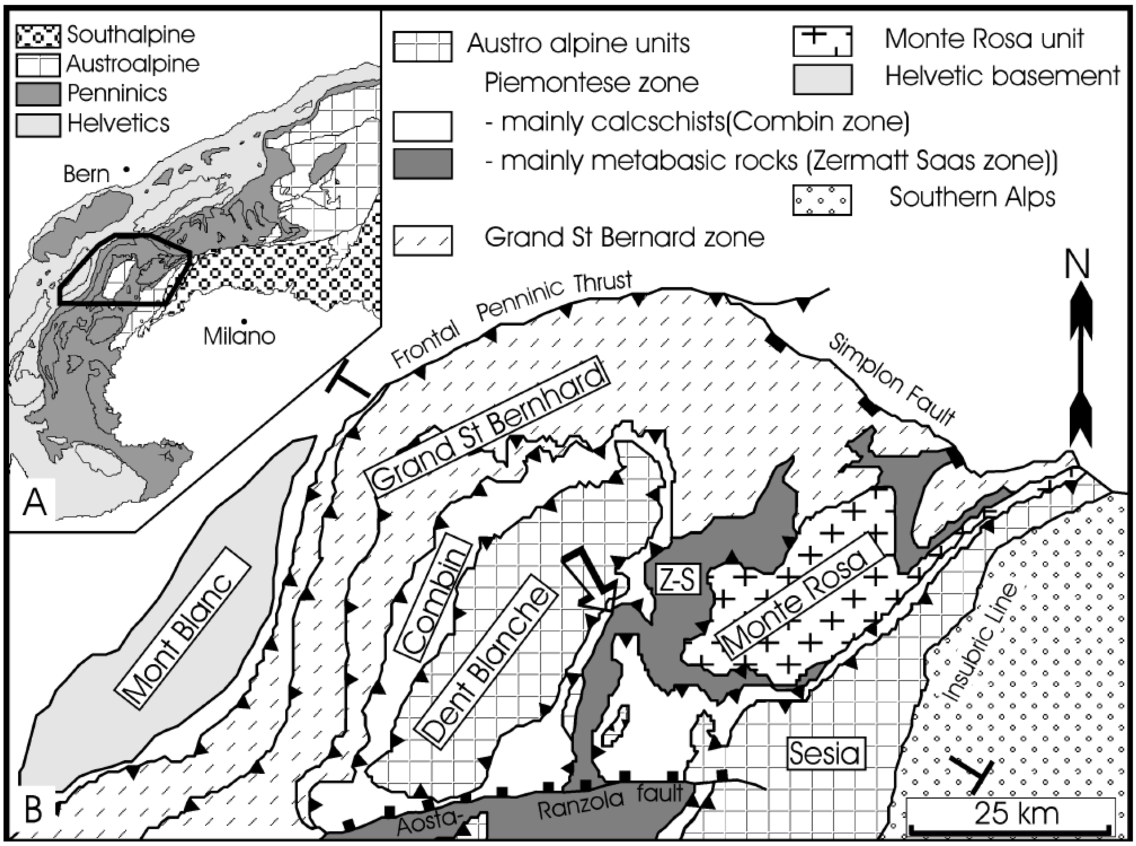

Fig.4.1: Location of the Lago di Cignana area (Val Tournanche) at the western margin of the Zermatt-Saas Zone(From Reinecke, 1998)

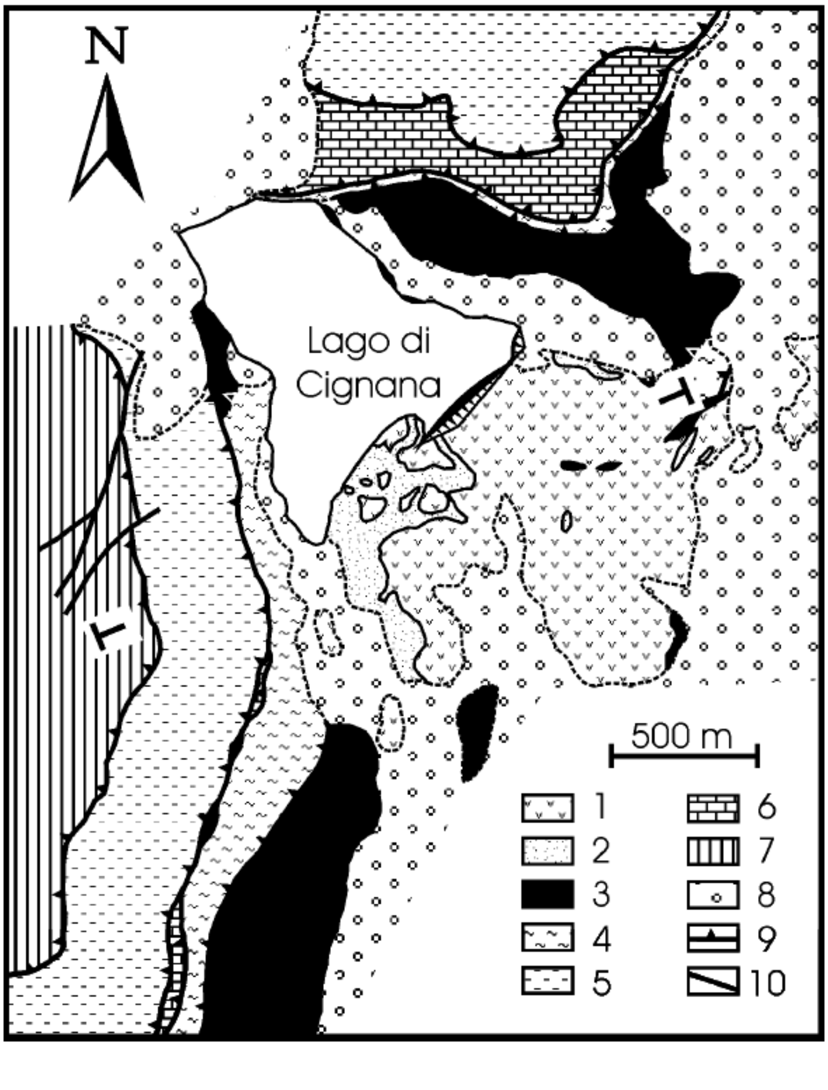

Fig.4.2: Simplified geological map of the vicinity ofLago di Cignana, Val Toumanche. (l)Serpentinized ultramafics, (2) coesite-eclogites and derived greenschists, (3) UHP metasediments, (4) greenschists interlayered with minor calc-schists (Combin Zone), (5) calc-schists with minor greenschists and marbles (Combin Zone). (6) dolomite-caicite marbles (exotic decollement sheet, cf. Dal Piaz, 1988), (7) Austroalpine (undifferentiated), (8) scree, moraine deposits and wetland. Solid lines = primary contacts; barbed wire = tectonic contacts; thick solid lines = normal faults. (From Van der Klauw et al., 1997)

