HKT-ISTP-2013 Himalayan Karakorum Tibet Workshop & International Symposium on Tibetan Plateau 2013

Field trip: Geological Highlights of the Western Alps

Bernd Lammerer, University of Munich

25. – 31. August 2013

Guide book

Panoramic view from Locanda del Diavolo (Les Druges - St. Marcel/Aosta) to the north

Participants

Bernd Lammerer (leader); Kornfeld, Daniela (technical assistance),

Yan, Maodu; Wang, Junbo; Cai, Yongen; Zhou, Shiyong; Sakai, Harutaka; Sakai, Rie; Yan, Bai; Wang, Yang; Spicer, Robert A.; Spicer, Teresa; Jessup, Micah; Zhou, Shiqiao; Rowley, David; Zhao, Junmeng; Ding, Lin; Cui, Liu; Zhou, Su; Gao (daughter of Cui Liu), Tianze; He, Wanfei;

Program

Sunday 25.8.2013 Route: Tübingen – Schaffhausen - Zürich – Flums (Tannenbodenalpe) – Flims
Program: Rheinfall of Schaffhausen, Helvetic nappes thrust over Molasse, Helvetic nappe stack of the Churfirsten. Total drive: 380 km. Accommodation: Flims

Monday 26.8.: From Flims by Cable car to the Cassonsgrat (2644m),
Program: Walk to see the Glarus Thrust and the Lochseiten-limestone, Flims landslide
Total walk: 5 h along gentle mountain trail. Total drive, 0 km. Accommodation: Flims

Tuesday 27.8.: Via Mala – San Bernardino - Bellinzona - Lugano - Morbio Superiore
Program: Penninic Gneiss nappes, Adula high pressure nappe, traverse through the Breggia gorge: Liasic to Cretaceous sequences with spectacular slump structures, Arzo quarries: breakup if the Carbonate-platform, crevasses in Triassic dolomite and implosion breccias with red clay matrix.
Total walk: 3 h. Total drive: 220 km. Accommodation: Lugano or Bellinzona or Locarno or Como

Wednesday 28. 8.: Finero - Cannobbio – Verbania – ins Aostatal – Valtournenche
Program: Insubric Linie**,** section through the Ivrea zone with peridotites and gabbros,
Total walk: : 4 h gentle mountain trail.Total drive 300 km. Accommodation ValTournenche

Thursday 29.8.: Ascent to the Lago di Cignana (2165 m),
Program: Ultra-high p-rocks with coesite and diamonds, nappes of the Zermatt -. Saas Fee zone, view to the Matterhorn. Total walk: whole days walk along mountain trail, total altitude difference:700 m up and down, Total drive**:** 0 km, Accommodation: ValTournenche

Friday 30.8.: Lac d´ Emosson über Mt.Blanc Tunnel - Chamonix
Program: Sesia Zone still high pressur rocks, antique Roman road, basement of the Mt. Blanc – Aiguille Rouge Massif and unconformity; Triassic quartzites with fossil tracks. Total walk: 5 h along mountain trails, total altitude difference:700 m up and down. Total drive**:** 170 km.
Accommodation: Martigny

Saturday: 31.8. Martigny – Tübingen
Program: Late Carboniferous Salvan-Dorenaz trough with fossil woods near Martigny
Total drive 440 km**:** back to Tübingen. **Arrival at Tübingen ~7.00 pm**

Accomodation:

Flims 25.-27.8.: Hotel Des Alpes CH-7018 Flims Waldhaus; Tel. +41 81 928 25 25
Fax +41 81 928 25 00; Coord: 32 T 522015.00 m E 5185903.00 m N

Locarno 27.-28.8.: Hotel du Lac; Via Ramogna 3; CH-6600 Locarno
Tel. +41 91 751 29 21; Coord: 484502.00 m E 5113009.00 m N

Valtournenche 28.-30.8.: Hotel Restaurant "AL Caminetto", Roma Street n°30
11028 Valtournenche (AO), Tel.e Fax 0039- 0166/92150
Coord . 393122.00 m E 5081200.00 m N

Martigny 28.-30.8.: Hotel Forclaz-Touring, Rue du Léman 15; CH-1920 Martigny
Tel. 0041-277222701, Coord: 351332.00 m E 5107453.00 m N

Field trip: Geological highlights of the Western Alps 25. – 31. August 2013
Bernd Lammerer University of Munich

Co-ordinates of the Stops in degree, minutes, seconds

1-1	Rheinfall Schaffhausen	N47 40 24.1 E8 37 11.6
1-2	Glarus thrust at Lochsite	N46 59 48.6 E9 05 37.2
1-3	Viewpoint Römerkastell	N47 07 17.4 E9 07 04.9
1-4	View Churfirsten/Tannenboden	N47 05 35.4 E916 45.7
Accommodation Flims-Waldhaus		
2-1	Viamala gorge and Buendnerschists	N46 39 47.5 E9 26 50.7
2-2	Zillis Church from the Year 1109	N46 38 03.6 E9 26 29.4
2-3	back to Flims and up Cassonsgrat	N46 52 42.0 E9 15 56.1
2-4	Cassons – flysch with nummulites	N46 53 01.5 E9 15 30.2
Accommodation Flims-Waldhaus		
3-1	View Suretta fold nappe hinge at Andeer	N46 36 54.1 E9 25 46.1
3-2	Suretta basement (Rofna porphyry)	N46 35 24.9 E9 25 26.6
3-3	Breggia gorge start (Morbio superiore)	N45 51 38.4 E9 01 17.3
3-4	Breggia gorge end	N45 51 02.7 E9 00 42.4
3-5	Arzo quarry	N45 52 53.2 E8 57 02.5
3-6	Verzasca valley Ponte di Salto	N46 15 35.9 E8 50 09.4
Accommodation Locarno		
4-1	Ivrea zone peridotite at cemetary	N46 06 29.8 E8 32 01.3
4-2	Finero Geo trail start	N46 06 29.1 E8 32 04.9
4-3	Finero Geo trail end	N46 06 17.0 E8 32 40.4
4-4	Roman road – Sesia high p zone	N45 36 08.0 E7 45 41.5
Accommodation Valtournanche		
5-1	Lago Cignana UHP - trail start	N45 54 30.8 E7 36 57.3
5-2	Finestra Di Cignana (2445) highest point	N45 53 25.8 E7 35 49.5
5-3	manganese schists with diamonds	N45 52 39.5 E7 35 34.0
5-4	eclogite with coesite	N45 52 42.1 E7 35 35.1
Accommodation Valtournanche		
6-1	Tunnel Mt. Blanc entrance	N45 49 04.7 E6 57 06.9
6-2	Lac D´Emosson Parking	N46 04 06.3 E6 56 12.8
6-3	start trail vieille Emosson	N46 03 55.0 E6 55 22.1
6-4	Post-Variscan unconformity	N46 03 13.8 E6 53 29.3
6-5	dinosaur tracks in quartzite	N46 02 54.4 E6 53 04.4
Accommodation Martigny		
7-1	Carboniferous of Dorenaz	N46 08 39.1 E7 02 44.5
7-2	View Dent de Morcles	N46 10 00.9 E7 02 04.7
Back to Tuebingen		

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.

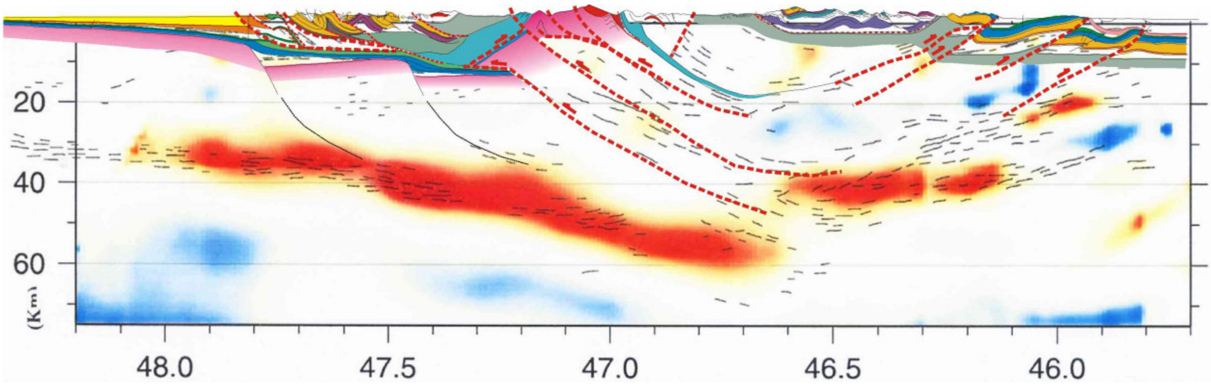


Fig. 1: The deep structures of the Eastern Alps. The red band at 30 – 60 km Depth reflects the lower crust, the pink band in shallow depth marks the top of the European basement and its uplift within the Tauern window.

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).

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

Stefan M. Schmid, **Geologisch-Paläontologisches Institut, Bernoullistr. 32, 4056 Basel**

(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.

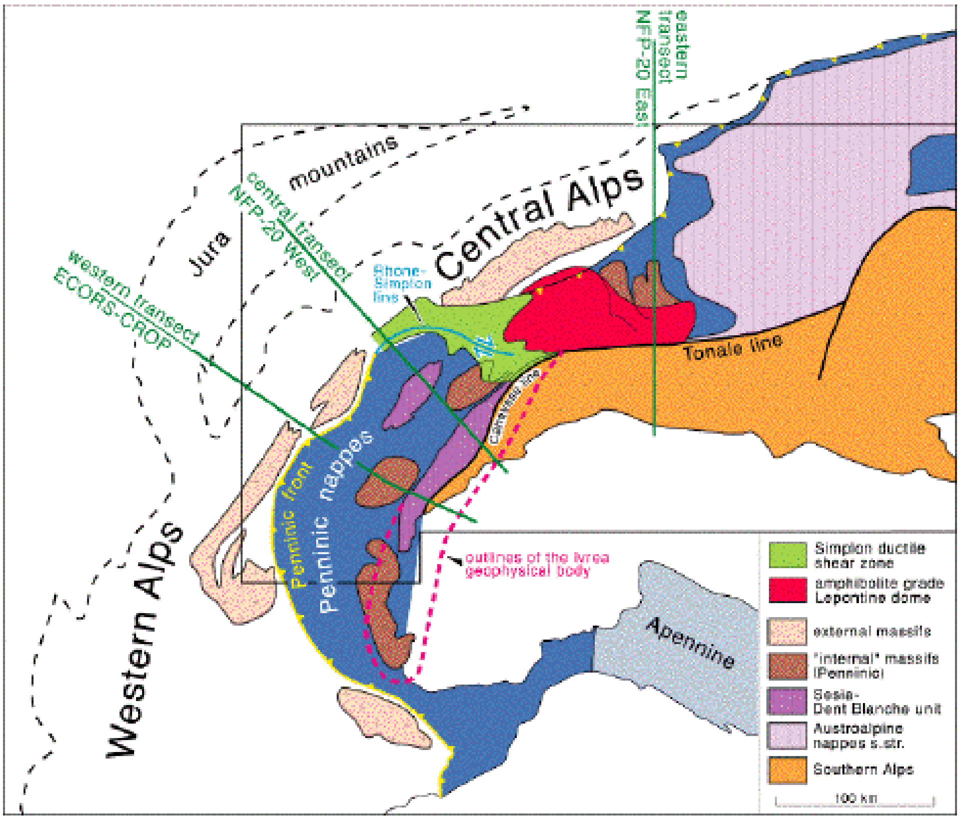


Fig. 1: Sketch map of the Alps, indicating locations of the three geophysical-geological transects depicted in Figure 3. From Schmid and Kissling (2000).

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).

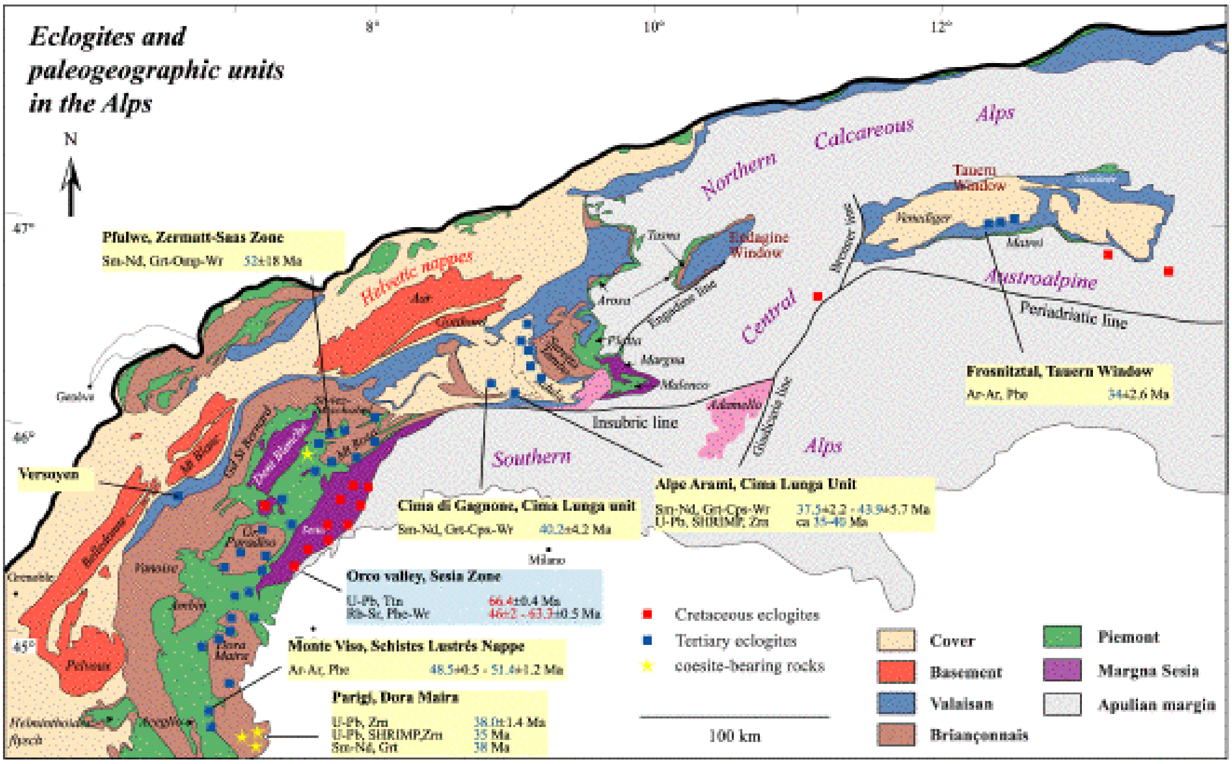


Fig. 2: Paleogeographic map of the Alps, indicating the present-day position of the major paleogeographic units of the Alps. Note that this is not a "tectonic" map in the traditional sense. After Froitzheim et al. (1996).

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).

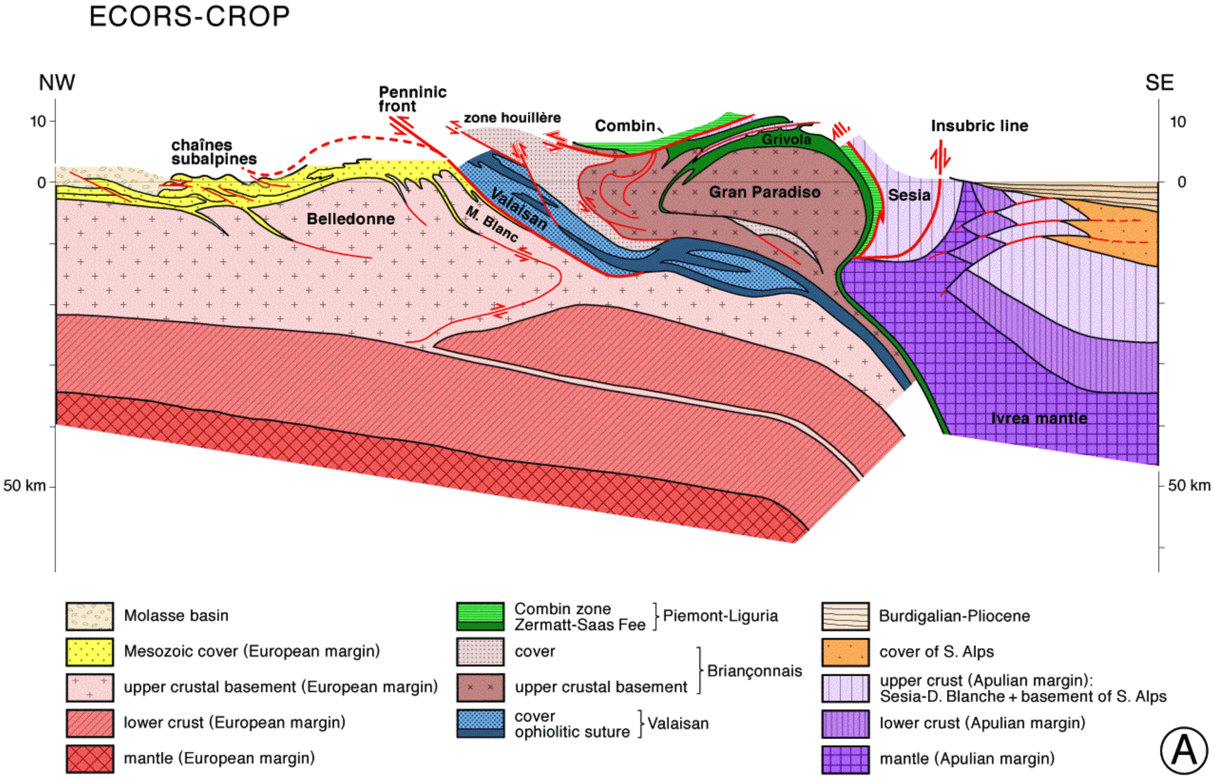

ECORS-CROP
NW
SE
10
0
50 km
chaînes subalpines
Penninic front
zone houillère
Combin
Grivola
Insubric line
Belledonne
M. Blanc
Valaisan
Gran Paradiso
Sesia
Ivrea mantle
Molasse basin
Mesozoic cover (European margin)
upper crustal basement (European margin)
lower crust (European margin)
mantle (European margin)
Combin zone
Zermatt-Saas Fee
Piemont-Liguria
cover
upper crustal basement
Briançonnais
cover
ophiolitic suture
Valaisan
Burdigalian-Pliocene
cover of S. Alps
upper crust (Apulian margin): Sesia-D. Blanche + basement of S. Alps
lower crust (Apulian margin)
mantle (Apulian margin)
A

Fig. 3a

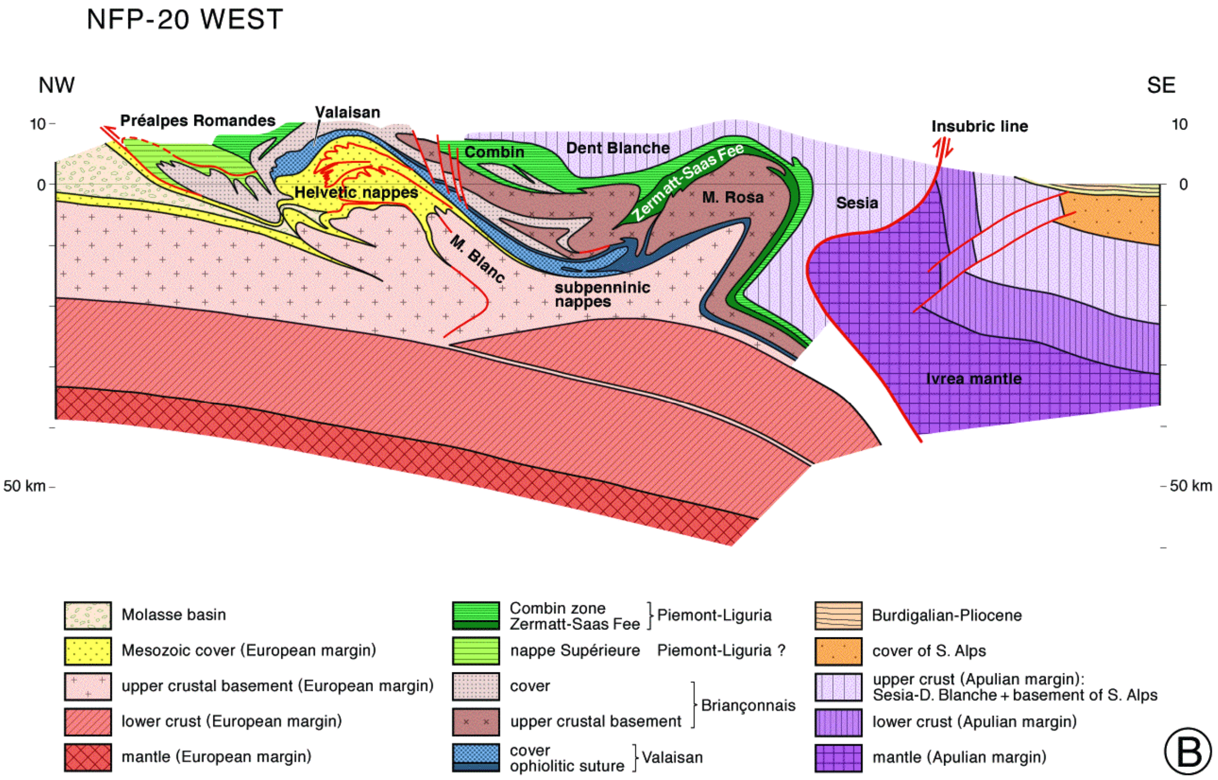

NFP-20 WEST
NW
SE
10
0
50 km
Préalpes Romandes
Valaisan
Combin
Dent Blanche
Zermatt-Saas Fee
Insubric line
Helvetic nappes
M. Rosa
Sesia
M. Blanc
subpenninic nappes
Ivrea mantle
Molasse basin
Mesozoic cover (European margin)
upper crustal basement (European margin)
lower crust (European margin)
mantle (European margin)
Combin zone
Zermatt-Saas Fee
Piemont-Liguria
nappe Supérieure
Piemont-Liguria ?
cover
upper crustal basement
Briançonnais
cover
ophiolitic suture
Valaisan
Burdigalian-Pliocene
cover of S. Alps
upper crust (Apulian margin): Sesia-D. Blanche + basement of S. Alps
lower crust (Apulian margin)
mantle (Apulian margin)
B

Fig. 3b

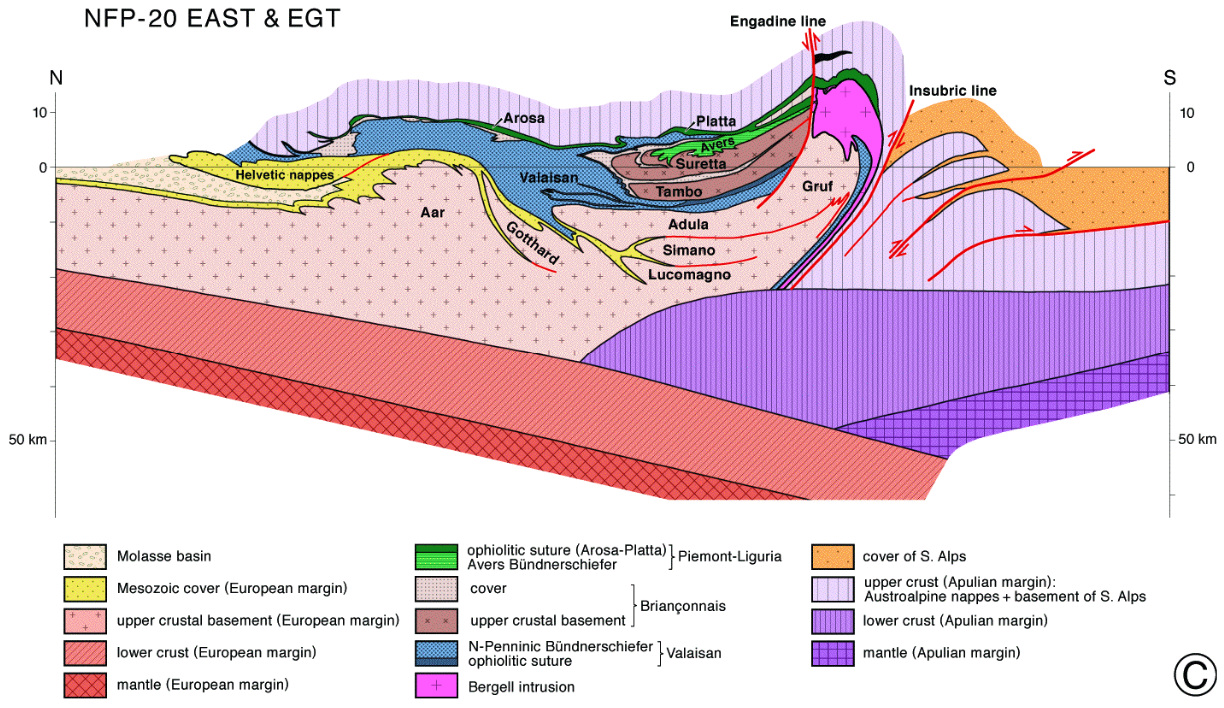


Fig. 3c

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.

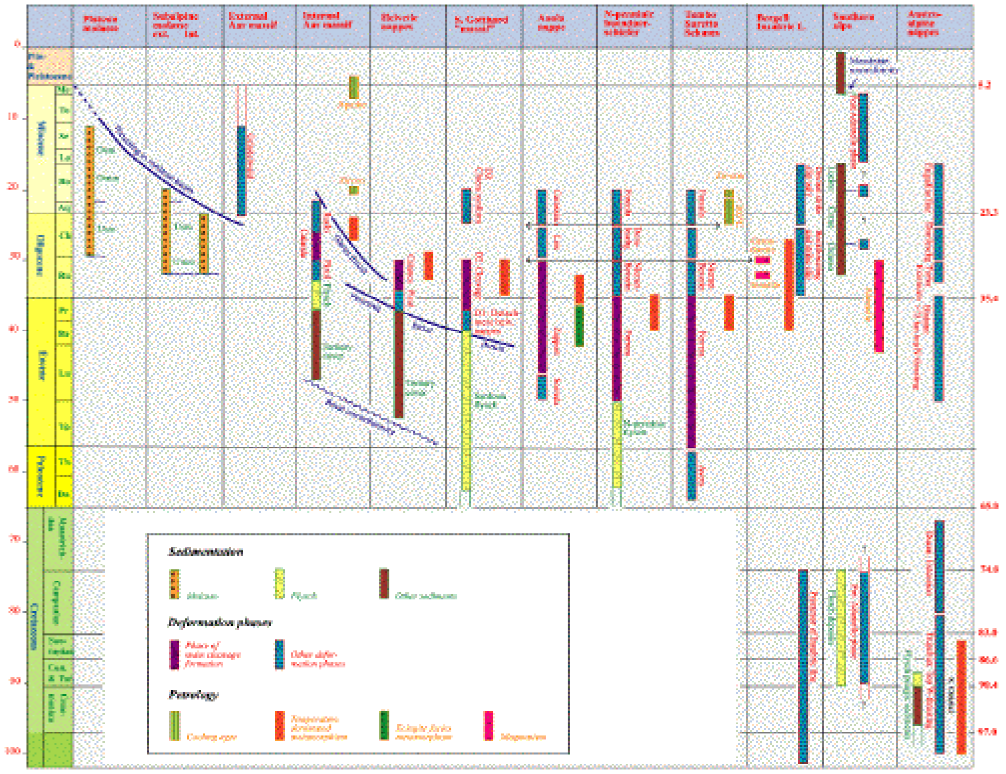


Fig. 4: Correlation table showing an attempt to date deformation phases and metamorphism along the eastern transect (NPF-20 East). For an extensive discussion of timing constraints see Schmid et al. (1997), for a brief overview, see Schmid et al. (1996). From Schmid et al. (1996).

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 Createceous ("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.

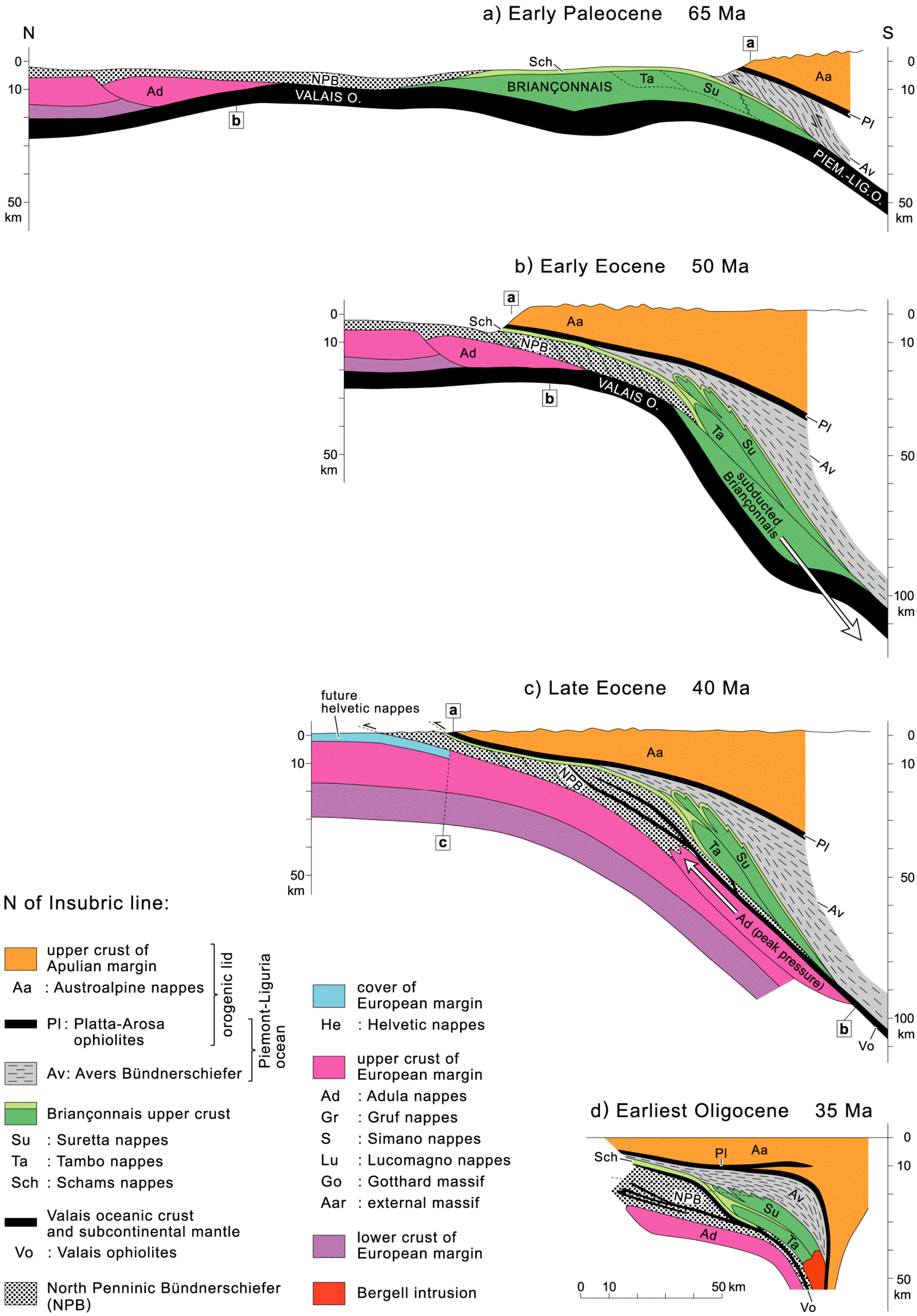

a) Early Paleocene 65 Ma
N
S
Sch
NPB
Ad
VALAIS O.
BRIANÇONNAIS
Ta
Su
Aa
Pl
Av
PIEM.-LIG.O.
a
b
0
10
50
km
b) Early Eocene 50 Ma
subducted Briançonnais
100
km
c) Late Eocene 40 Ma
future helvetic nappes
c
Ad (peak pressure)
Vo
d) Earliest Oligocene 35 Ma
0 10 50 km
N of Insubric line:
upper crust of Apulian margin
Aa : Austroalpine nappes
Pl : Platta-Arosa ophiolites
Av: Avers Bündnerschiefer
orogenic lid
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).

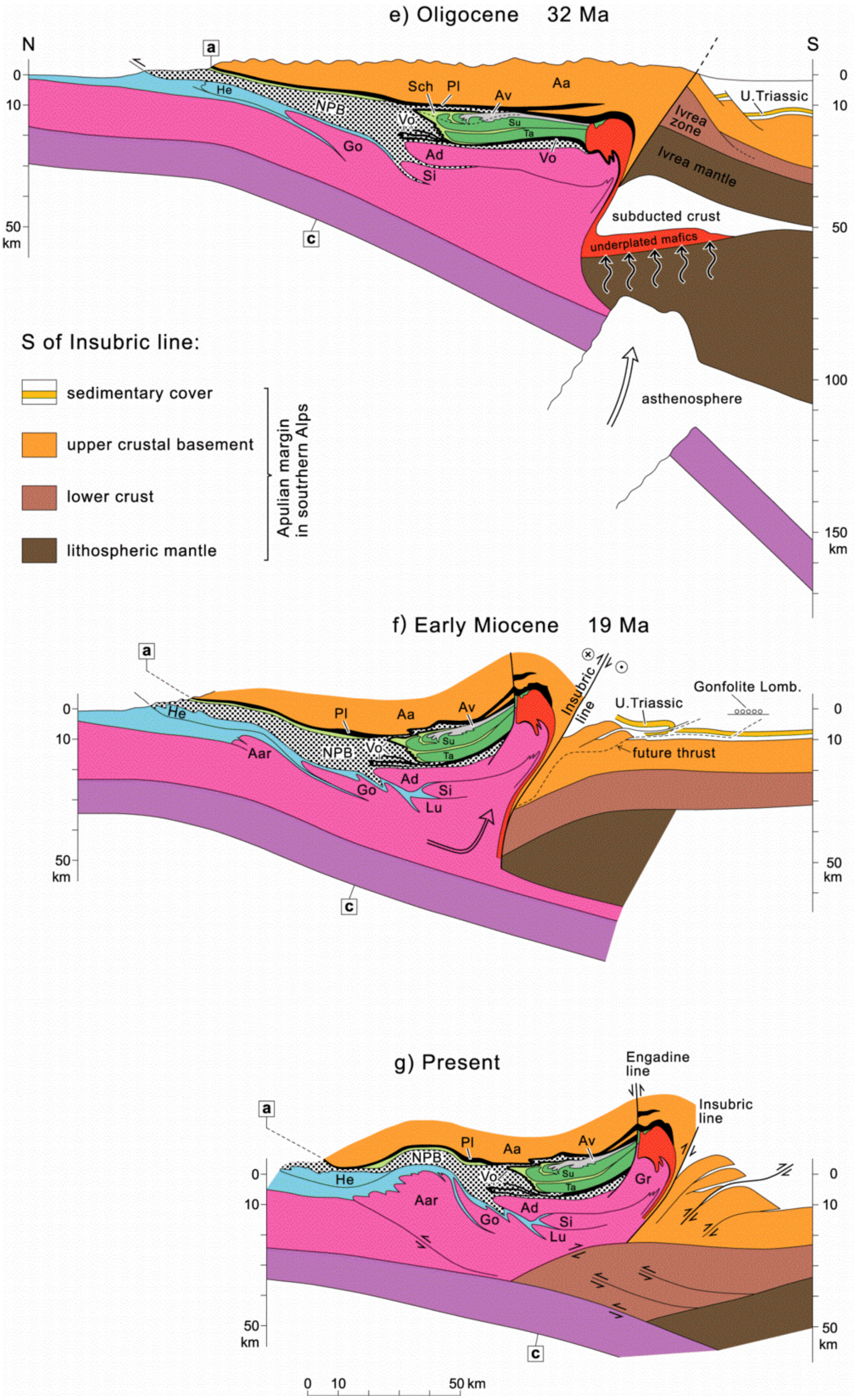


Fig. 5: Scaled and area balanced sketches of the kinematic evolution of the eastern Central Alps from early Tertiary convergence (a-b) to collision (c) and post-collisional shortening (d-g). From Schmid et al. (1996)

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.

