Results of NRP 20

Deep Structure of the Swiss Alps

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Cover photograph

The photo of the Swiss Alps in the Valais region displays (from left/east to right/west) the peaks of Monte Rosa, Breithorn, Matterhorn, Dent Blanche and Weisshorn.

The summit areas of Matterhorn, Dent Blanche and Weisshorn are erosional remnants of a large thrust sheet of crystalline basement derived from the Adriatic microplate, a spur of the African plate. This thrust sheet overlies a nappe complex that evolved from the subduction of the Piemont ocean and that extends from Breithorn to the lower slopes of Matterhorn, Dent Blanche and Weisshorn. Monte Rosa is carved out of crystalline basement pertaining to a still lower thrust sheet derived from the Briançonnais continental fragment. This entire nappe pile overlies the nappe stack derived from the Eurasian plate, which is buried deeply beneath the area of the photograph, as was imaged by seismic experiments of NRP 20.

14 Rifting and collision in the Penninic zone of eastern Switzerland

S. M. Schmid, O. A. Pfiffner & G. Schreurs

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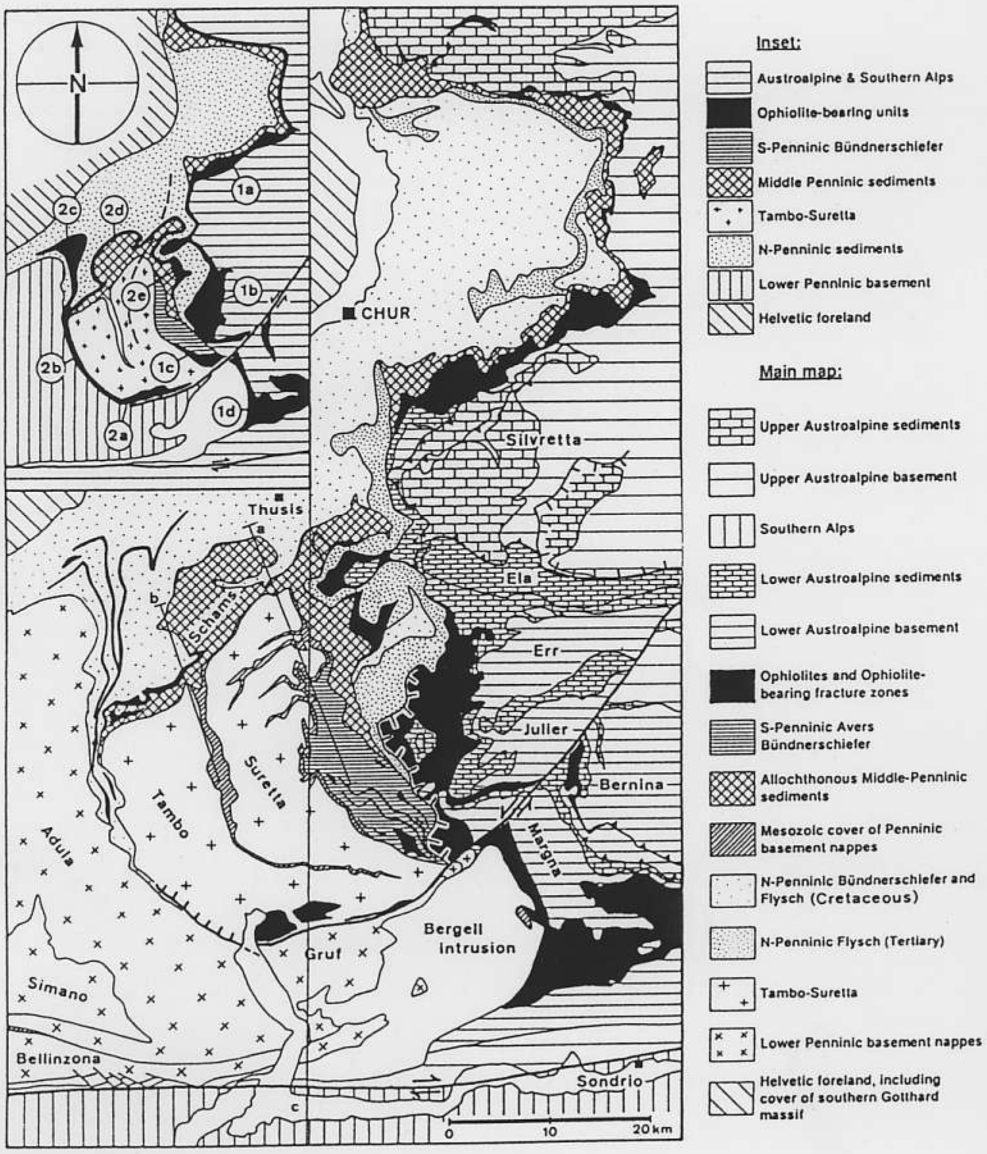


Figure 14-1
Tectonic map of eastern Switzerland, modified after Schmid et al. (1990). Encircled numbers in inset refer to ophiolite-bearing units (black) derived from the S-Penninic or Piemont-Liguria ocean (units labelled "1" and from the N-Penninic or Valais ocean (units labelled "2"). 1a: Arosa; 1b: Platta; 1c: Lizun and Avers; 1d: Malenco; 2a: Chiavenna; 2b: Misox zone; 2c: ophiolites within N-Penninic Bündnerschiefer; 2d: Areua-Bruschghorn; 2e: Martegnas. Profile traces a and b refer to Figures 14-10a and 14-10b, respectively. Profile trace c refers to Figure 14-2.

14.1 General geological introduction

The southern part of the eastern traverse of the NRP 20 reflection profile E 1 (see chapter 9) crosses the Penninic domain of eastern Switzerland (Figure 14-1). The following brief introduction into the geological setting intends to familiarize the reader with the study area. Inevitably, such an introduction is partly based on new findings presented in more detail later on.

Compositionally, the Penninic structural domain comprises:

(1) A stack of pre-Triassic crystalline basement units consisting of the Adula, Tambo and Suretta nappes within our area of interest. The two higher amongst these three basement nappes largely preserved their autochthonous to parautochthonous Mesozoic cover (Tambo cover or Splügen zone, Suretta cover) and contain relics of pre-alpine (Variscan and Late Paleozoic) magmatism, deformation and metamorphism. Hence they basically represent basement-cover flakes derived from upper continental crust, albeit strongly overprinted by Alpine deformation and metamorphism (mostly greenschist facies). The Misox zone definitely does not simply represent the cover of the Adula nappe but forms the telescoped southern continuation of detached sedimentary units described under (2). The Adula basement nappe, in contrast to Tambo and Suretta nappes, is very intensely sliced into sheets of quartzo-feldspathic basement interleaved with very thin eclogite facies mafics and Mesozoic metasediments.

2) Allochthonous cover sheets are found in front of, above and below the Tambo-Suretta pair. The Middle Penninic or Briançonnais cover slices are characterized by carbonate-rich platform sediments with reduced thickness (Schams, Falknis-Sulzfluh nappes). The bulk of the rather monotonous calcareous shales and arenites (N-Penninic Bündnerschiefer and Flysch, deposited in the Valais trough) are found in front of the Tambo-Suretta pair. Their extension into the footwall of the Tambo nappe forms the major part of the Misox zone while the extension into the hangingwall of the Suretta nappe is known as Arblatsch flysch.