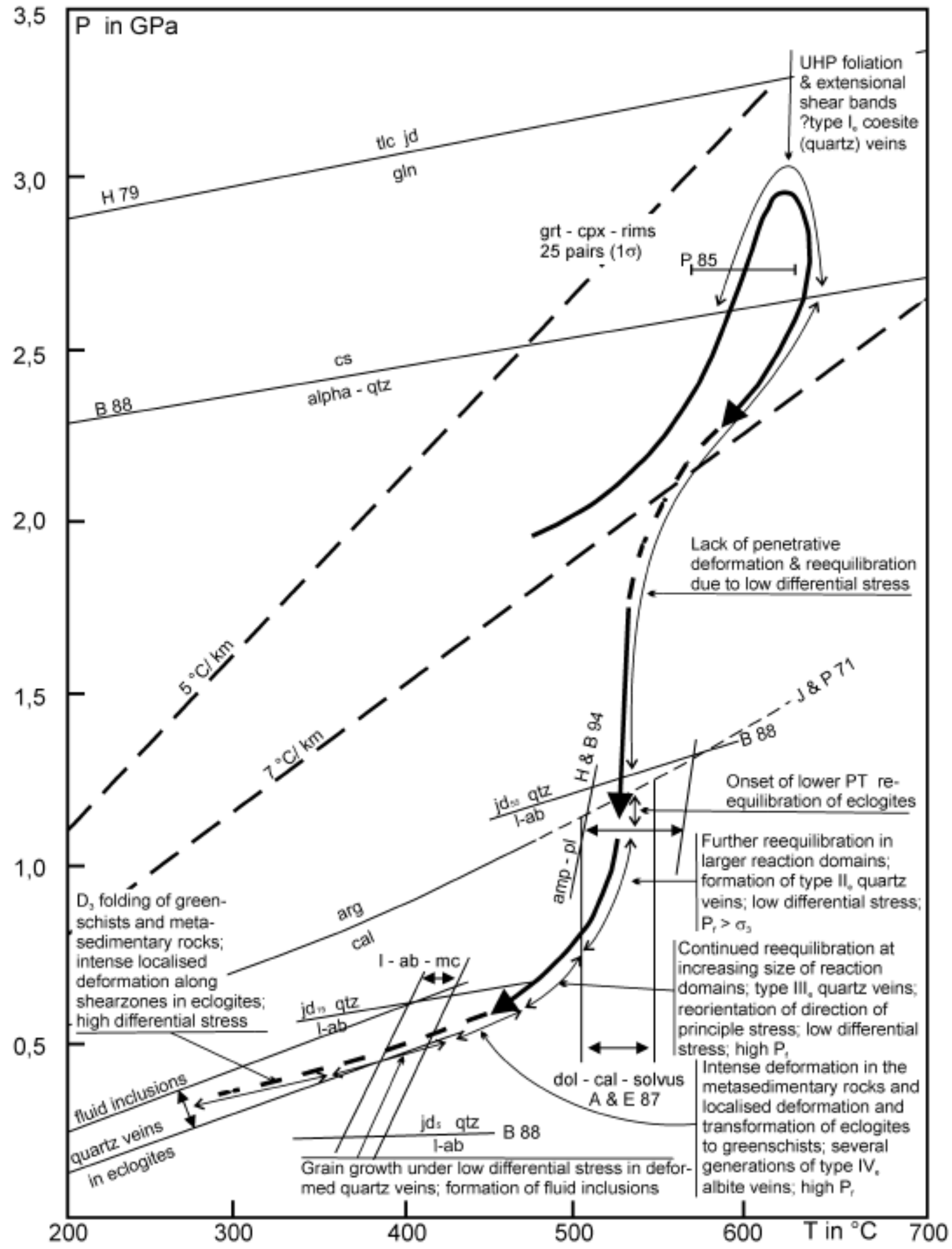

Fig.4.3; P-T path for the exhumation of UHPM rocks at Lago di Cignana, western Alps,Italy. Geotherms are based on the average rock density of 2.7 g/cm3. Continuous segments of the P-T path were derived from absolute and relativ thermobarometric methods applied to compositional zoning and inclusion pattems of garnet, combined with compositional information from other zoned matrix phases. (From van der Klauw et al., 1997)

METASEDIMENTS

The eclogite-facies metabasic and quartzose metapelitic rocks are thought to represent a coherent section through a segment of former oceanic crust (layer l and 2a), because:

1. both the metasedimentary and metabasic rocks have undergone the UHP metamorphism;
2. relic pillow structures, sometimes visible in fresh eclogites, as well as the sedimentary and geochemical characteristics of the overlaying metasediments, i.e. their non-clastic, siliceous composition and interlayering with ferromanganese, nodule-bearing metasediments match the features of recent oceanic crust;
3. major and trace element compositions of the eclogites have the characteristics of N-MORB (Beccaluva et al., 1984; Pfeiffer et al./1989; van der Klauw, in prep.).

ECLOGITES

In the field, the medium-grained, foliated eclogites locally show narrow glaucophane-garnet-rich layers and streaks wrapping around metre-sized eclogite lenses and pods. These structures are interpreted as relics of former pillows. The eclogites display a variable degree of lower-pressure re-equilibration ranging continuously from almost "fresh" sections containing less than 10-15 vol. % decomposition products to areas with statically re-equilibrated greenschist-facies assemblages which lack relics of the UHP stage, but still retain the eclogitic texture. Transitions between these extremes may occur over a distance of less a metre. Degree and areal extent of the lower-pressure overprint appears to be related to fluid advection on albite-actinolite-chlorite-epidote- filled veins and quartz-veins. In the eclogites, the UHP stage is represented by the assemblage: garnet (rim) + omphacite + glaucophane ± phengite + coesite + clinozoisite/zoisite + rutile + fluorapatite + pyrite ± dolomite.

Coesite rarely occurs as tiny inclusions in matrix omphacite and in narrow, inclusion- poor garnet rims. It is commonly inverted to quartz. Garnet is characterized by a continuous growth zoning with pyrope component increasing and spessartine, almandine and grossular components decreasing towards the rim (Fig. 2.4.1).

Poikiloblastic garnet cores have inclusions of omphacite, rutile, paragonite, clinozoisite and monocrystalline, almost unstrained quartz. Omphacite inclusions in garnet tend to be less jadeitic than matrix omphacite (Fig. 2.4.2). The Powell (1985) and Krogh (1988) expressions for the Fe-Mg garnet-clinopyroxene thermometer applied on 14 garnet (rim)-omphacite (matrix)- pairs in two eclogite samples yield 613134" C and 578±36'C, respectively (at P = 27 kbar). Exchange temperatures caiculated from compositions of omphacites included in garnet (cores) are on average by 30'C lower than peak temperatures. The major part of garnet growth obviously occurred on the prograde path in the stability field of a-quartz some tens of C below the thermal peak. Only with its last segment the P-T trajectory crossed the quartz-coesite transition and allowed growth of a final garnet rim under UHP conditions. In contrast to garnets from metasedimentary rocks, the eclogitic garnets do not show compositional reversais within the outermost growth zones. This indicates (hat garnet growth terminated near the thermal peak.

Metamorphic overprinting of the eclogites while cooling in the rising slab followed a complex (and not yet understood) pattern that commenced with the growth of titanite rims around matrix rutile, growth of topotactic Ca-Na-amphibole on glaucophane and around garnet, and growth of more Fe3+ rich rims on pre-existing clinozoisite/epidote. More advanced stages of overprinting show the replacement of omphacite and glaucophane by Ca-Na-amphibole-albite-symplectites emanating from grain boundaries and fractures. These features of intermediate-high-pressure decomposition grade into a greenschist-facies overprint that is characterized by the progressive resorption of eclogitic garnet and its pseudomorphous replacement by chlorite ± epidote ± biotite, the replacement of rutile by ilmenite ± titanite, and of Ca-Na-amphibole and paragonite by actinolite + albite and albite, respectively.

METASEDIMENTS

Pale green to silvery, medium-grained garnet-phengite-quartz-schists and quartzites are most common among the metasediments that overlay the eclogites with an apparent thickness of a few tens of metres in the area S and SW of the lake. The dominant foliation dips to NW-W. In structurally lower parts of the metasediments, dose to the contact with the eclogites, one observes dm- to m-thick layers and boudins of garnet-rich quartzites, garnet-clinopyroxene-quartzites, as well as highiy oxidized manganiferous rocks (piemontite-phengite-quartz-schists and -quartzites and aegirinejadeite-epidote- phengite-quartzites) which may reflect hydrothermal activity and/or hydrogenous precipitation m the pelagic Sediments near the former basaltsediment interface. Due to major viscosity contrasts between the different rock-types, the contacts between the eclogites and the metasediments are always disturbed.

Garnet-phengite-quartz-schists

Relics of the UHP stage are sparse and difficult to recognize in thin section, because(in contrast to the eclogites) the UHP matrix assemblage of the garnet-phengite-quartz-schists has been largely erased, i.e. the matrix minerals now mainly reflect the greenschist-facies overprint. From compositional zoning of garnet and phengite, inclusion relationships and pseudomorphous replacement textures the following UHP assemblage is inferred: garnet I + phengite I + coesite + epidote/clinozoisite or zoisite ± dolomite ± glaucophane/crossite + apatite + rutile + tourmaline.

Garnet I in high-Fe2+ bulk compositions forms continuously zoned grains (up to 4 mm) with Mg and Fe increasing and Ca and Mn decreasing from core to rim. Inclusions of Fe-rich chloritoid, paragonite and of monocrystalline, unstrained quartz in garnet I cores are considered as relics of the prograde high-pressure path. Small inclusions of partially transformed coesite or of polycrystalline, highiy strained quartz, pseudomorphous after coesite, rarely occur in Mg-rich garnet cores.

Lower pressure decomposition products in the matrix of the metapelitic schists and quartzites comprise two generations of less magnesian, Ca- or Ca-Mn-richer garnet II and III, titanite rims on rüttle, and growth of less siliceous phengite 11, paragonite, chlorite, biotite, albite, epidote and caicite. Na-amphibole is pseudomorphed by intergrowths of biotite + Ca-Na-amphibole/ actinolite + albite and phengite II may be almost isovolumetrically replaced by epidote or albite. The latter observations suggest the effective redistribution of matter via the fluid beyond the boundaries of single phase domains at the late-stage overprint.