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).

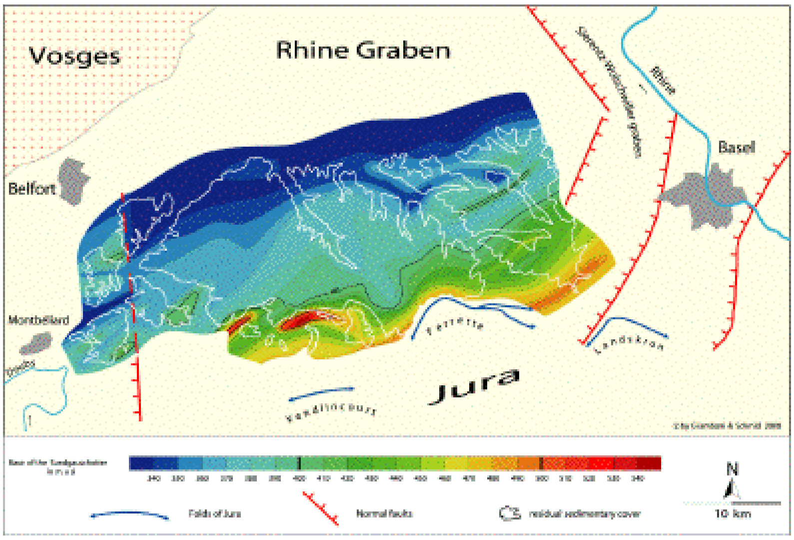


Fig. 6: Contour map of the base pf the Pliocene (3.2-2.6 Ma) Sundgau gravel deposits W of Basel. Assuming an originally planar base of these braided river deposits, the contours indicate folding with an amplitude of up to 200m. This indicates substencial tectonic movements during the last 3 Ma, post-dating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

References

Beaumont, C., Fullsack, P.H. and Hamilton, J., 1994. Styles of crustal deformation in compressional orogens caused by subduction of the underlying lithosphere. Tectonophysics, 232, 119-132.

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.

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
Field trip itinerary

Sunday 25.8.2013
Route: Tübingen – Schaffhausen - Zürich – Flums (Tannenbodenalpe) – Flims
Program: Rheinfall of Schaffhausen, Helvetic nappes thrust over Molasse, Helvetic nappe stack of the
Churfirsten. Total drive: 380 km. Accommodation: Flims

Stop 1: Rheinfalls of Schaffhausen

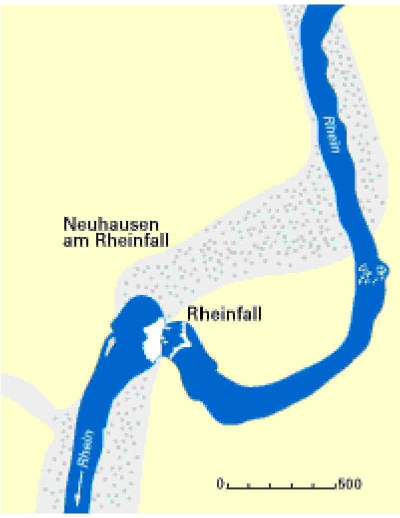
Formation of the Rhine Falls:

1. Original course of the Rhine

Up to the beginning of the Riss glacial period (approx. 200'000 years ago) the Rhine River flowed from Schaffhausen in a westerly direction through the countryside of Klettgau. This former Rhine valley was then filled up again with alpine debris.

2. First change of the course of the Rhine

During the Riss glacial period (approx. 120'000 years ago) the Rhine by Schaffhausen was diverted southwards and formed the Rhine Riss period gorge. Today the basin outlet below the falls is this actual channel, which was filled up again debris.



3. Second change of the course of the Rhine river and formation of the Rhine Falls

During the last glacial period, the so-called Würm glacial period, the Rhine was forced to change its route and formed a wide bow towards the south till it reached its present day bed of hard Malm limestone, above the falls. About 15'000 years ago the Rhine Falls were formed when the hard 'Malm' limestone at the top changed to the easily eroding debris of the Riss period channel. The Rhine Fall Rocks form the remains of the original steep sloping limestone flanks.**Source:** Franz Hofmann, Geologie und Entstehungsgeschichte des Rheinfalls, in: Neujahrsblatt der Naturforschenden Gesellschaft Schaffhausen Nr. 39/1987. Illustration: Typographic Schaffhauser Nachrichten

Stop 2: Glarus thrust near Schwanden at Lochsite: The Glarus thrust (Swiss Tectonic Arena

Sardona) 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 calledTschingelhörnerbetween Elm and Flims (in the same cliff is a natural hole called the Martinsloch).

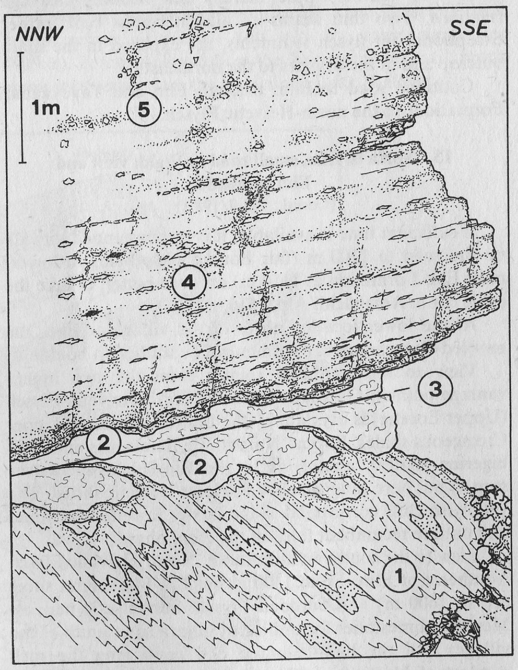


Fig. 6. Section at the Lochseite near Schwanden (GL). Essentially after Alb. Heim (1919) and St. Schmid (1974).

Fig. 6. Coupe de la Lochseite, près de Schwanden (GL). Essentiellement selon Alb. Heim (1919) et St. Schmid (1974).

1 = Eocene Flysch
2 = Lochseiten limestone (limestone mylonite)
3 = Fault gouge
4 = Green «Verrucano» conglomerates with strongly flattened components
5 = Purple «Verrucano» conglomerates with less deformed components

Stop 3 Kerenzerberg pass (or in the village of Filzbach): view of the Alpine front (Coord.: 726.8/220.1) Hotel Römercastel.

General Situation of the Walensee and Seez valleys (stops 17 to 21)

The Helvetic nappes plunge to the north, on the southern flank of the Amden-Wildhaus synform (continuation of the Obersee synform). The gentle southern slopes are essentially dip-slopes, whereas the steep walls of the Churfirsten-Alvier ränge to the north offer a magnificent cross- section of the nappes. The Walensee valley runs W-E, very obliquely to the strike; the Seez valley, with its NW-SE direction, is more nearly perpendicular to the strike. A branch of the Rhine glacier flowed through this valley, at least during the last two glaciations.

The following nappes lie above the Subalpine Molasse:

1. Wägeten slice, only in the town of Weesen.

2. Glarus nappe, Verrucano to Middle Jurassic, only south of the lake.

3. Mürtschen nappe. South of the lake: Verrucano to Middle Jurassic; north of the lake: Upper Jurassic, Cretaceous (relatively thin

and mainly calcareous), Eocene. A small slice near Weesen may belong to the front of Mürtschen nappe.

4. Axen (or Gonzen) nappe. South of the lake: Verrucano, Triassic, Lower Jurassic (the separations between nappes 2-4, at the level of the roof of the Verrucano, are only quite shallow). North of the Seez valley: Jurassic, basal Cretaceous shales. In contrast to the Situation west of the river Linth, the Axen nappe does here not comprise Cretaceous formations; these have been stripped away and now lie further north, in the folds of the Säntis range. The seemingly stratigraphic succession of Axen Jurassic and Säntis Cretaceous is a case of cover Substitution: the Cretaceous rocks are of much more southerly origin than the underlying Jurassic ones.

5. Säntis nappe: only Cretaceous, thickening southwards; upper part of the Churfirsten-Alvier range and a klippe southwest of Filzbach. Here, the nappe is not subdivided into three digitations as it is in the Schwyz Alps; the Cretaceous of Alvier is equivalent to that of the Räderten and Drusberg digitations. The Churfirsten and Alvier group has been described in a stupendous monograph by Amold Heim (1910-1917; maps 1907 and 1917). See also scale projections in R. Helbling (1938).

Viewpoint from Hotel Römercastel, Filzbach: view of the Alpine front (Coord.: 726.8/220.1)

Upper Oligocene conglomerates of Schäniserberg and Speer (1950 m, highest Molasse mountain of Switzerland). Jurassic limestones and Taveyannaz sandstones of the north-Helvetic Wägeten slice form the hill of Chapfenberg, in the little town of Weesen. Quarries to the right of the town are in Cretaceous and Eocene limestones of a small anticline, possibly a frontal fold of the Mürtschen nappe. It is overthrust by the front of the Säntis nappe: Kieselkalk (large quarry), the well-bedded Drusberg Formation and the massive cliffs of the Schrattenkalk (scar of a recent, 1974, rockfall). The broad synform of Amden is in Upper Cretaceous rocks, directly overlain by Penninic Flysch. The limestone mountain of Mattstock, due north, is not directly connected to the north flank of the Amden synform. The Amden syncline corresponds to the Obersee syncline, but it is offset for about 2 km to the north, by a left-lateral tear-fault in the lower Linth valley.

Formerly, the river Linth flowed direetly into Lake Zürich, and its alluvions caused a rise of the level of Walensee; the whole area was swampland. Under the direction of Hans-Conrad Escher, the Linth was deviated into Walensee (1811), where it has since built a small delta; somewhat later, the course of the river was straightened between Walensee and Lake Zürich. In the Sediments of Walensee (i.a. turbidites), the 1811 event is well marked.

Stop 4: Tannenbodenalp: view of the Churfirsten range (Coord.: 739.8/217.3)

Large parking lot in front of the cable-car Station.

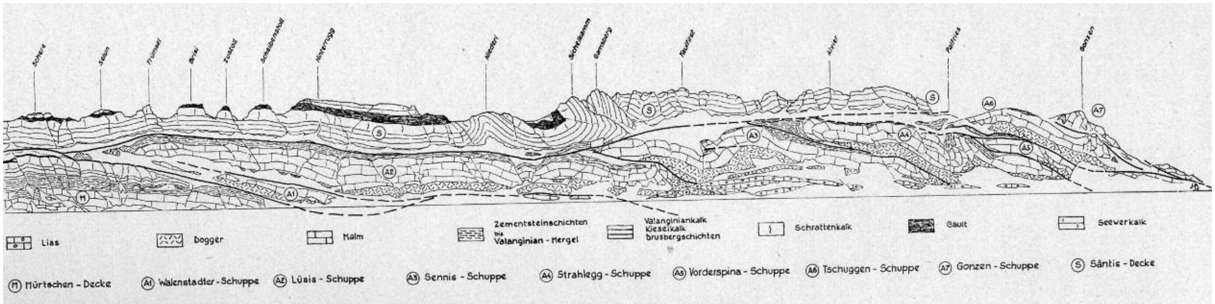
On the eastern side of the parking lot, the boundary between the red mudstones (slates) of the Upper Verrucano and the quartzitic Mels sandstones (Lower Tnassic or lower Middle Triassic) is well exposed. Note the bleaching of the topmost Verrucano slates. Walk about 400 m to the southeast, for the splendid view of the Churfirsten and Alvier ranges (Fig. 9; section in R. Trümpy 1969).

To the left, the Mürtschen nappe, where the calcareous Cretaceous formations have been deformed together with their Jurassic substratum m the Quinten antiform. Above Walenstadt, successively older formations of the Mürtschen nappe are truncated by the Axen thrust towards the south-east. The tectonic style of the Axen nappe is here very different from that of the type locality, but comparable to the structures north of Urnerboden (p.215). Its dark- coloured Lower Jurassic limestones crop out in the low hills near the valley bottom; their folds do not correspond to those of the Upper Jurassic limestones. The Quinten limestones, with a thin-bedded member in their middle part, are underlain by the yellowish shales of the Schilt Formation and overlain by the dark shale/limestone sequence of the Zementstein Formation. Their seven slices, with incomplete anticlinal and synclinal links, are well visible (the limestones in the hamlet of Tscherlach belong to the Walenstadt slice rather than to the Mürtschen nappe as suggested in Fig. 9). The Cretaceous formations on the crest of Churfirsten and Alvier are not afiected by the structures of the Axen nappe. The beautifui synclinal fold on Sichelkamm might correspond to that of Brünnelistock (stop 15).

Note that the relative displacement between the Jurassic (Axen) structural level and the Cretaceous (Säntis- Churfirsten) one decreases to southeast; in the little mountain of Fläscherberg, beyond the Rhine, the contact is almost stratigraphic. The facies change within the Churfirsten and Alvier chain is equally spectacular: thickening of the Palfris-Vitznau shales and especially of the Kieselkalk to the southeast, replacement of the thick-bedded, shallow-water Betlis limestones by the thin-bedded, hemipelagic Diphyoides limestones, encroaching of shale interbeds into the massive Schrattenkalk. The view encompasses part of the proximal European margin of the Lower Cretaceous Tethys.

Return to Flums and follow the road to Sargans. Platy sandstones of the uppermost Verrucano are quarried on the hill of Tiergarten. Between Mels and Sargans, the low watershed to the Rhine valley is crossed.

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)

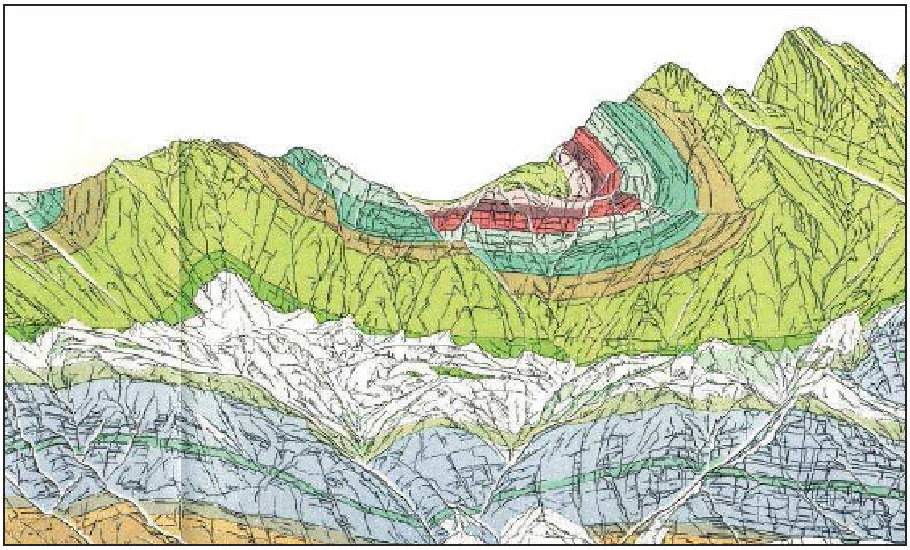


Abbildung 23: Geologische Interpretation der Sichelkammfalte: rostorange = Bommerstein-Formation mit Molser-Member, braun = Reischiben-Formation, blauviolett = Quinten-Formation, olivgrün = Zementstein-Formation, weisslichgrün = Palfries-Formation, grün = Betlis-Formation, senfgelb = Kieselkalk-Formation,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

Monday 26.8.:
From Flims by Cable car to the Cassonsgrat (2644m),
Program: Walk to see the Glarus Thrust and the Lochseiten-limestone, Flims landslide
Total walk: 5 h along gentle mountain trail. Total drive, 0 km. Accommodation: Flims

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 di 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

Bas den Brok, Oliver Jagoutz (2000): The Glarus thrust an erosional unconformity? An old idea of Otto Ampferer revisited. **Abstract for 17th Swiss Tectonics Studies Group Meeting Mar. 31 - April 1 2000** (Geologisches Institut ETH, Sonneggstrasse 5, 8092 Zürich)

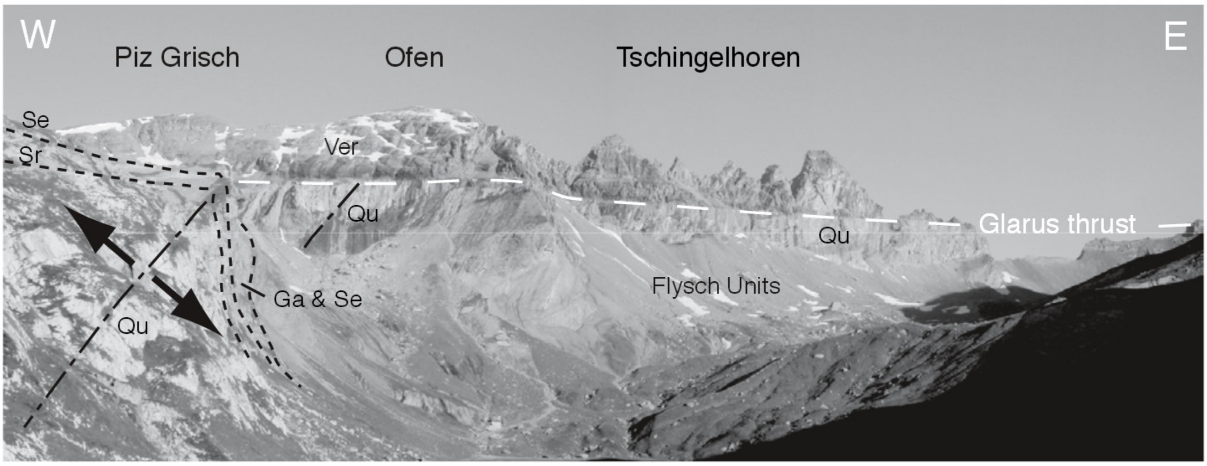
The well-known Glarus thrust at the base of the Helvetic nappes in Eastern Switzerland is a very remarkable structure. The thrust is only 20 to 200 cm wide, yet cuts as a single plane (straight, smooth, extremely sharply cut) over an area of probably more than 40 km through intensely folded and foliated Infrahelvetic rocks. It is difficult to imagine how this could have happened. How can such a single plane develop in mechanically very anistropic rocks such as the Infrahelvetics must have been at the time that the thrust developed? One could imagine that the thrust developed by ductile shear localisation within a large-scale ductile shear zone, but the Infrahelvetic structures (the folds and foliation) do not show largescale bending inwards towards parallellity with the thrust plane. Such a bend is only apparent locally within the upper meter of the Infrahelvetic rocks. At a larger scale, the Infrahelvetic foliation does not appear to be affected by the thrust. At some localities the foliation even bends into the wrong direction. How then could this spectacular thrust have developed?

According to Otto Ampferer (1934) the Glarus thrust originally was an erosional unconformity. According to him, the Infrahelvetic foot wall was already eroded before the Helvetic nappe complex was thrusted onto it ("Reliefüberschiebung"). This idea was immediately rejected by Jakob Oberholzer and Albert Heim (1934) and has not received much attention anymore since then. However, at least at

first sight, the idea of an erosional unconformity would provide a nice explanation for the sharply cut nature of the Glarus thrust.
What if Otto Ampferer *was* right? Uplift and a major erosional event must have taken place *after* Calanda phase thrusting, folding and foliation development, and *before* the entire Helvetic Glarus nappe complex (including the already Calanda-phase folded and foliated Glarus s.s, Mürtschen, and Säntis nappes ...) was thrusted over the Infrahelvetic basement. When could this have happened? Is there time for uplift, erosion and a second major thrusting phase?

The youngest sediments overthrusted by the Glarus nappe complex are of Acquitanian age (~22-23 Ma, Lower Freshwater Molasse). Within the Molasse basin, major, basin-wide Burdigalian age erosional events are reported to have occurred around 20 and 18 Ma. Enhanced subsidence of the Molasse basin has taken place between 17 and 14 Ma indicative of renewed tectonic activity, immediately followed by the deposition of the massive conglomeratic deposits of the Upper Freshwater Molasse (~14-11 Ma). These events are consistent with an Upper Oligocene Calanda phase deformational event, an Early-Miocene erosional event, and a Middle Miocene (Helvetian) thrusting event during which the entire Glarus nappe compex could have thrusted over an eroded Infrahelvetic basement. It seems that a possible erosional origin of the Glarus thrust deserves further attention to see, for example, whether available radiometric age data and the different metamorphic events are in agreement with it, and to think of how we could proove whether or not Ampferer was right.

Ampferer, O. (1934) Über die Gleitformung der Glarneralpen, *Sitzungsberichte der Akademie der Wissenschaften in Wien, Mathem.-naturw. Klasse, Abteilung I, Band 143,* S. 109-121.
Oberholzer, J., Heim, Alb. (1934) Zu Otto Ampferers „Gleitformung der Glarneralpen“ und „Flimser Bergsturz“, *Eclogae. geol. Helv. 27,* 507-516.



Tuesday 27.8.: Via Mala – San Bernardino - Bellinzona - Lugano - Morbio Superiore
Program: Penninic Gneiss nappes, Adula high pressure nappe, traverse through the Breggia gorge: Liasic to Cretaceous sequences with spectacular slump structures, Arzo quarries: breakup if the Carbonate-platform, crevasses in Triassic dolomite and implosion breccias with red clay matrix. Total walk: 3 h. Total drive: 220 km. Accommodation: Lugano or Bellinzona or Locarno or Como

Fig.: Scetch of J.W. Goethe

Stop 1: 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.

We continue to the world heritage of Ticino and Monte Generoso.

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),

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).

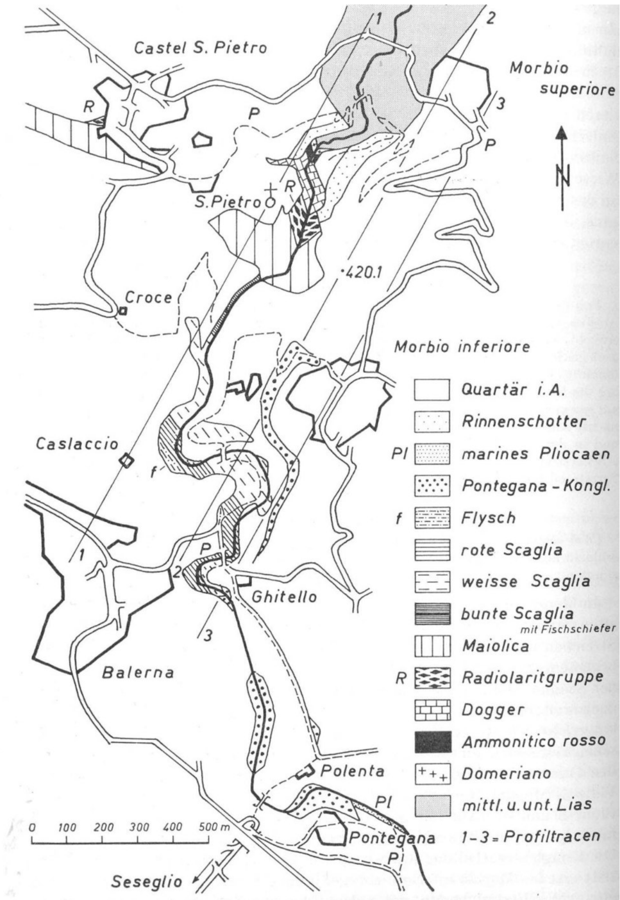
These sediments, known as "Grenzbitumenzone", contain up to 40% organic carbon. This horizon contains a rich fauna of vertebrates with abundant reptiles and fish. A thick succession of dolomites and limestones was formed under Sabkha-type conditions during the Late Triassic (Dolomia Principale, Calcare 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.



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

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 Grge: layers extremely rich in fossils, remains of underwater landslides, witness to climatic changes, and proof of volcanic eruptions.

Stop 3: Mendrisio-Breggia valley-Chiasso

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

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.

Guide books

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

Compiled by Helmut Weissert Dept. of Earth Sciences, ETH-Z CH-8092 Zürich, Switzerland

The Stops (By D. BERNOULLI)

Study of the Early Lias basin-fill of the Generoso basin and of the Jurassic-Cretaceous pelagic sequence of the Lombardian Alps in the Breggia valley. Basal part of the Oligo-Miocene clastics (south-Alpine "molasse") near Chiasso.

From Mendrisio along the flexure of Mendrisio which separates the folded and thrust Lower Lias siliceous limestones of Monte Generoso to the north from moraine-covered Cretaceous marls and flysch of the gentle hills to the southwest to Morbio Superiore and into the Breggia valley.

Large submarine slump in the Lombardian Siliceous Limestones (Along road from Caneggio to Bruzella, coord.: 723.55/081.65)

The Lombardian Siliceous Limestones are a sequence of dark grey, well-bedded, spongolithic siliceous calcilutites with bands and nodules of replacement chert, of Hettangian to Early Pliensbachian age. In the Generoso basin they are up to 4000 m thick with a large fraction transported by gravity flow related to contemporaneous down-faulting of the basin along the Early Jurassic Lugano fault (cf. Fig. 37 in Introduction, Part A). There are many submarine slumps, pebbly mudstones and olistholites comprising older and penecontemporaneously displaced shallow-water lithologies, often associated with graded caicirudites and calcarenites. A large rotated Stack of strata is well exposed along the roadside, with soft-sediment deformation on a smaller scale in the associated spongolithic limestones below and above.

Stop 3: Back to Morbio Superiore (722.9/080.0); walk down to the Breggia river.
Lower to Middle Jurassic pelagic Sediments of the Lombardian zone (Coord.: 722.50/079.95)

For the general Situation see Figure 26, for detailed description see Figure 27. In the Middle to Late Lias, with the opening of the oceanic Tethys, rifting, subsidence and sedimentation rates diminished. Submergence of the increasingly starved distal continental margin resulted in the deposition of a pelagic sequence whose facies was mainly determined by ongoing synsedimentary faulting and increasing water depth, reflected in the solution facies of the pelagic carbonates. The uppermost Lombardian Siliceous Limestones are light grey hemipelagic limestones with bands and nodules of granulär chert, preferentially replacing burrows, and interbedded silty micaceous maris (Molino Member). The overlying Morbio Formation (Domerian p.p. and lowermost Toarcian) consists of an alternation of pelagic limestones and marls, the limestones becoming more nodular to conglomeratic and burrowed up-section.

Both, the Morbio Formation and the overlying nodular marls and marly limestones of the ROSSO Ammonitico Lombarde (Toarcian p.p.) contain numerous ammonites whose aragonitic tests were only dissolved after burial. In the overlying limestones and maris with pelagic bivalves (Upper Toarcian-Lower Bajocian) ammonites become rare up-section and finally occur only in redeposited strata (*in Fig. 27); minor submarine slumps (S! in Fig. 27), associated with graded skeletal calcarenites, exclusively contain pelagic limestone clasts and planktonic organisms indicating submergence of the source areas. Burrows are of Chondrites- and Zoophycos-type. Part of the section is repeated in a large, approximately 10 m thick submarine slump overiain by a graded bed levelling the irregularities along the top of the slump. Transport was to the southeast according to fold axes. There are some additional minor slumps in the overlying pelagic and turbiditic siliceous limestones (Middle Jurassic p.p.).

For continuation of section walk back to Morbio Superiore and continue by bus to cement factory a few hundred meters down-stream.

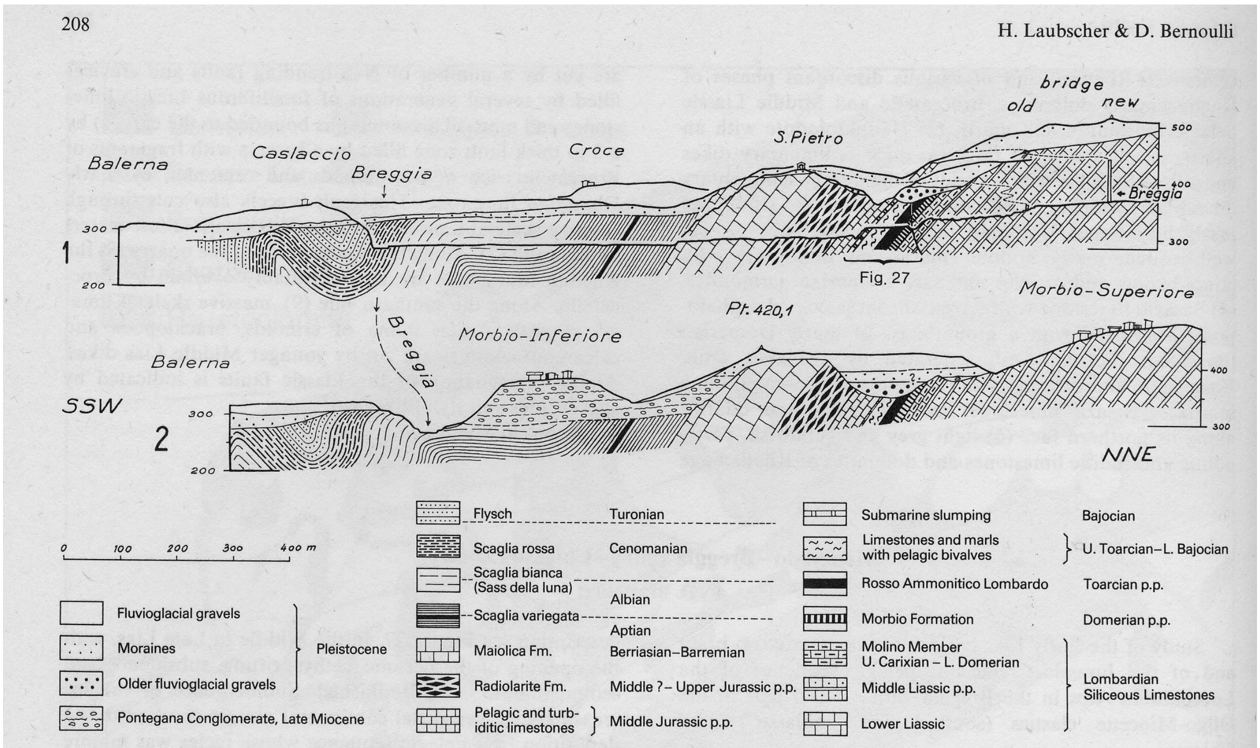


Fig 26 Geologie profile through the Mesozoic sequence ofthe Breggia valley (after L. Vonderschmitt 1940, modified).

Middle Jurassic to Cenomanian pelagic sediments and Cenomanian-Turonian flysch of Lombardian zone (Quarry SACEBA, coord.: 722.3/079.5)

In the Breggia gorge, a continuous section from the Middle Jurassic pelagic and turbiditic limestones to the base of the Upper Cretaceous flysch is exposed (Fig. 26).
The Middle Jurassic partly turbiditic pelagic limestones contain pelagic bivalves and radiolarians and gradually pass into dark red to dark grey and black radiolarites with little carbonate only and partly recrystallized to vitreous chert with silty-argillaceous partings. The radiolarians are sometimes size-sorted, they appear to be redeposited by dilute turbidity or bottom currents. Upwards the parallel-bedded radiolarites pass into first dark-coloured, then brick-red knobby radiolarites (undulating beds of vitreous chert with thin margins of granulär chert and micrite and with interbedded siliceous argillites). Up-section these pass into red argillaceous limestones with chert and frequent aptychi (ROSSO ad Aptici). Three thin intercalations of pure montmorillonite are interpreted äs devitrified cineritic tuffs.

The contact to the overlying Maiolica Formation (here Bemasian to Barremian) is sharp: the lowermost part of the Maiolica Formation (uppermost Tithonian) is missing, probably removed by submarine slumping which is frequent in the lower part of the formation. The change from radiolarian ooze to almost pure coccolith ooze - a regional phenomenon - is interpreted as an increase in the production of calcareous nannoplankton which lowered the calcite compensation depth. Radiolaria are concentrated in winnowed layers that were silicified during diagenesis and gave rise to layers of chert. Black shales intercalated between the heavily burrowed (Chondrites, Planolites, Zoophycos) grey coccolith oozes of the uppermost Maiolica record episodes of stagnant bottom waters during the Barremian. The top of the Maiolica is also sharp and riddled with burrows that contain glauconite and limonitic material. It is overlain by about 180 m of Aptian to Cenomanian marls and marly limestones (Scaglia variegata, bianca and rossa) with rich faunas of planktonic Foraminifera. In the uppermost Cenomanian, the hemipelagic Sedimentation was interrupted by the onset of flysch Sedimentation: the uppermost hemipelagic marls ofthe Scaglia Rossa Formation are interbedded with graded lithic sandstones and gradually pass into a sequence of silty marls and bituminous calc-schistes alternating with graded lithic arenites. The latter contain mainly clasts ofMesozoic Sediments of south-Alpine origin.