3) Ophiolite-bearing units shown in Figures 14-1 and 14-2 contain, amongst other rock types, rocks which unambiguously represent former oceanic crust (e.g. Platta, Martegnas). Mafic and ultramafic lithologies whose origin is not entirely clear occur as components of mélange zones (Areua-Bruschghorn mélange) between N-Penninic Bündnerschiefer and Schams nappes or as imbricates, preferably at the base of individual tectonic-stratigraphic subunits within the N-Penninic Bündnerschiefer (Figure 14-2).

Parts of the Tambo and Suretta basement nappes suffered pre-Alpine as well as Alpine metamorphism (polycyclic basement). Post-Variscan (Permo-Carboniferous) granitoids and volcanics (Truzzo granite, Rofna porphyry) underwent one (Alpine) metamorphic event only (monocyclic basement). Judging from their Mesozoic cover, the Tambo and Suretta basement flakes represent the uppermost few kilometers of continental crust, paleogeographically part of the Briançonnais platform (Middle Penninic). This rise started to individualize in the early Middle Jurassic as the result of oblique rifting and drifting. A more external basin, the Valais or N-Penninic trough is characterized by very thick Jurassic and Cretaceous turbiditic sequences (Bündnerschiefer) grading into Late Cretaceous to Eocene flysch. Interlayered prasinites and serpentinites are interpreted here as tectonically emplaced remnants of oceanic crust. Classically, they have been described as stratiform intrusions into the Bündnerschiefer, which were deposited onto continental crust ("Adula"-Bündnerschiefer). While in the latter view the Briançonnais simply represents distal European margin, the former interpretation results in a more complicated paleogeographic scenario with two oceanic domains separated by the continental Briançonnais platform. An oceanic origin for the S-Penninic or Piemont-Liguria basin (e.g. Platta unit) is undisputed.

Alpine convergence presumably started in "Mid"-Cretaceous times in the S-Penninic domain. Upper Cretaceous W to WNW-directed thrusting and metamorphism is documented within the Austroalpine nappes (Thöni 1986, Schmid & Haas 1989, Froitzheim et al 1994), synchronous with the suturing of the Austroalpine nappes with the Arosa-Platta ophiolitic unit (Ring et al. 1988). However, the extent of deformation at the southern margin of the Briançonnais platform during the Cretaceous as well as the timing of collision between the Austroalpine and the Briançonnais will have to be discussed.

The main deformation affecting the more external mid-Penninic rise, the Valais trough and the distal European margin (hence the entire nappe pile below the Platta unit of Figure 14-2) did not occur before the Tertiary according to our results (for differing point of views, calling for Cretaceous deformation and metamorphism in this nappe pile, particularly for the Adula nappe, see Hunziker et al. 1989, Ring 1992 a & b). Hence, the present day structure of the Penninic structural zone depicted in Figure 14-2 is essentially the result of Tertiary orogeny. Late Cretaceous uplift and cooling, associated with severe extension (Froitzheim 1992) rendered the Austroalpine units virtually undeformable during Tertiary orogeny: Austroalpine and Platta units behaved as an orogenic lid and were thrusted N-wards over the viscously deforming Middle and North Penninic units (Schmid et al. 1990, Froitzheim et al. 1994). Stacking into basement and cover nappes is the result of foreland tectonics in the distal European margin, and (if the N-Penninic domain represents former oceanic crust), accretionary wedge formation leading to collision of the distal European margin with the Briançonnais platform.

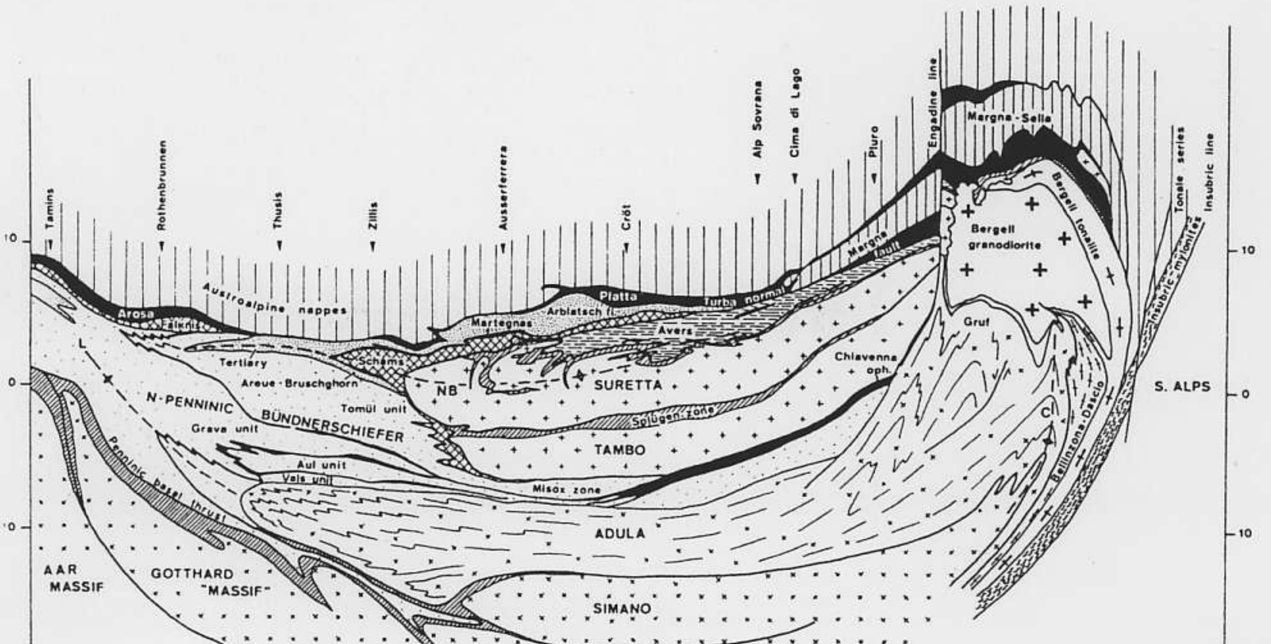


Figure 14-2

N–S profile through the Penninic zone of eastern Switzerland, drawn after the seismic line E1 and the integrated cross section of the Eastern Traverse fully presented and discussed in Chapters 9 and 22. Axial traces indicated refer to the Lunschania antiform (L), the Niemet-Beverin fold (NB) and the Cressim antiform (C). Profile trace indicated in Figure 14-1, trace c.

Intense post-nappe folding locally inverted the stacking order of previously detached units. This strong overprint is associated with post-collisional shortening starting in the Oligocene. It is important to stress the fact that in the external zones (Helvetics, Southern Alps) substantial post-collisional shortening went on at least until the end of the Miocene, while our area of interest was largely consolidated by the end of the Oligocene.

The foregoing introduction showed that a palinspastic reconstruction of the Penninic domain in its paleographical sense is crucial for a better understanding of tectonic processes such as rifting and collision. An answer to the following two questions is fundamental for understanding the tectonic evolution along the NRP 20 Eastern Traverse: What is the original position of the Schams nappes in the nappe pile before post-nappe refolding, and, what is the nature of the Valais trough (oceanic or distal European margin)?