Oxidized manganiferous quartzose schists and quartzites:

Two major rock types can be distinguished on the basis of mineral content and bulk composition:

L Garnet-phengite-epidote-Na-Pyroxene-quartzites with the low-variance UHP assemblage:
garnet I + phengite (Si3.4) + coesite + manganian epidote + aegirine-jadeite/
chloromelanite + hematite + rutile and rare manganoan dolomite, dravite, apatite and (?) aragonite.

2. Mn-Al-rich piemontite-phengite-quartz-shists that were previously described by Bearth (1967), Dal Piaz et al. (1979) and Reinecke (1991) from one occurrence ca. 100 m SE of the southern wing of the dam. Two principal low-variance UHP assemblages could be recognized (Reinecke 1991; in prep:):

Aug. 13 Tursday: Aosta – Tunnel Grand St. Bernard - Martigny (75 km)

2- Salvan Dorenaz basin: Permo – Carboniferous basin sediments, sedimentology of clastic sediments (alluvial fans, playa lakes, braided river systems)

(Guide: Andreas Wetzel and Petra Veselá)

Accomodation: Camping Martigny "Les Neuvilles" Rue du Levant 68 1920 Martigny Tel.: 027 722 45 44 Fax: 027 772 35 44

http://www.unil.ch/webdav/site/igp/shared/stampfli_research/field_trips/field_trip1.pdf

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 platetectonics 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 Hunsuperterrane (Stampfli 2000). Depending on their former location, preVariscan 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 Rousses, 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 highamphibolite 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 volcaniclastic 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 shortlived 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 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 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 Itype 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. Flyschtype 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/drifting 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 highgrade 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, Visean 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 Streel, 1980). UpperCarboniferous 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 & Streel (1980) dated Lower Carboniferous sedimentary rocks for the first time using palynology (Late Visean acritarchs), thus allowing a clear separation from the nearby upper Carboniferous deposits.

These detrital rocks consist of metamorphozed 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 ± Amp ± 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 SALVANDORÉNAZ BASIN)

They are remarquably 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 rightlateral strike-slip, transtensive fault (Brändlein et al., 1994). Other steeply dipping N-S to NE-SW 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. Illitecrystallinity 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; Badertscher and Burkhard, 1998).

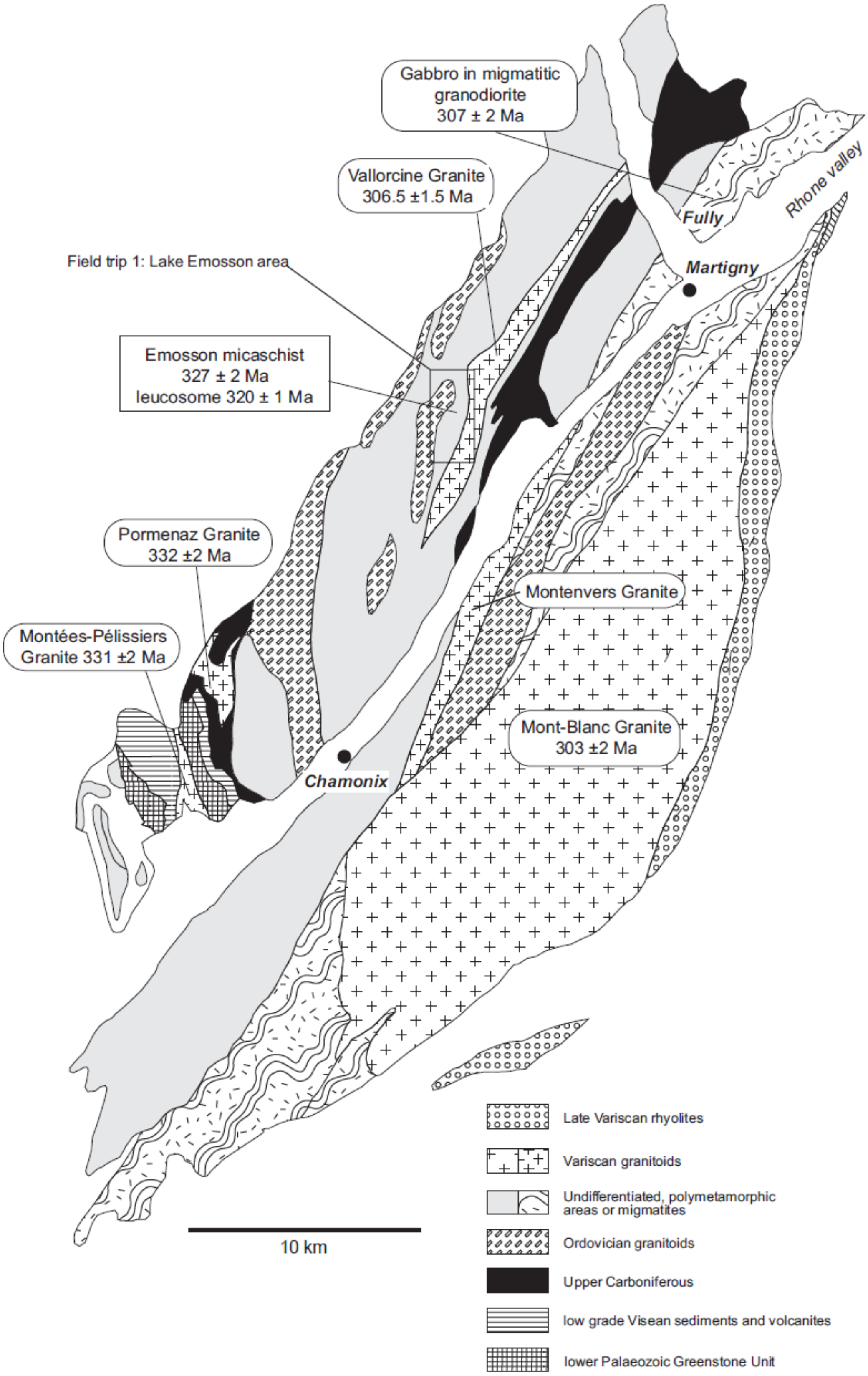

Fig. 23 - Simplified geological map of the Aiguilles Rouges-Mont Blanc massifs, after von Raumer et al.(1999).