The eastern bank of the Breggia gorge, south of Morbio Inferiore is formed by a high cliff constituted by the Pontegana Conglomerate, an ill-sorted alluvial deposit composed of yellow, bleached boulders of Lombardian Siliceous Limestones deposited m a canyon which was eroded when the base level of erosion was lowered during the Messinian salinity crisis of the Mediterranean. South of the Breggia gorge younger Pliocene clays with a rich marine fauna were found to onlap onto the Pontegana Conglomerate.

From the Breggia gorge through Pleistocene and Recent alluvial deposits to Chiasso.

Basal part of post-orogenic Oligo-Miocene clastics (Rio della Maiocca, Chiasso, coord.: 723.95/078.50)

The (unconformable) contact between the Mesozoic Sediments and the south-Alpine "molasse" (Gonfolite Group) is no longer exposed. The base of the sequence is formed by the Chiasso Formation, a more than 50 m thick sequence of grey silty, strongly bioturbated mudstones withrare interbeds of graded feldspathic sandstones and siltstones. This late Middle to early Late Oligocene formation, zone NP 24 (= 32 my), was deposited in the upper bathyal zone (500-1000 m) according to benthonic Foraminifera. It is overlain with a sharp but conformable contact by the Como Formation (1000 m), pooriy stratified, inverse to normally graded conglomerates with occasional pebble imbrication. Lenses of sandstone suggest amalgamation of the organized conglomerates. The association with deeper marine formations below and above and the facies of the conglomerates suggest channel deposition in a submarine fan.

The pebbles of the Como conglomerates are mainly crystalline basement rocks derived from the Southern Alps and the Austroalpine nappes. Sedimentary components (nummulitic limestones, Mesozoic of Southern Alps) occur near the base of the formation, but disappear up-section.
400 m above its base, the Como Formation contains large boulders of the post-orogenic Bregaglia granitoids, radiometrically dated as 28 my old.

Although the Gonfolite Group overlies the previously deformed Mesozoic and Early Tertiary sequence of the Southern Alps with an unconformity, the main deformation of the latter occurred in Middle to Late Miocene times: near Chiasso the beds of the Gonfolite are strongly tiltedand now dip 55° to the south.

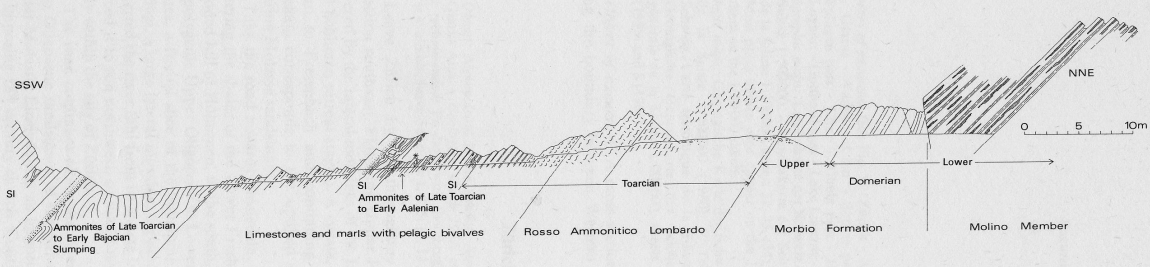


Fig. 27. Stratigraphic section of the Pliensbachian to Middle Jurassic pelagic sequence in the Breggia valley (after D. Bernoulli and F. Wiedenmayer 1967, modified).

Fig. 27. Stratigraphisches Profil der pelagischen Abfolge vom Pliensbachian bis zum Mittleren Jura im Breggia-Tal (nach D. Bernoulli und F. Wiedenmayer 1967, abgeändert).

Stop 4 Arzo quarries: Early Lias synsedimentary tectonics in the quarries north of Arzo (Coord.: 717.35/082.25)

Early Lias synsedimentary rifting (Fig. 25) in the Southern Alps resulted in tilted fault-blocks with a pronounced basin and swell topography. Its effects are most conspicuous along the eastern border of the Early-Middle Lias Arbostora high. This is characterized by a thin and incomplete sequence, whereas east of the N-S running boundary fault of Lugano, in the Generoso basin, approximately 1000 m of Rhetian shallow-water sediments and up to 4000 m of Early to Middle Lias spongolitic limestones and gravity flow sediments were deposited. The reduced sequence of the eastern margin of the Arbostora swell and the associated tectonic features are well exposed in the quarries north of Arzo.

Late Triassic (Norian) intertidal to shallow subtidal dolomites (Hauptdolomite Formation) are overlain by a thin sequence of Rhetian limestones, dolomites and marls (Tremona series), with thicknesses varying from one faultblock to another (0-70 m). These are unconformably overlain by Early Lias micritic and skeletal limestones with brachiopods, crinoids and calcareous sponges (Broccatello), which in turn are followed by a few meters of red, massive, condensed pelagic limestones with cephalopods and crinoid roots with ferromanganese coatings (Besazio limestone, Pliensbachian) and/or red, well-bedded marly limestones with ammonites and crinoid ossicies (Domerian). Submarine faults affected the Sedimentation of this sequence which reflects eventual submergence while the adjacent basins continued to subside. The Norian Hauptdolomite Formation, the Rhetic limestones and the Broccatello are cut by narrow grabens, fault zones and crevices which are filled by younger Sediments and complex polyphase breccias (Macchia Vecchia) with up to six phases of Sediments in which the older phases occur äs clasts of the host formations, whereas the younger ones occur äs matrix, forming a complex network of discordant sedimentary ("neptunian") dikes intruded frorn above.

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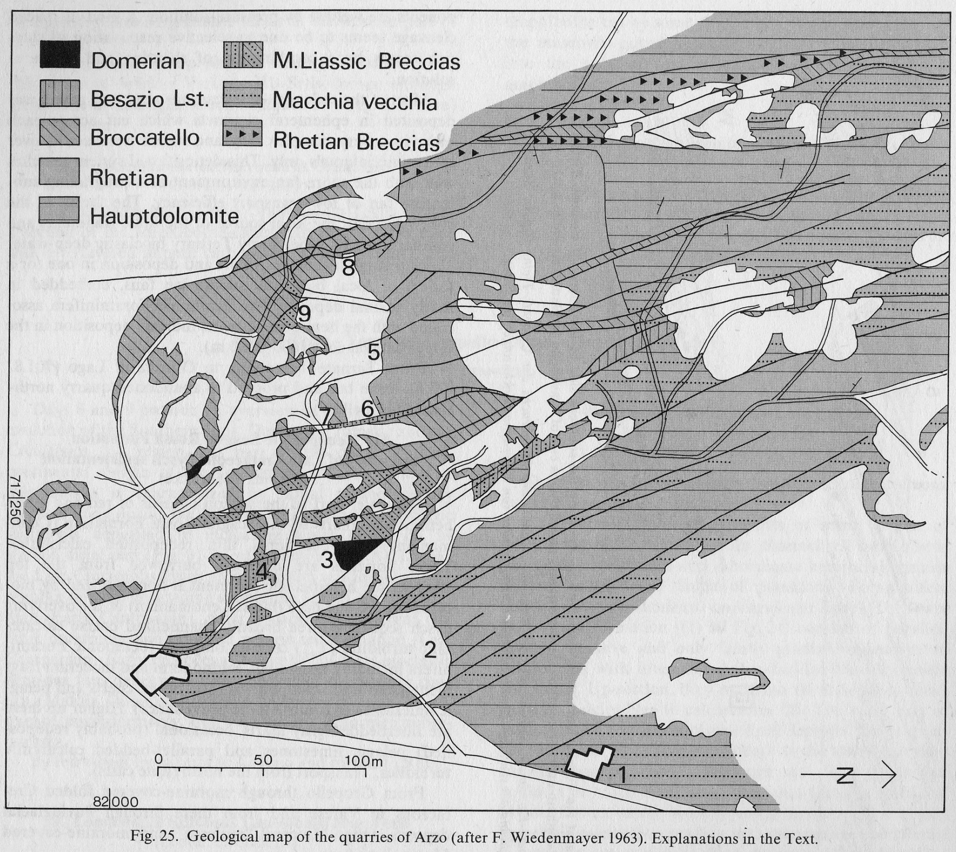


Fig. 25. Geological map of the quarries of Arzo (after F. Wiedenmayer 1963). Explanations in the Text.

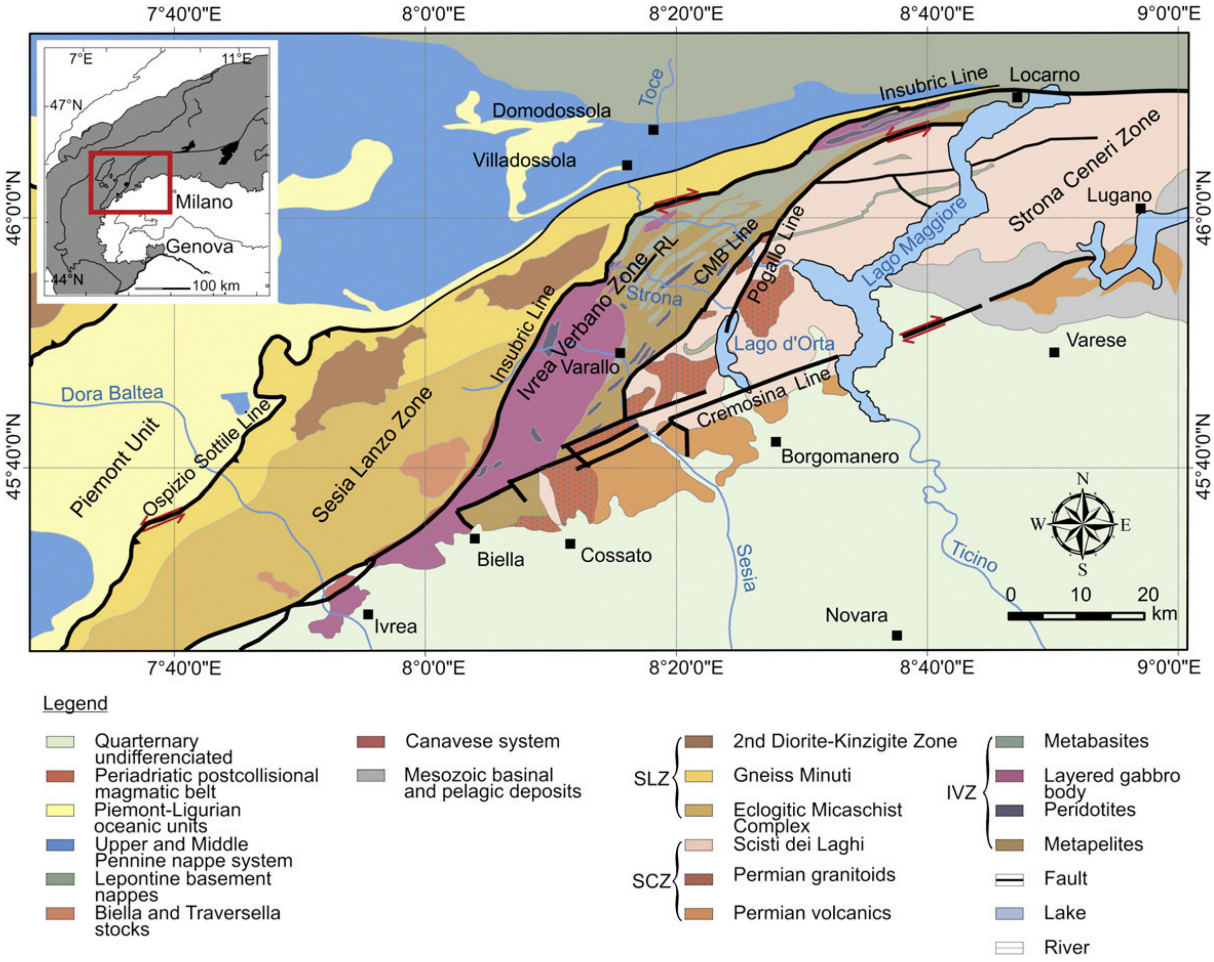
Fig. 25. Geologische Karte der Steinbrüche von Arzo (nach F. Wiedenmayer 1963). Erklärungen im Text.

Fig. 25. Geological map of the quarries ofArzo (after F. Wiedenmayer 1963).

Visit of the quarries (numbers refer to locations in Fig. 25);
(l) Quarry in the Macchia Vecchia breccias displaying in situ brecciated Hauptdolomite (algal stromatolites and flat-pebble breccias) with infills of white calcitic cements (? Rhetian) and of various discordant phases of Rhetian marly dolomites, Broccatello and Middle Liassic pelagic limestones and marls. (2) Hauptdolomite with an anastomosing System of thin, cm-thick sedimentary dikes and sills. Along the western wall of the quarry sedimentary dikes of Broccatello, of Besazio limestone and of Domerian marly limestones. (3) Small fault-bounded block of red, well-bedded, marly nodular maris and limestones with crmoid roots and ossicies and rare Domerian ammonites. (4) Besazio limestone with ferromanganese-coated cephalopods, brecciated with a groundmass of marly Domerian limestones. To the west, separated by a Liassic fault, Broccatello and breccias with fragments of Broccatello in a matrix of Besazio limestone. Proceed to the large quarry; along its northern face (5) light grey and yellowish, partly oolitic and ruditic limestones and dolomites of Rhetian age are cut by a number of N-S-trending faults and crevices filled by several generations of fossiliferous Liassic limestones and marls. This outcrop is bounded to the east (6) by a 3 m thick fault zone filled by a breccia with fragments of Broccatello rieh in brachiopods and cemented by marly Domerian limestone. This fault breccia also cuts through massive skeletal limestones of the Broccatello a few meters to the south (7). Along the western side of the quarry (8) the Rhetian limestones are unconformably overiain by Broccatello. Along the southern side (9), massive skeletal limestones with a rieh fauna of crinoids, brachiopods and calcareous sponges are cut by younger Middle Lias dikes. Alpine rejuvenation of the Liassic faults is indicated by slickensides and stylolitized surfaces.

Late Triassic (Norian) intertidal to shallow subtidal dolomites (Hauptdolomite Formation) are overlain by a thin sequence of Rhetian limestones, dolomites and marls (Tremona series), with thicknesses varying from one faultblock to another (0-70 m). These are unconformably overlain by Early Lias micritic and skeletal limestones with brachiopods, crinoids and calcareous sponges (Broccatello), which in turn are followed by a few meters of red, massive, condensed pelagic limestones with cephalopods and crinoid roots with ferromanganese coatings (Besazio limestone, Pliensbachian) and/or red, well-bedded marly limestones with ammonites and crinoid ossicies (Domerian). Submarine faults affected the Sedimentation of this sequence which reflects eventual submergence while the adjacent basins continued to subside. The Norian Hauptdolomite Formation, the Rhetic limestones and the Broccatello are cut by narrow grabens, fault zones and crevices which are filled by younger Sediments and complex polyphase breccias (Macchia Vecchia) with up to six phases of Sediments in which the older phases occur äs clasts of the host formations, whereas the younger ones occur äs matrix, forming a complex network of discordant sedimentary ("neptunian") dikes intruded frorn above.

Wednesday 28. 8.:
Finero - Cannobbio – Verbania – ins Aostatal – Valtournenche
Program: Insubric Linie**,** section through the Ivrea zone with peridotites and gabbros,
Total walk: : 4 h gentle mountain trail.Total drive 300 km. Accommodation ValTournenche



R. Wolff *, I. Dunkl 1, G. Kiesselbach 2, K. Wemmer 3, S. Siegesmund 4 2012,: Wolff, R., et al., Thermochronological constraints on the multiphase exhumation history of the Ivrea-Verbano Zone
of the Southern Alps, Tectonophysics (2012), doi:10.1016/j.tecto.2012.03.019

Abstract: The Ivrea-Verbano Zone of the western Southern Alps (NW Italy) exposes a well-preserved tilted sectionacross the lower continental crust, making it a key region for studying deep crustal and exhumation processes. This paper refines the cooling and exhumation history of the Ivrea-Verbano Zone using K/Ar dating of mica and illite-rich fault gouges as well as zircon fission track and (U-Th)/He thermochronology. The adjacent Strona-Ceneri Zone, Sesia-Lanzo Zone and Lower Penninic nappes are included in the study to derive a broader picture of the low-temperature history of the area. In the Strona profile of the Ivrea-Verbano unit the biotite K/Ar, zircon fission track and (U-Th)/He geochronometers show well preserved, but unusually wide partial retention zones. The youngest ages, representing the formerly deepest position, are situated along the Insubric Line. The main foliation of the Ivrea-Verbano Zone is vertical. The exhumation of the Ivrea-Verbano Zone, which section has a horizontal position on the surface now – took place in three steps. During Jurassic time the Ivrea-Verbano Zone was exhumed to a shallow to mid-crustal position by continental-scale extension. In this displacement the Pogallo Line probably played a dominant role. The studied section occupied an oblique position with a calculated angle of ca. 15 to 23° in the Jurassic. Later the Ivrea-Verbano Zone

experienced a minor cooling event in the Late Eocene (~38 Ma zircon fission track ages) that was probably related to thrusting and erosion. The final exhumation towards the surface took place in the mid-Miocene as documented by the ca. 14Ma zircon (U-Th)/He ages and a 12.8Ma K/Ar fault gouge age. The magnitude and the high rate of final exhumation suggest orogen-parallel extension as a driving force,which iswidespread in the Alps in the Lower to Middle Miocene and is most probably connected to orogenic collapse.

1. Introduction

The Ivrea-Verbano Zone (IVZ) spreads between Locarno in Switzerland and Ivrea in the Southern Alps of Italy. The IVZ (Fig. 1) is interpreted as a tilted section of the lower continental crust of the Southern Alps,whichmakes it a significant region for examining a profile through the continental crust to lenses of the upper mantle (e.g. Berckhemer, 1969; Fountain, 1976; Fountain and Salisbury, 1981; Mehnert, 1975; Rabbel et al., 1998; Weiss et al., 1999). Separated by the Cossato-Mergozzo-Brissago (CMB) Line the Strona-Ceneri Zone (SCZ) is part of the upper crust of this profile. The Southern Alps were affected by the Permian high temperature metamorphism and magmatism, but not by the regional Alpinemetamorphism. Separated by the Insubric Line the Penninic units including the Sesia-Lanzo Zone (SLZ) are the units, which were underthrust below the Southern Alps during the Alpine orogeny. The adjacent units in the Ivrea area differ significantly in their tectonometamorphic histories. During the Late Cretaceous (80 to 65 Ma, Babist et al., 2006) the SLZ underwent high pressure metamorphism at a depth of 60 km (Compagnoni, 1977; Dal Piaz et al., 1972), while the IVZ and the SCZ were already cooled below greenschist facies conditions (Zingg et al., 1990). Due to rotation and exhumation along the Insubric Line and minor fault zones, these units with very different time-temperature histories are now adjacent to each other. The emplacement kinematics and timing of the tectonic movements, which lead to the vertical position of the main foliation is still a subject of ongoing debates (Handy and Zingg, 1991; Handy et al., 1999; Hurford et al., 1991; Schmid, 1993; Schmid et al., 1987; Siegesmund et al., 2008). According to Schmid et al. (1987; 1989), the rotation of the IVZ into its vertical position adjacent to the Insubric Line is post-Oligocene and related to the Oligocene back-thrusting, which was synchronous with the back-folding of the Central Alps combined with a minor sinistral strike-slip component. The backthrusting and the differential uplift of the SLZ led to mylonitization under greenschist-facies conditions along the Insubric Line. Based on the data of Hurford (1986), Schmid et al. (1987) concluded that backthrusting started between 30 Ma and 23 Ma.

Paleomagnetic analyses in the IVZ (Schmid et al., 1989) suggest independently a post-Oligocene rotation of 60° around a horizontal axis striking parallel to the Insubric Line in a clockwise sense viewed to the north. According to Zingg et al. (1990), the IVZ rotated during the Oligocene-Miocene dextral transpression into its present attitude. The rotation of the IVZ was possibly due to the combination of the inherited Paleozoic crustal setting and the change from early Mesozoic extension to Cretaceous shortening, which caused a SE dipping of the MOHO.

Boriani and Giobbi (2004) argued, however, that the basement of the western Southern Alps is not a tilted section. Their main arguments are: in the Early Permian, the IVZ to the SCZ were already in contact through the CMB Line. Later the Permian Baveno granite intruded the SCZ at a depth of about 4 km (Boriani et al., 1990b).

Because the Baveno pluton shows a layering, dipping less than 20° (Boriani et al., 1990a), the IVZ and the SCZ have not been tilted more than 20° in post-Permian time. Furthermore, the SCZ was at least in part exhumed in Permian times. Additionally, vertical dykes with chilled margins intruded in the Permian along the CMB Line. These dykes are still vertical, they have not been rotated. However, these dykes are debated rather being concordant sills than dykes (Handy and Streit, 1999). The interpretation of the Ivrea-Verbano Zone as a tilted section is criticized in favour of a trans-tensional model (Boriani and Giobbi, 2004).

Handy et al. (1999) considered that banding and foliation of the IVZ were sub-horizontal prior to the Paleogene period. Insubric faulting and brittle folding accompanied the rotation of this crustal section. However, while the IVZ was rotated by around 90°, the SCZ underwent only minor to moderate Alpine rotation. Brittle reactivation of the Pogallo Line likely accommodated the differential rotation of the crustal blocks.

Rutter et al. (2007) argued that the IVZ formed a large monoclinal kink during the Alpine orogenesis, in which the main compositional banding was tilted to the vertical. Siegesmund et al. (2008) interpreted the lower break in slope of the zircon fission track ages along a profile in the Val Strona di Omegna at 50 Ma. This break emerges during or immediately after the rotation of the IVZ, because the part closest to the Insubric Line has been exhumed already tilted at this time.
The cooling and exhumation history of the Ivrea-Verbano Zone is important for the proper understanding of the timing and magnitude of the Mesozoic extension and the Paleogene to Neogene shortening events. In this paper we present new mica K/Ar, zircon fission track (ZFT) and zircon (U-Th)/He (shortly ZHe) ages. The ZHe method is sensitive to the low-temperature events in the shallow crust. Combined with the other thermochronometers these new age constraints are suitable for setting up a refined model on the cooling and exhumation history of the IVZ.

2. Geological setting

The Southern Alps are characterised by south verging fold and thrust belts (Schmid et al., 2004). According to the seismic profiles (ECORS-CROP, Roure et al., 1996) this fold and thrust belt continues towards the south below the Po plain and up to the area of Milan. The 'Seengebirge' (Schmid, 1967) spreads between Biella and Lago di Como in the western part of the Southern Alps and is also calledMassiccio dei Laghi (Boriani and Sacchi, 1973). It combines the Ivrea- Verbano Zone and the SCZ, which is also named Serie dei Laghi.

The Ivrea-Verbano Zone comprises metapelites, quartz-feldspar gneisses with garnet, biotite, sillimanite and graphite. In older literature the metapelites are divided into 'kinzigites' and stronalites'; the metamorphic grade of the IVZ decreases from granulite grade in the NW towards the high-temperature amphibolite grade in the SE (Schmid, 1967). Because of this decrease in metamorphism, together with the pressure gradient (Henk et al., 1997), the paleomagnetic evidences and the lenses of upper mantle, the IVZ is interpreted as a 90° tilted and shortened section. The Mafic complex at the SW part of the IVZ is of Permian age (288 Ma; Pin, 1986; Peressini et al., 2007) and represents an example of lower crustal underplating (Fountain, 1989). This range of ages brackets the growth of the Mafic complex, from its inset by sporadic pulses to the final caldera collapse (Sinigoi et al., 2011). The Early Permian thermal overprint caused by the intrusion of the Mafic complex obscured the ages of the earliest events of the IVZ (Henk et al., 1997; Peressini et al., 2007).

Continuing to the southeast the Strona-Ceneri Zone dips 15°–30° SE and represents the former upper crust capped by Permian volcanites (Handy et al., 1999). These volcanites are interpreted as part of a supervolcano. The volcanic field underwent the collapse of a caldera; the plumbing system is represented by both the IVZ Mafic complex and the coeval granitic plutons (Quick et al., 2009). The SCZ comprises a series of metasedimentary schists and gneisses with occasional amphibolite sheets, which are cut by orthogneisses of Ordovician protolith age. During the Carboniferous uplift, retrograde greenschist facies paragneisses are formed. In the SCZ large granite plutons and mafic-intermediate stocks and dykes were emplaced during the Permian. These so-called Graniti dei Laghi form a batholith system, comprising the Valle Mosso (Valsessera-Biellese), Alzo- Roccapietra, Quarna, Mottarone-Baveno, and Montorfano plutons (Boriani and Giobbi, 2004). A shallow intrusion depth of less than ~4 km is inferred by miarolithic cavities in the Baveno granite, which can only develop under low pressure conditions (Boriani et al., 1990b). In addition, Boriani et al. (1990b) suggested that the SCZ was, at least in part, exposed in the Permian. Accordingly, the Baveno granite has already cooled down in the Permian. The contact between the miarolithic and non-miarolithic granites indicate, that the Baveno pluton has been rotated 15°–20° eastwards after its emplacement (Boriani et al., 1990b). The Permian emplacement is constrained by Pinarelli et al. (1988) with Rb-Sr whole-rock ages of 276±7 Ma and Schaltegger and Brack (2007) with zircon U-Pb ages of 281±1 Ma.

Siegesmund et al. (2008) dated the Baveno granite to 291 Ma and 284 Ma with biotite K/Ar. In contrast, Pinarelli et al. (1988) reported younger Rb/Sr biotite ages (170 Ma) close to the CMB Line. This younger age could be due to a thermal event in the Middle Jurassic. In Mesozoic times the SCZ was dissected by extensional faults, also the CMB Line and the Pogallo Line were reactivated (Boriani

et al., 1990b). During the alpine compression the SCZ was affected by Sverging cataclastic overthrusting (Boriani et al., 1990b).
The Sesia-Lanzo Zone is located north of the Insubric Line and belongs to the distal part of the Adriatic continental margin lithosphere. It represents the highest nappe of the western alpine belt and is located on top of the Penninic nappes, which crop out northwest of the SLZ. According to Compagnoni (1977), the SLZ comprises three basement nappes, which underwent eclogite-facies conditions due to subduction during the Late Cretaceous between 80 Ma and 65 Ma (Babist et al., 2006). The exhumation was isothermal to a depth of 20 km (Babist et al., 2006). Thereafter in the Eocene the exhumation proceeded together with extensional shearing. The high pressure rocks cooled slowly from 40 Ma to 30 Ma (Babist et al., 2006) and were intruded by shallow granitic plutons (Biella and Traversella plutons, 30 Ma; Krummenacher and Evernden, 1960), when they were already exposed to erosion at the surface. Only the NE part of the SLZ remained below 10 km depth until its exhumation due to back-folding and thrusting of the Insubric Line in the Tertiary and brittle faulting in the post-Oligocene (Babist et al., 2006).

The Periadriatic lineament marks the northern and western boundary of the Southern Alps and the southern limit of the Tertiary metamorphism, excluding the segment south of the Tauern window (Ahrendt, 1980). The protoliths tectonically incorporated into this mylonitic belt are derived from the IVZ, SLZ and from the Permo- Mesozoic sediments of the Canavese Zone in the study area (Ahrendt, 1980). Its segments are called from west to east, the Insubric- (Canavese- and Tonale-), Giudicarie-, Pustertal- and Gailtal Lines (Schmid et al., 1987). The shear zone at the Insubric Line accommodated the back-thrusting synchronously with the back-folding of the Central Alpine nappes. This was followed and partly synchronous with the dextral strike slip of ~100 km along the E–W striking Periadriatic lineament (Schmid and Kissling, 2000; Schmid et al., 1987). In the study area this ~1 km thick greenschist-facies mylonite belt is called the Insubric Line, which separates the SLZ and the IVZ (Fig. 1). Wemmer (1991) determined the last reactivation of the Insubric Line at Rimella to 22 Ma±0.7 Ma by K/Ar fault gouge dating. The total NW side-up displacement between the Oligocene and present was 10 to 20 km, which means the SLZ was uplifted with respect to the IVZ (Hurford, 1986; Schmid et al., 1987).

The IVZ and the SCZ were adjacent since the Early Permian and shared the same metamorphic history (Boriani et al., 1990a). The shear zone dividing the IVZ and the SCZ is the Cossato-Mergozzo-Brissago Line (Boriani and Sacchi, 1973). The CMB Line forms the main contact southwest of the Val d'Ossola, where the Pogallo Line breaks away from the lithological contact. This Permian contact is a subject of ongoing debates (Boriani et al., 1990a; Handy, 1987; Handy et al., 1999; Mulch et al., 2002). Boriani et al. (1990a) and Mulch et al.(2002) define the CMB Line as the major tectonic contact between the IVZ and the SCZ, while in earlier literature the Triassic Pogallo Line was defined as a contact (Hodges and Fountain, 1984; Schmid et al., 1987). The CMB Line is a sub-vertical high temperature ductile shear zone that was active until the Early Permian. The mylonitic shear activity went along with the intrusion of gabbro-dioritic bodies called Appinites (Boriani and Giobbi, 2004). The vertical displacement along the CMB Line is negligible (Handy et al., 1999). The Pogallo Line displaces the CMB Line north of the Valle d'Ossola (Mulch et al., 2002). It is interpreted as a low angle normal fault due to crustal thinning (Hodges and Fountain, 1984), which was active between 210 Ma and 170 Ma (Zingg et al., 1990). The extension from Late Triassic to mid Jurassic correlates with the opening of the Piemont Ocean (Zingg et al., 1990). The northern part of the Pogallo Line formed under amphibolite facies conditions, whereas at the same time, to the south, shear occurred in greenschist facies conditions (Handy, 1987). The fault plane of the Pogallo Line strikes NNE and dips 80° towards the WNW (Zingg et al., 1990). It is concordant with the attitude of the kinzingites of the IVZ but discordant with the SCZ (Boriani et al., 1990a).

To the south the Cremosina Line crosscuts the CMB Line. It was suggested, that this fault zone was active during Carboniferous to Permian times by Boriani et al. (1990a) and Giglia et al. (1996). However, according to the evolution of the Sesia magmatic system, the Permian Graniti dei Laghi were emplaced besides and separated by the Cremosina line. Thus the activity has to be shifted into

post-Permian time (pers. Comm. Quick and Sinigoi). It was reactivated as a dextral strike-slip fault during post-Oligocene times (Boriani et al., 1990a).
In the north the Ospizio Sottile fault locally separates the Piemont Unit and the SLZ. This sinistral fault zone was reactivated from Miocene to the present, which was triggered by the escape of the Penninic Unit from the Simplon detachment (Bistacchi et al., 2001). Hunziker et al. (1992) presented a collection of the available geochronological data of the Central and Western Alps. In this papertheir data set is used and complemented with the geochronological results fromthe following studies: Jäger and Faul, 1960; Krummenacher and Evernden, 1960; Carraro and Ferrara, 1968;McDowell and Schmid, 1968; McDowell, 1970; Wagner and Reimer, 1972; Hunziker, 1974; Frey et al., 1976; Zingg et al., 1976;Monié, 1985; Oberhänsli et al., 1985; Diamond and Wiedenbeck, 1986; Hurford, 1986; Stähle et al., 1986; Stöckhert et al., 1986; Hurford et al., 1991; Wemmer, 1991; Inger et al., 1996; Vance, 1999; Keller et al., 2005;Malusà et al., 2005;Malusà et al., 2006; Siegesmund et al., 2008.

6. Discussion

6.1. Areal distribution of cooling events The Ivrea area west of Locarno comprises units of different tectonometamorphic and exhumation histories. The units are divided by three major faults: the CMB Line, the Pogallo Line and the Insubric Line, which were active during different geologic periods. The juxtaposition of the SLZ, the IVZ and the SCZ, happened at different times. For an overview of the data, the results in this study were grouped together with the AFT ages from Wagner and Reimer (1972), Hurford (1991) and Keller et al. (2005) into three units of different cooling histories, called ‘cooling patterns’ (Fig. 8).

6.1.1. Pre-Alpine cooling pattern
The pre-Alpine cooling pattern spreads within the SCZ between Verbania, the Lago Maggiore, the Lago d'Orta, the Baveno granite and the CMB Line. The first group of apparent cooling paths is defined by the K/Ar mica ages older than 225 Ma, ZFT ages of 225–145 Ma and ZHe ages of 215–130 Ma (Fig. 8). The most characteristic feature of this cooling pattern is that the ZHe ages were not reset during the Alpine orogeny. Hence, this unit stayed in thermal conditions below the closure temperature of the ZHe thermochronometer (ca. 185 °C according to Reiners et al., 2004) since at least 130 Ma.

6.1.2. Eo-Alpine cooling pattern
The Eo-Alpine cooling pattern spreads in the IVZ and the northern SCZ, between the Lago Maggiore and the Insubric Line. The K/Ar mica
ages range between 300 Ma and 130 Ma, the ZFT ages between 150 Ma and 35 Ma and the ZHe ages between 75 Ma and 15 Ma. Hence, this unit stayed below the closure temperature of white mica since 140 Ma. The Alpine orogeny has only reset the ZFT and Zhe ages.

6.1.3. Tertiary cooling pattern
The Tertiary cooling pattern is located in the SLZ. The K/Ar mica ages range from 80 Ma to 20 Ma, the ZFT ages between 45 Ma and 20 Ma and the ZHe ages between 24 Ma and 14 Ma. This unit has been affected by Tertiary metamorphism, followed by a fast cooling. These patterns mark coherent areas in the western Southern Alps and indicate the migration of the exhumation through time (see map sketches in the right panels of Fig. 8). However, the ‘patterns’ outlined above show internal trends and they are not homogeneous. Thus, the apparent cooling paths need a more detailed evaluation. In the discussion below a more refined exhumation history of the Ivrea area is presented based on the new low-temperature geochronological data.