The results of detailed studies carried out in conjunction with the NRP 20 project concentrated on an area occupied by the frontal part of the Tambo and Suretta nappes, including the allochthonous Schams cover slices wrapped around the front of these basement nappes. Since stratigraphic-sedimentological and structural-tectonic arguments are not independent from each other a combined interdisciplinary approach was chosen. For clarity of presentation, however, the results on the rifting and drifting stage will be presented before the results on Alpine convergence. But the reader should bear in mind that the stratigraphic-sedimentological data presented first heavily rely on the results of the structural analysis and vice versa. For details on the work carried out in connection with NRP 20 the reader is referred to the results of three PhD theses (Mayerat 1989, published in modified form under Mayerat Demarne 1994, Rück 1990, published1995, Schreurs 1990, published 1995) and unpublished diploma theses (Adler 1987, Schegg 1988, Pauli 1988 at ETH Zürich, and Hitz 1989, Dalla Torre 1991, Christen 1993 at the University of Berne). Previous summaries have been published by Pfiffner et al (1990), Schmid et al. (1990) and Schreurs (1993). This summary will additionally make an attempt to place these detailed studies into a larger context by comparing them with published and ongoing work in neighbouring areas in order to arrive at a larger scale paleogeographic and kinematic picture.

14.2 Mesozoic rifting and drifting

14.2.1 Introduction

Amongst the major paleogeographical domains found in the Alps (Helvetic, Penninic, Austroalpine-Southalpine) the Penninic domain is the most problematic one. To regard the Penninic domain as simply representing an oceanic suture zone between the European foreland (Helvetic domain) and the overriding Apulian plate (Austroalpine-Southalpine) would be a gross oversimplification. The Penninic nappes in the sense of a series of nappes overriding the Helvetic nappe system and found in the footwall of the Austroalpine nappe system do not represent a single paleogeographic domain. These Penninic nappes are made up of three categories: (1) basement nappes and their autochthonous to parautochthonous cover, (2) detached Mesozoic cover nappes or slices (e.g. Schams nappes, Bündnerschiefer and Flysch) whose derivation from individual basement nappes or from nappe-dividing Mesozoic thrust zones ("roots") is controversial, and (3) ophiolite-bearing units.

A Penninic paleogeographical domain, corresponding to the Penninic nappe system, is hard to define for several reasons. An extension of the former northern passive continental margin (Helvetic) into units occupying a structural position within the lower parts of the Penninic domain in the structural sense is to be expected (e.g. Milnes 1974, Subpenninic units). This raises serious problems in terminology because of historical reasons. Unfortunately, no clear distinction is usually made between "Penninic" in the paleogeographical and tectonic sense, a heritage of Argands "Embryonaltektonik" (Argand 1916), associated with extreme cylindrism regarding paleogeographical domains which, according to him, predetermine future nappes in "embryonic" form. Unfortunately, the breakthrough of modern paleogeographic reconstructions based on the concepts of plate tectonics (passive continental margin formation, ocean floor spreading, accretionary wedge formation) did not lead to a corresponding change in the traditional terminology, which still reflects the concept of a geosyncline.

Paleogeographical reconstructions, apart from nomenclatural problems, are also hampered with other major problems: (1) many sediments are ill-dated; (2) basement-cover relationships are obscured by décollement of sedimentary cover slices whose position in respect to basement nappes is often a matter of debate (e.g. the "Schams dilemma", see Trümpy and Haccard 1969, Trümpy 1980), in some cases corresponding basement units are missing altogether; (3) polyphase penetrative deformation often associated with large scale refolding of a previously emplaced nappe pile (e.g. Milnes 1974) prevents simple paleogeographic reconstructions; (4) intense deformation associated with metamorphism makes stratigraphical and sedimentological studies extremely difficult.

This chapter reports some progress made regarding the paleogeographical evolution of an important part of the NFP 20 Eastern Traverse. In the framework of the NFP 20 project stratigraphic-sedimentological investigations primarily aimed at (1) rocks of the Schams nappes (Rück 1990, Schmid et al. 1990), in close conjunction with structural work (Schreurs 1990) and (2) similar cover rocks presently found at the front of the Tambo nappe investigated by combined structural and stratigraphic work (Mayerat 1989, Pfiffner et al. 1990). The results of these investigations will be discussed in the context of older investigations (e.g. Falknis-Sulzfluh nappes, Gruner 1981) and very recent investigations (e.g. Bündnerschiefer; Steinmann et al. 1992, Steinmann 1994) in neighbouring sedimentary realms with the aim of deriving a larger scale paleogeographical reconstruction crucial for understanding the Alpine orogeny.

14.2.2 Stratigraphy and sedimentology of the Schams nappes

14.2.2.1 Paleogeographic and tectonic units in the Schams nappes

In order to discuss stratigraphy and sedimentology of the Schams nappes, or better the Schams cover slices, it is convenient to subdivide the paleogeographical realm covered by the Schams slices into roughly E-W trending units and subunits depicted in Figure 14-3 a and b (Rück 1990). This subdivision is based on a combined structural and sedimentological approach (Schmid et al. 1990). Units and subunits primarily denote paleogeographical entities, but abrupt facies changes control the position of décollement horizons and, consequently, the structural subdivision of the Schams nappes (Figure 3c and d). These facies changes are most pronounced in the Middle Jurassic to Lower Cretaceous strata. The reconstruction along a N–S paleogeographical profile is based on a detailed structural analysis (Schreurs 1990).

Presently, subunit 1b (hemipelagic basin) may be directly linked with subunit 1a (proximal slope characterized by the Vizan breccia) around a F1 fold hinge whose location is predetermined by an abrupt facies change (Figure 14-3). Both subunits 1b and 1a roughly correspond to the Gelbhorn nappe of Streiff (1939, 1962), Jäckli (1941) and Neher in Streiff et al. (1971/1976). Over large distances the Gelbhorn unit 1 has been detached along the Carnian evaporites, older sediments and basement have been left behind in a yet unkown position at the base of the Tambo basement nappe. Only where this décollement horizon is missing due to Mesozoic erosion (locally in subunit 1a) have older sediments and basement slivers (Taspinit and Nolla basement, Figure 14-4) also been incorporated into the Schams cover slices.

Only locally are F1 hinges between Gelbhorn unit 1 and Tschera-Kalkberg unit 2 preserved. Subunit 2a (partly but not entirely identical with the Tschera nappe of Streiff, 1939) simply represents the Upper Triassic and younger cover of subunit 2b (Gurschus-Kalkberg nappe of Streiff et al., 1971/1976), 2a and 2b being detached from each other during early F1 imbrication along the Carnian evaporites. The basal detachment of the Schams cover slices in general changes from the Carnian evaporites (Gelbhorn unit 1) to the base of the Triassic carbonates (unit 2b), both décollement levels being at a similar depth according to the reconstruction in Figure 14-3c. This geometry is strongly controlled by the paleogeography: in parts of subunit 1a, and particularly in subunit 2c (Figure 14-3a,b), Carnian evaporites are missing due to predepositional erosion at the base of the Middle to Upper Jurassic breccia formations; consequently, décollement has to occur at the base of the Triassic carbonates. Subunit 2c (partly corresponding to the Wissberg nappe of Kruysse 1967, see also Pauli 1988) locally exposes a coherent sequence from the base of the Triassic carbonates into the Upper Cretaceous. Large parts of this subunit, however, are strongly dismembered by Alpine tectonics, in particular by large scale boudinage of the competent Middle Triassic carbonates.