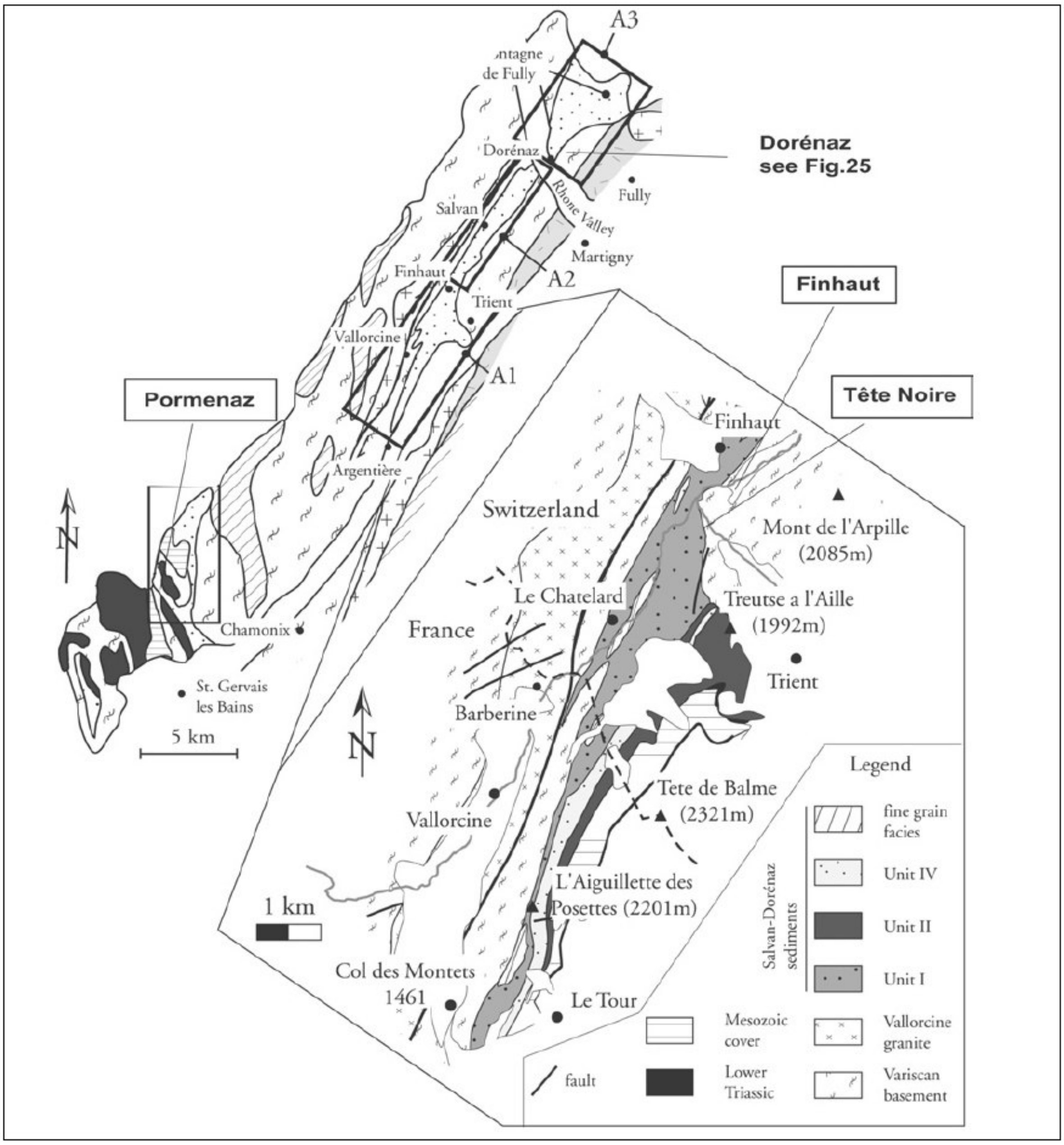

Fig. 24 - (from Capuzzo 2000, Fig. 13) Schematic map of the Aiguilles-Rouges massif (modified after Brändlein et al., 1994). and distribution of sedimentary facies in Late Carboniferous sediments. (with indication of localities

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, sandfilled 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 volcaniclastic 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 paleoflow from NE (Unit II) to SW (Unit III) may be attributed either to drainage reversals during a 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 coarseningupward 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 volcaniclastic layers found within sediments of Unit II and III testify for high-explosive volcanic eruptions from distant volcanic centers at 295 +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).

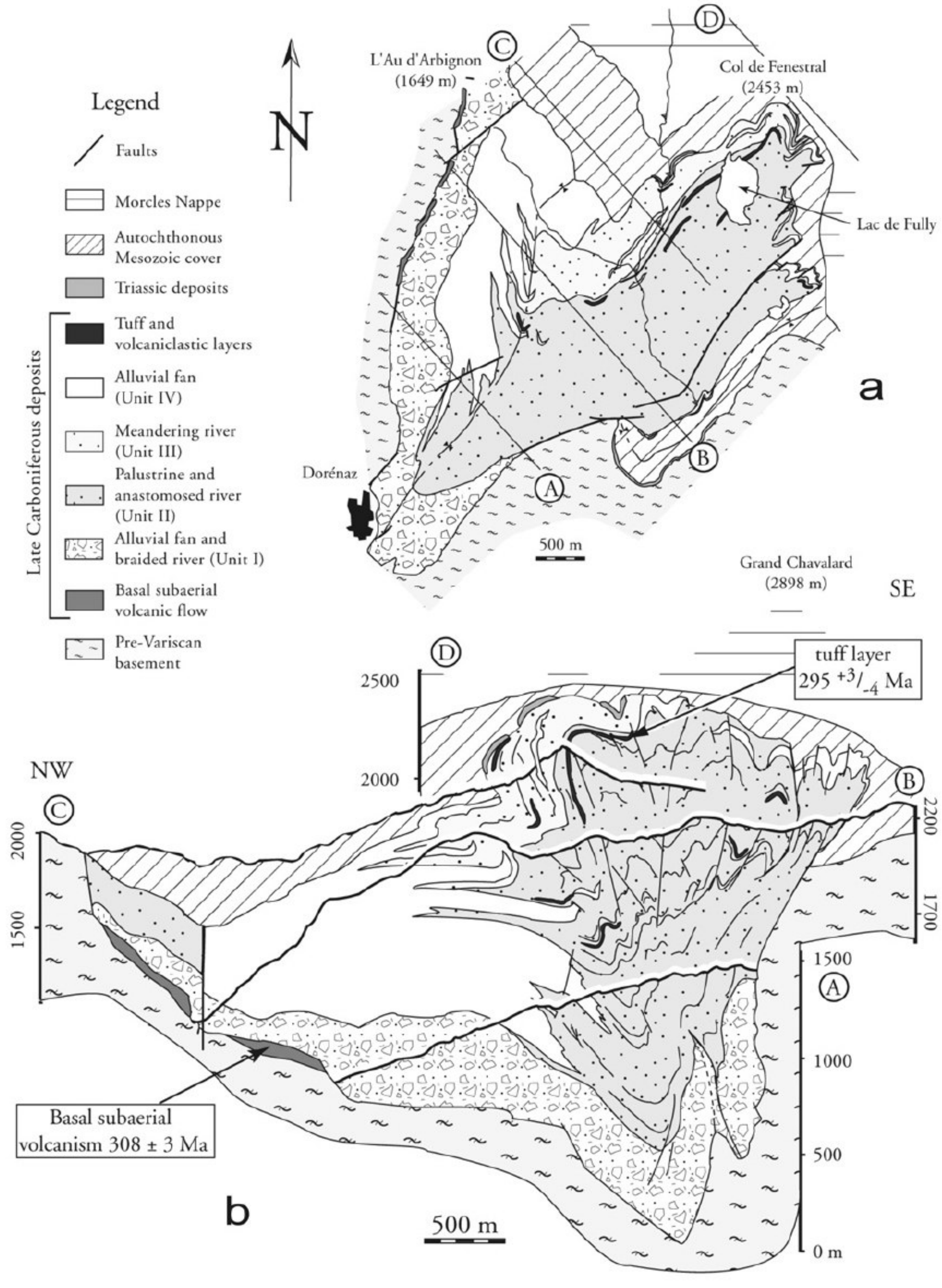

Fig. 25 (Fig 4.3 in Capuzzo 2000) : a) Geological map of the northern areas of the Salvan-Dorénaz syncline. The four lithological units that fill the basin are schematically reported, as the location of the volcanic and volcanigenic layers; b) Multiple cross-sections of the northern areas of the Salvan-Dorénaz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