6.2. Major steps and thermal conditions in the post-Permian evolution
6.2.1. Post-Permian cooling
The presented dataset contains information about the cooling and exhumation history of the Ivrea area after the Permian underplating event. This history is based on the tectonic movements of the main units along the major contacts, the Insubric Line and the CMB Line. The dataset shows a joint exhumation history of the IVZ and the SCZ after the Permian underplating without any offset in the

data crossing the CMB Line (Fig. 9). Neither the monotonic increasing trend of K/Ar (Siegesmund et al., 2008), nor the trend of ZHe ages indicate abrupt changes at the crossing of the CMB Line.

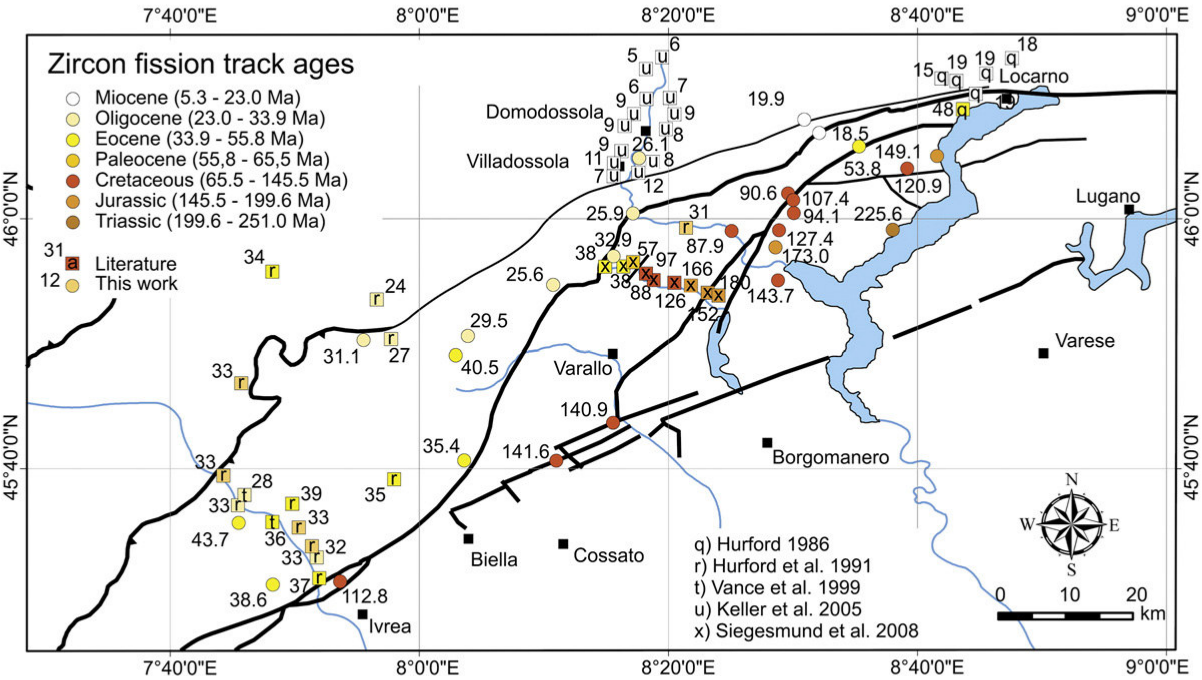
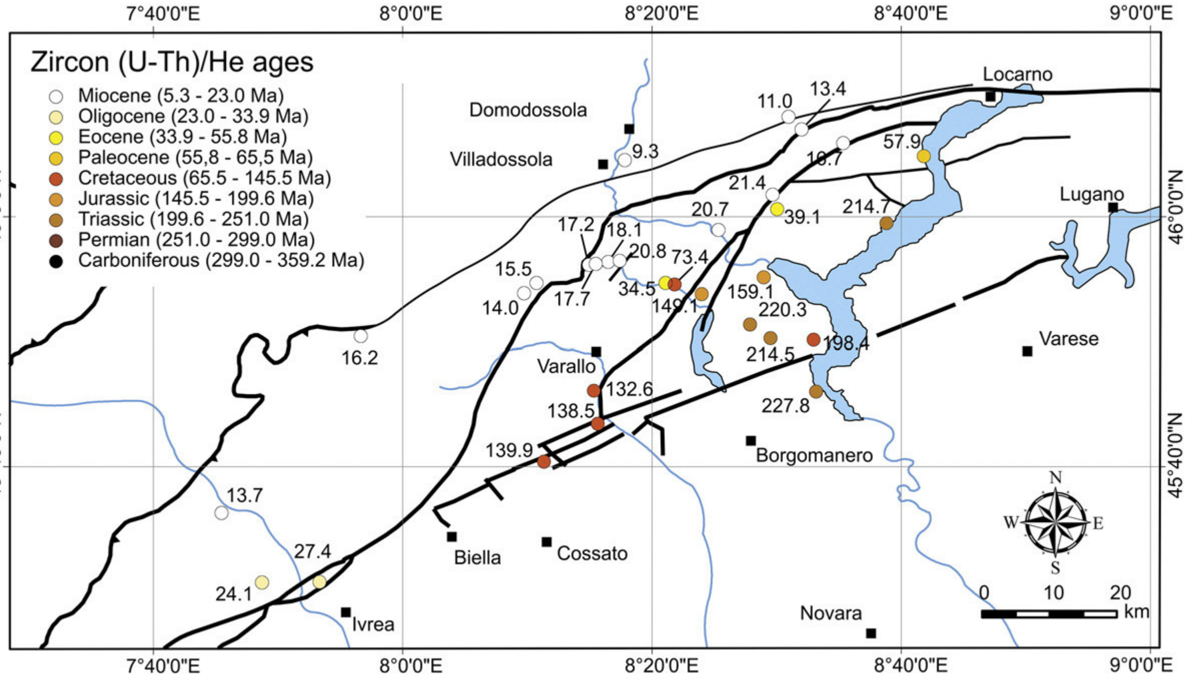


Fig. 5. Compilation of zircon fission track ages in the study area. Circles: samples analysed in this work; squares: cited from other studies.

6.2.2. Metamorphic conditions along the Pogallo Line The Pogallo Line displaces the CMB Line and is interpreted as a low angle normal fault due to crustal thinning (Hodges and Fountain, 1984), which was active between 210 Ma and 170 Ma (Zingg et al., 1990). The extensional activity from Late Triassic to Middle Jurassic coincides with the opening of the Piemont Ocean (Zingg et al., 1990). Metamorphic conditions along the Pogallo Line change from ductile (north of the Val d'Ossola) to brittle deformation further to the south (Handy, 1987). The multi-method thermochronology allows a more precise determination of the temperature conditions along the Pogallo Line (Fig. 10). The cooling paths of the samples in the northern and southern parts of the Pogallo Line yield temperatures of 290 to 350 °C and 220 to 310 °C, respectively, in the time span of the fault activity.

6.2.3. Insubric phase

The main exhumation and tilting of the IVZ was governed by the activity of the Insubric Line. Units which suffered alpine metamorphism and those which did not (Ahrendt, 1980) are now adjacent after the 10 to 20 km of post-Early Oligocene vertical uplift along the Insubric Line (Frey et al., 1974).



FFig. 6. Distribution of the new zircon (U-Th)/He ages.

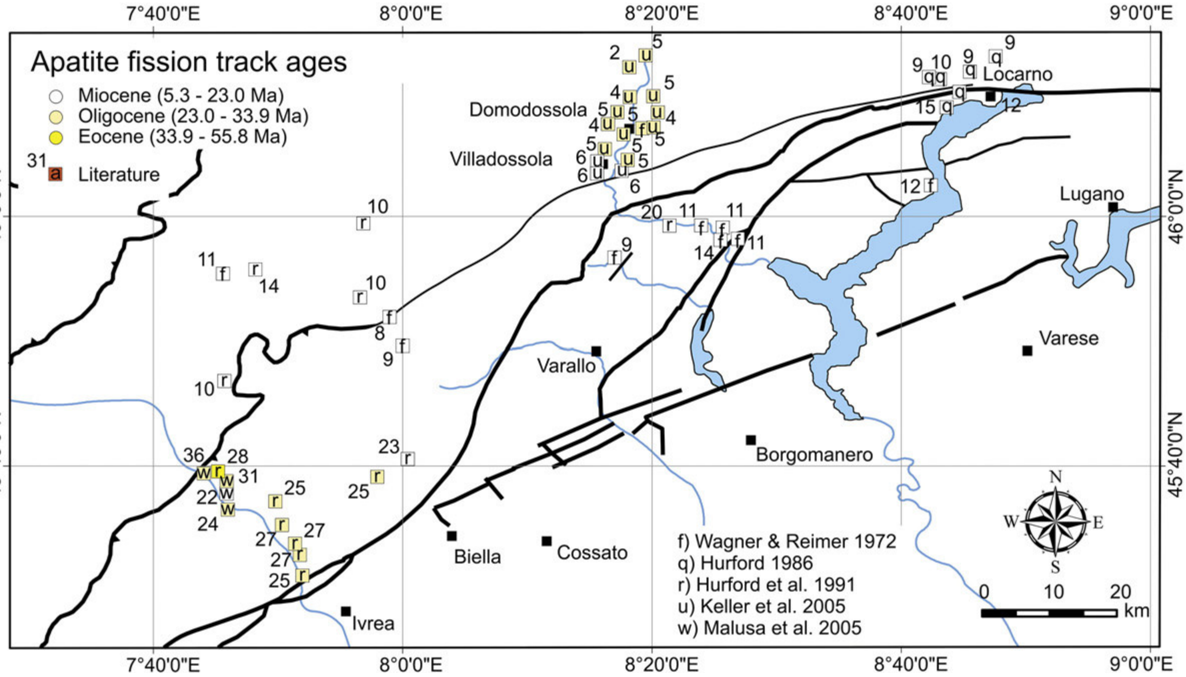


Fig. 7. Compilation of apatite fission track ages in the study area.

According to Schmid et al. (1987; 1989), the rotation of the IVZ occurred during the back-thrusting event, which was synchronous with the back-folding of the Central Alps and combined with a minor sinistral strike-slip component. The

back-thrusting and the differential uplift of the SLZ led tomylonitization under greenschist facies conditions along the Insubric Line. Based on the data of Hurford (1986), Schmid et al. (1987) concluded that the backthrusting started between 30 Ma and 23Ma. Later dextral strike slip related to transcurrent displacements also lead to mylonitization. The K/Ar fault gouge ages along the Insubric Line bracket this activation history between the Miocene and the Eocene. The ages along the Insubric Line range between 36.7±1.0 Ma and 12.8±0.7 Ma west and east of Locarno, respectively (this paper).

They complement the data from Wemmer (1991) yielding 22.9±0.7 Ma at Rimella, Zwingmann and Mancktelow (2004) yielding 19.0±0.5 Ma and 23.5±0.5 Ma east of Bellinzona and Zingg et al. (1976) ranging between 37.8±2.2 Ma and 19.5±3.7 Ma ages between Biella and Valle d'Ossola. The following section is focussing on a profile perpendicular to the general strike of the IVZ through the Val Strona.

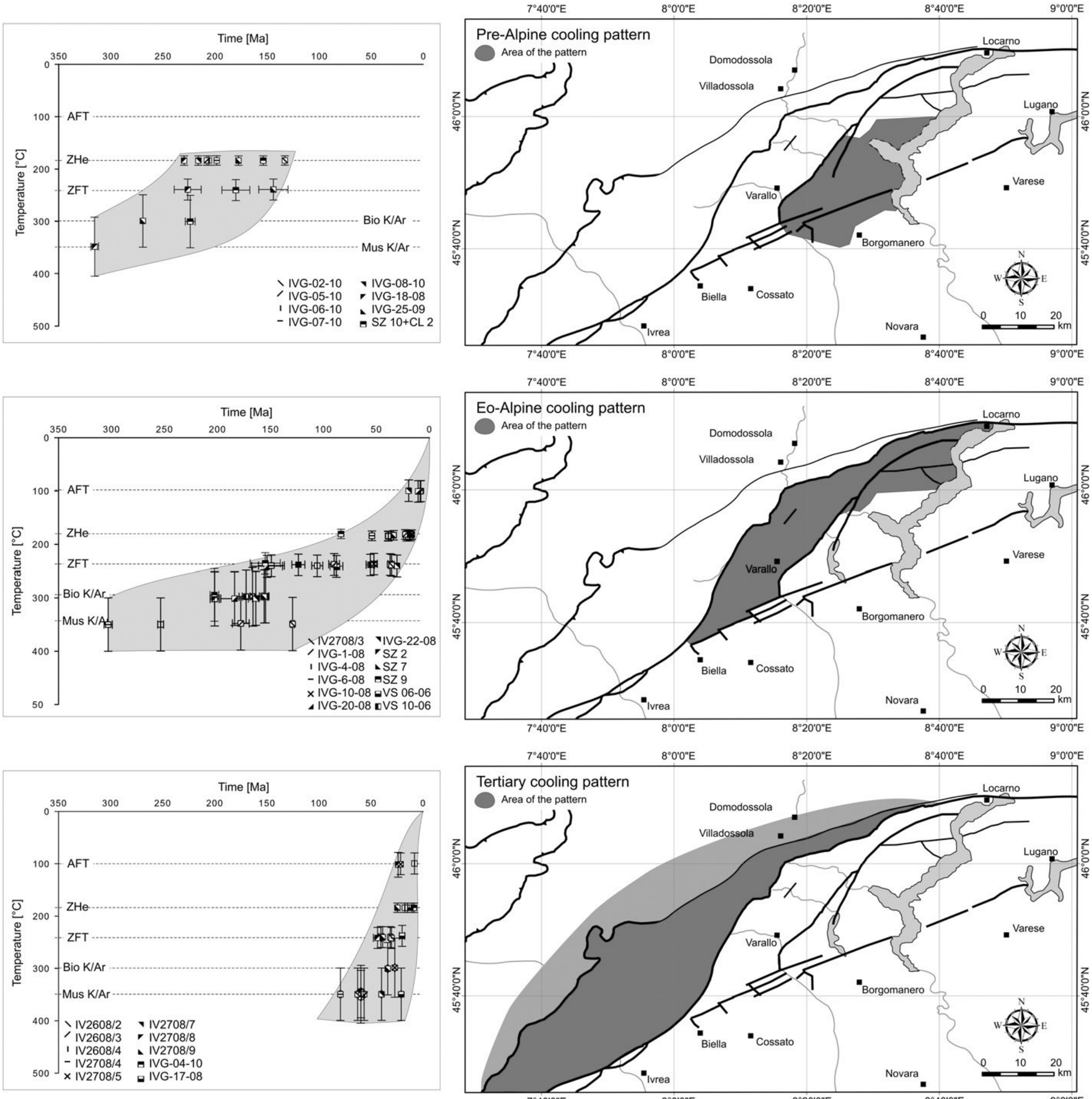


Fig. 8. (Left) The samples were analysed with multi-method geochronology and were grouped according to their apparent cooling paths. (Right) The geographic spread of the three different grouped samples is marked on the map. (A) Pre-Alpine cooling trend: mica K/Ar ages, ZFT and ZHe ages are typically older than 150 Ma. (B) Eo-Alpine cooling trend: the mica K/Ar ages range from Variscan times to the Jurassic, the ZFT ages are typically Cretaceous and Tertiary, while the ZHE ages are exclusively Tertiary. (C) Tertiary cooling trend: the oldest mica K/Ar ages are Late Cretaceous, but all other thermochronometers having lower closure temperatures yield Tertiary ages. Shaded area is unconfined. Data from this study but the AFT ages are from Wagner and Reimer (1972), Hurford (1991) and Keller et al. (2005).

6.3. Exhumation history of the Ivrea zone using the newthermochronological data The new thermochronological data on the Strona profile (A–B on Fig. 2) allows a closer examination of the exhumation history. Fourteen zircon (U-Th)/He ages were generated along a ca. 30 km long, nearly strait NW–SE profile (Fig. 9). In combination with 30 biotite K/Ar and nine zircon fission track ages by Siegesmund et al. (2008), all three geochronometers show a monotonous rejuvenation trend towards the Insubric Line in the north-western direction (Fig. 9). No detectable offset in the ages occurs when crossing the Rosarolo shear zone (Siegesmund et al., 2008), the CMB Line and the Pogallo Line. The biotite K/Ar ages range from 335 Ma to 156 Ma. The ages increase steadily with increasing distance from the Insubric Line, but the slope is low at the beginning and the end and steeper in between 8 and 25 km. The ZFT ages range between 180 Ma and 38 Ma. The two nearest samples to the Insubric Line (at 0.2 and 3 km distances) show similar ages, then the ages increase rapidly to 180 Ma at 13 km distance. The ZHe ages vary between 228 Ma and 14 Ma. The first seven samples in between 0 and 8.5 km distance have ages of 14, 15.5, 17.2, 17.7, 18.1, 20.8 and 35.5 Ma. The following ages increase up to 220 Ma at 19 km distance and then show similar ages again up to the most distal sample at 30 km distance.

The monotonic increase of the ages of the different thermochronometers across the Ivrea-Verbano Zone is interpreted as exhumed and tilted partial reset zones (Fig. 9). These individual ages do not express cooling events or episodes of mineral growth; they are the result of the equilibrium between radioactive decay and the diffusion of decay products (or annealing of fission tracks). Such static partial reset zones develop everywhere in the Earth's crust along profiles having a vertical component. A complete vertical profile with a monotonic transition of ages is composed of three sections (e.g. Fitzgerald et al., 1991): (i) the oldest ages are situated in the upper part of the profile, and expressing the cooling after the oldest thermal event documented by the given thermochronometer. (ii) In the middle section the ages gradually decrease downward. Here the ages were developed in temperature conditions where the products of the radioactive decay only partially disappear by annealing or by diffusion. This transitional zone is defined as the partial annealing zone (PAZ; Fitzgerald et al., 1991) in fission track thermochronology and as the partial retention zone (PRZ; House et al., 1999) in argon and (U-Th)/He thermochronology. (iii) The youngest ages are situated at the base of the profile. If the bottom of the profile is still in the temperature zone where the decay products of the given thermochronometer reset ‘geologically instantaneously’, then the youngest ages are zero ages. If the youngest ages at the bottom of the profile are non-zero ages, then they express the age of cooling from the former total reset zone to lower temperature conditions, where the decay products are already stable. These youngest data typically register valuable information about the age and rate of the latest exhumation event (Fitzgerald et al., 1991; Kamp and Hegarty, 1989).

In the Strona profile the upper and lower asymptotes of the age curves (Fig. 9) match well with the major known thermotectonic events of the region. The oldest biotite K/Ar ages in the SE of the profile covering the SCZ point to the Carboniferous and clearly mark the age of the Variscan metamorphism, which was a determinant period in the development of the Southern Alpine metamorphic basement (e.g. Boriani and Villa, 1997). The youngest part of the biotite K/Ar age trend towards the Insubric Line is not completely developed, but it is obvious that the ages are converging to ca. 150 Ma. This age actually matches well with the period of relaxation of the isotherms after Early-Mid Jurassic rifting (Stampfli et al., 2002) that caused thinning and increased heat flow in the entire Southern Alps (Bertotti et al., 1997).
The zircon fission track ages indicate a well developed lower part of a fossil partial annealing zone, but unfortunately the oldest (topmost) part of the section was not preserved. However, the oldest ages of the ZHe thermochronometer gives a minimum age of ca. 200 Ma for the oldest ZFT ages because the fission tracks closure temperature is higher than the ZHe closure temperature (e.g. Reiners et al., 2004). The youngest ages are around 38Ma, indicating a rapid cooling event in the Late Eocene. This event can be related to the so-called ‘pre-Oligocene thrusting’ phase outlined by Schmid et al. (1989) and documented by numerous field observations and geochronological data in theWestern Alps and in the area of Bellinzona (Ahrendt, 1980; Nagel et al., 2002). The zircon (U-Th)/He data indicate a

completely preserved partial retention profile with well documented young and old asymptotes. In the upper part of the profile the thermal conditions allowed complete helium retention in the zircon crystals during the last ca. 200 Ma. Accordingly, this section occupied a shallowposition in the upper crust, above the 100 °C isotherm (for explanation see Fig. 1a in Reiners and Brandon, 2006). This age coincides well with the extension in Early Jurassic time (Hodges and Fountain, 1984; Lemoine and Trümpy, 1987; Santantonio and Carminati, 2010; Zingg et al., 1990). This heating resulted in the reset of the ZHe thermochronometer in the Baveno granite that cooled down already in Permian time after its emplacement in shallowdepth. In the young part of the profile (close to Insubric Line) the ZHe ages converge to ca. 14Ma. These ages indicate that the final exhumation only took place in the Middle Miocene. Along a ca. 5 km section in the vicinity of the Insubric Line the ZHe ages are very uniform. Thus, the final exhumation from the total reset thermal conditions (>~185 °C) took place at a high slip rate.

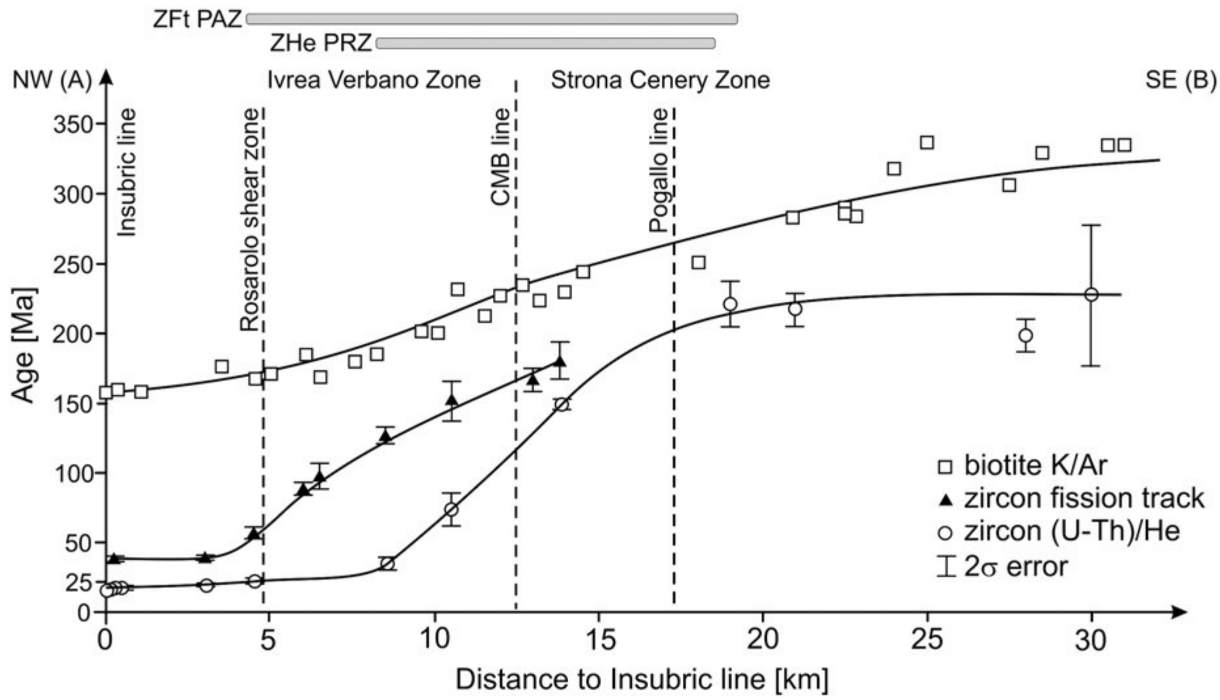


Fig. 9. The trend of biotite K/Ar, ZFT and ZHe ages along the Strona profile (A–B on Fig. 2) between the Insubric Line and Lesa at the Lago Maggiore at ca. 30 km distance (K/Ar and ZFT analysis from Siegesmund et al., 2008). Empty squares: biotite K/Ar, filled triangles: ZFT, empty circles: ZHe. The 2σ uncertainty of the Cenozoic ZHe ages is smaller than the symbol.

6.4. Pre-exhumation position of the Ivrea-Verbano Zone

According to the geothermometry and geobarometry data, the Ivrea-Verbano Zone experienced a ca. 90° horizontal axis rotation; however, the timing of this movement is debated (for more details refer to the Introduction and Geology above). The new facts that can be expressed from the thermochronological data and which are relevant for the detection of the pre-exhumation position of the IVZ will be summarized below. The widths of the detected partial reset zones play a key role in this reconstruction. In the Strona profile the zircon fission track partial annealing zone is as wide as 15 km, while the zircon helium retention zone is ca. 10 km wide (Fig. 9). Worldwide there are only a few complete documentations available on the thickness of the static partial reset zones. Zaun and Wagner (1985) reported the first indications of a zircon PAZ in a borehole. The Urach deep drill hole in southern Germany could not penetrate the entire PAZ, but according to the downhole trend of ZFT ages the thickness of the zone is not more than ca. 2.5 to 3 km. Stockli et al. (2000) studied a naturally exhumed and tilted PAZ in the western Basin and Range province and found a similar thickness for the zircon PAZ, ca. 3 km, while Bernet (2009) documented a thickness of ca. 4 km. For the thickness of the static PRZ of the ZHe thermochronometer Stockli et al. (2000) and Wolfe and Stockli (2010) determined a thickness of ca. 2.5 km. The partial reset zones of all these studied sites are thinner than the detected thicknesses of the PAZ and PRZ in the Strona profile. Since the zircon helium partial retention zone is perfectly preserved in the Strona profile, a robust calculation with the results of the Zhe method can be undertaken. Modelling of the ZHe ages along a vertical profile have been performed using the HeFTy (Ketcham, 2005). The parameters used for the modelling are:

– Radius of zircon grains: 55 µm(average of the calculated radius for crystals measured in the Strona profile),
– Uranium content of zircon grains: 390 ppm (median of the U concentration measured in the dated crystals).
– The assumed thermal histories are composed of three or four parts:

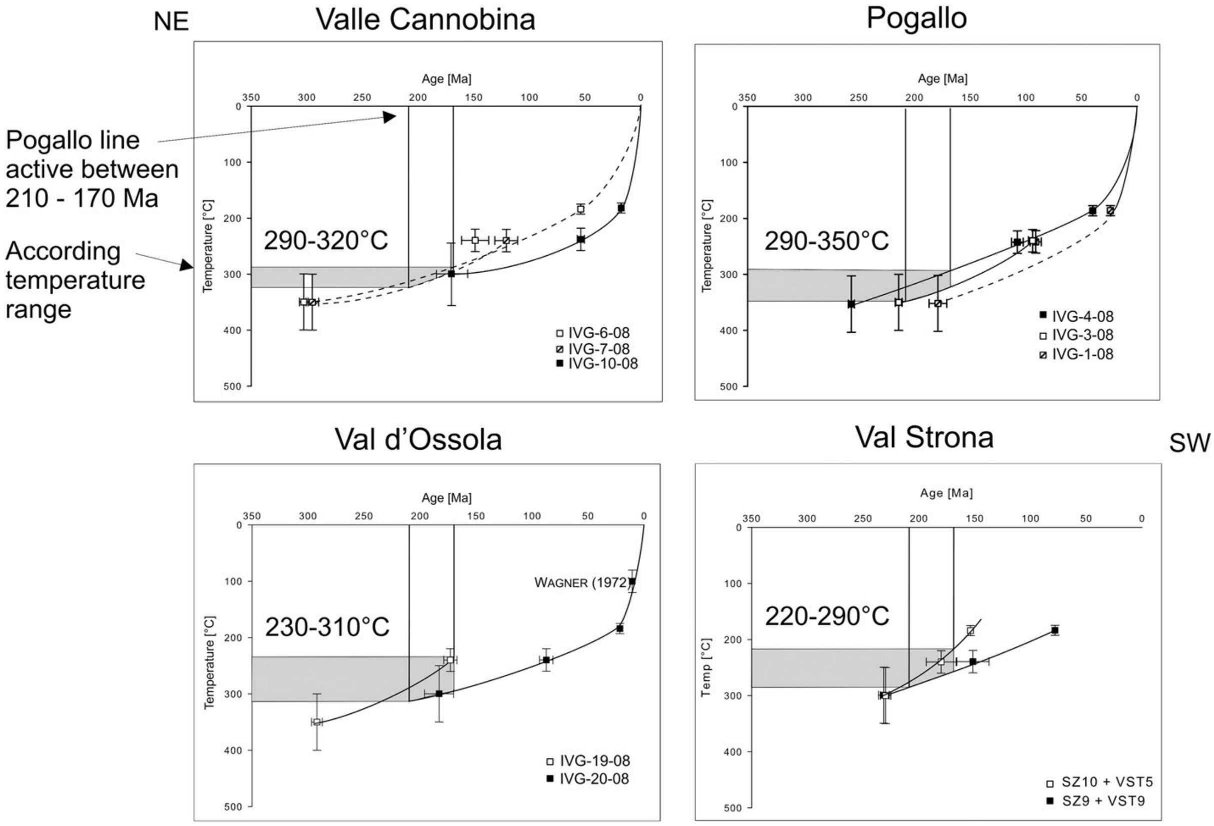


Fig. 10. Apparent cooling paths from samples along the Pogallo Line sorted from the northeast (top left) to the southwest (bottom right). These multi-method thermal histories can be used to determine the temperature (coloured in grey) of the dated rock volumes during the activity period of the Pogallo Line (210 Ma–170 Ma). The temperature decreases from 290 °C to 350 °C in the northeast and ranges 220 °C to 310 °C in the southwest.

(i) Thermal stagnation since 200 Ma until the onset of rapid cooling (argument: LateMesozoic passivemargin sedimentation proves that there was no significant modification in the post-Early Jurassic thermal regime of the Southern Alps during Late Mesozoic time).

(ii) Short and rapid cooling event in the Eocene or Miocene or a combination of two cooling events (argument: these cooling events are well constrained by the new data).

(iii) Slow cooling since the major Tertiary exhumation (argument: apatite FT ages indicate a moderate post-14 Ma exhumation rate (see compilation of apatite ages in Fig. 7). During the modelling the age and rate of cooling events were optimized by multiple iterations in order to obtain the best match of the calculated age profile and the measured ZHe age profile across the IVZ. The best approximation of the age trend (especially in the young part of the Strona profile) was acquired by a two-step cooling model.

In these modelling runs the thermal conditions in the uppermost part of the lithosphere were considered as constant between 200 Ma and 38 Ma. Afterwards an episode of ca. 30 °C cooling until 30 Ma took place, where a final, quick cooling step exhumed the profile to surface temperatures between 15 and 12 Ma. For this assumed thermal history the modelling resulted in a partial retention zone between ca. 120 and 180 °C. The conversion of this temperature range to depth depends on the geothermal gradient. The only available geothermal gradient in the studied tectonic unit derives from the geothermometric data determined on the amphibolite to granulite facies metamorphic rocks along the Strona Valley profile (Henk et al., 1997). These data yielded a geothermal gradient of ca. 22°C/km (calculated from Fig. 11a), but we should consider that the metamorphic paragenesis was formed in Permian time during the thermal climax. If we review the factors determining the post- Permian geothermal gradient of the western Southern Alps in the surrounding area of the IVZ then we should keep in mind two major facts: (i) Jurassic rifting increased the heat flow, but the relaxation should happen in a few 10 Ma, thus this event had no long lasting effect on the ca. 200 to 38 Ma time span. (ii) The proportion of mafic lithologies in the Ivrea body is high and in this Mafic complex the radioactive element content is low (Peressini et al., 2007). Thus, the near-surface radiogenic heat production is reduced in this structural unit, and that the plausible geothermal gradient can be estimated to be between 20 and 30 °C/km during the late Mesozoic and early Cenozoic times.

Assuming this gradient the calculated thickness of the ZHe PRZ should be 3 to 5 km, which is close to the detected thickness of the exhumed static partial retention zones in the case studies mentioned above (e.g. Stockli et al., 2000). The detected helium partial retention zone in the Strona profile is much thicker than the calculated thickness of the retention zone assuming a 20 to 30 °C/km geothermal gradient (Fig. 11b). The best match of measured and calculated profiles can be achieved by assuming a thermal gradient of 7.7 °C/km. This figure is an unrealistic value in an orogen. The problem is probably rooted in the assumed simple geometry of the vertical axis rotation. There are two possible solutions:

(1) The wide PRZ can result from the modification of the geometry of the Ivrea body by stretching. This deformation should happen after the formation of the entire 'normal width' of the zircon PRZ, and under brittle conditions, since the closure temperature of ZHe chronometer is below 200 °C. However, field observations do not prove such an overall and homogeneously distributed stretching. Moreover, the assumption of stretching would also contradict the 'normal' (~22 °C/km) geothermal gradient detected by the Permian mineral assemblages along the profile.

(2) The alternative – andmore plausible – solution is the assumption that the exhumed part of the Ivrea body is intact, it has not experienced significant penetrative deformation in brittle conditions, but its initial position before the Miocene exhumation was not vertical. If it occupied an oblique position, dipping at low angle, before the final exhumation, then the partial retention zone could have developed in an apparently stretched length (see Fig. 11c). If a 20 or 30 °C/km pre-exhumation geothermal gradient is assumed, then the subsurface angle of the currently exhumed and practically horizontal Strona profilewas 23° or 15°, respectively. The Ivrea body should be inclined at a low angle already in the Jurassic, because the development of the zircon PRZ has started ca. 200Ma ago (see the oldest ZFT and ZHe ages in Fig. 9). Themost plausible process for this tilting at lowangle of the IVZ is the normal faulting partly along the Pogallo Line in the Late Triassic to Middle Jurassic (Hodges and Fountain, 1984; Manatschal et al., 2007; and see Fig. 7 in Zingg et al., 1990).