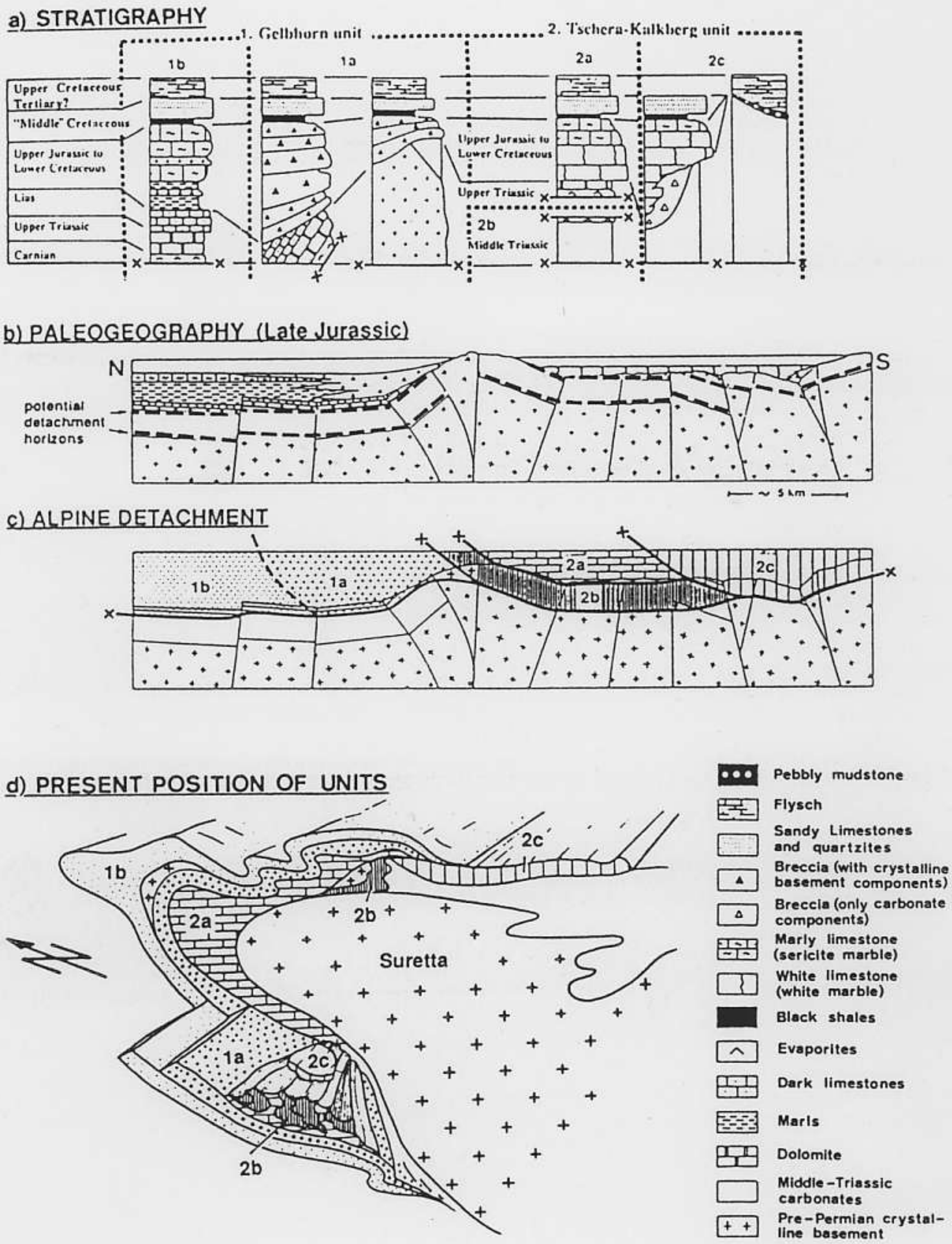


Figure 14-3
Schematic overview of stratigraphy and structure of the Schams nappes after Rück (1990).
a) Stratigraphy of the Schams units.
b) Sketch of the paleogeography during the Late Jurassic epoch.
c) Position of future Alpine detachments.
d) Sketch of the present day cross section through the Schams nappes.

14.2.2.2 Lithostratigraphy of the Schams nappes

Correlations of the formations and series schematically illustrated in Figure 14-4 are mainly based on lithologies which at least locally occur in all the subunits. These are in particular the Tumpriv-Serie reaching from the Carnian evaporites into the Lower Jurassic (fossil occurrences in the Rhaetian and the Pliensbachian, Wilhelm 1933) and, additionally, the Ölquarzit and Plattensandstein (erroneously referred to as Platten-"quarzit" by Streiff et al. 1971/1976) embedded in anoxic black shales. The latter lithologies exhibit a high content in organic carbon and show striking lithological similarities to well dated formations in the Falknis nappe (Allemann 1957 dated the top of the Plattensandstein, locally referred to as "Gault", to be of Cenomanian age). Rück (1990) correlates these lithologies with the "Mid"-Cretaceous anoxic event (Jenkyns 1980). There are no further age determinations for the intervening formations. Continuation of sedimentation into the Tertiary is to be expected for parts of the Schams nappes from the presence of dated Tertiary sediments in the neighbouring Arblatsch flysch (Lower Eocene, Ziegler 1956, Eiermann 1988) and in the Falknis nappe (Paleocene, Allemann 1957).

Correlations across subunits 1a and 1b are facilitated by the presence of distal turbidites of the Vizanbrekzien-Serie in subunit 1b. Similar interfingering is also observed across the boundary between subunits 1a and 2a (Figure 14-4). The Untere Sericitmarmor is common to subunits 1a and 2a. Its age is probably Latest Jurassic to Early Cretaceous because it overlies the Tschera-Marmor-Serie (white marbles and carbonate breccias) found in subunits 2a,b and c, considered to be of Upper Jurassic age (based on facies analogies to the Sulzfluhkalk of the Sulzfluh nappe and Upper Jurassic limestones in the Préalpes Médianes), and because it is covered by "Mid"-Cretaceous formations. The Vizanbrekzien-Serie locally encompasses the entire age bracket between Early Jurassic (post-Pliensbachian) and Mid-Cretaceous times. It locally cuts down section due to pre-depositional erosion, in places by more than 600 m (maximum thickness of the Middle Triassic carbonates) and down to the pre-Triassic basement. The actual thickness of the Vizanbrekzien-Serie is extremely variable and may reach 250 m (corresponding to 500 m after retrodeformation according to a strain analysis carried out by Schreurs 1990). The minimum thickness of the post-Carnian cover in subunit 1b amounts to about 400 m, intensive layer-parallel cleavage suggests that this is only a fraction of the original thickness. In subunit 2 the thickness of the post-Triassic sediments is highly variable and may reach about 100 m, this again being an absolute minimum (due to intensive cleavage formation). Because of these large uncertainties the thicknesses in Figure 14-3 and 14-4 are only approximately scaled. The horizontal scale in Figure 14-3 is estimated according to an area balance (24 km^2 are occupied by Schams nappes and related units at the front of the Tambo nappe in a N–S section) and a very rough estimate of the average thickness of the Schams cover (800m). According to this estimate the extreme facies variations across the various subunits occur over a lateral distance of only some 30 km.