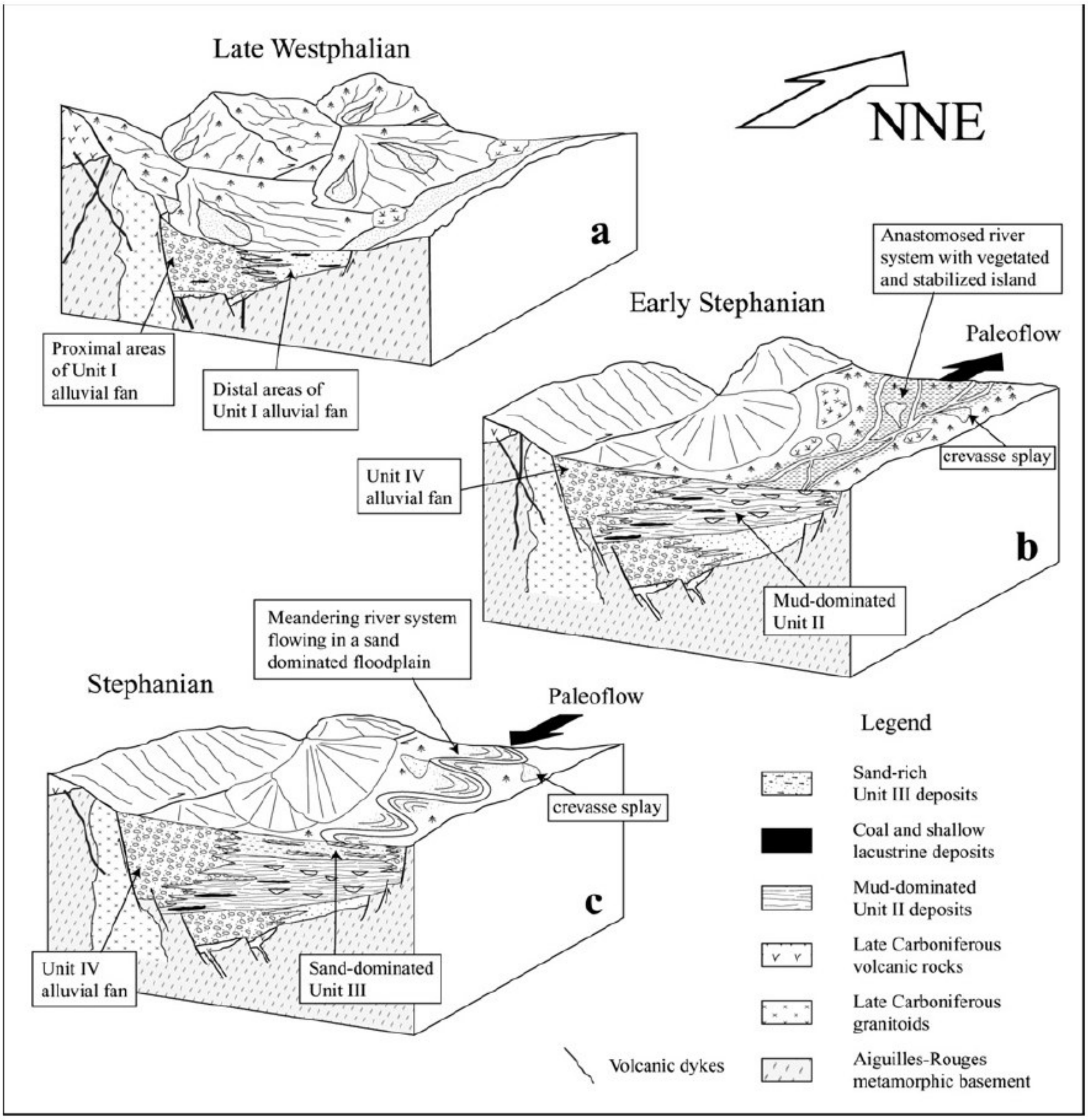

Fig. 26 - (from Capuzzo 2000, Fig. 2.10) Schematic block diagrams representing various depositional environments during theevolution of the Late Carboniferous Salvan-Dorénaz basin. a) formation of the basin induced by asymmetric subsidence along western bounding faults, which also favoured the emplacement 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.

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° orientation in the Aiguilles Rouges - Mont Blanc area against N10° 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 fracture patterns appear in specific lithologies and Alpine foliation is expressed as a faint neoformation 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 neoformation of chlorite, garnet and/or epidote is observed along joint surfaces.

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]: SALVANDORÉNAZ 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 anatectic 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]: SALVANDORÉNAZ 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 couloured fine grained channels underline the quickly changing situation between the deposits in the channel and the very coarse-grained boulder beds.

Aug 14 Friday: US group: **Martigny – Zurich Kloten Airport or Munich (550 km)**
German group: Martigny – Trient - Lac D`Emosson _ Martigny (70 km) -
2- Aiguille Rouge basement, Triassic transgression, dinosaur footprints in Triassic quartzites
Guide: Andreas Wetzel and Petra Veselá
Accomodation: Camping Martigny "Les Neuvilles"

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 positionof the Mesozoic sediments, which rest horizontally on top of the basement rocks in the middle of the massif (Aiguille de Belvédere, 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 isvisible 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 theclasts. 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 themylonites 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 metapeliticinterlayers;
(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 boudinshaped 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 kmlong 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 ± diopside ± 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 eucocratic 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 strikeslip 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%) ± muscovite ± 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 nonmigmatitic metapelites, for which a lower peakmetamorphic 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 stilpmnomelane 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.

Aug. 15. Saturday: Martigny – Sion – Chalet Anke Friedrich

geology of Wallis area

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

Philippe Thélin

I: INTRODUCTION

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

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 Turtmanntal (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 Turtmanntal to a trail of lenses in the Vispertal 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 prasinites are common, is considered to be Permo- Carboniferous (Bearth 1972 and 1978).

1.2.2 The Pontis Nappe

The Pontis nappe includes a polymetamorphic basement, a monometamorphic Permo-Carboniferous basement and Triassic fragments of a Subbrianҫonnais cover. The polymetamorphic basement which forms the main part of the nappe is fragmented, due to Alpine tectonics, into three distinct units:

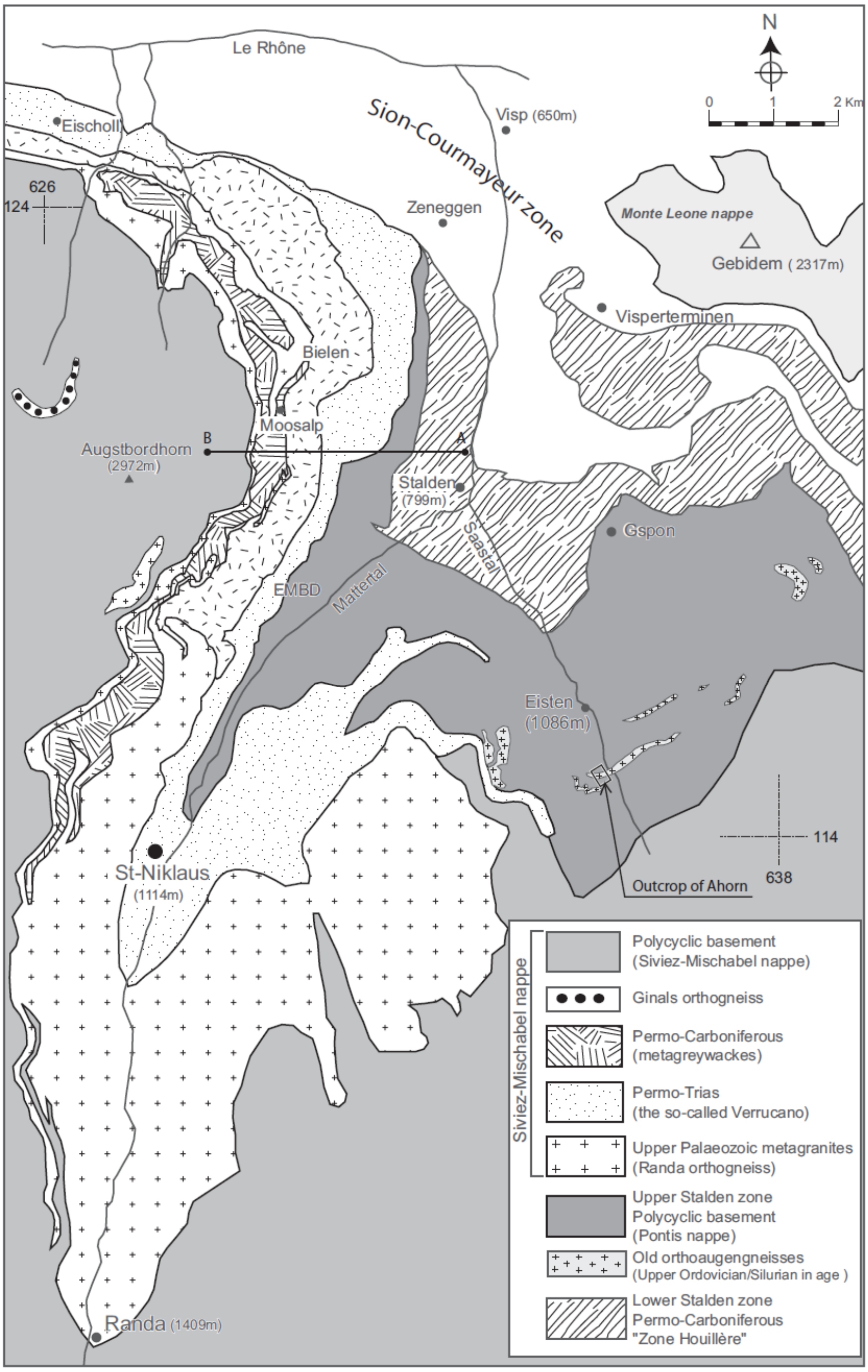


fig.32 - Simplified geological sketch map of the southern Visp area (Mattervalley and Saasvalley). Modified after Thélin 1987. A-B: location of the lithostratigraphic cross-section (Fig. 33).