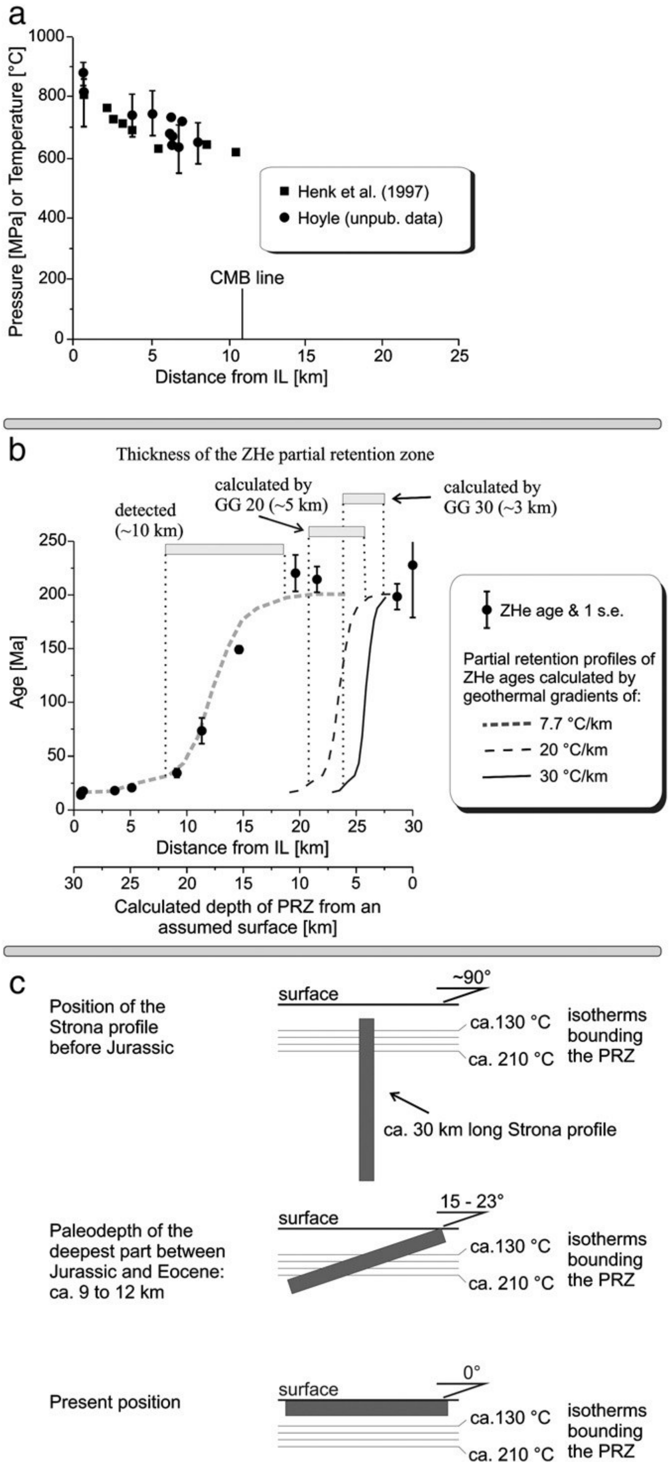


Fig. 11. (a): Peak metamorphic temperature along the Strona profile (from Henk et al., 1997, and Hoyle, unpubl.1999, this laboratory). (b): Measured zircon (U-Th)/He ages plotted across the Ivrea-Verbano Zone and the calculated shape of the ZHe partial retention profiles assuming that (1) the originally vertical profiles are rotated to a horizontal position, (2) the top of the profile is at ca. 30 km distance from the Insubric Line and (3) different geothermal gradients. Assuming a geothermal gradient of 7.7 °C/km results in the best-fit of the measured ages and the calculated curve. Details of the modelling are given in the text. The thickness of the detected PRZ and the calculated ones using gradients of 20 and 30 °C/km (GG 20 and GG 30) are indicated at the top of the figure. (c): Simplified sketch showing the position of the currently exhumed Ivrea-Verbano Zone between the Jurassic and Eocene. Assumption of this position is necessary to get the stretched partial retention zone demonstrated in (b). Angles indicate the dip relative to the present orientation of the profile.

6.5. Driving forces of the Miocene exhumation in the western Southern Alps

The new results indicate that a significant horizontal axis rotation of the IVZ took place in the Jurassic and acquired its final position only in the Middle Miocene, considerably later than the earlier models postulate: Permian rotation was assumed by Boriani and Giobbi (2004) and Oligocene by Schmid et al. (1987, 1989). The low gradient in the ages close to the Insubric Line proves that the final, Middle Miocene exhumation of the IVZ took place at a high slip rate. Only a few apatite FT ages are known from the region, but they are close to the youngest ZHe ages measured along the Insubric Line (14 to 9 Ma AFT vs. 14 Ma ZHe; see Fig. 7). This minor age difference indicates very quick cooling and subsequently minor post-Middle Miocene erosion.

Such rapid exhumations of flat-lying mid-crustal sections may typically be generated by extensional processes (Coney, 1980; Wernicke and Axen, 1988). It generates quick cooling and exposes the formerly buried static partial retention zones of the different thermochronometers (Dokka et al., 1986; Fitzgerald, et al., 1991; Stockli et al., 2000).

We propose that this newly dated Middle Miocene exhumation phase in the western Southern Alps was also linked to extensional tectonics. The detected flat, oblique pre-exhumation position of the Ivrea body also fits this assumption. During the evolution of the Alpine chain, the building up of the relief occurred in the Eocene in the Western Alps and in the Oligocene in the Eastern Alps. The orogenic collapse followed the development of the thickened crustal structure and resulted in several manifestations of orogen-parallel extensions in the Miocene. In theWestern Alps numerous brittle faults indicate extension in theMiocene (e.g. Champagnac et al., 2006; Surace, 2004; Surace et al., 2011). In the Central Alps the Miocene extension along the Simplon fault is well documented under both ductile and brittle conditions (e.g. Grosjean et al., 2004; Mancktelow, 1992). In the Eastern Alps the significant Middle Miocene stretching was localized mainly along the Brenner and Katschberg low angle faults (e.g. Frisch et al., 2000). The major displacement pulse along the Lavanttal faults also took place between 14 and 12Ma (Wölfler et al., 2010). In the eastern continuation of the Eastern Alps the Miocene extension is much more pronounced: the western Pannonian Basin experienced significant vertical thinning in the same time (Tari et al., 1999). In the eastern Southern Alps before the Late Miocene development of the current montaneous relief the southern Dolomites were buried by molasses sediments. In the Langhian (16–13 Ma) a rapid deepening of the Venetian basin to bathial conditions took place (Mellere et al., 2000).

Thus, this study concludes that the detected time range of the final exhumation of the IVZ matches well to the overall orogen parallel extension of the entire Alps (Mancktelow, 1992; Ratschbacher et al., 1989; Selverstone, 2005). Furthermore it may also be a result of the continental scale extension.

7. Conclusions

Based on the new results from this study and combined with previously published thermochronologic data, the understanding of cooling and exhumation of the Ivrea-Verbano Zone, as well as the adjacent SCZ, SLZ and Lower Penninic nappes, can be significantly refined. The upper part of the Ivrea-Verbano Zone and the SCZ had already occupied a near-surface position (below ca. 100 °C) in the Early Jurassic. In contrast, during the Late Cretaceous (80 to 65 Ma; Babist et al., 2006) the SLZ suffered high pressure metamorphism at ca. 60 km depth (Compagnoni, 1977; Dal Piaz et al., 1972). These different histories are illustrated as cooling patterns, which mark intact areas in the western Southern Alps and indicate the migration of exhumation through time (Fig. 8). The Ivrea-Verbano Zone yields the most complete and coherent geochronological data set, and the evolution of this block can be summarized in the following steps:

− The presently exposed part of the Ivrea-Verbano Zone reached its thermal climax in the Permian (Henk et al., 1997; Peressini et al., 2007; Sinigoi et al., 2011). The geobarometric and geothermometric data indicate a Permian geothermal gradient of ca. 22 °C/km (Henk et al., 1997).

− During the Triassic only some thermal relaxation and subsidence may be considered, without significant modification in the thermal structure of the Southern Alpine crust.

− The Early Jurassic rifting increased the heat flow all over the Southern Alps. This heating resulted in the reset of the ZHe thermochronometer in the Baveno granite that already cooled down in the Permian after its emplacement at shallow depth.

− The Jurassic extension uplifted the Ivrea-Verbano body to a shallow to -mid crustal position. The currently exposed section in the Strona Valley occupied an oblique position of ca. 15 to 23° tilt already at the beginning of the Jurassic. The southeastern part of the section was in a near surface position, while the belt along the Insubric Line was in ca. 9–12 km depth, corresponding to temperature conditions of ca. 300 °C and marked by the partial reset of the biotite K/Ar, ZFT and zircon helium ages.

− After Early Jurassic rifting the Mesozoic thermal history of the Southern Alps was presumably static; steady state thermal conditions can be considered for the time span from 200 Ma to ca. 38 Ma. Late Mesozoic passive margin sedimentation proves that there was no significant modification in the post-Early Jurassic thermal regime of the Southern Alps during Late Mesozoic time.

− During the Eocene a rapid exhumation took place. The deepest part of the currently exposed Ivrea-Verbano Zone (ca. 3 km long section) cooled from the ZFT total annealing zone (from ca. 300 °C) at ~38 Ma. This event was probably related to a Paleogene thrusting phase affecting the entire Western Alps (Schmid et al., 1989).

− The final exhumation and tilting of the Ivrea-Verbano Zone is reflected by 14 Ma ZHe ages and can be related to a rapid and pronounced Middle Miocene event. The latest documentation of these movements is the 12.8 Ma illite K/Ar age obtained on fault gouges of the Insubric Line. Apatite fission track data (Hurford, 1991; Wagner and Reimer, 1972) do not show any trend across the IVZ, indicating that further exhumation took place 'en bloc'.

− The final exhumation phase cannot be excluded as being related to the overall orogen-parallel stretching of the Alps. Moreover, this
large-scale process is responsible for the exhumation of the Ivrea body from the shallow and oblique position that was already occupied in the Early Jurassic.

Supplementary data related to this article can be found online at doi:10.1016/j.tecto.2012.03.019.

Thursday 29.8.:
Ascent to the Lago di Cignana (2165 m),
Program: Ultra-high p-rocks with coesite and diamonds, nappes of the Zermatt -. Saas Fee zone, view to the Matterhorn. Total walk: whole days walk along mountain trail, total altitude difference:700 m up and down, Total drive**:** 0 km, Accommodation: ValTournenche

On this day we will visit the UHP Lago di Cignana Unit, hosting exceptionally well preserved UHP assemblages in both eclogites and early-post rift quartzitic metasediments, formed during Eocene subduction of the Jurassic oceanic crust. We will observe eclogites with lawsonite pseudomophs, phengite and glaucophane, and Mn-rich quartzitic metasediments with garnet-, omphacite-, and zoisite-bearing layers. The recent discovery of micro-diamonds in garnets from the Mn-rich quartzites attests to the great depth reached by the Lago di Cignana Unit during subduction. The Lago di Cignana (Cignana Lake) is located in a beautiful high mountain landscape not far from the Matterhorn visible during the uphill driving.

Schistes Lustres-Lago di Cignana HP/UHP Metasedimentary Rocks as a Reference Suite for Study of Chemical Cycling in Subducting Sediments

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General Information

Metapelitic rocks of the Schistes Lustrés in the Cottian Alps, Italy (peak metamorphic conditions of 350-500 °C, 1.2-2.0 GPa), and the UHP Lago di Cignana locality (Valtournenche, Italy; ~600 °C, 3.4 GPa; coesite- and microdiamond-bearing), preserve records of prograde devolatilization in their mineral modes and chemistry, contents of volatiles and more fluid-mobile elements, and stable isotope compositions. Together, these units allow evaluation of devolatilization history in metasedimentary rocks that experienced high-P/T prograde paths similar to those experienced in most modern subduction zones. Calculated prograde dehydration histories using the Perple-X database indicate that significant (average of 20%) dehydration occurred over the temperature interval of 450 to 510 °C largely related to the breakdown of chlorite and, to a lesser extent, carpholite. Ion microprobe analyses of phengites show subtle decrease in B concentration, with increasing grade, but impressive uniformity across grade of other trace elements thought to be relatively fluid-mobile (Ba, Cs, and the somewhat less fluid-mobile Ba). Across grade, whole-rock samples show uniform ratios of the concentrations of N and the more fluid-mobile elements B and Cs to the concentrations of the less mobile K_2O, Rb, and Li, with only minor hints of Cs loss in some of the highest grade samples. Whole-rock _^{15}N shows a hint of shift to higher values in the highest grade rocks at Lago di Cignana. Small amounts of loss of N into H_2O-rich fluids released from these rocks could have produced minor shift in _^{15}N, with the related shift in whole-rock N concentration masked by the large degree of heterogeneity inherent with the sedimentary protoliths. Partitioning of B and Cs from white micas into H_2O-rich fluids produced by chlorite breakdown could similarly have produced the subtle decreases in the concentrations of these elements noted in some high-grade samples. Neoblastic tourmaline in higher-grade rocks likely sequestered some fraction of the B lost from micas, resulting in a lack of obvious whole-rock B loss to accompany the up-grade trend of decreasing B concentrations in phengite. This tourmaline shows core-to-rim decrease in _^{11}B consistent with growth during small amounts of progressive B loss from phengites. Taken together, the whole-rock and SIMS data presented here demonstrate impressive retention, during prograde forearc devolatilization in a cold subduction environment, of components thought to be relatively fluid-mobile (particularly N, B, Li, Ba, and Cs). Despite some minor uncertainties regarding the degree to which exhumation-related recrystallization, in particular at the higher grades, resulted in obscured prograde trends, the Schistes

Lustrés/Cignana traverse can serve as a reference suite for examining the effects of subduction-zone devolatilization on selected trace element and isotopic tracers.

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Jurassic formation and Eocene subduction of the Zermatt–Saas-Fee ophiolites: implications for the geodynamic evolution of the Central and Western Alps

Daniela Rubatto; Dieter Gebauer; Mark Fanning

The Zermatt–Saas-Fee ophiolites (ZSFO) are one of the best preserved slices of eclogitic oceanic crust in the Alpine chain. They formed during the opening of the Mesozoic Tethys and underwent subduction to HP/UHP conditions during Alpine compression. A cathodoluminescence-based ion microprobe (SHRIMP) dating of different zircon domains from metagabbros and oceanic metasediments was carried out to constrain the timing of formation and subduction of this ophiolite, two fundamental questions in Alpine geodynamics. The formation of the ophiolitic sequence is constrained by the intrusion ages of the Mellichen and the Allalin metagabbros (164.0 ± 2.7 Ma and 163.5 ± 1.8 Ma) obtained on magmatic zircon domains. These data are in line with the maximum deposition age for Mn-rich metasediments which overlie the mafic rocks at Lago di Cignana (161 ± 11 Ma) and at Sparrenflue (ca. 153–154 Ma). An Eocene age of 44.1 ± 0.7 Ma was obtained for whole zircons and zircon rims from an UHP eclogite and two metasediments at Lago di Cignana. One of the Eocene zircons contains a rutile inclusion indicating formation at HP conditions. As the temperature and pressure peak of these rocks nearly coincide, the Eocene zircons probably constrain the age for the deepest subduction of the ZSFO. This Eocene age for the UHP metamorphism implies that the ZSFO were subducted later than the Adriatic margin (Sesia-Lanzo Zone) and before the Late Eocene subduction of the European continental crust below Apulia. A scenario with three subduction episodes propagating in time from SE to NW is proposed for the geological evolution of the Central and Western Alps.

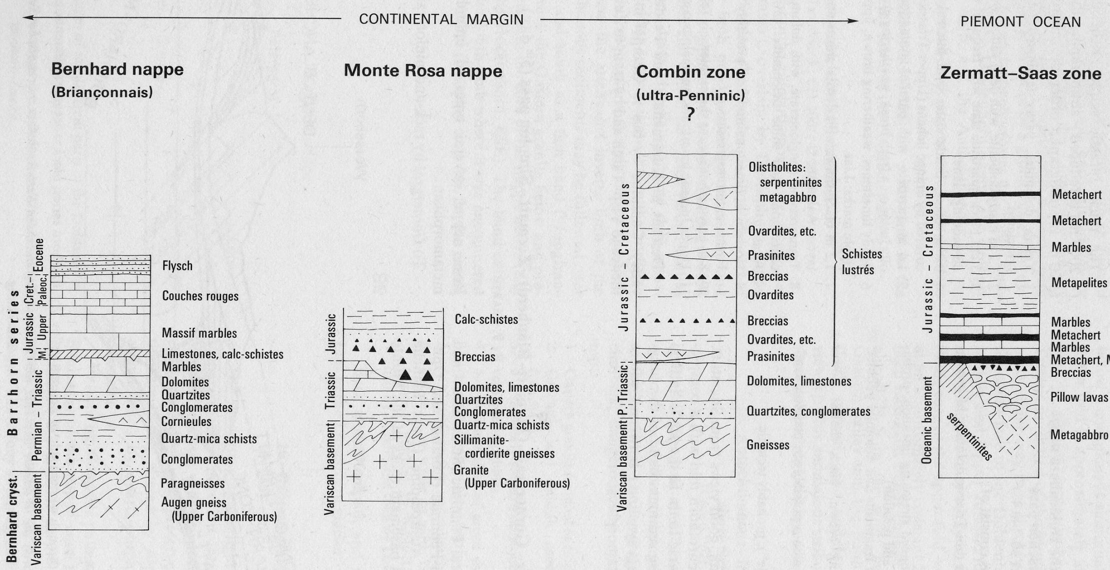

Fig. 17. Stratigraphic columns of the tectonic units in the Zermatt area.

Fig. 17. Stratigraphische Profile der tektonischen Einheiten von Zermatt.

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HP metamorphic belt of the western Alps

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The understanding of the subduction-related processes benefited by the studies of the high-pressure (HP) metamorphicrocks from the western Alps. The most stimulating information was obtained from the inner part of the western Alpine belt, where most tectonic units show an early Alpine eclogite-facies recrystallisation. This is especially true for the Austroalpine Sesia Zone and the Penninic Dora-Maira massif. From the Sesia zone, which consists of a wide spectrum of continental crust lithologies recrystallised to quartz-eclogite-facies mineral assemblages, the first finding of a jadeite-bearing meta-granitoid has been described, supporting evidence that even continental crust may subduct into the mantle. From the Dora-Maira massif the first occurrence of regional metamorphic coesite has been reported, opening the new fertile field of the ultrahigh-pressure metamorphism (UHPM), which is now becoming the rule in the collisional orogenic belts.

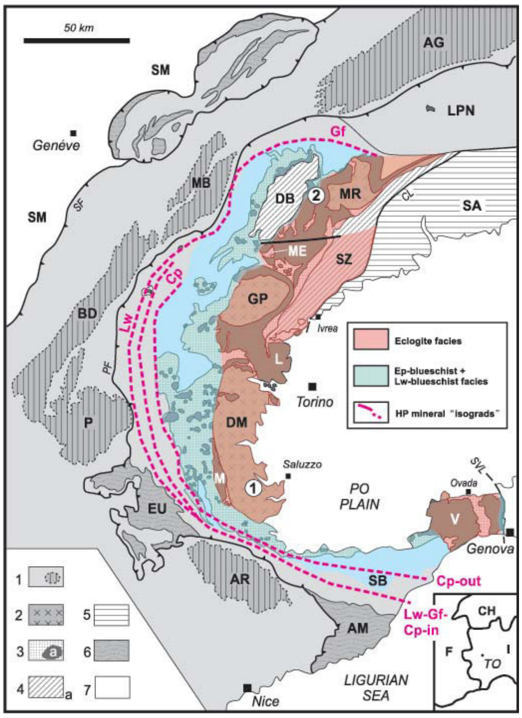

Figure 1 Simplified structural sketch-map of the Western Alps.
1: Jura, Helvetic Domain and external Penninic Domain. The dashed line contours the External Crystalline Massifs (AR: Argentera, P: Pelvoux, BD: Belledonne, MB: Mont Blanc- Aiguilles-Rouges, AG: Aar-Gotthard). SB: Grand St. Bernard Zone, LPN: lower Penninic nappes. 2: Internal Crystalline Massifs of the Penninic Domain (MR: Monte Rosa, GP: Gran Paradiso, DM: Dora-Maira). 3: Piemonte Zone (L: Lanzo Massif; M: Monviso Massif; V: Voltri Massif) and a) main meta-ophiolite bodies. 4: Austroalpine Domain (DB: Dent Blanche nappe, ME: Monte Emilius, SZ: Sesia Zone). 5: Southalpine Domain (SA). 6: Helminthoid Flysch nappes (EU: Embrunais-Ubaye, AM: Alpes maritimes). 7: Swiss Molasse (SM), Po Plain and Piemontese-Ligurian Tertiary basin. CL: Canavese line; SVL: Sestri-Voltaggio line; SF: Subalpine frontal thrust; PF: Penninic thrust front. (1) UHP Brossasco-Isasca Unit, Dora-Maira Massif; (2) UHP Lago di Cignana unit, upper Valtournenche, Piemonte Zone. Gf-in, from Bocquet (1974); Lw-in, and Cp-in and Cp-out from Goffé and Chopin (1986). In the inset: CH: Switzerland; F: France; I: Italy; TO: Torino.

Introduction

The western Alps extends from the Sestri-Voltaggio tectonic Line, which separates it from the non-metamorphic rocks of the Apennine chain, to the Lower Penninic Nappes of the Lepontine dome (LPN, Figure 1). On the internal side, the western Alps are bounded by the Quaternary post-orogenic clastic deposits of the Po Plain up to about the latitude of Torino, and from there northwards by the Canavese tectonic Line (CL), the SW extension of the Insubric Line (or Periadriatic lineament), which separates the pre-Alpine domain of the Southern Alps (Ivrea Zone + Strona-Ceneri Zone) from the western Alpine chain reworked during the Alpine orogeny.

The HP belt of the western Alps comprises most tectonic units of the Penninic and Austroalpine domains (Figure 1). The Penninic Domain is a heterogeneous realm, which consists of both continent- and ocean-derived tectonic units (see Dal Piaz et al., this issue). The continent-derived units are (from the internal concave side toward the external convex side): the *Austroalpine Sesia-Lanzo Zone—Dent Blanche nappe system*, the "*Internal Crystalline Massifs*" of Monte Rosa (MR), Gran Paradiso (GP) and Dora- Maira (DM), and the *Briançonnais Zone* (or Grand Saint Bernard nappe system), which overthrust the Helvetic Domain: this tectonic boundary is known as "Penninic Thrust Front" (PTF) (Figure 1). The units derived from the Mesozoic Tethys ocean make up the *Piemonte Zone*, also named "Zone of calc-schist (French: "*schistes lustrés*") with meta-ophiolite", which consists of a number of thrust sheets with different high-pressure metamorphic recrystallisations (Figure 1).

The Austroalpine domain includes the Sesia-Lanzo zone (in the following referred to as the Sesia Zone) and the Dent Blanche Nappe system, which are fragments of Variscan granulite to amphibolite facies continental crust intruded by late-Variscan granitoids, derived from the Southalpine (or Insubric) plate. The "Eclogitic Micaschist Complex" (EMC) of the Sesia zone and the Monte Emilius klippe of the Dent Blanche nappe are the best preserved examples of continental crust basement recrystallised under eclogite-facies conditions (Compagnoni, 1977; Dal Piaz et al., 1983).

Due to the widespread occurrence of unaltered HP metamorphic rocks, the western Alps have been the ideal area for the study of this type of metamorphism. Contributions to the knowledge of the HP metamorphism go back to the end of 19th century. The first eclogite-facies metapelite (named "eclogitic micaschist") was reported by Stella (1894) from the EMC of the Sesia zone and later studied by Franchi (1900, 1902), who also described the reaction jadeite+quartz⇒albite (Franchi, 1902); the first eclogitised pillowed basalts were described by Bearth (1959) from the meta-ophiolites of the Piemonte zone from the Zermatt valley; the first eclogite-facies jadeite-bearing metagranite was reported by Compagnoni and Maffeo (1973) at Mt. Mucrone from the EMC of the Sesia zone; the first coesite in continental crust was reported by Chopin (1984) from the southern Dora-Maira massif; and the petrologic importance of Mg- Fe-carpholitess for the blueschist-facies was first recognised by Goffe and Chopin (1986), studying Brianconnais lithologies previously considered unimportant for geobarometric estimates. And finally, let's remember that the high-density, hard and tough Neolithic stone implements, excavated from all over the western Europe, are made of eclogite and jadeitite derived from the HP meta-ophiolites of the Piemonte zone (Compagnoni et al., 1996).

Regional distribution of the HP metamorphism

Since Bearth's (1962) pioneering work, a number of attempts were made to trace the "isograds" of the HP metamorphism (for a review see Desmons et al., 1999). Many published and unpublished petrologic data were first summarised in the *Metamorphic Map of the Alps* (sheet 17 of the 1:1,000,000 *Metamorphic Map of Europe*, edited by Zwart, 1973) and then in the 1:500,000 *Map of the Alpine Metamorphism* in the *New Metamorphic Map of the Alps* (Frey et al., 1999).

Broadly speaking, at a regional scale the HP metamorphism in the western Alps includes from E to W eclogite-facies, epidote- and lawsonite-blueschist facies, and lawsonite-albite-chlorite facies rocks (Figure 1). The quartz-eclogite facies units prevail, but two small coesite-eclogite facies units have been recognised: the Brossasco-Isasca unit (BIU) from the southern Dora-Maira massif (Figures 1 and 2) and the Lago di Cignana unit from the Piemonte zone (Figure 1). This large-scale metamorphic zoning is considered as evidence showing that the northern European plate was subducted below the Adrian (or Insubric) plate (Ernst, 1971; Dal Piaz et al., 1972).

"Isograds" for most significant blueschist-facies minerals, such as glucophane-in (Gf-in), lawsonite-in (Lw-in), and carpholite-in (Cp-in)- and-out (Cp-out) have been traced in the western Alpine belt, and provide a large-scale idea of their mineral zoneography in the most external Pennine zone (Figure 1). However, the increase of detailed petrographic studies showed that the HP mineral distribution is really much more complicated than originally supposed. For example, in the Aosta valley, the tectonometamorphic setting of the Piemonte zone consists of a quartz-eclogite facies unit or composite lithotectonic unit ("Zermatt-Saas zone": Bearth, 1967) overlain by an epidote- blueschist facies unit or composite lithotectonic unit association ("Combin zone"). The main tectonic contact between the two zones is marked by the presence of a thin coesite-eclogite facies meta-ophiolitic unit ("Lago di Cignana Unit": Reinecke, 1991) and of several Austroalpine continental crust slices, showing an Alpine quartz eclogite-facies overprint. This complex tectonometamorphic setting indicates the presence of a polyphase tectonic evolution, involving both compressional and extensional large-scale processes.

Figure 2 illustrates the southern Dora-Maira massif, where a tectonic thrust sheet with Alpine coesite-eclogite facies overprint (Brossasco-Isasca Unit: T≈750˚C, P≈3.5 GPa) is sandwiched between two quartz-eclogite facies units (San Chiaffredo and Rocca Solei units: ≈550˚C, ≈1.5 GPa), bounded in turn by an epidote-blueschist facies unit (Pinerolo unit: ≈450˚C, ≈ 0.8 GPa) and a quartz-eclogite-facies unit (Dronero-Sampeyre unit: ≈550˚C, ≈1.5 GPa) (Compagnoni and Rolfo, 2003 with ref. therein).

Peculiarities of the HP metamorphism

The Alpine eclogite-facies recrystallisation is especially well developed in the Eclogitic Micaschist Complex of the internal Sesia Zone, in the Monte Emilius klippe of the Dent Blanche nappe, and in the meta-ophiolites of the Piemonte Zone, while relics of eclogite-facies lithologies occur in the Internal Crystalline Massifs of Monte Rosa, Gran Paradiso, and Dora-Maira (Figures 1 and 2). Most units have been recrystallised under quartz-eclogite-facies conditions, except for the two small units of Brossasco-Isasca (BIU) (T=730 ˚C and P≥3.3 GPa) from the southern Dora-Maira Massif (Figure 2) and Lago di Cignana (T≈600 ˚C and P=2.6–2.8 GPa) from the Piemonte zone, which contain coesite relics.

HP metamorphism in the oceanic lithosphere

Historically important are the pillow metabasalts of the Piemonte zone (Zermatt Saas zone: see later on) described by Bearth (1959) for the Zermatt valley, in which the dry pillow core was converted to a coarse-grained bimineralic eclogite assemblage, whereas the hydrous rim was replaced by a glaucophane-rich rock. In the Monviso ophiolite of the Cottian Alps (Figure 1), a complete section of oceanic crust is exposed, including poorly deformed isotropic to layered cumulus gabbros, massive to pillowed basalts, and basaltic dykes, which recrystallised under eclogite-facies conditions (Schwartz et al., 2000).

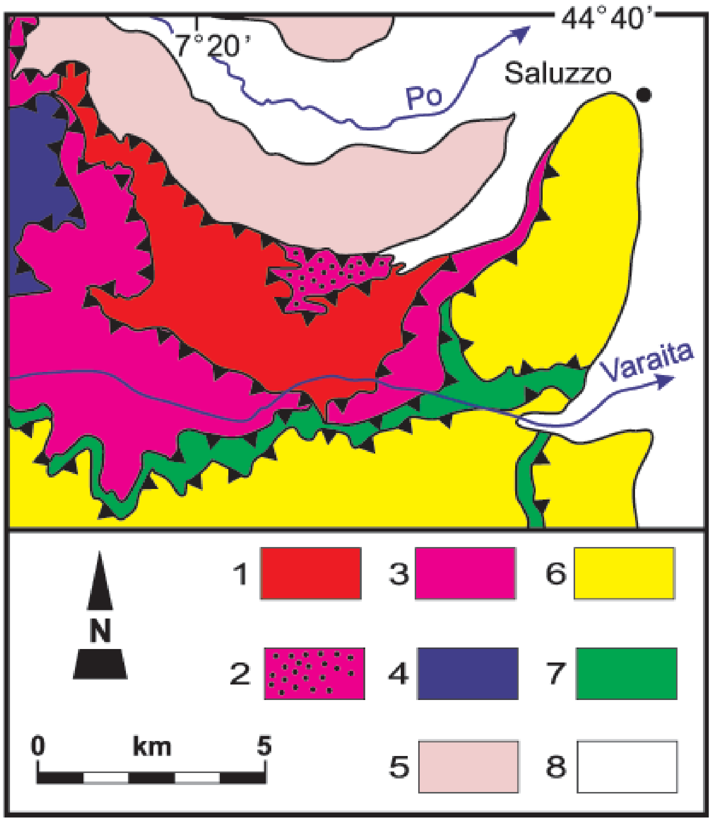

Figure 2 Structural sketch map of southern Dora-Maira Massif. Continent-derived Dora-Maira Massif: Coesite-eclogite facies Brossasco-Isasca Unit (1); Quartz-eclogite facies San Chiaffredo Unit (2); Quartz-eclogite facies Rocca Solei Unit (3); Pre-Alpine basement and Permo-Triassic cover unit (4); Epidote-blueschist facies Pinerolo Unit (5); Quartz-eclogite facies Dronero-Sampeyre Unit (6); Ocean-derived Piemonte Zone: Epidoteblueschist facies metaophiolites and oceanic metasediments (7); Quaternary alluvial deposits (8).

The Lanzo ultramafic massif (Figure 1) is the largest portion of upper mantle peridotite exposed in the western Alps, which experienced Alpine eclogite facies metamorphism with P in excess of 2.0 GPa and T= 550˚C. Its central less-deformed portion still consists of spinel-plagioclase lherzolite and minor harzburgite—so fresh that the high-T petrology and deformation may be studied (Boudier, 1976), and gabbro dykes are locally found which may preserve the original Ca-rich plagioclase. By contrast, its marginal portion has been converted to sheared serpentinite, which contains the eclogitefacies assemblage: antigorite, metamorphic olivine (Fe-richer than peridotitic olivine), clinohumite (mostly red-brown titanian clinohumite) magnetite, Fe-Ti alloys±diopside±Mg-chlorite±apatite. Similar mineral assemblages may be found in the antigorite serpentinite of the whole eclogite-facies internal Piemonte zone, where olivine+Ti-clinohumite+Mg-chlorite± apatite metamorphic veins occur, which in part at least developed at the expense of primary igneous dykes rich in ilmenite (Scambelluri and Rampone, 1999).

Poorly deformed metagabbros are the best-preserved ophiolitic lithologies, since the coarse grain-size and dry composition favoured preservation of the igneous protolith. Let's first mention the Allalin metagabbro (Bearth, 1967) from the Swiss Valais, in which all reaction steps from the original olivine gabbronorite to a coarse-grained eclogite may be observed (Meyer, 1983). The poorly-deformed metagabbros are characterised by a more complex mineral association than the sheared gabbros, because pseudomorphous reactions develop after each igneous mineral and coronitic reactions at the original boundaries between igneous minerals, respectively. For example, in coronitic Mg-Al-metagabbros, diopside-rich omphacite+talc develops after igneous clinopyroxene, jadeite-rich omphacite+garnet+ zoisite±quartz after plagioclase, and omphacite+talc+Naamphibole after olivine. In coronitic Fe-Ti-metagabbros ferrian omphacite develops after igneous clinopyroxene, omphacite+ garnet after plagioclase, and Na-amphibole+rutile after the hornblendeilmenite intergrowth (Messiga and Scambelluri, 1988). In the pervasively deformed portions, Fe-Ti-metagabbros recrystallised to omphacite+garnet+rutile±glaucophane±epidote assemblages, whereas Mg-Al-metagabbros recrystallised to omphacite+garnet+rutile±chloritoid± talc±chlorite assemblages. In the so-called "smaragdite" metagabbros, the original igneous Cr-diopside has been replaced by bright-green "smaragdite" (i.e., a Cr-bearing omphacite±tremolite± talc), plagioclase by jadeite-rich pyroxene+ zoisite, olivine by talc±tremolite, and ilmenite by rutile.

Mineral assemblages and mineral compositions are closely related to bulk rock chemistry: for example, in gabbros with high Mg/(Mg+Fe) ratio (such as Mg-Al-gabbros), Mg-Al minerals (such as chlorite and/or chloritoid) form, whereas the bimineralic omphacite+garnet (+rutile) eclogite assemblage is produced only in Al-poor, Fe-Ti-rich gabbros. Similarly, garnet is almandine-richer in Fe- Ti-metagabbros, whereas it is pyrope-richer in Mg-Al-metagabbros.
Lawsonite-eclogites are extremely rare in the western Alpine belt and confined to some small tectonic units of Liguria. On the contrary, the widespread occurrence of lozenge-shaped whitemica+ paragonite pseudomorphs up to about two cm-long indicates that porphyroblastic lawsonite was common during prograde eclogite- facies metamorphism.
In eclogites, veins locally occur, which indicate that the eclogite- facies metamorphism developed in the presence of a hydrous fluid phase. However, the fluid flow was limited during eclogitefacies metamorphism and fluid was mainly released by devolatilisation reactions of dense hydrous silicates (such as lawsonite) or during plastic crystal flow (Philippot, 1993).
Associated with eclogites derived from Fe-Ti-gabbros, very minor felsic rocks (plagiogranite or sodagranite) are locally found, which exceptionally occur as km-sized bodies (Castelli et al., 2002 with ref. therein): they are the only lithology among meta-ophiolites containing the association jadeite+quartz±garnet.