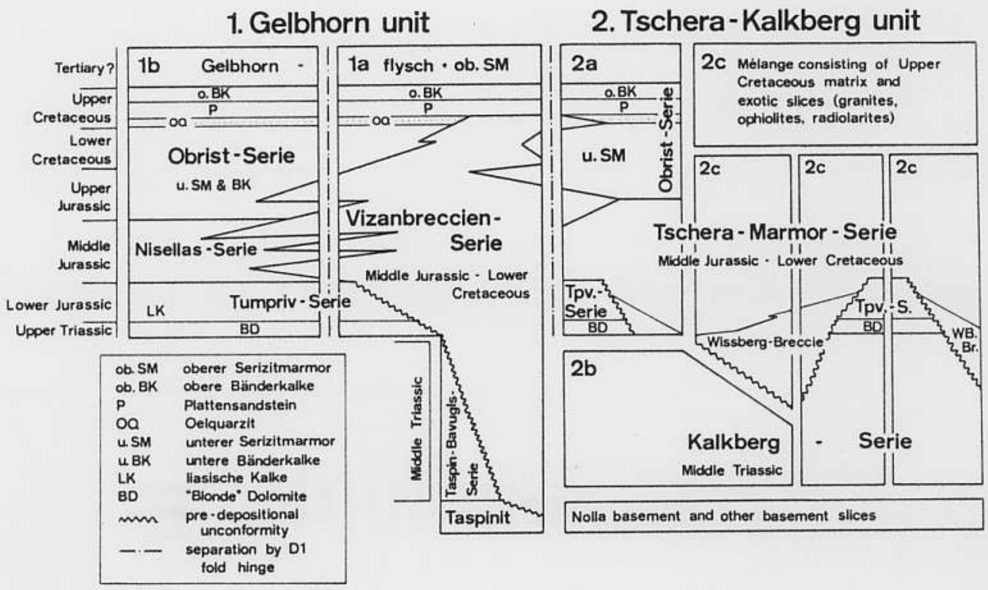


Figure 14-4
Lithostratigraphy of the Schams nappes, after Rück (1990)

14.2.2.3 Basin evolution of the Schams nappes

Information on the nature of the **pre-Triassic basement** is crucial for speculations about the present-day position of the basement of the detached Schams sediments. Such information is found in thin detached basement slivers (Taspinit and Nolla basement) and from breccia components in the Vizanbrekzien-Serie. Microstructural investigations indicate that the locally preserved pre-Alpine foliation in the Taspinit and Nolla basement formed under upper-greenschist to amphibolite grade conditions (Schreurs 1990). Hence the Rofna porphyry, typical for the entire front of the Suretta nappe, and similar "monocyclic" basement lithologies (i.e. lithologies which did not undergo a pre-Alpine tectonometamorphic event) in the Tambo nappe can be excluded to have formed the "homeland" of the Schams sediments. The erroneous statement in Schmid et al. (1990), that Vizanbrekzien-Serie pebbles resemble the Rofna porphyry was based on macroscopic appearance. Also excluded as source area for crystalline pebbles is the top of the Suretta and Tambo nappes since they are in most places sealed by their own Mesozoic cover. Detachment of the Schams nappes from the base of the Tambo and Suretta thrust sheets is extremely unlikely for structural reasons. The Adula nappe is separated from the Schams nappes by the N-Penninic Bündnerschiefer of the Misox zone and can therefore be ruled out as well.
In conclusion, it seems that most of the sediments of the Schams nappes have been originally deposited on polycyclic continental basement (i.e. basement which underwent pre-Alpine deformation and metamorphism) which was almost completely subducted during Alpine orogeny. This basement is only preserved in the form of thin slivers within the Schams nappes and it is unlikely that the large basement nappes depicted in Figure 14-2 (Adula, Tambo, Suretta) can be considered as representing the homeland of the detached Schams sediments.

The **early stages of subsidence** led to the deposition of a thick (max. 600 m) Middle Triassic carbonate platform, the Upper Triassic dolomites above the Carnian evaporites exhibiting a very reduced thickness and intercalations of clay formations identical with the Helvetic Quartenschiefer ("Carpathian" facies). In contrast to the Upper Austroalpine and part of the Lower Austroalpine passive continental margin the Lower Jurassic limestones indicate a transition from neritic to open marine hemipelagic conditions and evidence for extensive rifting of pre-Toarcian age is missing.

The **rifting and/or transcurrent faulting stage** starts immediately after the Pliensbachian, possibly in the Toarcian represented by black shales containing the first distal turbidites of the Vizanbrekzien-Serie in subunit 1b. These black shales grade into carbonate bearing shales and marls, virtually undistinguishable from the N-Penninic Bündnerschiefer (Nisellas-Serie). In subunit 1a, however, the base of the Vizanbrekzien-Serie, where conformable on the Tumpriv-Serie, immediately overlies Liassic limestones. Locally, these Liassic limestones are Gryphäa-bearing and covered by a thin veneer of black shales (Toarcian?). It is important to note that while conformable parts of the base of the Vizanbrekzien-Serie indicate the initiation of rifting and/or transcurrent faulting in or immediately after the Toarcian, the top of the Vizanbrekzienserie (typically represented by basement derived arkoses) locally grades conformably but abruptly into the Plattensandstein of "Mid"-Cretaceous age (at Piz Vizan, Rück 1990). Hence the entire age interval from Mid-Jurassic to "Mid"-Cretaceous is represented by the Vizanbrekzien-Serie. The exact age of intermediate lithostratigraphic levels remains unkown.
The depositional geometry of the Vizanbrekzien-Serie, sediment transport directions and the geometry of the basal unconformity (only schematically sketched in Figure 14-5) have been interpreted in terms of a transpressive rather than distensive regime (see Rück 1990 and Schmid et al. 1990 for further details). About 50% of the Vizanbrekzien-Serie is made up of gravity flow and rock-fall deposits, indicating a very strong relief and near-source deposition almost exclusively to the N of a roughly E–W trending fault escarpment (Figure 14-6). This fault escarpment delimits the N end of a platform represented by unit 2. It is depicted as sinistrally transpressive in Figure 14-6, the sense of shear being purely based on the large scale plate tectonic framework. The remainder of the resediments is made up of proximal turbidites.