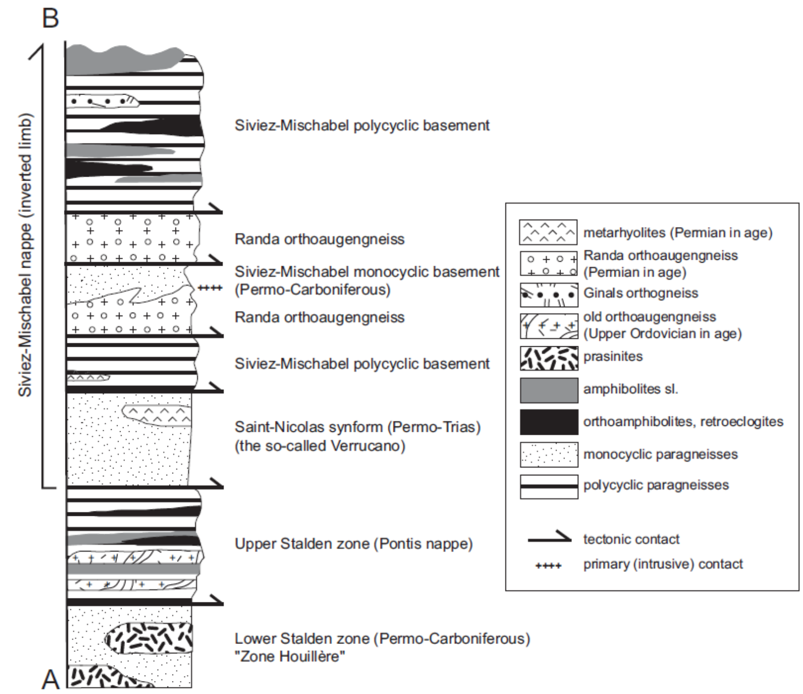
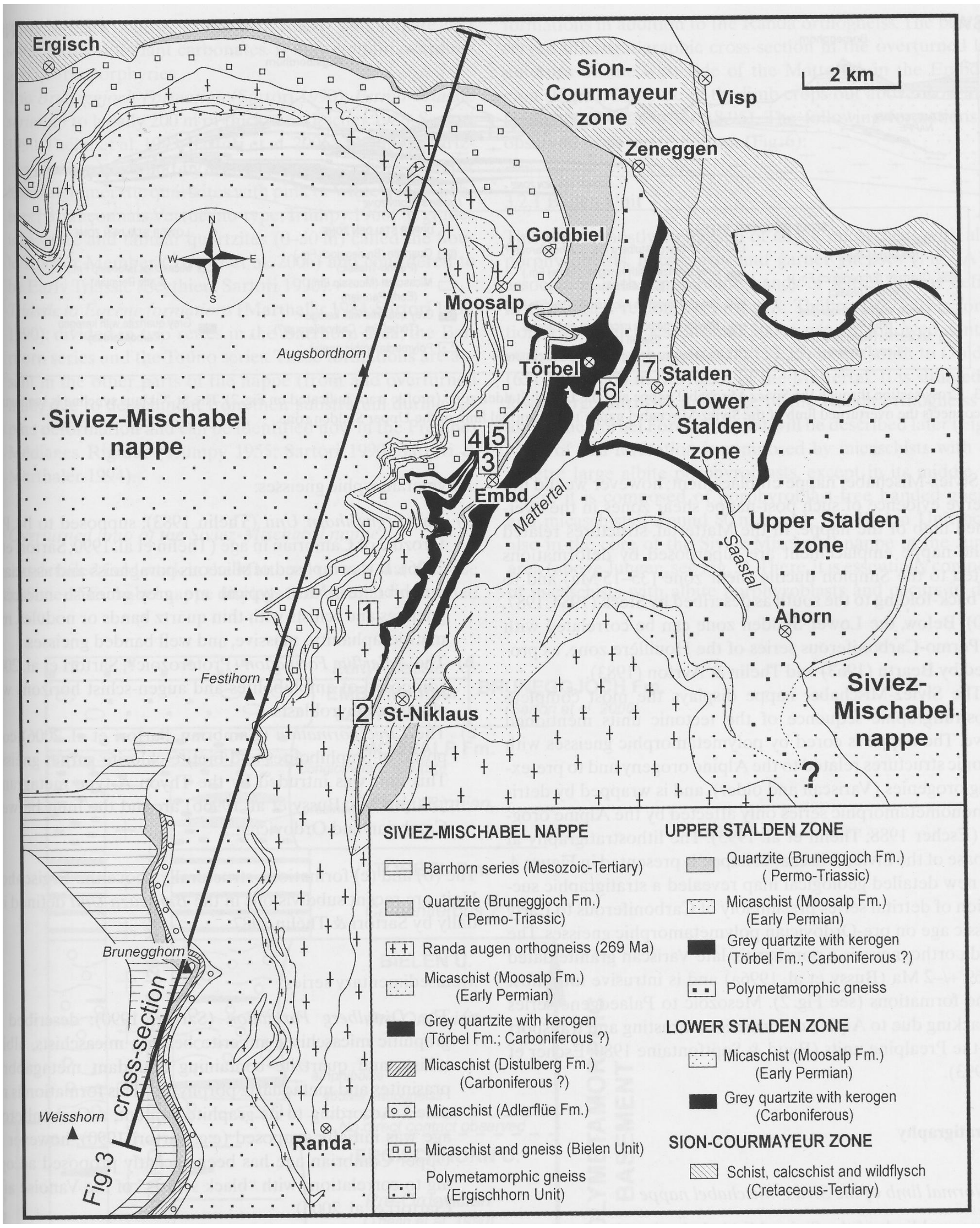
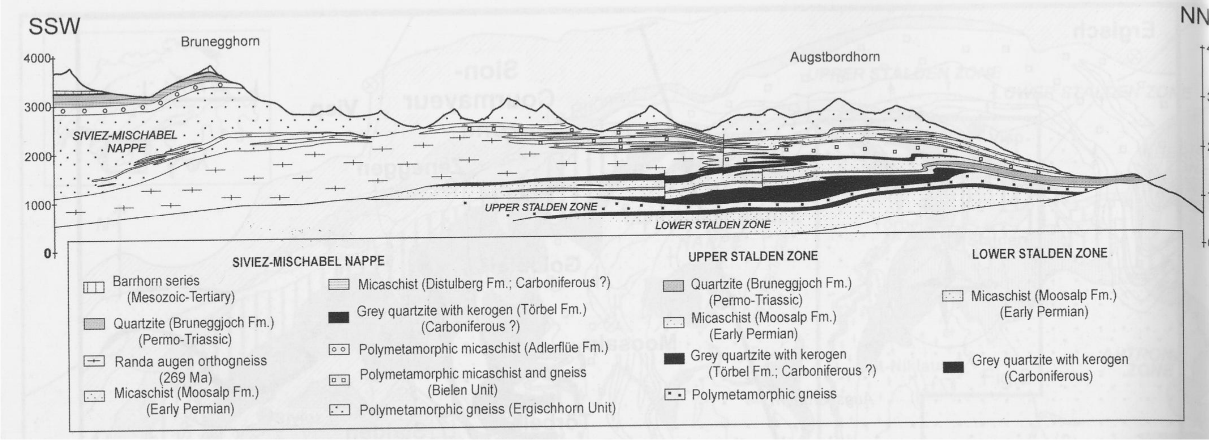


Fig.33 - Synthetic lithostratigraphic cross-section of the Middle Penninic units between ca. Stalden (799m) and Augstbordhorn (2972m). The vertical scale is arbitrary.

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; Thélin 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 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 coworkers suggest the following petrogenetic interpretations: (a) the mafic and ultamafic 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 (SiO_2<48wt.%), emplaced around 475Ma.


Ergisch
Sion-
Courmayeur
zone
Visp
2 km
Zeneggen
Goldbiel
Moosalp
Augsbordhorn
Törbel
7
Stalden
6
Lower
Stalden
zone
Siviez-Mischabel
nappe
4
5
3
Embd
Mattertal
Saastal
Upper Stalden
zone
Ahorn
1
Festihorn
2
St-Niklaus
Siviez-
Mischabel
nappe
?
Bruneggghorn
cross-section
Weisshorn
Fig.3
Randa
SIVIEZ-MISCHABEL NAPPE
Barrhorn series (Mesozoic-Tertiary)
Quartzite (Bruneggjoch Fm.) (Permo-Triassic)
Randa augen orthogneiss (269 Ma)
Micaschist (Moosalp Fm.) (Early Permian)
Grey quartzite with kerogen (Törbel Fm.; Carboniferous ?)
Micaschist (Distulberg Fm.) (Carboniferous ?)
Micaschist (Adlerflüe Fm.)
Micaschist and gneiss (Bielen Unit)
Polymetamorphic gneiss (Ergischhorn Unit)
UPPER STALDEN ZONE
Quartzite (Bruneggjoch Fm.) (Permo-Triassic)
Micaschist (Moosalp Fm.) (Early Permian)
Grey quartzite with kerogen (Törbel Fm.; Carboniferous ?)
Polymetamorphic gneiss
LOWER STALDEN ZONE
Micaschist (Moosalp Fm.) (Early Permian)
Grey quartzite with kerogen (Carboniferous)
SION-COURMAYEUR ZONE
Schist, calcschist and wildflysch (Cretaceous-Tertiary)


SSW
NN
Brunegghorn
Augstbordhorn
4000
3000
2000
1000
0
SIVIEZ-MISCHABEL NAPPE
UPPER STALDEN ZONE
LOWER STALDEN ZONE
SIVIEZ-MISCHABEL NAPPE
Barrhorn series (Mesozoic-Tertiary)
Quartzite (Bruneggjoch Fm.) (Permo-Triassic)
Randa augen orthogneiss (269 Ma)
Micaschist (Moosalp Fm.) (Early Permian)
Micaschist (Distulberg Fm.; Carboniferous ?)
Grey quartzite with kerogen (Törbel Fm.) (Carboniferous ?)
Polymetamorphic micaschist (Adlerflüe Fm.)
Polymetamorphic micaschist and gneiss (Bielen Unit)
Polymetamorphic gneiss (Ergischhorn Unit)
UPPER STALDEN ZONE
Quartzite (Bruneggjoch Fm.) (Permo-Triassic)
Micaschist (Moosalp Fm.) (Early Permian)
Grey quartzite with kerogen (Törbel Fm.; Carboniferous ?)
Polymetamorphic gneiss
LOWER STALDEN ZONE
Micaschist (Moosalp Fm.) (Early Permian)
Grey quartzite with kerogen (Carboniferous)