In lithologies of unusual compositions belonging to the layered gabbros sequence, Cr-Mg-chloritoid and other Cr-bearing minerals have been found, most likely formed at the expense of chromite-rich bands of the cumulus layered gabbros (Messiga et al., 1999). Rodingites, derived from both gabbroic or basaltic dykes, are ubiquitous within most serpentinised peridotites and have a typical paragenesis of diopside, vesuvianite, ugranditic garnet, epidote, and Mg-chlorite; however, the local presence of Alm-rich garnet and omphacite indicates that some of them formed during HP metamorphism.

In the calc-schists and associated metasediments, eclogitefacies assemblages are more difficult to identify because typically they are more easily retrogressed. However, for suitable compositions, the eclogite assemblage appears to be phengite+Alm-rich garnet+ rutile± paragonite±zoisite±chloritoid±porphyroblastic lawsonite, mostly replaced by paragonite+epidote pseudomorphs. Noteworthy is also the Mn-deposit of Praborna, Saint Marcel (Aosta Valley), where a number of eclogite-facies Mn-bearing minerals or mineral varieties have been described (Martin and Kienast, 1987).

HP metamorphism in continental crust
The best preserved portion of eclogitised continental crust in the western Alps is the Eclogitic Micaschist Complex of the Sesia Zone. In this complex, derived from a Variscan amphibolite-facies basement intruded by late-Variscan granitoids (Compagnoni, 1977), the whole spectrum of continental lithologies from paragneiss to orthogneiss, from marble to metabasic rock, may contain one or more quartz-eclogite facies minerals, such as garnet, Na-pyroxene (omphacite and jadeite), high-celadonite phengite (3T polytype), paragonite, glaucophane, zoisite, and chloritoid. Well-preserved relics of Variscan crystalline basement also occur in the Internal Crystalline Massifs. The best examples of undeformed lithologies are the coronitic metagranitoids, which are exposed in the central Gran Paradiso massif, at Monte Mucrone in the "Eclogitic micaschist Complex" (EMC) of the Sesia zone, and in the Brossasco-Isasca unit (BIU) of the Dora-Maira massif. In all three units, metabasic rocks have been converted to the eclogite assemblage, but metagranitoids show significant differences: in the Gran Paradiso metagranitoids (about 400–500 ˚C and about 1.0 GPa) the igneous structure and mineralogy are mainly preserved; in the Monte Mucrone metagranitoids, Sesia Zone (about 550 ºC and 1.6–1.8 GPa) igneous quartz and K-feldspar are preserved, but plagioclase is pseudomorphically replaced by jadeite+zoisite+quartz± kyanite±K-feldspar; in the BIU metagranitoids, Dora-Maira massif (about 750 ˚C and 3.5 GPa) igneous quartz is also replaced by a granoblastic polygonal aggregate of metamorphic quartz, inverted from former peak pressure coesite (Biino and Compagnoni, 1992).

However, the most interesting continental crust lithology of the western Alpine HP belt is the "silvery micaschist", which occurs as cm- to m-thick lens-like layers within the orthogneiss of the Internal Crystalline Massifs and is interpreted as a metasomatic rock formed at the expense of a granitoid protolith along shear zones (Compagnoni and Hirajima, 2001, with ref. therein). This rock, which is important petrologically because its phase relationships can be modelled in the relatively simple KMASH system (Chopin, 1981), is usually referred to as 'whiteschist" for its characteristic kyanite+talc (or talc+phengite) assemblage, indicative of pressures in excess of about 1.0 GPa (Schreyer, 1977). The best-known and unique white schist is that from the Dora-Maira massif first described by Chopin (1984), which contains in addition to kyanite, phengite, talc, rutile, and accessories, a pale-pink idioblastic pyrope (up to 98 mole% of the Mg-Al end-member). Pyrope crystals are up to 20 cm across and host, in addition to abundant kyanite, unusually large coesite relics.

A number of new minerals (such as bearthite, ellenbergerite and phosphellenbergerite) or minerals of unusual composition (such as magnesiodumortierite, magnesiochoritoid, magnesiostaurolite), suggestive of UHPM conditions, have been found in the pyrope crystals (Chopin and Ferraris, 2003 with ref. therein).

Age of HP metamorphism
Up to the end of the 1980s, most radiometric ages relevant to the HP metamorphism from the western Alps, recognised as indicative of an Early-Alpine or Eoalpine metamorphic event fell into the range of 140–60 Ma, further subdivided into an early eclogite-facies stage (140–85 Ma) and a later blueschist-facies and cooling stage (85–60 Ma) (for a review see: Hunziker et al., 1992). The younger Tertiary ages (around 50 Ma) were interpreted as the evidence of two episodes of HP metamorphism (Monie et al., 1989).

The first work, which suggested the possibility of a Tertiary age for the eclogite-facies metamorphism from the western Alps, was that by Tilton et al. (1991, with ref. therein) who dated U-Pb on minerals from the UHPM Brossasco-Isasca Unit (BIU), southern Dora-Maira massif (DMM). However, the major change in the age determination of the eclogite-facies metamorphism in the western Alps was brought about by the *in-situ* U-Pb dating of high closure-temperature minerals, such as zircon, with the Sensitive High Resolution Ion-Microprobe (SHRIMP), assisted by cathodoluminescence imaging (Rubatto et al. 2003, with ref. therein). The most complete geochronological work was done on the UHPM rocks of the southern Dora-Maira massif, previously studied by Tilton et al. (1991) (Gebauer et al., 1997 with ref. therein): the new U-Pb ages, consistently fell at the Eocene-Oligocene boundary (~35 Ma) and were later confirmed by the results from Lu- Hf dating of garnet (Duchene et al., 1997).

At present, with the exception of the Cretaceous age (~ 65 Ma) of the quartz eclogite-facies metamorphism of the Sesia zone (Rubatto et al. 2003), other U-Pb SHRIMP ages on the western Alps eclogites are all Tertiary (35–45 Ma).

P-T-t paths of the HP metamorphism

The tectonic units of the western Alps recrystallised under eclogitefacies conditions, independently of their peak P conditions, consistently show a similar P-T-path, which is characterised by two thermal peaks. The first thermal peak, which corresponds to the high- to ultrahigh-pressure climax, is followed by significant decompression coupled with moderate cooling. The second thermal peak is at low-P (about 0.4–0.6 GPa), and corresponds to the greenschist-facies event of the Alpine literature. It is followed by significant cooling coupled with moderate decompression. In Figure 3, the P-T paths of the most significant units of continental crust (i.e. Monte Rosa, Sesia zone and BIU) are reported. From their comparison, it is evident that all the P-T paths are clockwise, and that the highest is the peak pressure, the tightest is the P-T loop. This feature, initially difficult to explain, is now believed to be the best evidence that the whole subduction/ exhumation process is much faster than formerly supposed. For example, subduction and exhumation speeds of about 2 cm/a have been suggested by Rubatto et al. (1999) for the EMC of the Sesia zone and by Rubatto and Hermann (2001) for the BIU of the Dora-Maira massif.

Conclusions

The existence of two age clusters, initially interpreted as evidence for two metamorphic events of Cretaceous and Tertiary ages, affecting the whole western Alps, have turned out to be the record of separate events in different units. Therefore, the different units were subducted and exhumed at different times, implying that at the same time subduction and exhumation were active in different portions of the orogenic belt. This diachrony of the HP and UHP metamorphism of the Western Alps gives rise also to a nomenclature problem, since the so-called early-Alpine or Eoalpine metamorphism, previously considered to be Cretaceous in age and subduction-related, is not a single coeval event throughout the western Alps.

The preservation in both ocean- and continent-derived units of poorly deformed lithologies, which preserve the protolith structure and even part of the primary mineralogy, indicates that at any scale the HP deformation (and metamorphic recrystallisation) mainly occurred along shear zones. This mechanism was favoured by both the relatively low fluid content in continental and oceanic lithologies and the unusually high speed of the whole subduction/exhumation process, which took place in a time span of only a few millions of years. The high speed of the tectonic processes, the scarcity of a free fluid phase, and the continuous cooling during exhumation also account for the local superb preservation of the HP peak mineral assemblages.

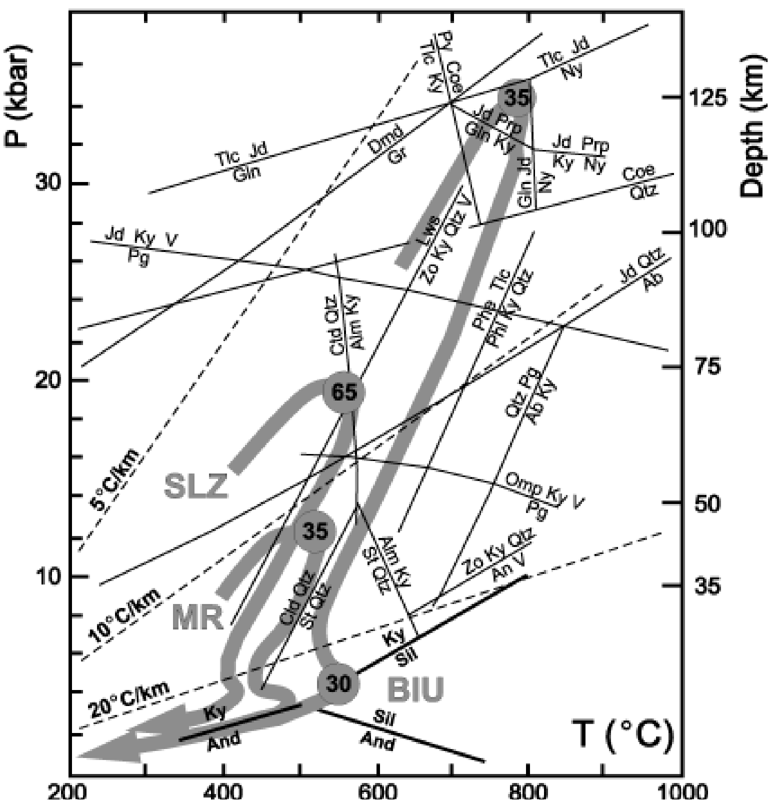

Figure 3 P-T-t paths of representative continental eclogite-facies units from the western Alps. BIU: UHPM Brossasco-Isasca Unit of Penninic Dora-Maira Massif (from Compagnoni et al., 1995, modified). SLZ: Austroalpine continental Sesia Lanzo Zone (from Tropper and Essene, 2002, modified). MR: Penninic Monte Rosa nappe of the Internal Crystalline Massifs (Borghi et al., 1996). Mineral abbreviations after Bucher and Frey (1994). Circled numbers are SHRIMP radiometric ages in Ma for BIU (Rubatto and Hermann, 2001), SLZ (Rubatto et al., 1999), and MR (Rubatto and Gebauer, 1999), respectively.

Acknowledgements
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References

Bearth P., 1959, Uber Eklogite, Glaukophanschiefer und metamorphe Pillowlaven. Schweiz. mineral. petrogr. Mitt., v. 39, pp. 267-286.
Bearth P., 1962, Versuch einer Gliederung alpinmetamorpher Serien der Westalpen. Schweiz. mineral. petrogr. Mitt., v. 42, pp. 127-137.
Bearth P., 1967, Die Ophiolithe der Zone von Zermatt-Saas Fee. Beitrage der geologische Karte der Schweiz, NF, v. 132, pp. 130.
Biino G. and Compagnoni R., 1992, Very-high-pressure metamorphism of the Brossasco coronite metagranite, southern Dora-Maira Massif, western Alps. Schweiz. mineral. petrogr. Mitt., v. 72, pp. 347-362.
Borghi A., Compagnoni R. and Sandrone R., 1996, Composite P-T paths in the internal penninic massifs of the western Alps: petrological constraints to their thermo-mechanical evolution. Eclogae geol. Helv., v. 89, pp. 345-367.
Bocquet (Desmons) J., 1974, Etudes mineralogiques et petrologiques sur les metamorphismes d'age alpin dans les Alpes francaises. These, Grenoble, 489 p.
Boudier F., 1976, Le massif lherzolitique de Lanzo (Alpes piemontaises): Etude structurale et petrographique. PhD Thesis, Univ. Nantes, 163 p.
Bucher K. and Frey M. (1994) Petrogenesis of metamorphic rocks. Springer-Verlag, 318 p.

Castelli D., Rostagno C. and Lombardo B., 2002, Jd-Qtz-bearing metaplagiogranite from the Monviso meta-ophiolite (Western Alps). Ofioliti, v. 27, pp. 81-90.
Chopin Ch., 1981, Talc-phengite, a widespread assemblage in high-grade pelitic blueschists of the western Alps. J. Petrol., v. 22, pp. 628-650.
Chopin Ch., 1984, Coesite and pure pyrope in high-grade blueschists of the Western Alps: a first record and some consequences. Contrib. Mineral. Petrol., v. 86, pp. 107-118.
Chopin Ch. and Ferraris G., 2003, Mineral chemistry and mineral reactions in UHPM rocks,. *in* A. Carswell and R. Compagnoni, eds, Ultrahigh pressure metamorphism. EMU Notes in Mineralogy, v. 5, Chapter 3.
Compagnoni R., 1977, The Sesia-Lanzo Zone: high pressure-low temperature metamorphism in the Austroalpine continental margin. Rend. Soc. Ital. Miner. Petrol., v. 33, pp. 335-374.
Compagnoni R. and Maffeo B., 1973, Jadeite-bearing Metagranites l.s. and Related Rocks in the Mount Mucrone Area (Sesia-Lanzo Zone, Western Italian Alps). Schweiz. mineral. petrogr. Mitt., v. 53, pp. 355-378.
Compagnoni R., Ricq-de-Bouard M., Giustetto R. and Colombo F., 1995, Eclogite and Na-pyroxenite stone axes of southwestern Europe: a preliminary petrologic survey, *in* B. Lombardo, ed, Studies on metamorphic rocks and minerals of the western Alps. A volume in memory of Ugo Pognante. Boll. Museo Regionale Sci. Nat. (Torino), v. 13, suppl. 2, pp. 329-359.
Compagnoni R. and Hirajima T., 2001, Superzoned garnets in the coesitebearing Brossasco-Isasca Unit, Dora-Maira massif, western Alps, and the origin of the whiteschists. Lithos, v. 57, pp. 219-236.
Compagnoni R. and Rolfo F., 2003, The most representative UHP metamorphic units in the world: The Western Alps, *in* A. Carswell and R. Compagnoni, eds, Ultrahigh pressure metamorphism. EMU Notes in Mineralogy, v. 5, Chapter 2.1.
Dal Piaz G.V., Hunziker J.C. and Martinotti G., 1972, La Zona Sesia-Lanzo e l'evoluzione tettono-metamorfica della Alpi nord-occidentali interne. Mem. Soc. Geol. Ital., v. 11, pp. 433-460.
Dal Piaz G.V., Lombardo B. and Gosso G., 1983, Metamorphic evolution of the Mt. Emilius klippe, Dent Blanche nappe, Western Alps. Amer. J. Sci., v. 283A, pp. 438-458.
Desmons J., Compagnoni R. and Cortesogno L., with the collaboration of Frey M., Gaggero L., Dallagiovanna G., Seno S. and Radelli L., 1999, Alpine metamorphism of the Western Alps: II. High-P/T and related pregreenschist metamorphism. Schweiz. mineral. petrogr. Mitt., v. 79, pp. 111-134.
Duchene S., Blichert-Toft J., Luais B., Telouk P., Lardeaux J.-M. and Albarede F., 1997, The Lu-Hf dating of garnets and the ages of the Alpine high-pressure metamorphism. Nature, v. 387, pp. 586-589.
Ernst W.G., 1971, Metamorphic zonation on presumably subducted lithospheric plates from Japan, California and the Alps. Contrib. Mineral.
Petrol., v. 34, pp. 43-59.
Franchi S., 1900, Sopra alcuni giacimenti di roccie giadeitiche nelle Alpi Occidentali e nell'Appennino ligure. Bollettino del Regio Comitato geologico d'Italia, v. 31, pp. 119-158.
Franchi S, 1902, Uber Feldspath-Uralitisierung der Natron-Thonerde-pyroxene aus den eklogitischen Glimmerschiefern der Gebirge von Biella (Graiische Alpen). Neues Jahrbuch fur Mineralogie, Geologie und Palaeontologie, v. 1902 (II), pp. 112-126.
Frey M., Desmons J. and Neubauer F., 1999, The new metamorphic map of the Alps, with 12 explanatory articles. Schweiz. mineral. petrogr. Mitt., v. 79, pp. 1-230.
Gebauer D., Schertl H.-P., Brix M. and Schreyer W., 1997, 35 Ma old ultrahigh- pressure metamorphism and evidence for very rapid exhumation in the Dora Maira Massif, western Alps. Lithos, v. 41, pp. 5-24.
Goffe' B. and Chopin Ch., 1986, High-pressure metamorphism in the western Alps: zoneography of metapelites, chronology and consequences. Schweiz. mineral. petrogr. Mitt., v. 66, pp. 41-52.
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. Memoires de Geologie (Lausanne), v. 13, pp. 1-59.
Martin S. and Kienast J.-R., 1987, The HP-LT manganiferous quartzites of Praborna, Piemonte ophiolite nappe, Italian Western Alps. Schweiz. mineral. petrogr. Mitt., v. 67, pp. 339-360.
Messiga B. and Scambelluri M., 1988, Comparison between two types of coronitic eclogites from the Western Alps: implication for a pre-eclogitic evolution. Schweiz. mineral. petrogr. Mitt., v. 68, pp. 225-235.

Messiga B., Kienast J.R., Rebay G., Riccardi M. P. and Tribuzio R., 1999, Crrich magnesiochloritoid eclogites from the Monviso ophiolites (Western Alps, Italy). J. metam.Geol., v. 17, pp. 287-299.
Meyer J., 1983, Mineralogie und Petrologie des Allalingabbros. PhD Thesis, Univ. Basel, 331p.
Monie P., Chopin Ch. and Philippot P., 1989, 39Ar/40Ar dating in the Dora Maira and Monviso massifs (Western Alps): evidence for two stages of eclogitic metamorphism. Terra Abstracts, v. 1, pp. 264.
Philippot P., 1993, Fluid-melt-rock interaction in mafic eclogites and coesitebearing metasediments: Constraints on volatile recycling during subduction. Chem. Geol., v. 108, pp. 93-112.
Reinecke T., 1991, Very-high pressure metamorphism and uplift of coesitebearing metasediments from the Zermatt-Saas zone, Western Alps. Eur. J. Miner., v. 3, pp.7-17.
Rubatto D. and Gebauer D., 1999, Eo/Oligocene (35 Ma) high-pressure metamorphism in the Monte Rosa nappe (western Alps): implications for paleogeography. Schweiz. mineral. petrogr. Mitt., v. 79, pp. 353-362.
Rubatto D., Gebauer D. and Compagnoni R., 1999, Dating of eclogite-facies zircons: the age of Alpine metamorphism in the Sesia-Lanzo Zone (Western Alps). Earth Planet. Sci. Lett., v. 167, pp. 141-158.
Rubatto D. and Hermann J., 2001, Exhumation as fast as subduction? Geology, v. 29, pp. 3-6.
Rubatto D., Liati A. and Gebauer D., 2003, Dating ultrahigh pressure metamorphism, *in* A. Carswell & R. Compagnoni, eds, Ultrahigh pressure metamorphism. EMU Notes in Mineralogy, v. 5, Chapter 7.
Scambelluri M. and Rampone E., 1999, Mg-metasomatism of oceanic gabbros and its control on Ti-clinohumite formation during eclogitization. Contrib. Mineral. Petrol., v. 135, pp. 1-17.
Schreyer W., 1977, Whiteschist: their compositions and pressure-temperature regimes based on experimental, field, and petrographic evidence. Tectonophysics, v. 43, pp. 127-144.
Schwartz S., Lardeaux J.M., Guillot S., Tricart P., 2000, Diversite du metamorphisme eclogitique dans le massif ophiolitique du Monviso (Alpes occidentales, Italie). Geodinamica Acta, v. 13, pp. 169-188.
Stella A., 1894, Relazione sul rilevamento eseguito nell'anno 1983 nelle Alpi Occidentali (Valli dell'Orco e della Soana). Bollettino del Regio Comitato geologico d' Italia, v. 25, p. 343-371.
Tilton G.R., Schreyer W. and Schertl H.-P., 1991, Pb-Sr-Nd isotopic behaviour of deeply subducted crustal rocks from the Dora Maira Massif, Western Alps, Italy — II: what is the age of the ultrahigh-pressure metamorphism? Contrib. Mineral. Petrol., v. 108, p. 22-33.
Tropper P. and Essene E.J., 2002, Thermobarometry in eclogites with multiple stages of mineral growth: an example from the Sesia-Lanzo Zone (Western Alps, Italy). Schweiz. mineral. petrogr. Mitt., v. 82, pp. 487-514.
Zwart H.J., 1973, Metamorphic map of Europe, scale 1:2,500,000. Subcommission for the Cartography of the Metamorphic Belts of the World. Leiden/ UNESCO, Paris.

Stop: 1 Lago di Cignana

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Mapping of oceanic crust with HP to UHP metamorphism: The Lago di Cignana unit (Western Alps).

Abstract: Lago di Cignana is an ultra – high-pressure (UHP) metamorphic unit in the Western Alps (reaching 590-630°C, 2.6-2.8 GPa) (Reinecke, 1991). The UHP slice is only hundreds of metres across and is extensively boudinaged (Fig. 7). Interlayers of metasediment and metabasite preserve structures and mineral assemblages that suggest a multi-stage and complex exhumation history (cf. Klauw et alii, 1997; Reddy et alii, 1999, 2003) beneath a km-scale extesional ductile shear zone in movement. The extensional structures shortened during their more recent history, forming regionally developed upright folds and dome and basin structures that decorate and emphasize the UHP boudins.

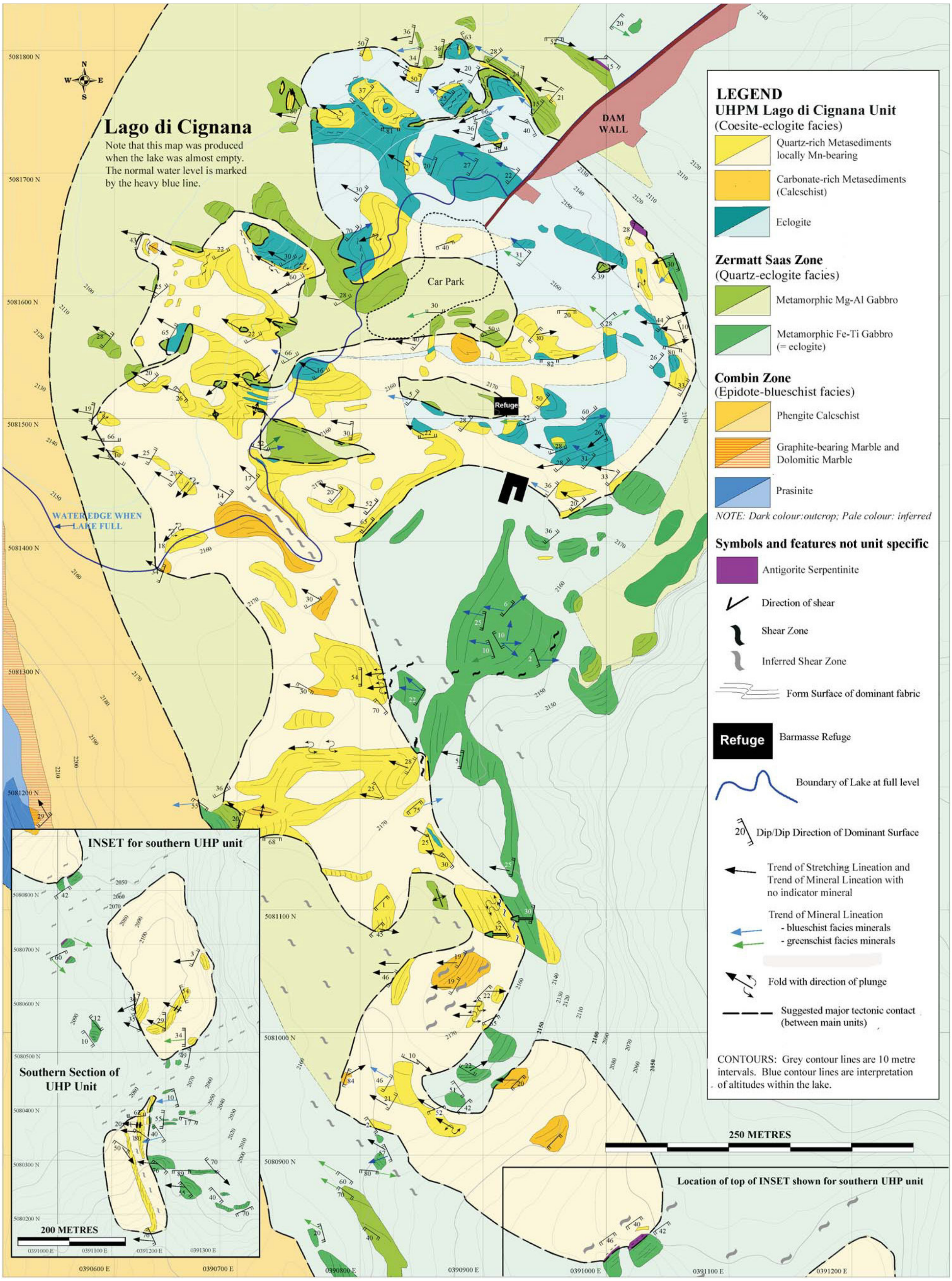


Fig. 1. Structural map of the region around the Lago di Cignana. A sliver of UHP coesite bearing lithologies occur where the Zermatt Saas Zone and the Combin Zone of the Piemonte meta ophiolites are juxtaposed. The contacts are all tectonic. Inset is the southern, topographically lower section of the UHP unit. The schistosity shown on the map is the amin regional schistosity associated with the boudin formation, although relict older and overprinting younger fabrics occur throughout the UHP unit.

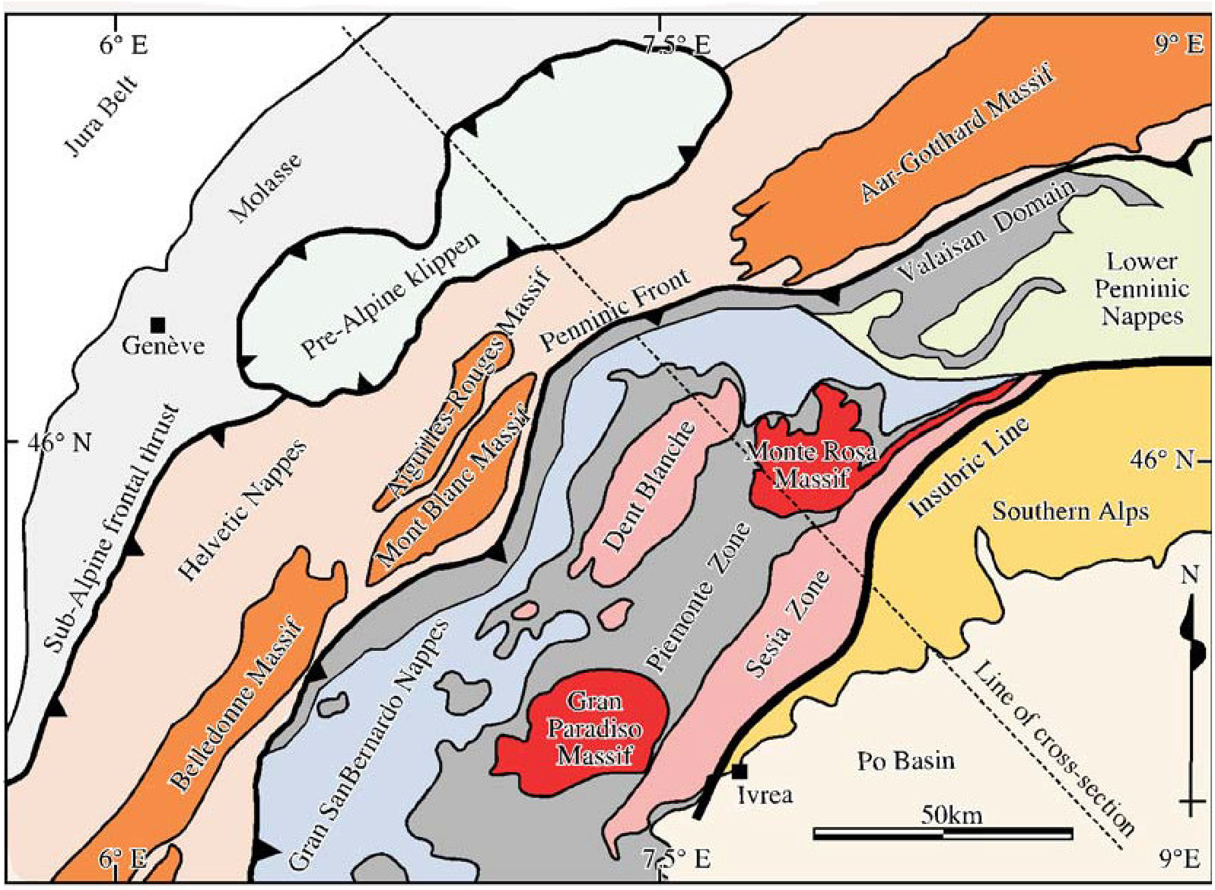


Fig. 2: Tectonic map of the Northwestern Alps.

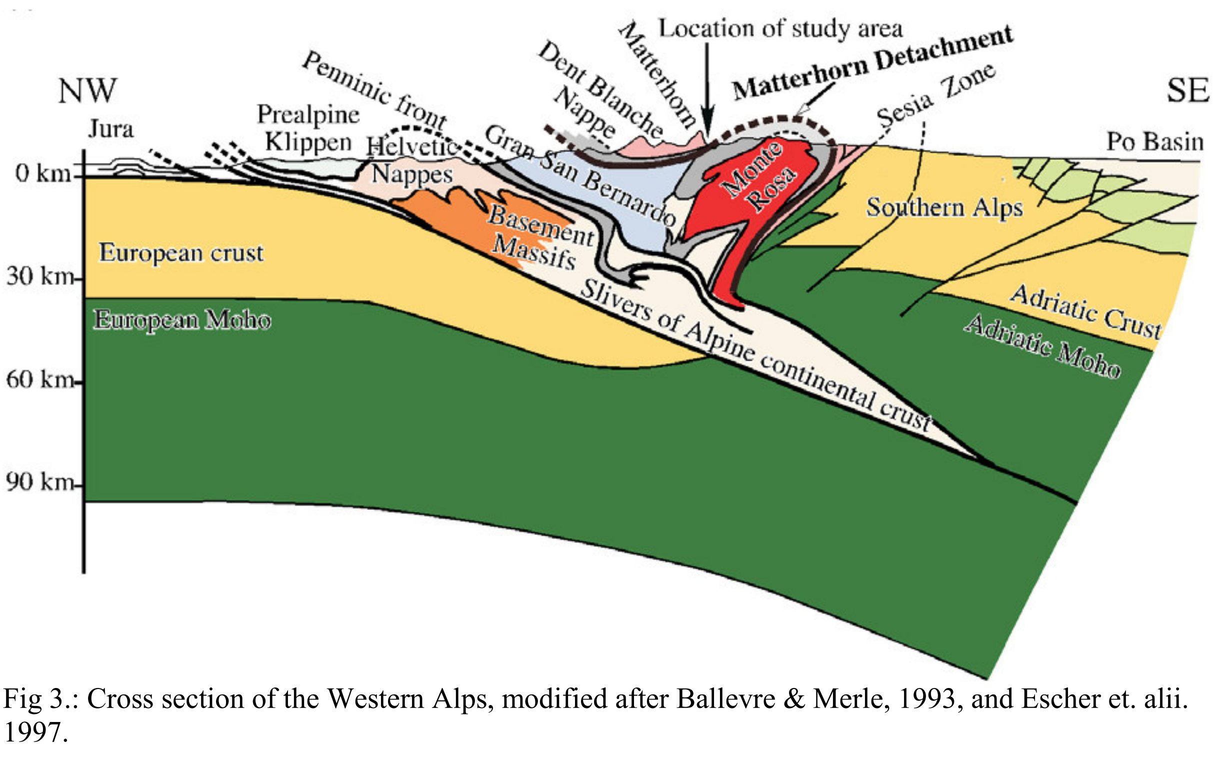


Fig 3.: Cross section of the Western Alps, modified after Ballevre & Merle, 1993, and Escher et. alii. 1997.

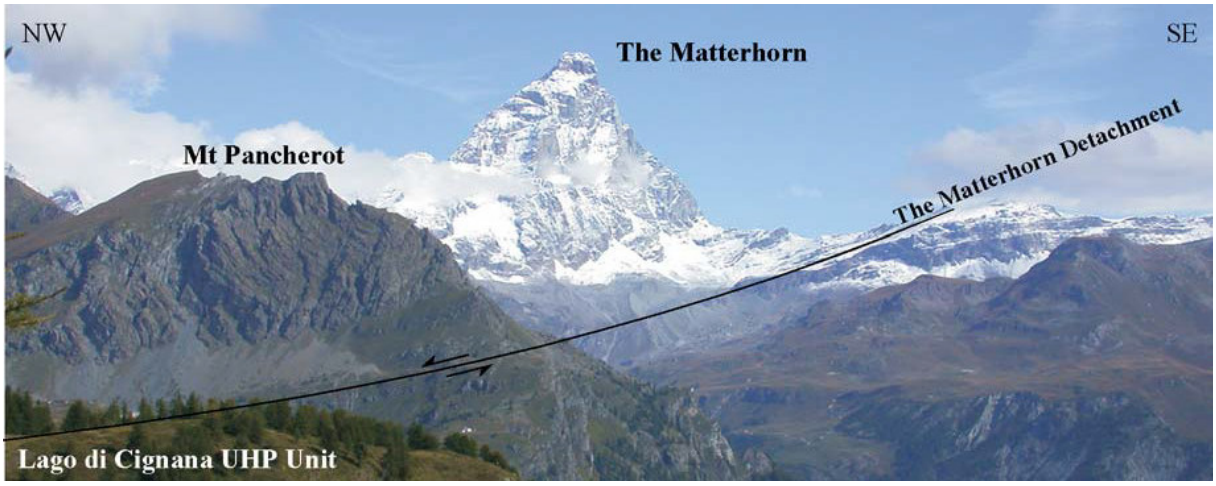


Fig 4: The Matterhorn detachment. The UHP units occur bebeath the detachment on top of the Zermatt Saas unit. In the hangingwall (e.g. Mt Pancherot) the Combin unit.