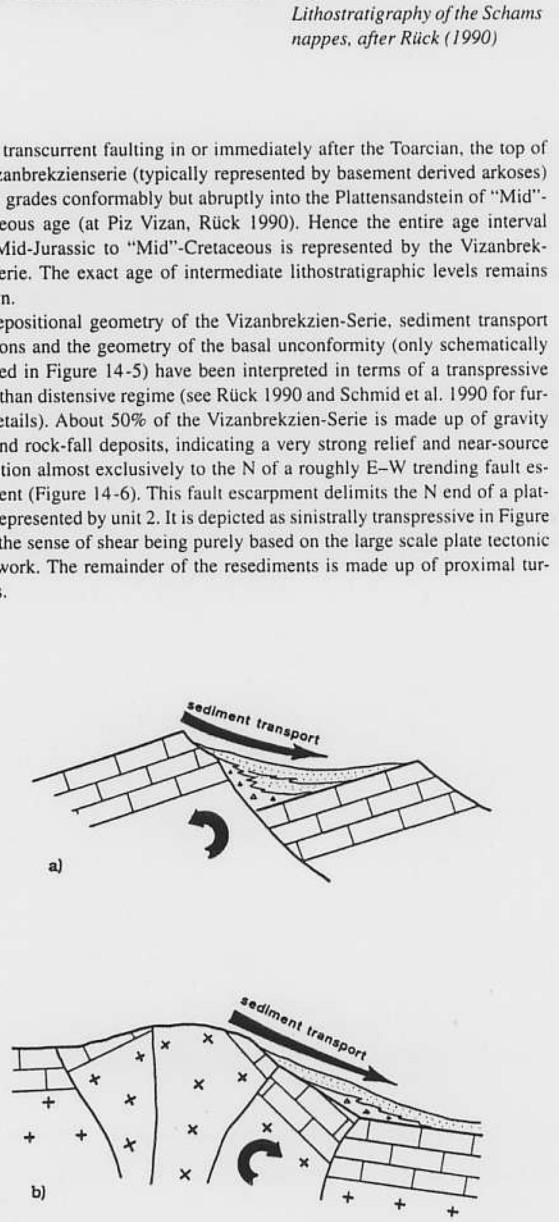


Figure 14-5
Sketch visualizing relations between tilt-direction of pre-rift sediments and sediment transport direction of syn-rift sediments for the case of (a) extension or transtension and b) transpression. After Rück (1990), for discussion see text and Rück (1990).

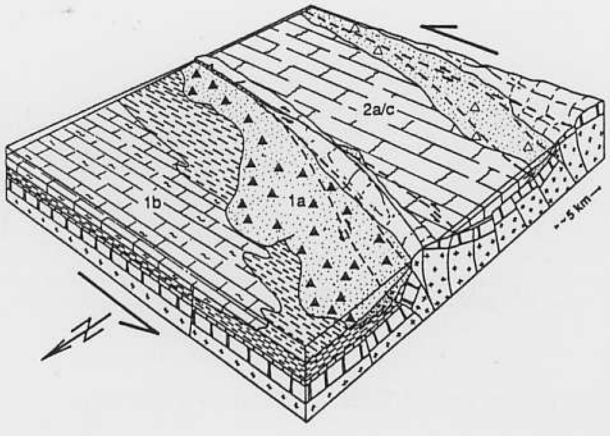


Figure 14-6
Block diagram of the paleogeography of the Schams nappes during Late Jurassic times, after Rück (1990). For symbols and numbers see Figure 14-3a.

Sedimentological data indicate that the fault escarpment remained active untill "Mid"-Cretaceous times (Rück 1990). Clast abundance generally indicates a tendency for upwards increasing amounts of crystalline basement and generally a thinning-fining-upward tendency is observed, occasionally in two cycles (Piz Vizan, Piz Tschera, Rück 1990). Contemporaneous tectonic activity is responsible for continuous supply of pebbles, almost exclusively derived from a local source (basement, overlying Middle Triassic carbonates and Tumpriv-Serie). This ongoing tectonic activity is directly manifested by Mesozoic cataclasites within the Taspinit basement (Bavugls area, see Rück 1990). Abundant neptunian dykes in basement and pre-breccia cover indicate extension and illustrate that this tectonic activity was not of a transpressive nature for the entire time interval covered by the Vizanbrekzien-Serie (Mid- Jurassic to "Mid"-Cretaceous). Large scale considerations also indicate that the Schams nappes were part of a passive continental margin setting associated with substantial crustal thinning in the Early to Middle Jurassic. It is not clear during which time interval the transpressive regime inferred from the depositional geometry (Figure 14-6) prevailed. Also, transpression may only be a local phenomen, restricted to the Schams area.
Contemporaneous sediments in unit 2 are typical platform sediments (Tschera-Marmor-Serie topped by local occurrences of Unterer Sericitmarmor and locally with the Wissberg breccia at the base). The Wissberg breccia is entirely composed of reworked Triassic and Jurassic carbonates and is, in contrast to the Vizanbrekzie, devoid of basement clasts. Direct transgression onto Middle Triassic carbonates and erosion of intervening lithologies (Tumpriv-Serie) is reminiscent of the Préalpes Médianes Rigides. In subunit 1b the already described Nisellas-Serie is topped by impure carbonates (mainly Unterer Sericitmarmor and untere Bänderkalke of the Obrist-Serie, identical with the so-called "Nivaigl-Serie" mapped in the E-Schams by Streiff et al. (1971/1976).

A **tectonically quiescent phase** before the onset of Alpine convergence is indicated by the "Mid"-Cretaceous lithologies which are identical in all subunits. Obviously, the ongoing tectonic activity which is responsible for the preservation of the extreme facies variations characteristic for older formations within the Schams nappes came to a halt by "Mid"-Cretaceous times. The resediments found in the Ölquartzit and Plattensandstein are very mature and are widespread over the entire Tethys (Weissert 1981, 1989).

The **Alpine convergence and accretionary stage** is represented by the ill-dated Gelbhorn "flysch" which contains the Obere Sericitmarmor, probably representing the Couches Rouges. Thin-bedded limestones, intercalated with shales form the remainder of this "flysch". Unit 2 is characterized by mélange formations, embedded in a Late Cretaceous matrix of pebbly mudstones (planktonic foraminifera described by Pauli 1988 and Neher in Streiff et al. 1971/1976). This mélange locally contains exotic slices of granite, ophiolites and radiolarites (subunit 2c), indicating the proximity to and/or accretion of the southernmost Briançonnais to the S-Penninic oceanic domain at this time. While it is possible that accretion and mélange formation started in Late Cretaceous time within the southernmost subunit 2c, it is possible that relatively undisturbed sedimentation (which does not exclude coeval accretionary wedge formation) continued into the Tertiary within unit 1b.

14.2.3 Stratigraphy of neighbouring sedimentary units

14.2.3.1 The sediments at the front of the Tambo nappe

Mayerat (1989) showed that large parts of these sediments, referred to as Areua-, Vignone- and Knorren-Zone by Gansser (1937) are made up of Schams cover slices and that the Schams nappes near Splügen may be directly traced across the Hinterrhein valley and into the area in the front of the Tambo nappe. The Areua basement slice, not being part of the Schams nappes, has a very reduced autochthonous cover (Permo-Carboniferous, basal Triassic) and is part of the Areua-Bruschghorn mélange (Schmid et al. 1990) enveloping the Schams nappes and continuous with the Martegnas mélange in the E-Schams. The Schams sediments in front of the Tambo nappe are in tectonic contact with the Areua mélange and have been detached within the Carnian evaporites. The lithologies found are diagnostic for both subunits 1a and 1b: Tumpriv-Serie, Vizanbrekzien-Serie (predominantly dolomitic and rarely with basement pebbles), Unterer Sericitmarmor, Plattensandstein, Oberer Sericitmarmor and Gelbhorn "flysch". Hence, only elements of subunits 1a and 1b are present, while subunit 2 is completely missing (no Middle Triassic nor Tschera-Marmor-Serie).