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 Mono- metamorphic 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" Barneuza 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 Barneuza unit are essentially pelitic. They are especially well developped in its middle part, forming augen-micaschists with albite porphyroblasts (Sartori and Thélin, 1987). Complete Alpine recrystallisation/ neoformation 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 enveloppe (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 ro cks (prasinites, 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 twostage 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 40Ar/39Ar 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” ORTHOGNEISSESES

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 (Bearth 1972; Thélin et al. 1993). They can also be traced equally well in the Berisal unit and in the Ruitor 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 megaclasts are single crystals (chessboard albite, low microcline, quartz) or augenshaped 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 biotitemuscovite (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 (Stype) 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 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. Highprecision U-Pb zircon dating yielded a subconcordant age of 500 +3/-4Ma. The Thyon metagranite is the third record of a Cambro- Ordivician alkaline magmatic activity in the Alps.

As A-type granitic magmatism is common in postorogenic 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 (Bearth 1964; Thélin 1987; Thélin et al. 1993; Thélin and Girod 2000). The RA derive from a subalkaline porphyritic granite, Permian in age (U/Pb zircon age: 269 Ma +/- 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 - 1cm) 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-rapakiwi mantles-Thélin 1987).

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, ie. 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 "exsolutions"; C2. Boudin of retroeclogite (detail). D. Narrow level of augengneiss, slightly folded, within the banded complex. E. "Old" orthoaugengneiss, Upper Ordovician in age (ca. 450 Ma.). Phyllonitic mylonites locally replace these orthogneisses.

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. ft5-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 a 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 tholeitic 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 slowspreading 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.

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 Zenneggen 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 ?).

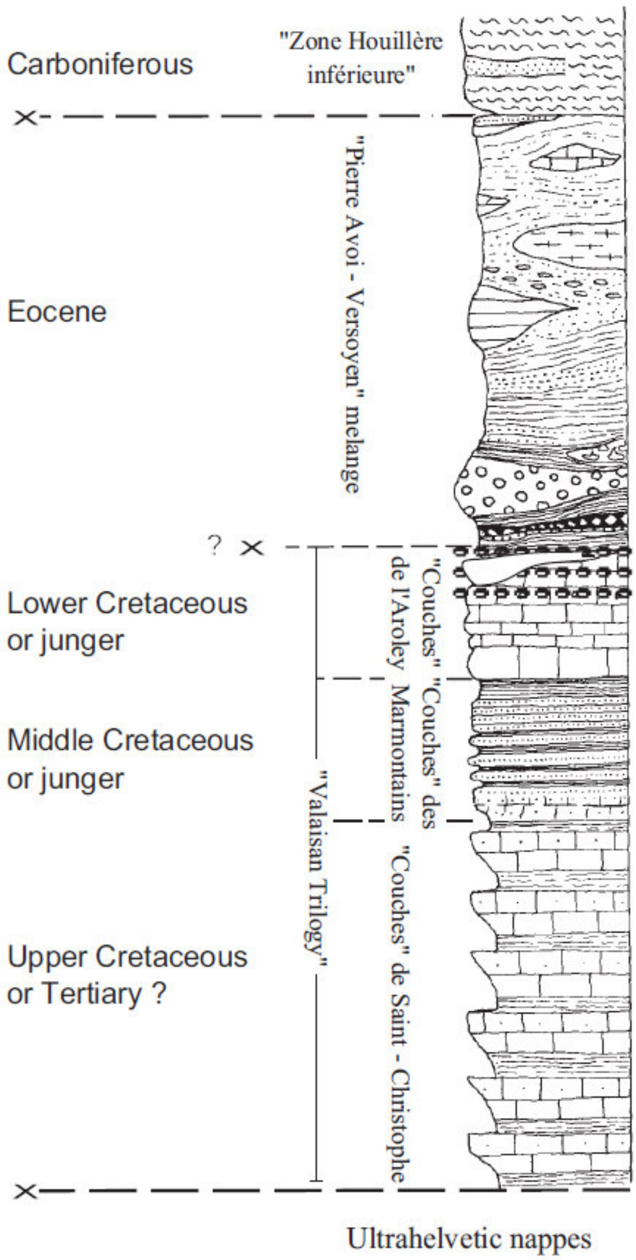


Fig. ft5-A - Synthetic Lithostratigraphy of the Sion-Courmayeur zone (modified from Jeanbourquin and Burri 1991).

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” orthoaugengneisses (ca. 25m), locally replaced by mylonitized bands, such as phyllonites; as described under 1.4.1. (fig. 34E).
(4): Sharp contact with two-mica and staurolitegarnet bearing paragneisses.

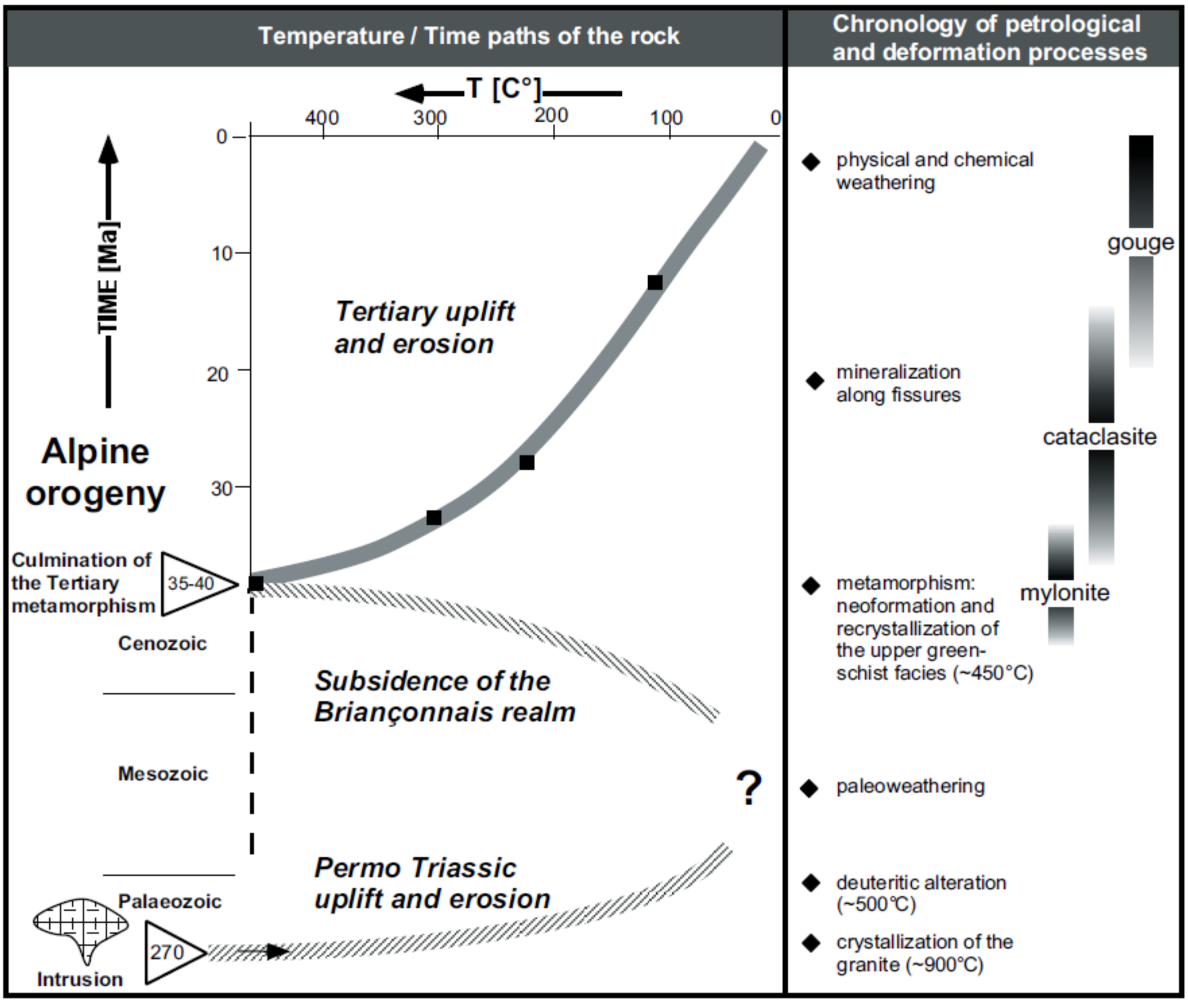


Fig. ft5-D - ***Thermal and petrological history of the Randa granite related to the main geological events and associated with the***
deformation processes (ductile to brittle). Slightly modified after Girod, 1999 and Thélin and Girod, 2000

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).