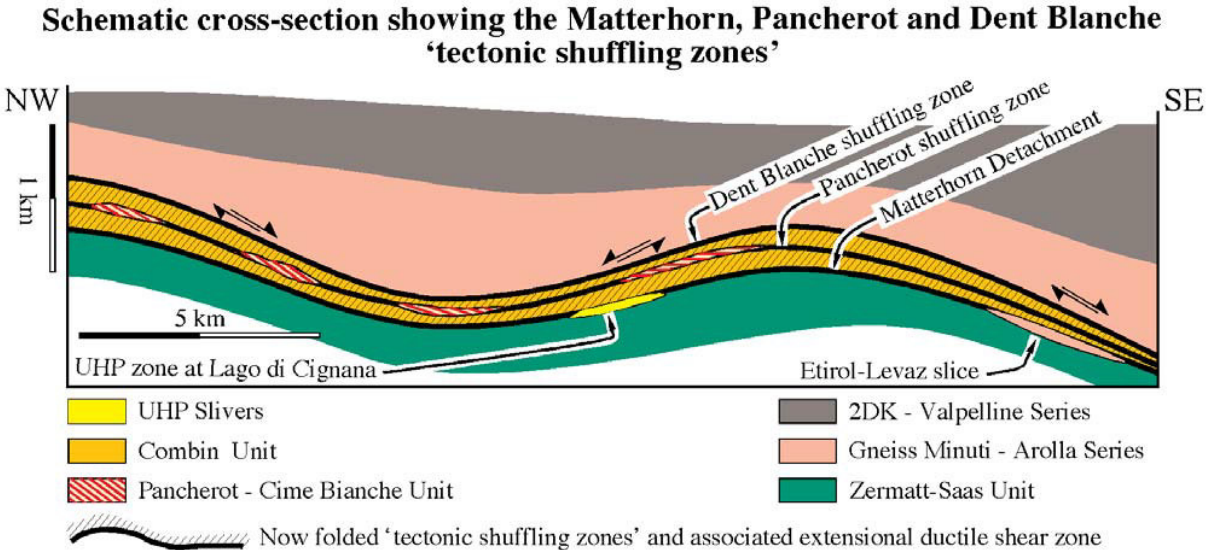


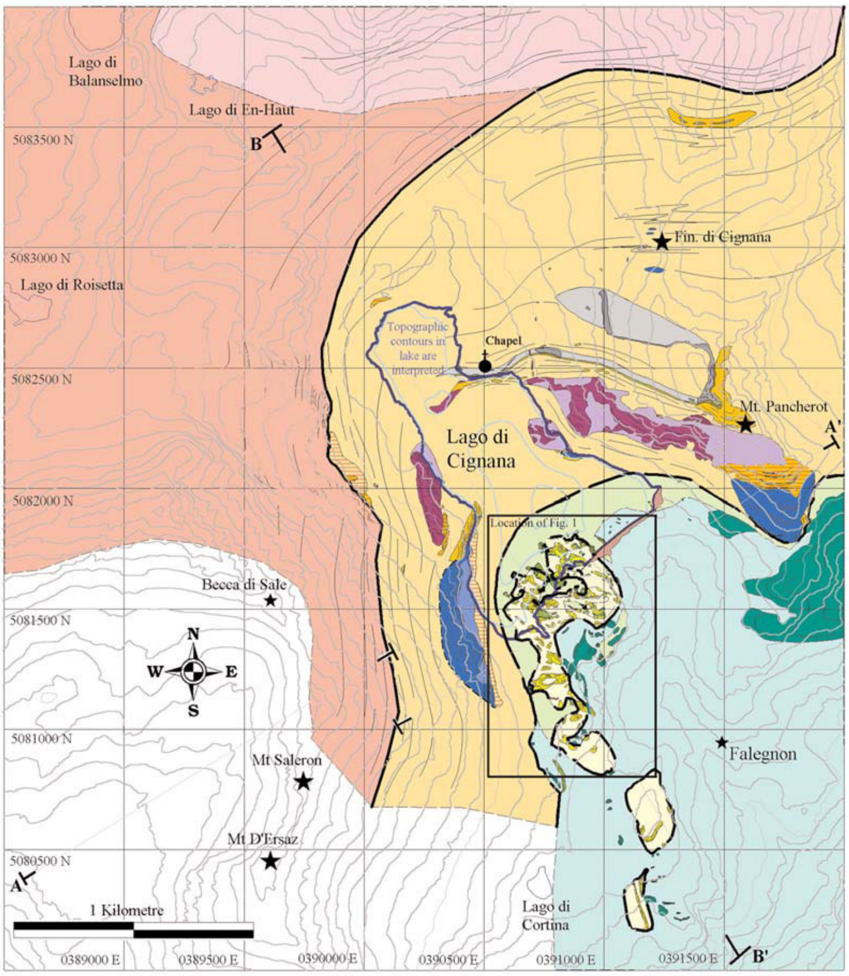
Fig. 6 - *The diagram shows structural relations as would have applied at the end of the extensional phase (at ca 36 Ma). Post ca 25 Ma the Alpine crust was folded to produce the present geometry (as shown in Fig. 3). The base of the extensional shear zone that affected the Combin Zone is cut off by the semi-brittle Matterhorn Detachment. A similar fault may exist at the Dent Blanche contact, where a narrow more intense shear zone is also observed. Tectonic shuffling zones may be the result of thrusts that have been re-activated as detachment faults/shear zones, stranding anastomosing lenses of higher pressure rocks. Crustal-scale boudinage during the later stages of movement may have overprinted adjacent parts of the shear zone differently, producing shear sense variation as shown. The shear zones at Lago di Cignana were affected by NW-directed sense of shear at this time, while zones to the southeast (e.g. Gressoney) were affected by SE-directed sense of shear (REDDY* et alii *2003).*

INTRODUCTION

The aim of this study was to map the exposure of UHP rocks at Lago di Cignana (Valle d'Aosta, NW Italy), one of the only two known coesite locations in the Alps, in order to determine the 3D geometry of lithologies, planar and linear fabrics, faults and folds. In this way, we intend to constrain the movement picture that led to the exhumation of these remarkable rocks, from as deep down as 100 km in the Earth up to their present location. This datum provides a 3D-time geometrical constraint that

allows us to test different models: some models suggest that exhumation is driven by local effects, namely the relative buoyancy of individual sheets of rock, while others suggest that the crust is torn apart by regional sub-horizontal extension, with a movement focussed in crustal-scale extensional shear zones. These periods of intense crustal extension may closely follow individual episodes of high-pressure metamorphic mineral growth.
In regions, where several distinct episodes of high-pressure metamorphism have taken place (such as around Lago di Cignana), it follows that several episodes of intense crustal shortening followed by intense crustal extension might be discerned. Such a history of tectonic mode switches, marking the end of stages in a sequence of "inversion cycles", would produce distinctive patterns of fabric and microstructural evolution, and characteristic outcrop-scale and map-scale geometries.

A team effort was required in order to understand this complex evolution, involving a group of individuals that combined the expertises of field mapping, structural geology, metamorphic petrology and geochronology. Key fabrics and microstructures were subsequently analysed under the microscope. Microprobe mineral analyses allowed geothermobarometric calculations, while the painstaking separation of minerals grown in specific microstructural events allowed the processing of geochronology to be initiated.
The production of a detailed geological map is the first and most important stage in this systematic and objective methodology, for it provides the framework on which all else is based. The preliminary results of this analysis are given on the following pages.
The UHP lens is located close to or at the upper boundary of the HP metamorphic Zermatt-Saas Zone, directly below the dominantly greenschist Lago di Cignana "HP"/"UHP" metamorphism facies Combin Zone (Figs. 7, 8, 12 and 13).



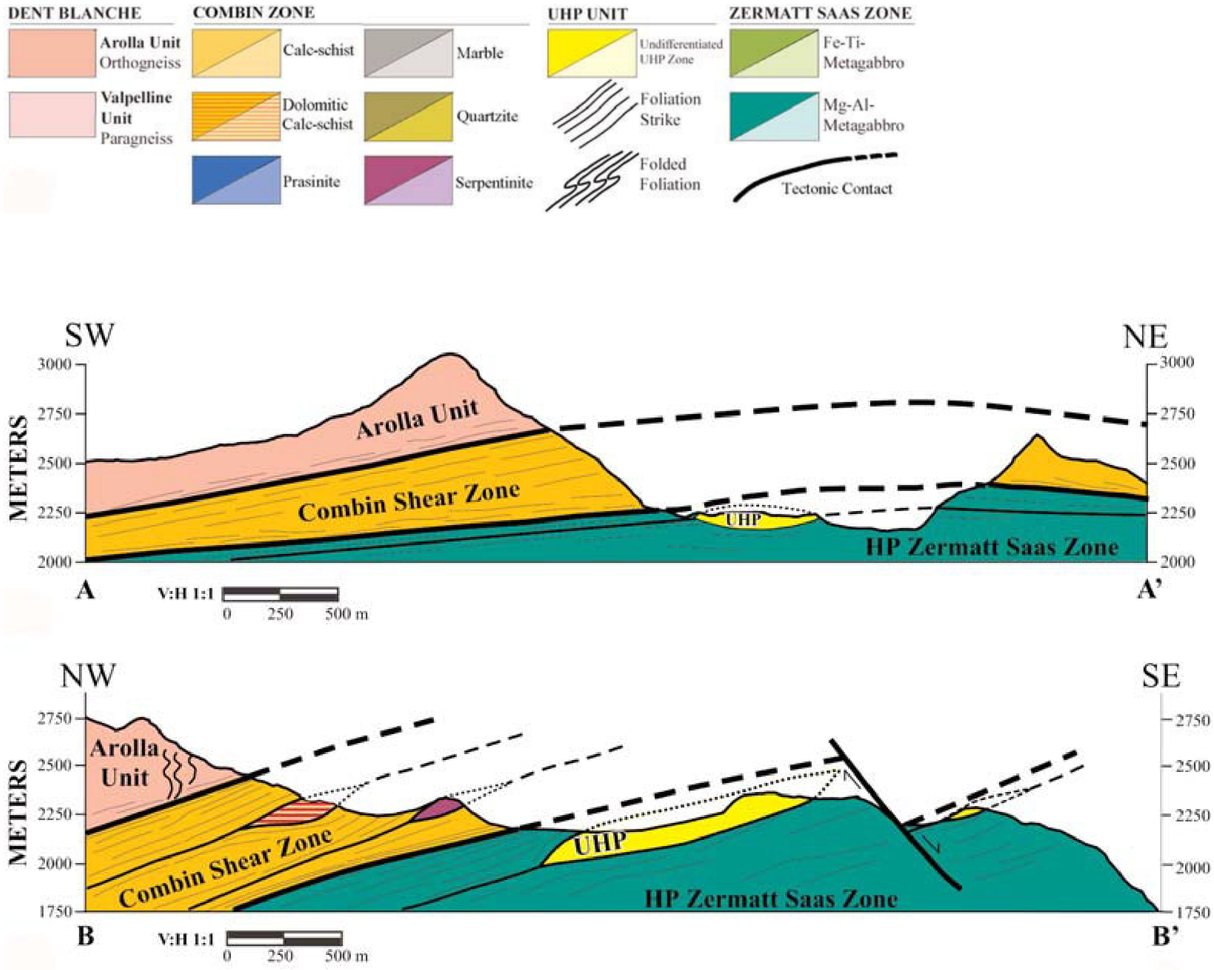


Fig. 7 - Geological map of the Lago di Cignana region. Legend shows map units, with outcrop in vivid colour and the inferred extent in the equivalent pale colour. The UHP Unit occurs in the uppermost structural levels of the Zermatt-Saas Zone, overlain by a thin, highly deformed slice of metagabbro. The structurally higher Combin Zone contains boudins of varying lithologies such as serpentinite, dolomitic calc-schist and prasinite. Marble and quartzite lenses occur in spatial association with the dolomitic lenses. The cross-sections are both perpendicular and parallel to the dominant NW/SE trending stretching lineation. The perpendicular section is along line A-A' on the map, while the parallel section is along line B-B'. Note that several different lithologies, distinguished in the map, have been grouped in the cross-sections.

The characters of the upper and lower contacts of the UHP lens are different, and rendered complex by overprinting deformational events (e.g. a NWtrending extensional event and late-stage NW/SE shortening). The UHP Unit is overlain by a dissected thin sheet of highly deformed Fe-Ti metagabbro from the Zermatt-Saas Zone, observed at the northern boundary of the UHP Unit (below high water level) and at several locations on the upper surface of the UHP lens (Figs. 17 and 18). The lower contact with the Zermatt-Saas Zone is not well defined and is difficult to locate. Zermatt- Saas metabasics are characterised by planar layering, while the UHP eclogitic metabasics are deformed boudins.

Tectonic slices derived from the Zermatt-Saas Zone thus enclose and wrap the UHP boudins (Figs. 1, 12, 13 and 18).
The upper tectonic contact of the UHP lens was later folded and cut by late-stage ductile and brittle deformation. Calcschist occurs within the UHP Unit, and is observed as a structurally higher

lithology, with tight upright folds, and commonly intensely sheared by young shear zones. This calc-schist may belong to the Combin Zone, caught up in the structurally higher regions of the UHP Unit, or it may have undergone the same deformational and metamorphic history as the quartz-rich metasediment layers.

The UHP lens preserves an earlier deformational history and direction of movement when compared to the fabrics in the overlying Combin Zone.

The Combin Zone localizes a major extensional shear zone, and the UHP lens has been exhumed beneath this structure. This shear zone extends throughout the region (Fig. 6, REDDY et alii, 2003) and can be extrapolated northward from Lago di Cignana and beneath the Matterhorn. A comparison of the sense of shear on shear zones between the Combin Zone and the UHP Unit suggests a multi-stage episodic exhumation of the UHP Unit. The low-angle faults formed during the late stages of movement, while the extensional structures shortened more recently (Fig. 3), forming regionally developed upright folds and dome and basin features at a km- to outcrop-scale.

GEOLOGICAL SETTING

The study area is exposed within the calc-schists and meta-ophiolites of the Piemonte zone, which consists of a pile of tectonic slices including both Alpine epidote-blueschist facies ('Combin Zone') and eclogite facies ('Zermatt- Saas Zone') metamorphic rocks. In the Valtournenche area, the ocean-derived rocks of the Piemonte Zone are sandwiched between the overlying Austroalpine Dent Blanche nappe and the underlying Penninic Monte Rosa nappe.

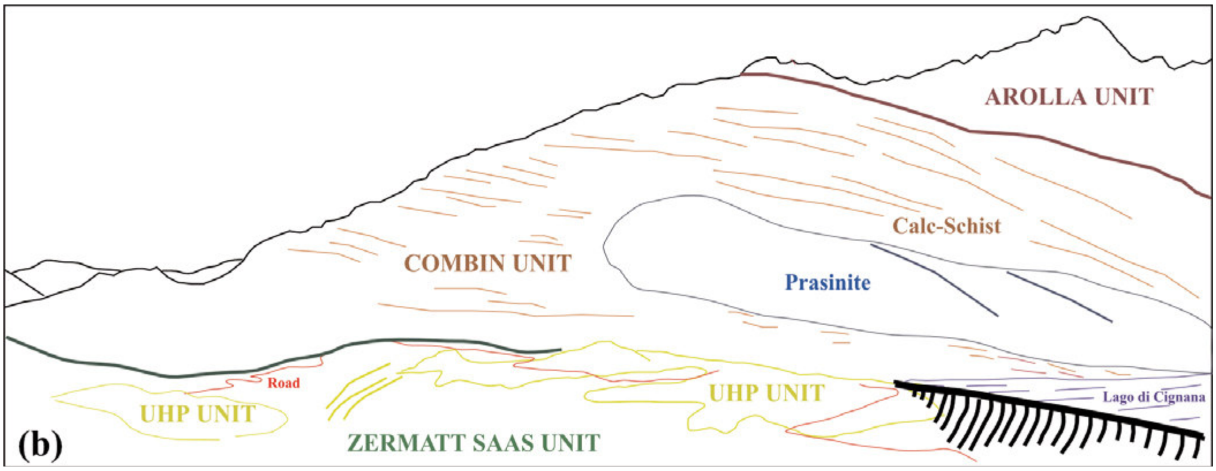


Fig. 8 - a) The sketch shows the main lithological units and major structural contacts between the units seen in photograph Fig. 8b).


b) Panorama photograph of the Arolla Unit on the high mountain peak and the Combin Zone directly below. The cliff in the centre of the photograph is a prasinite boudin within the Combin Shear Zone: the contact with the UHP Unit is directly below this boudin. Complex slivers of
prasinite, calc-schist and serpentinite occur at the tectonic contact with the underlying UHP Unit. The UHP Unit is on the grassed area in the foreground, from left to centre of the picture, as well as in the tree area on the far left hand side. The Zermatt-Saas Zone is the cliff region in the centre base of the photograph. Photograph is looking west.

Fig. 9 - A major mylonite zone is located on the eastern side of the lake, marking the tectonic contact between the Zermat-Saas Zone and the overlying Combin Zone. Eclogite facies assemblages are still identifiable in the Zermatt-Saas rocks, although retrogression has occurred. (E:0391805 N:5081755)

Fig. 10 - An intense shear zone marks the base of the Combin Zone. On the western side of the lake, slivers of calc-schist (brown lens at the centre of the photograph) are interleaved into a prasinite boudin (the pale blue outcrop). Intense shearing and serpentinite occur around the calcschist sliver. David Giles stands at the centre for scale. (E:0390440 N:5081727)

Fig. 11 - The UHP unit occurs as three distinct lenses, all surrounded by Zermatt-Saas metagabbro. A mylonitic greenschist facies shear zone is located between the main UHP lens and the lens at Cortina Lake, suggesting movement between these two lenses at a late-stage in their ductile history. The fault dips to the southeast, with complex crenulations occurring in the chloritic zone. (E: 03901000 N: 5080816)

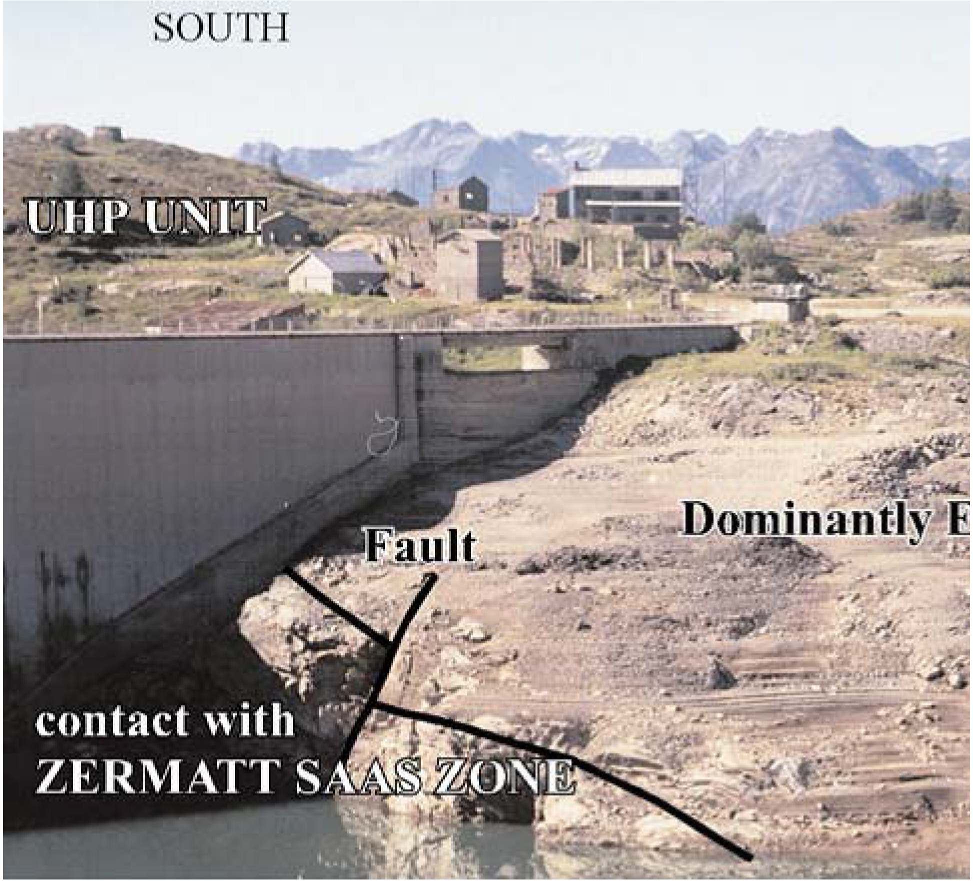


Fig. 12 - Lago di Cignana was emptied of water in mid-2001, and the project was undertaken while the lake was empty and the UHP Unit was fully exposed. Many of the mounds that can be seen in the photograph are UHP boudins with flat-lying schistosity on the upper surface and steep to moderate dipping schistosity on their terminations. The cliff below the dam wall comprises mainly Fe-Ti metagabbro belonging to the Zermatt-Saas Zone, immediately overlain by the lens of UHP eclogitic metabasites and metasediments. The darker boudins on the left side are metabasites. Metasediments dominate the right-hand side of the photograph and the wooded, grassy hill. Photo taken facing SSE.

Lago di Cignana "HP"/"UHP" metamorphism

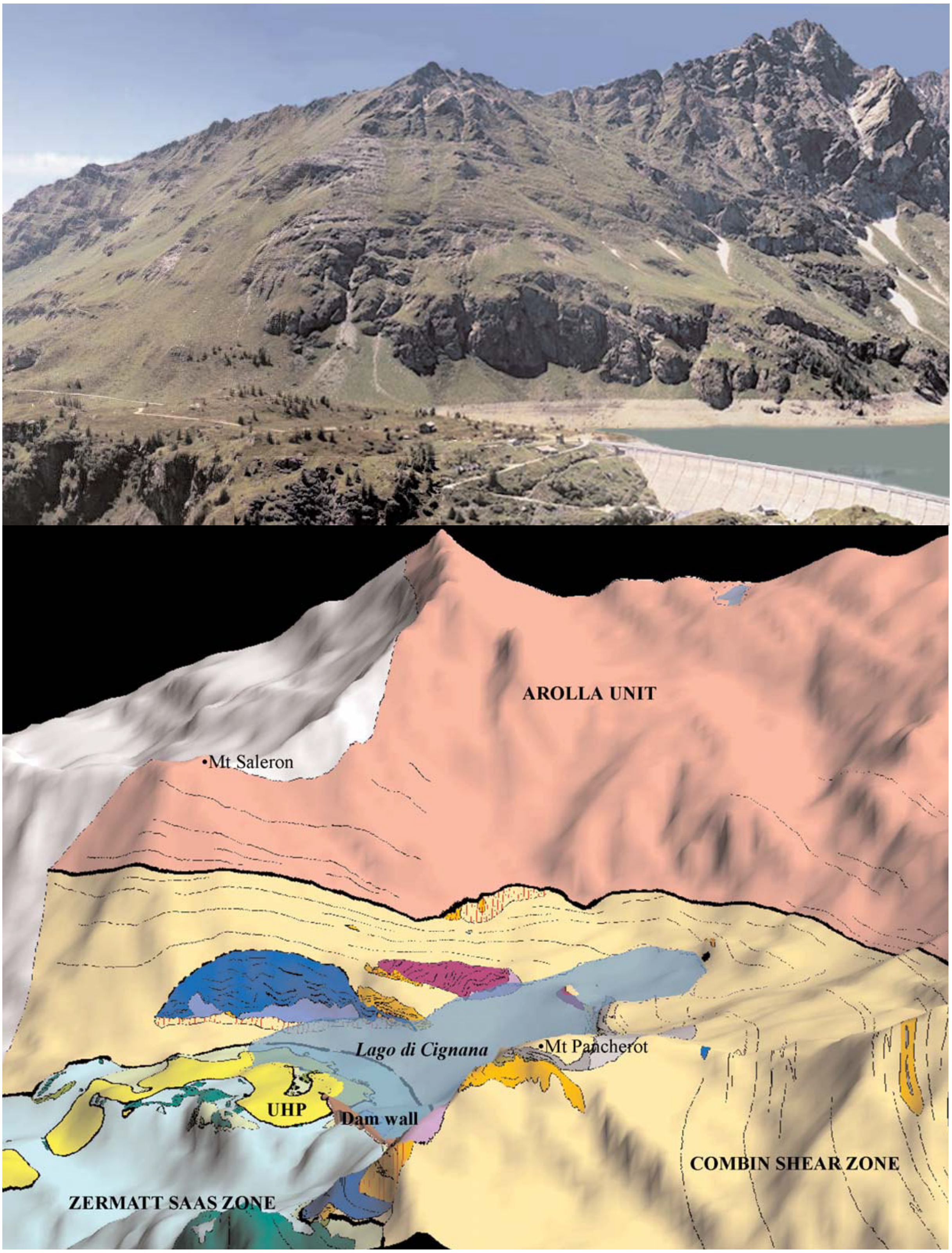


Fig. 13 - 3D image of the geology of the Lago di Cignana area. The picture was produced using GOCAD software. The 3D topography is based on the topographic contours from geological maps (Figs. 1 and 7). The lake has been filled in order to provide a more realistic perspective. The bold black lines are tectonic contacts. Note the planar contact with the Arolla Unit. The fault/shear zone truncates recumbent folds formed earlier. Colours match the legend for the geological maps (Figs. 1 and 7). The white area has not been mapped.

Fig. 14 - The quartz-rich metasediments of the UHP unit were isoclinally folded prior to the deformational event that produced the recrystallization and formation of boudins. An axial planar cleavage associated with this folding event can be detected. FOV ~3 metres. (E: 0390805 N: 5081285).

Fig. 15 - An example of the boudins that characterize the UHP unit. This eclogitic metabasite boudin is bounded by flat-lying schistosity on the upper and lower boundaries with steep-dipping schistosity at the ends of the boudin. The boudin is ~1.5 metres across.

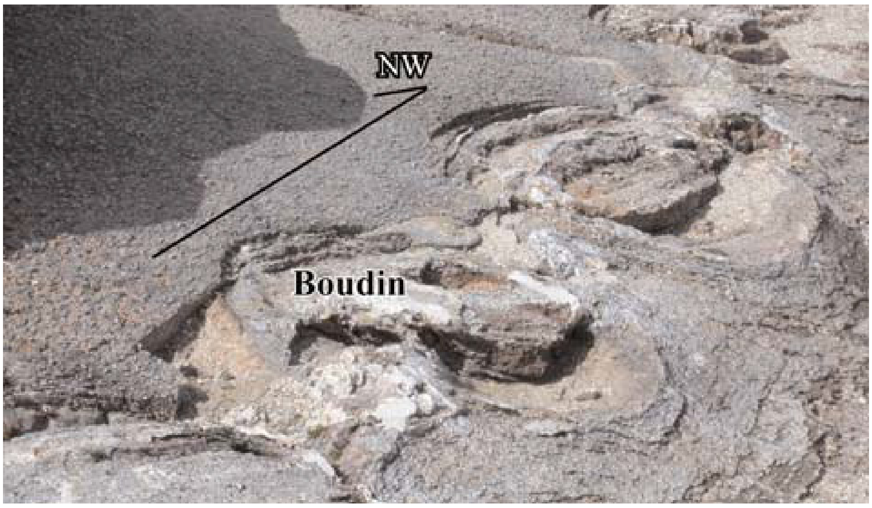


Fig. 16 - Foliation boudinage in the quartz-rich metasediment. Abundant garnets are found immediately below the overlying tectonic contact with the Zermatt Saas Zone. The trend of the boudinage is NW/SE, parallel to the dominant stretching lineation. FOV ~4m. (E:0390724 N: 5081418).

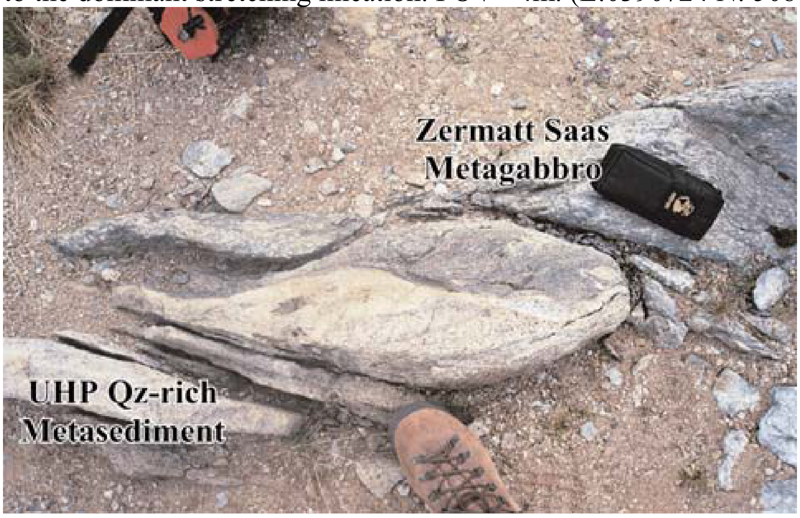


Fig. 17 - A tectonic contact between UHP metasediment and Zermatt-Saas Zone Fe-Ti metagabbro is overprinted by boudinage and a compressional event, observed in the structurally highest region of the UHP Unit. (E:0390712 N: 5081505).

Fig. 18 - Zermatt-Saas Fe-Ti metagabbro overlying UHP metabasites. Shearing is most intense immediately adjacent to this tectonic contact. A late-stage, NW/SE compressional event has caused metre-scale folding, here downwarping the overlying Zermatt-Saas metagabbro sheet.

FIELD IDENTIFICATION OF LITHOLOGIES AND FABRICS

The methodology used in the field by structural geologists and metamorphic petrologists relies essentially on the identification of fabrics and lithologies (in this case, in respect both to protoliths and their deformed and metamorphosed equivalents). This requires field work to focus on a range of different scales, from microscale to map-scale, as it is impossible to carry out such a study without such wide-ranging attention to detail.

This page focuses on illustrating the fabrics and lithologies that can be observed at Lago di Cignana in order to enable readers of the Volume to readily identify critical aspects during a visit to this classic field location. Protoliths can be hard to identify because of the variable extent of overprinting metamorphic mineral growth, in some cases due to multiple episodes. More important are the effects of deformation, varying from minor modification of the original fabrics to extreme stretching and total disruption of original textures and early fabrics. Key features are illustrated in individual photographs, and described in figure captions.

The photographs are presented according to their position in individual tectonometamorphic slices. The highest structural levels are in the upperplate, above the Matterhorn Detachment and its precursor extensional shear zone. Here remnants of the Dent Blanche nappe - recumbently folded and later sheared augengneisses of the Arolla Unit (Fig. 19) - are found. Lithologies caught in the underlying extensional shear zone are illustrated in Figs. 20 to 22. The UHP Unit and the Zermatt-Saas Zone are in the lower plate (Figs. 23 to 36).

Note that the UHP slice is intercalated with slices of the Zermatt-Saas Zone (Figs. 1 and 7). The uppermost tectonic slice of the lower-plate is a deformed metagabbro belonging to the Zermatt-Saas Zone. This tectonic slice lies above the anastomosing boudins of the ultrahigh pressure unit. Because the detachment slices through anastomosing boudins and tectonic slices, the Combin Zone may also be in contact with the UHP slice locally.

The Matterhorn Detachment and the extensional Combin Shear Zone appear to be responsible for the exhumation of the UHP slice and of the HP rocks of the Zermatt-Saas Zone. Fabrics and mineralogy reveal that the Combin Shear Zone has operated through blueschist facies conditions until greenschist facies metamorphic conditions were reached. Intense fabrics are developed. For a detailed description of metamorphic petrology for UHP lithologies refer to COMPAGNONI & ROLFO (1999).

AROLLA UNIT:

In order to examine the contact with the Arolla Unit one needs to climb the steep grass slopes and cliffs west of Lago di Cignana. Samples vary from metagranitoids to augen mylonites, with platy foliations anastomosing around feldspar clasts. These gneissic fabrics are themselves crenulated, with gentlydipping axial planes. Larger-scale recumbent folds can be discerned in the rock walls, that may be equivalent in timing to these structures. Shallowly-inclined (detachment?) faults and intense ductile shear zones at the contact cut through these folds and the earlier formed gneiss fabric.