Mayerat (1989) somewhat artificially separated these Schams cover slices from a dismembered unit referred to as Knorren mélange. This mélange largely consists of Schams lithologies and is in direct tectonic contact with the front of the Tambo basement nappe. In Val Vignone (Motta da Caslasc) this mélange contains a gneissic breccia directly overlying a basement slice, consisting of Permo-Carboniferous breccias and paragneisses. The analogies with similar contacts at coordinates Bavugls (E-Schams, Rück 1990) are striking and indicate the presence of slices of subunit 1a in the Knorren mélange at the front of the Tambo nappe.
Recent field work (Schmid unpublished) revealed that the Andrana zone (topmost cover slice of the Misox zone, Strohbach 1965) contains typical Tumpriv-Serie, Unterer Sericitmarmor and Plattensandstein (along the San Bernardino motorway at 736 600 / 144 400). This suggests that the Schams nappes can be traced southwards and along the base of the Tambo nappe as far as near the village Mesocco. On the one hand this excludes the front of the Tambo nappe as a possible site of detachment for the Schams cover slices. On the other hand it makes the hypothesis for rooting the Schams nappes in the Splügen zone (Mayerat 1989), i.e. at the top of the Tambo nappe, virtually impossible.

14.2.3.2 Splügen zone and Suretta cover

Most previous workers considered the age of the autochthonous cover of the Suretta nappe (excluding the Avers Bündnerschiefer) and of the sediments of the Splügen zone to be of pre-Jurassic (Triassic) age. In the case of the Splügen zone, however, Baudin et al. (1993) suggest that part of this cover, represented by carbonate breccias (similar to the Wissberg breccia in the Schams nappes), white calcite marbles, yellow sericite marbles, black shales and calcareous schists is of Jurassic and Cretaceous age. These authors found that Permo-Triassic conglomerates ("Verrucano"), volcanoclastics ("Rofna"-type) and basal Triassic sandstones are widespread at the base of the Splügen zone, representing the parauthochthonous cover of the Tambo nappe (partly identical with the Bardan zone of Strohbach 1965). These basal slices, together with the Andossi zone in the hangingwall, containing the post-Triasssic lithologies mentioned above, form the Splügen zone as a single unit characterized by complex imbrication of the cover of the Tambo nappe.
The Triassic quartzite of the Suretta Mesozoic cover is transgressive on a polycyclic basement (Timun unit) or on monocyclic basement (Rofna porphyry and "Verrucano"-type sediments). While parts of the dolomite and calcite marble alternations topping the basal quartzite are undoubtedly of Triassic age (close facies analogies to the Middle Triassic of the Tschera-Kalkberg unit in the Schams nappes) there is strong evidence for the presence of younger sediments. At several locations in the Suretta nappe, breccias contain angular components of dolomite and calcite marbles and, locally, crystalline basement overlies Triassic carbonates. To the south (Val Madris) these breccias contain large boulders and can be interpreted as (submarine) rock fall breccias (Hitz 1989). At some localities these breccias are in direct contact with the pre-Triassic basement and contain meter-size fragments of basement rocks (Dalla Torre 1991) suggesting that the Triassic quartzites and carbonates had been eroded locally and/or that the breccia-basement contact is a synsedimentary normal fault associated with Jurassic rifting. A peculiar relationship between basement, carbonates and breccias can be studied on the E side of Val Ferrera. In the Piz Grisch area (Christen 1993) and E of Campsut (Pfiffner unpublished, Figure 14-7), large (100m size) blocks of pre-Tri-

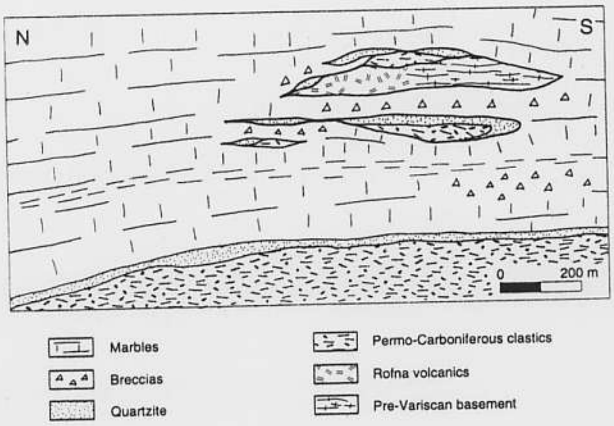


Figure 14-7
Evidence for Jurassic normal faulting in the Suretta nappe.
Lenses of pre-Triassic basement and Triassic quartzites are embedded in Jurassic breccias and Triassic carbonates (geologic sketch of the slope E of Campsut).

assic basement and quartzites embedded in a matrix of breccias occur as elongate lenses within Triassic carbonates. Alpine imbrication alone, even if considering the polyphase structural evolution, cannot be responsible for the observed structures. The large elongate basement blocks are more likely the result of synsedimentary faulting, which was subsequently obliterated by polyphase Alpine folding and thrusting.

The local unconformities leading to the direct transgression onto basal Triassic quartzite or pre-Triassic basement indicate substantial predepositional erosion caused by emersion following rifting. The large blocks of pre-Triassic basement and Triassic quartzites embedded in Jurassic breccias point to synsedimentary normal faulting and, associated, high relief. Both features are typical for the rifting and/or transcurrent faulting stage during the Jurassic and the Early Cretaceous as also found in the Schams nappes.

Hence, it appears that the cover of both Tambo and Suretta nappe exhibits strong similarities to the Tschera-Kalkberg unit of the Schams nappes, where the Tschera-Marmor-Serie (including the Wissberg breccia) is often transgressive onto the Middle Triassic. In many places the Tumpriv-Serie is missing due to post-Triassic erosion and hence lithologies typical for the Lower and Middle Jurassic are frequently absent. Tectonically emplaced cargneule horizons (presumably of Carnian age) and yellow dolomites are the only relics of the Tumpriv serie in the Tambo and Suretta cover. In view of the paleogeographic reconstruction of the Schams nappes (Figure 14-3) and the derivation of the Schams nappes below the Tambo basement nappe it is suggested that the cover of the Suretta and Tambo nappes represents the direct southern continuation of the Schams nappes, analogous to the Barrhorn Serie in the Western Swiss Alps which represents a paleogeographical equivalent of the Médianes Rigides, still autochthonous on the pre-Triassic basement (Sartori 1990).

The Avers Bündnerschiefer must be considered allochthonous with respect to the underlying Triassic carbonates and Jurassic rift breccias. Milnes & Schmutz (1978) considered the local occurrence of the Jurassic breccias, which they interpreted as having a tectonic origin, to be indicative of a thrust separation. However, recent mapping (Hitz 1989, Christen 1993 and field work by Pfiffner) showed that (Triassic) cargneules overly in places (Jurassic) breccias at this contact. In other places small scale imbrications of Bündnerschiefer and basement or Triassic carbonates point to a tectonic contact. The Avers Bündnerschiefer unit has been accreted to the Austroalpine orogenic lid and thrusted over the Suretta cover during an early stage of Alpine convergence. The presence of radiolarian cherts (Nievergelt pers. comm.), together with ophiolitic slivers within the Avers Bündnerschiefer (Oberhänsli 1977) suggests a S-Penninic origin of the Avers Bündnerschiefer. They should no longer be considered to represent the post-Triassic cover of the Suretta nappe.