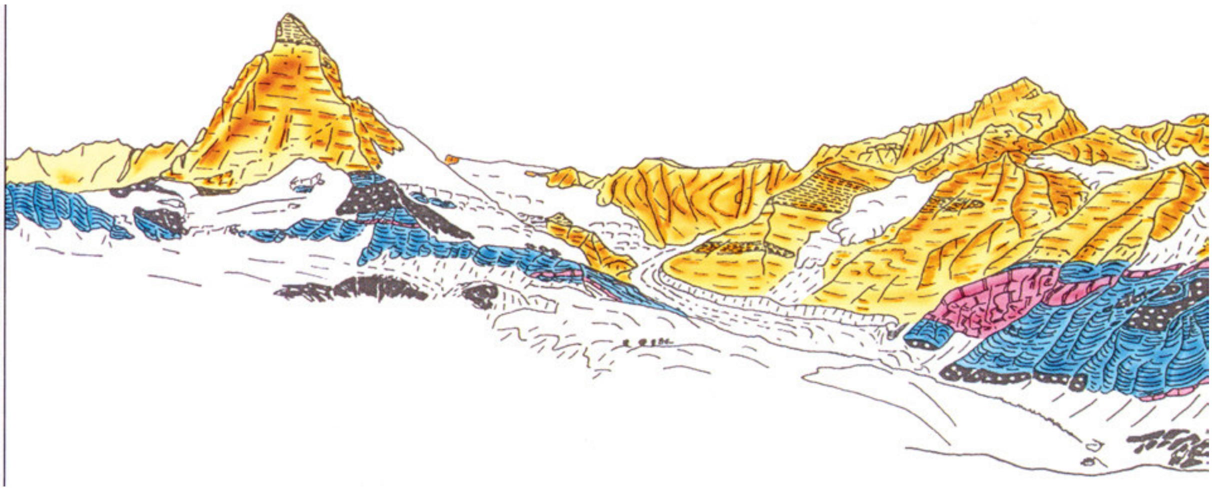
In fact, during spring 1991, several rock falls affected the sector of Randa: 1°) on April 18th, 20.106 m3, mainly of orthogneiss, collapsed at the Grossgufer location; 2°) on April 23rd, and on May 9th 10.106 m3, mainly of paragneiss, collapsed in the same sector. Each phase of rock fall was associated 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 Mattervispa river was stopped by the talus slope.

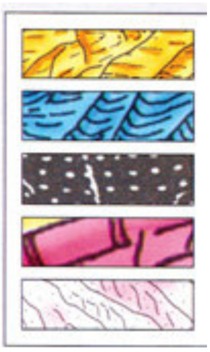
In order to avoid the rise of a lake that would have, in the long term, embedded the village of Randa, it was decided to dig a deviation channel. The case study of Randa shows the complexity in evaluating the triggering factors that have led to the collapse of the cliff. As shown by the “Centre de Recherche 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.

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.

Matterhorn Dent Blanche





Variscan basement Dent Blanche nappe Adriatic plate back-stop:

Sediments Tsaté nappe Accretionary prism

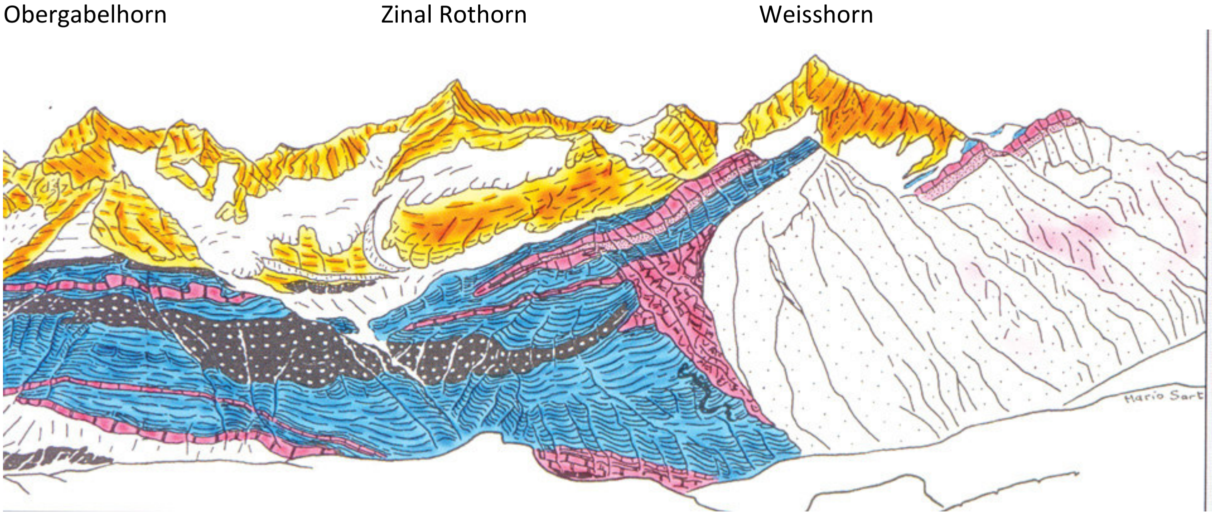
Ophiolites Tsaté nappe Accretionary prism

Mesozoic marine sediments of the Briançonnais margin(Siviez-Mischabel nappe)

Variscan basement of the Briançonnais: European (Iberic) plate

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).

Obergabelhorn Zinal Rothorn Weisshorn



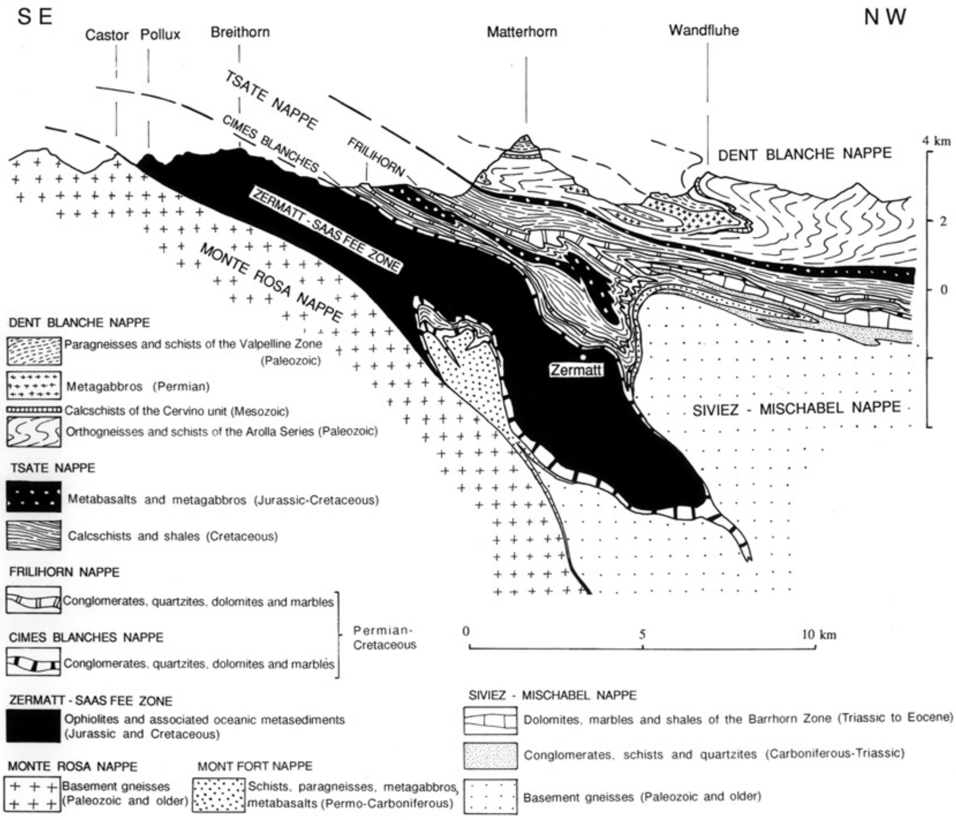


Fig. ft5-E2 - Simplified geologic section showing the relation between the Monte Rosa, Siviez-Mischabel, Zermatt-Saas Fee, Tsate and Dent Blanche nappes in the Zermatt region (Escher et al, 1997).

Aug. 16. Sunday: Back to Munich (550 km)