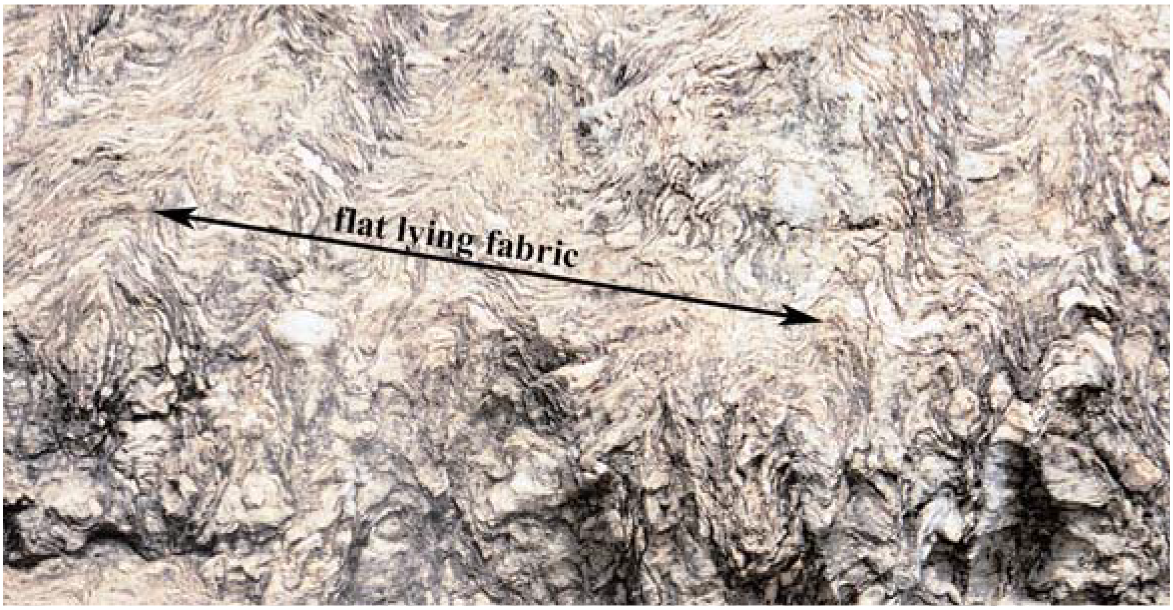


Fig. 19 - Crenulations of an older mylonitic fabric in boulders of augengneiss that have fallen from the Arolla slice down the steep slopes besides Lago di Cignana. FOV 15cm across.

COMBIN ZONE:

The Combin Zone is dominated by calcschist, with several large boudins of serpentinite and prasinite (Figs. 7 and 12). Structurally above these boudins is a sequence of quartzite, marble and dolomite (Fig. 20). The calcschist in the structurally higher regions of this zone is increasingly graphitic and more intensely deformed closer to the overlying tectonic contact with the Arolla Unit. Intensity of deformation also occurs adjacent to its lower boundary (Figs. 21 and 22). Intensity of shearing varies remarkably throughout the Combin Zone.

Fig. 20 - Marble and dolomite boudins occurring in the central region of the Combin Zone and locally mylonitized. FOV ~1.3 m.

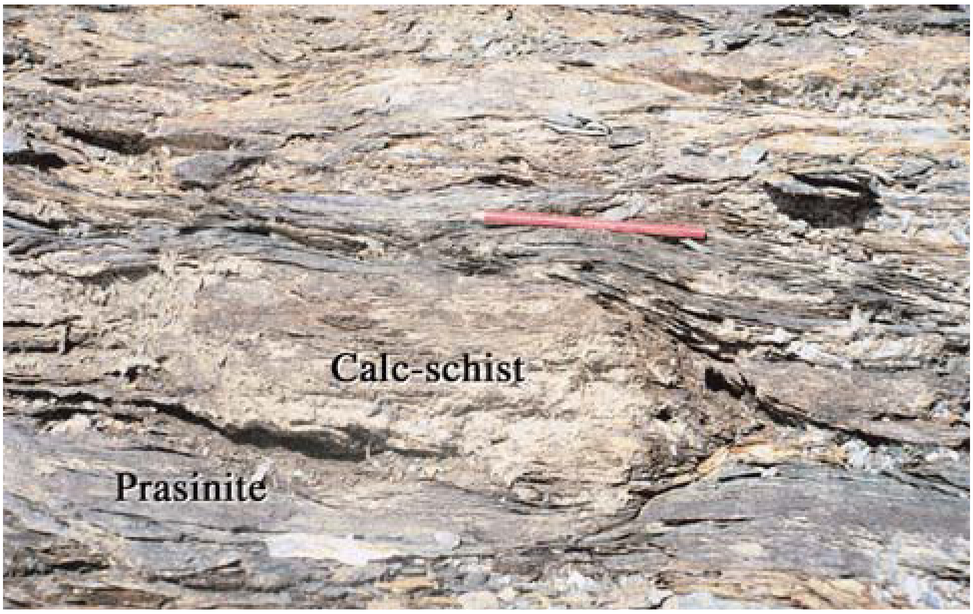


Fig. 21 - Interleaved calc-schist and prasinite intensely sheared at the tectonic contact between the Combin Zone and the underlying Zermatt-Saas metaophiolite. FOV 56 cm.

Fig. 22 - Calc-schist immediately adjacent to the lower tectonic contact of the Combin Zone that has undergone extreme shearing, and was later locally kinked or crenulated. Low mica content produces a gneissic character. FOV 32 cm.

ZERMATT SAAS ZONE:

The Zermatt-Saas Zone consists of an ophiolite sequence, deformed and metamorphosed later under eclogite and blueschist facies conditions. There are occurrences of suspected pillow-basalts (some more deformed and retrogressed than others), large volumes of serpentinite, and numerous lenses of Fe-Ti or Mg-Al metagabbros (Figs. 23 and 24). Thin basalt dykes (now mainly replaced by chlorite) - a sought after stone for sculpture - occur. Bands of serpentinite, rodingite and talc often mark metasomatic zones associated with faulting, and/or late-stage shear zones. Serpentinites often contain porphyroblasts of Ti-clinohumite that are easily mistaken for garnet. The lithologies and fabrics are generally flat-lying and occur in cliff faces below the dam wall at the lake.

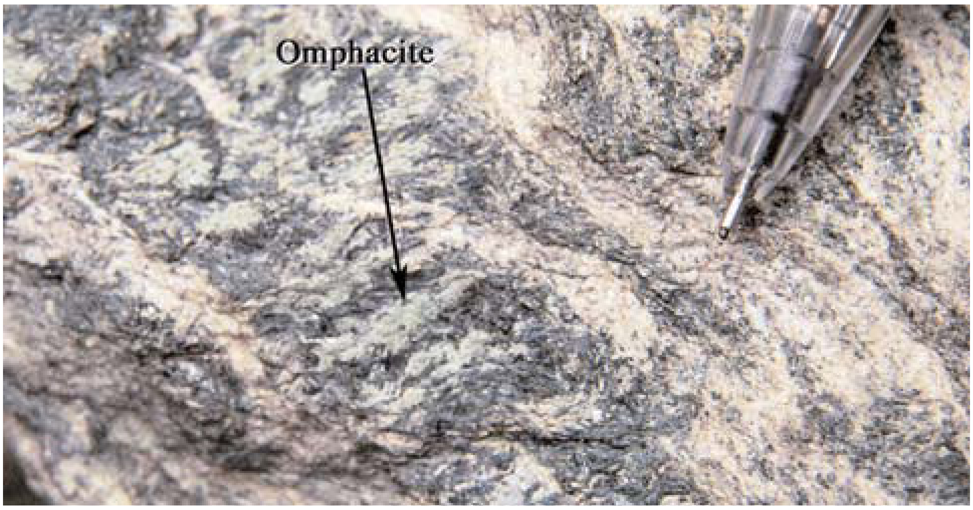


Fig. 23 - Fe-Ti metagabbro occurs within the Zermatt Saas Zone. These rocks best preserve original igneous textures in spite of the quartz-eclogite and greenschist facies overprints. This sample has not undergone intense shear and has preserved omphacite within small lenses. Early Na-amphibole and later albite overprint the eclogitic assemblage. Location in a low-strain zone east of the UHP unit (E:0390940 N: 5081740). FOV 6 cm.


Fig. 24 - A Mg-Al metagabbro within the Zermatt-Saas Zone, converted to quartz-eclogite/blueschist mineral assemblages. This lithology has similar characteristics to the eclogites from the UHP unit and it can be difficult to differentiate between these two units. Microstructural analysis and geothermobarometric estimates provide definitive tests. FOV ~8 cm.

THE ULTRA-HIGH PRESSURE (UHP) UNIT

The UHP unit is a narrow sequence of lenses (Fig. 7) that has undergone UHP metamorphism. It is less than 150 m in thickness and is characterised by a sequence of boudins ranging from ~100 m to 1 cm in length. Some of the boudins have been truncated by late-stage, high-angle normal faults or related late-stage shear zones. Others have been almost totally or totally scraped away by glaciation. The UHP Unit is located within the structurally highest levels of the Zermatt-Saas Zone, overlain only by a thin, highlydeformed slice of Zermatt-Saas metagabbro. The UHP Unit consists of metasediments and metabasites, with much variation occurring within these lithologies (Figs. 25 to 36). It is difficult to differentiate between the metabasites of the UHP unit and the Zermatt-Saas Zone without P-T estimates or the occurrence of coesite. Boudins of eclogite in the UHP Unit have been found to be coesite-bearing. Adjacent, flat-lying, highly-stretched sheets of eclogite appear to be structurally part of the Zermatt-Saas Zone.

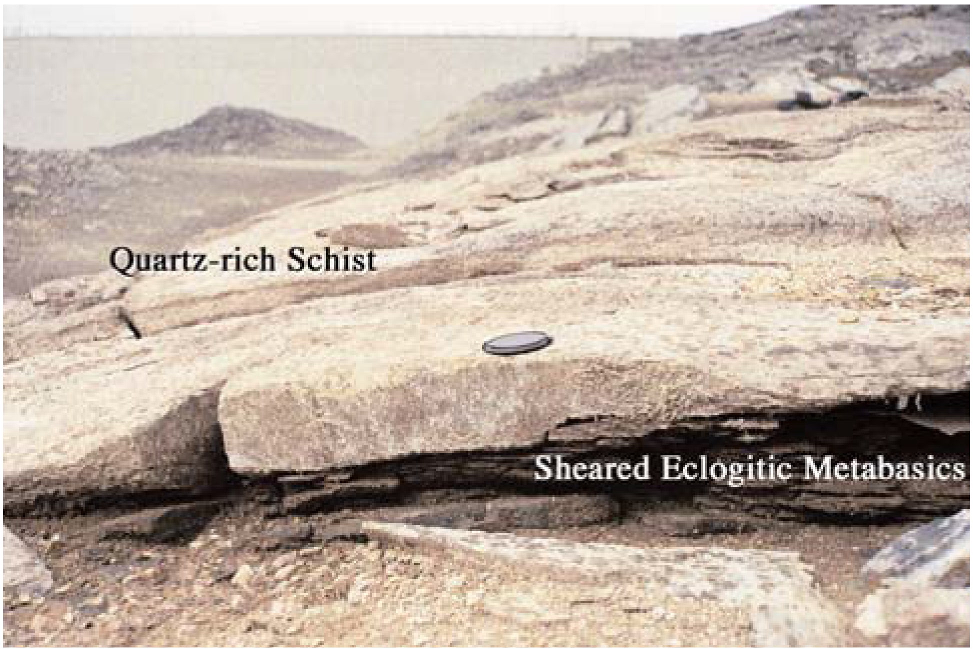


Fig. 25 - Quartz-rich metasediments can occur as thin layers and/or boudins (up to 10s metres in length) within a eclogitic metabasic matrix. The most intense fabrics occur within metabasic material (as seen above). Younger, small-scale shear zones may mantle these boudins and may truncate them at their ends, usually with steep dips (up to 80°). Schistosity within the boudins is usually flat lying. FOV ~80 cm.

Fig. 26 - Boudins of metabasic eclogitic material occur at the lower boundary of the UHP metasediments. The boudins vary in the degree of deformation, some being more round than elongated. This boudin is one of a sequence that occurs on the southern side of the dam wall, below the access road. Boudin is ~1.5 m long.

Fig. 27 - Calc-schist is located in the structurally highest regions of the UHP unit, occurring as dolomitic and micarich layers, often observed within boudin-like isoclinal infolds. Calc-schist is the dominant lithology in the Combin Zone and it is possible that the calc-schist within the UHP unit comprises slices of Combin calc-schist deformed against the UHP unit during earlier stages of the operation of the shear zone. FOV ~1.4 metres.

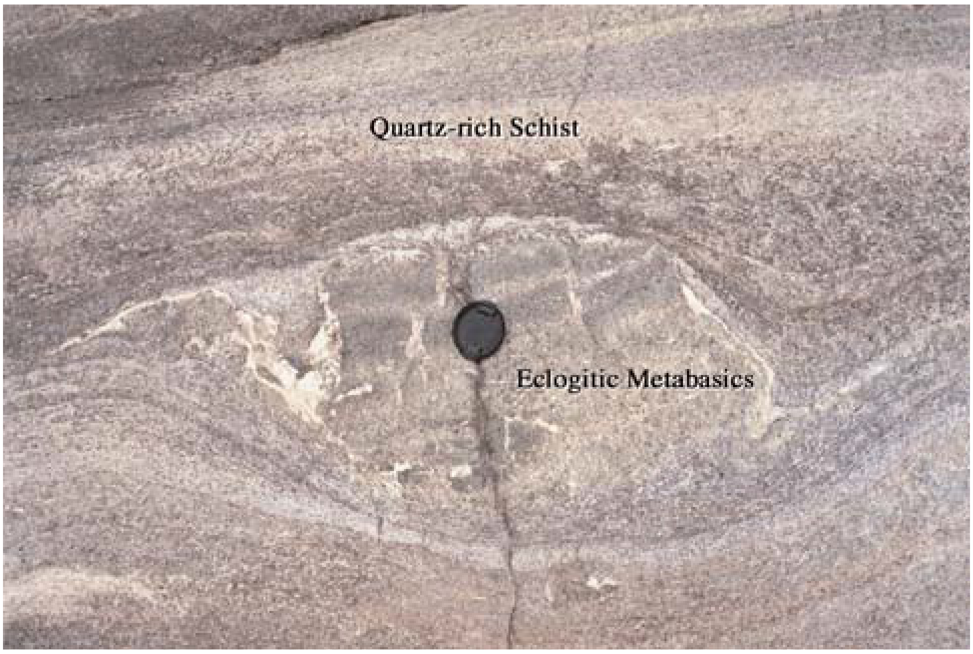


Fig. 28 - The structurally lower levels of the quartz-rich metasediments of the UHP Unit (in the section near the dam wall) have fine, mafic-rich layers at the base of the sediment beds. Metabasic eclogitic boudins also commonly occur within these metasediments. The sense of shear on the boudins rendered more complex by later shortening. FOV ~80 cm.

Fig. 29 - Typical character of the quartz-rich metasediments. The main variation that can occur is in the inclusion, character and abundance of garnet. Note in this sample that the intensity of shear increases towards the top and that the abundance of garnet decreases with the increase in intensity of shearing. FOV ~9 cm.

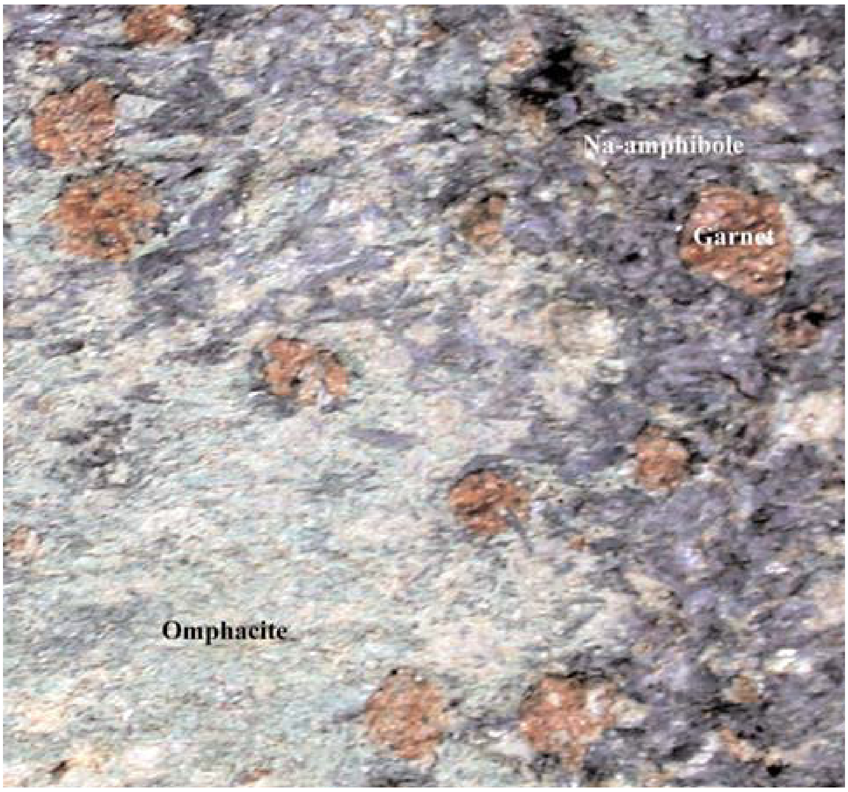


Fig. 30 - This eclogite from near the dam wall has been almost completely converted to blueschist facies metamorphic minerals (garnet porphyroblasts, glaucophane laths, zoisite plus phengite). Such rocks provide clear evidence of separate episodes of mineral growth, first under eclogite facies conditions, and later under blueschist facies conditions. Petrographic examinations reveal that a period of deformation occurred after each mineral growth event. The garnet porphyroblasts have grown under static metamorphic conditions. FOV ~3 cm.

Fig. 31 - Small-scale boudins of eclogitic assemblages are preserved within a pervasively blueschist fabric, providing evidence of its earlier metamorphic history. The fabric formed in a HP shear zone. S-C bands commonly occur throughout the outcrop. FOV ~6 cm.

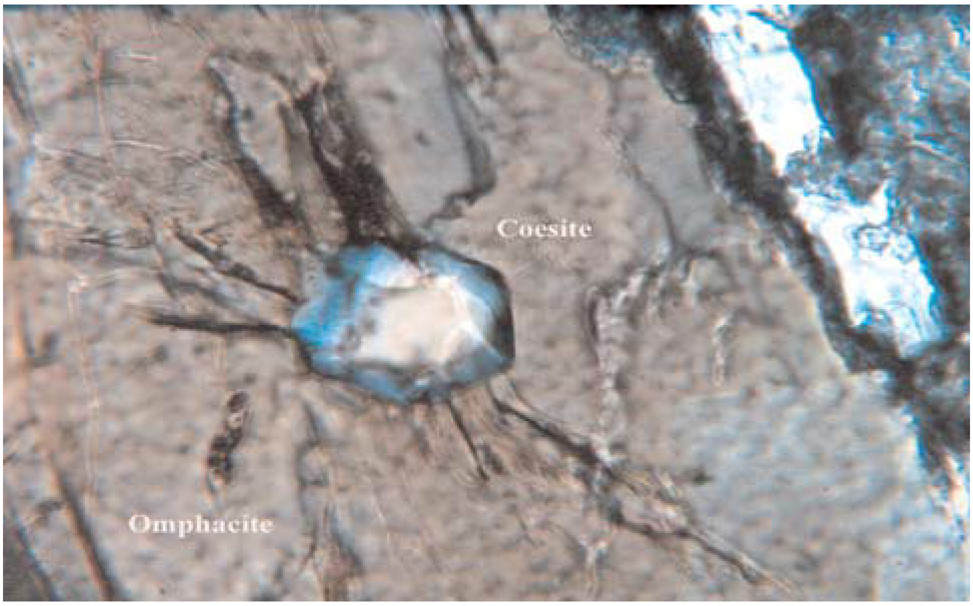


Fig. 32 - The UHP unit at Lago di Cignana has been found to be a coesite-bearing unit. Both the metasediments and metabasic rocks have been found to represent UHP metamorphic conditions. This photomicrograph shows a coesite grain within an omphacitic pyroxene. FOV ~200 µm.

Fig. 33 - The garnet porphyroblasts within the quartzrich metasediments have often retrogressed to chlorite. The centre of larger garnet porphyroblasts is often preserved. FOV ~6 cm.

Fig. 34 - Mn-rich, quartz-rich metasediments occur in several locations in the UHP Unit. This site is located in the outcrop in the centre of the car park, adjacent to the dam. Another site is found on the southern side of the dam wall. These could be metamorphosed manganiferous cherts. FOV ~4 cm.

Fig. 35 - There is a significant variation in the metabasic rocks of the UHP Unit. Distinct outcrops with omphacite + garnet assemblage, or Na-amphibole + garnet regions occur. The overprinting blueschist assemblage is more commonly observed. Variations in deformational style can give these rocks quite different appearances. FOV ~8 cm.

Fig. 36 - The metabasic eclogites of the UHP Unit sometimes show strong fabrics in omphacite; the one shown here has a single fabric and overprinting shear bands. Two generations of garnets can also be observed. FOV ~4 cm.

CHARACTER OF THE MOVEMENT DIRECTION OF THE COMBIN SHEAR ZONE AND EXHUMATION OF UHP ROCKS

The character and relative timing of the exhumation of the UHP rocks at Lago di Cignana can be determined by structural analysis and geochronology of the Combin Shear Zone and Matterhorn Detachment. Movement indicators across the Combin Shear Zone are complex, with regions preserving multiple stages of overprinting while other regions are structurally simpler, due to either the pervasive overprinting of intense younger events and/or a lack of reactivation of some regions during younger deformational events.

The dominant movement sense and complexity of fabric development within the UHP unit differs to that observed in the overlying Combin Shear Zone and the underlying Zermatt Saas Zone. This is due to the UHP unit being formed at greater depths with a different exhumation history, at least in the initial stages of exhumation compared to the adjacent tectonic slices. The oldest fabric discernible in the UHP unit is defined by eclogite facies minerals: however, sense of movement of this event cannot be positively defined at Lago di Cignana. It is also difficult to know if the UHP event is associated with this early fabric or whether this is an overprinting eclogite facies fabric. Tectonic juxtaposition of the "HP" slices from the Zermatt-Saas Zone against the "UHP" slices is most likely to have occurred during this eclogite facies event. The tectonic regime at this point in the history cannot be constrained, but fabrics are intense and later recumbently folded, and this geometry might reflect a switch from overall crustal extension to overall crustal shortening, for example. This sheared early fabric is overprinted by multiple generations of S-C bands, all of which are suggested to represent extensional regimes. The older S-C bands show a top-to-the-east sense of movement but this is overprinted by a largerscale and more dominantly developed S-C bands with top-to-the west sense of movement. These younger dominant S-C bands often appear as conjugate, the timing may or may not be synchronous.

The dominant movement direction in the UHP unit is in eclogite/blueschist facies with a top to the west-north-west to north-west direction. Fabrics formed in this event cut and overprint the earlier formed tectonic contact between the Zermatt-Saas and Combin Zone. Boudins and foliation boudinage with recrystallisation during deformation characteristically developed at this time.

A switch in movement direction is then observed with a south-east sense of shear. This is a less significant event formed at greenschist facies conditions, often associated with garnet growth. This is the final stage of ductile exhumation observed across the Combin Shear Zone.

Regional-scale structural analysis suggests the existence of a large-scale shear zone, based on variation of strain intensity (cf. REDDY et alii 2003). The geometry of this shear zone, now folded on the lithospheric scale (Fig. 3) suggests at least initially that shear sense was east-southeast directed. The pattern of later overprinting suggests a combination of tectonic mode switches and/or intense stretching and crustal-scale boudinage during the later history of exhumation (Fig. 6).

CONCLUSIONS

A narrow slice of rock that has undergone UHP conditions is exposed on the southern shore of the Lago di Cignana (Figs. 5 and 7), in the Valtournenche, in NW Italy. This sliver of rock has undergone a complex exhumation history and is now located interleaved between the upper structural boundary of the Zermatt-Saas Zone and the lower structural boundary of the Combin Zone. The maximum metamorphic grade recorded by the Zermatt-Saas Zone is high-pressure (HP) eclogite facies, whereas the now dominantly greenschist facies Combin Zone may locally contain medium-pressure (MP) blueschist facies relicts.

In detail, the UHP slice is structurally located in the lower boundary of a km-scale shear zone containing sheared and retrogressed rocks from the Combin Zone. Nevertheless, in the underlying UHP tectonic slice relict, UHP minerals and their fabric and microstructures are preserved. The UHP slice is a nest of boudins within boudins, and comprises metasediments and metabasic rocks. Locally, the UHP slice is overlain by a single tectonic slice of highly sheared Zermatt-Saas eclogitic metagabbro.

This has been transected by the Matterhorn Detachment, so that locally the UHP slice may be in direct contact with the Combin Shear Zone. The UHP slice is preserved as a thin tectonic slice because the locus of the later extensional structures has not precisely followed the trajectory of older thrusts. Fabrics and mineralogy reveal that the Combin Shear Zone has operated through blueschist facies conditions until greenschist facies metamorphic conditions were reached. Intense fabrics are developed, commensurate with large shear strains associated with significant horizontal relative displacement.

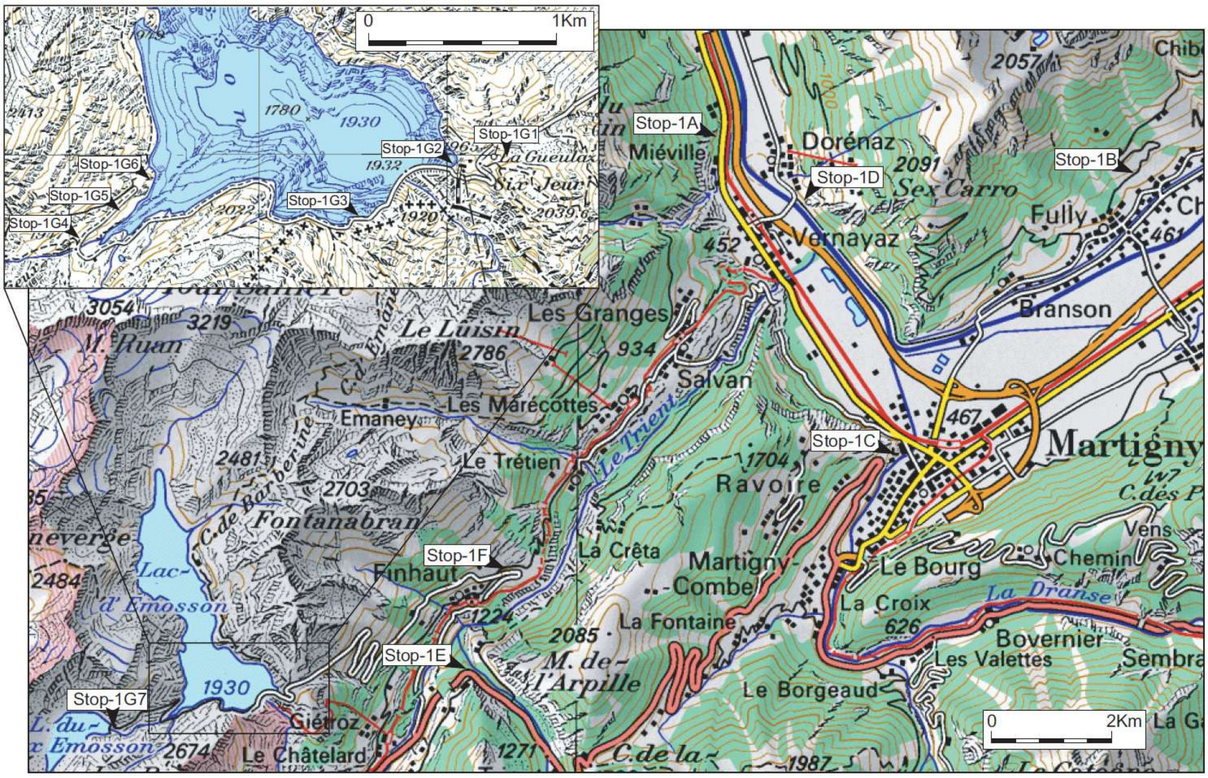
The structural geology of the area suggests a history of large-scale overthrusting, during which the UHP rocks were emplaced over the HP rocks of the Zermatt-Saas Zone. Tectonic inversion subsequent to the period of HP metamorphism led to the Alpine orogen being subjected to large-scale (roughly NW-SE-directed), horizontal stretching. This led to the formation of orogen-scale, extensional shear zones, of which the Matterhorn Detachment could be one of the most important manifestations.

The Matterhorn Detachment and its precursor, the extensional Combin Shear Zone, appear to be responsible for the exhumation of the UHP slice and of the HP rocks of the Zermatt-Saas Zone. FORSTER et alii BALLÈVRE M., & MERLE O. (1993) – The Combin Fault: compressional reactivation of a Late Cretaceous-Early Tertiary detachment fault in the Western Alps. Schweiz. Mineral. Petrogr. Mitt., 73: 205-227.

REFERENCES

COMPAGNONI R., & ROLFO F. (1999) - Characteristics of UHP pelites, gneisses, and other unusual rocks. Intern. Geol. Review, 41: 552-570.
KLAUW S.N.G.C., VAN DER REINECKE T. & STOCKHERT B. (1997) - Exhumation of ultrahigh-pressure metamorphic oceanic crust from Lago di Cignana, Piemontese zone, western Alps: the structural record in metabasites. Lithos, 41: 79-102.
REDDY S.M., WHEELER J. & CLIFF R.A. (1999) - The geometry and timing of orogenic extension: an example from the Western Italian Alps. J. Metam. Geol., 17: 573-589.
REDDY S.M., WHEELER J., BUTLER R.W.H., CLIFF R.A., FREEMAN S., INGER S., PICKLES C. & KELLEY S.P. (2003) - Kinematic reworking and exhumation within the convergent Alpine orogen. Tectonophysics, 365: 77-102.
REINECKE T. (1991) - Very-high-pressure metamorphism and uplift of coesite-bearing metasediments from the Zermatt-Saas zone, Western Alps. Eur. J. Mineral., 3: 7-17.

Friday 30.8.:
Lac d´ Emosson über Mt.Blanc Tunnel - Chamonix
Program: Sesia Zone still high pressur rocks, antique Roman road, basement of the Mt. Blanc – Aiguille Rouge Massif and unconformity; Triassic quartzites with fossil tracks. Total walk: 5 h along mountain trails, total altitude difference:700 m up and down. Total drive**:** 170 km.
Accommodation: Martigny



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

Topic: Variscan metamorphism/anatexis and Alpine overprint of sedimentary and igneous lithologies.
Outline: The Lake Emosson area is one of the best sites to observe the polymetamorphic basement of the Alpine External Massifs. The oldest lithologies are upper Proterozoic to lower Palaeozoic

sediments and volcanites, which were intruded by Ordovician granitoids (orthogneisses), before all rocks underwent Variscan metamorphism. The latter was of high amphibolite facies grade and locally induced partial melting (see field-trip introduction). Metamorphism was accompanied by long-lasting deformation with superposition of at least three fold generations, the last one being of kilometer scale. The latter is identified by ubiquitous Z-, S- and M-shaped parasitic folds, which allow distinction between adjacent anti- and synforms. Alpine metamorphism reached low greenschist facies grade. All units are unconformably overlain by sandstones of Triassic age, hosting the famous saurian footprints (Demathieu and Weidmann 1982) in the Vieux Emosson Lake area. The Alpine dome-like structure of the massif is underlined by the position of the Mesozoic sediments, which rest horizontally on top of the basement rocks in the middle of the massif (Aiguille de 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 is visible behind the small chapel facing the restaurant (coord. 561310/102000). It is characterized by a 1 m thick brecciated and silicified band, with fibrous quartz crystallized radially all around the clasts. This spectacular texture has been interpreted as a result of hydraulic brecciation during the shallow level intrusion of a fluid-saturated granitic magma (Genier, 2000).

Mylonite-Zone (stop 1G2)

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

Metasedimentary units (stop 1G3)

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

(I)
a unit of graywackes with metapelitic interlayers;

(II)
a mixed unit composed mainly by metapelites with some thin metagraywacke layers, one quartzite horizon, one layer of carbonates (appearing mostly as large boudins), and hosting one or two amphibolite layers;

(III)
a third unit characterized by finely banded metagraywackes and metapelites with a rusty patina. Many detailed observations are found in von Raumer (1983), von Raumer and Schwander (1985), Schulz & von Raumer (1993), Dupasquier (1996), Schmocker (1996), Fracheboud (1997), and Marquis (1997).

A comparison with other European lithostratigraphic sections brings convincing evidence that these lithologies have a Late Proterozoic to Early Palaeozoic age (see field-trip introduction).

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

Metapelites and metagraywackes (stop 1G5) record a Barrowian-type of metamorphism with early [biotite-staurolite-kyanite-garnet] assemblages evolving towards [biotite-garnet-sillimanite] parageneses. Joye (1989) locally observed sillimanite and cordierite in strongly sheared rocks among the mylonites. The thermal peak has been dated at 327 ± 2 Ma (U/Pb on monazite, Bussy et al. 2000). Late stages are quartz segregation lenses with K-feldspar and andalusite. Such a sequence of parageneses does not necessarily represent a continuous PT-path, but could reflect two distinct events, i.e. an early-Variscan high pressure phase and a late-Variscan, more temperature dominated phase. Geochemical data from different localities indicate that metagraywackes carry the fingerprints of an active margin setting (Bhatia 1983), resulting from the erosion of quartzitic lithologies or acidic volcanites (Roser and Korsch 1988). This agrees well with our general interpretation of shelf sediments located at the Gondwana active margin (see introduction).

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

Metabasites (stop 1G6) appear as 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 TiO_2 and V-contents (>0.5% and > 100 ppm, respectively) suggests a volcanic or subvolcanic origin (Pfeifer et al. 1989). The original rocks were interpreted (von Raumer et al. 1990) as transitional MORB volcanites typical for continental rift zones at the onset of ocean floor spreading. Their age could be either Cambrian or Ordovician. Some of the largest amphibolite bodies are accompanied by leucocratic garnet-bearing gneisses, containing large crystals of staurolite and kyanite, the latter with reaction rims of cordierite and hercynite, interpreted as a HT-decompression reaction (Dupasquier, 1996). The origin of these leucocratic rocks remains unexplained, but they could result from dehydration melting of amphibolites comparable to those observed in the Lake Cornu area (von Raumer et al. 1996).

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

Alpine deformation

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

Saturday: 31.8.
Martigny – Tübingen **Program:** Late Carboniferous Salvan-Dorenaz trough with fossil woods near Martigny Total drive 440 km**:** back to Tübingen. **Arrival at Tübingen ~7.00 pm**

STOP 1D - DORÉNAZ QUARRY [569630/110330]: SALVA N -DORÉNAZ LAT E CARBONIFEROUS SEDIMENTARY BASIN For location look Fig. above from Friday.

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.
Back to Tübingen


Cassonsgrat, in front of the Glarus thrust


Slump within the Breggia Gorge


Val Tournanche, ascent to Lago di Cignana


Ascent to Lac d´Emosson

Over the Lac d´Emosson

Bernd with Matterhorn from the South