14.2.3.3 Falknis, Sulzfluh and Tasna nappes

The Cretaceous sediments of the Falknis, Sulzfluh and Tasna nappes, in particular the Gault of "Mid"-Creaceous age, show close similarities to their equivalents in the Schams nappes. The Jurassic lithologies, however, differ substantially from those found in the Schams nappes (Gruner 1981 and references therein). Only the Upper Jurassic Sulzfluh-Kalk of the Sulzfluh nappe (Allemann 1957, Ott 1969) is virtually identical with the Tschera-Marmor-Serie of the Schams nappes. The Jurassic sediments are subject to strong facies variations also within and amongst the Falknis, Sulzfluh and Tasna nappes. The Falknis nappe, detached along black shales of Lower Jurassic (Toarcian?) age, lacks proximal breccia input during the Middle Jurassic (distal turbidites and background sedimentation, Gruner 1981). However, during a short time interval (Late Kimmeridgian? to Early Tithonian) basement-rich breccias (Falknis-Brekzie), very reminiscent of the Vizanbrekzien-Serie in terms of sedimentary transport mechanisms, indicate a short-lived important pulse, followed by the deposition of pelagic and detrital limestones at the Jurassic-Cretaceous boundary. Lithologies similar to those of the Falknis nappe are found in the tectonically dismembered Tasna nappe (Gruner 1981). There, however, the most important breccia horizon is of Cretaceous age (Gürler 1982).

We do not share the view of Gruner (1981) regarding the paleogeographical position of these three nappes at the Austroalpine margin, based on differing views about the structural evolution of the Schams nappes. Instead we regard these nappes as part of the Briançonnais domain, together with the Schams nappes. Retrodeformation of the Schams F2 event confirms the old view of Haug (1925), revived by Streiff (1962), and suggests a direct link between Falknis-Sulzfluh nappes and E-Schams nappes due to a S-closing fold in the Avers valley. The facies arguments proposed by Gruner (1981), based on the differing petrology of pre-Alpine basement components in the breccias, are not conclusive for deriving the paleogeographical position during the Jurassic and merely imply petrologically different basement sources for the Falknis and the Vizan breccia. The close neighbourhood of the Falknis and Sulzfluh nappes with their contrasting lithologies is very reminiscent of the rapid facies changes in the Schams nappes. The stacking order (Sulzfluh over Falknis nappe) is that of the W-Schams nappes (Tschera-Kalkberg over Gelbhorn unit), supporting the notion of a S-closing megafold between E-Schams (inverted stacking order) and Falknis-Sulzfluh nappes. The southern continuation of the Falknis-Sulzfluh nappes into the Avers valley is strongly disrupted and thinned and the hinge in the Avers valley is affected by subsequent normal faulting along the Turba mylonite zone (Nievergelt et al. 1996). The thickness of Jurassic-Cretaceous sediments reaches about 1000m in the Falknis nappe but is substantially less in the Sulzfluh nappe. Assuming an average thickness of the Mesozoic cover of about 800m for both Falknis and Sulzfluh nappes and measuring the area occupied by these two nappes in a N–S-section (31 km^2), i.e. subparallel to the movement direction during Tertiary orogeny, gives a width of 39 km for the depositional area of the Falknis-Sulzfluh sediments measured in a N–S-direction. Since the Falknis-Sulzfluh nappes structurally represent the N continuation of the Schams nappes (their depositional width in a N–S direction was earlier estimated to some 30km), the width of the Briançonnais platform (not including the Tambo- and Suretta sediments) in a N–S-direction may be estimated to be in the order of 70 km.

14.2.3.4 N-Penninic Bündnerschiefer and ophiolites

In the W-Schams and in front of the Tambo nappe the Areua-Bruschghorn mélange zone (Mayerat 1989, Rück 1990, Schmid et al. 1990) forms the basal thrust zone of the Schams cover slices, overriding the N-Penninic Bündnerschiefer. Around a large scale post-nappe fold (Niemet-Beverin fold) this mélange zone directly connects with the Martegnas mélange (Figure 14-2), which is part of an overturned stack of tectonic units in the E Schams (Rück 1990, Schmid et al. 1990). There, the Martegnas mélange separates the Schams nappes from the Tertiary Arblatsch flysch, originally part of the N-Penninic flysch zone. Only due to post-nappe refolding is the Arblatsch flysch at present in a structurally higher position in respect to the Schams nappes (Figure 14-2).

The continuous Areua-Bruschghorn-Martegnas mélange zone consists of ophiolitic remnants whose derivation from oceanic crust is, contrary to other mafic and ultramafic units within the N-Penninic Bündnerschiefer, beyond any doubt. Serpentinites, gabbros, pillow lavas, radiolarian cherts, pelagic limestones are widespread near Piz Martegnas (Streiff et al. 1971/1976, Eiermann 1988, Schmid et al. 1990) and imbricated with basement slivers (i.e. Areua basement) and sediments of the Schams nappes (i.e. Tumpriv-Serie, Vizanbrekzien-Serie, Tschera-Marmor-Serie). While the ophiolitic remnants are virtually undistinguishable from S-Penninic ophiolites (Platta-Arosa) the non-ophiolitic constituents are different in these two mélange zones (Schams lithologies in the Martegnas mélange versus Austroalpine lithologies in the Platta-Arosa mélange).

The bulk of the N-Penninic Bündnerschiefer and flysch sediments is structurally below the Areua-Bruschghorn-Martegnas mélange and the Schams nappes. These Bündnerschiefer are imbricated into four major units (Nabholz 1945, Steinmann et al. 1992, Steinmann 1994) which are, from top to bottom (Figure 14-2): the Tomül unit (equivalent to the backfolded Arblatsch flysch reaching into the Lower Eocene (Ziegler 1956, Eiermann 1988), the Grava unit (probably continuous with the Prättigau Bündnerschiefer and flysch also reaching into the Tertiary (Nänny 1948), the Aul marble unit (pre-

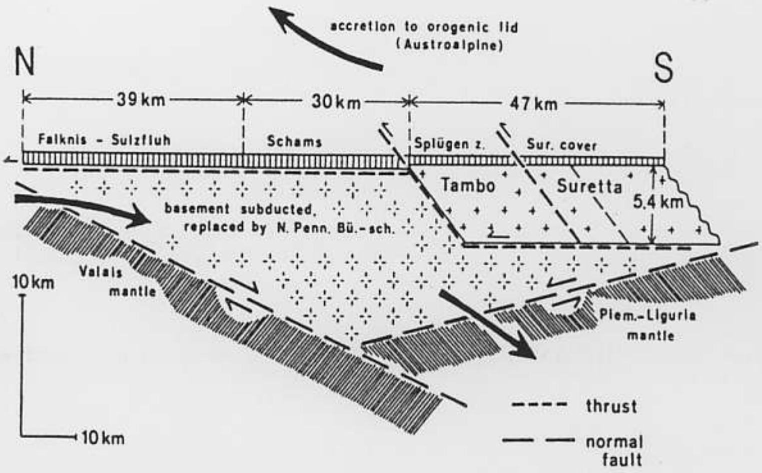


Figure 14-8
Very schematic restoration showing (1) the dimensions of crustal flakes derived from the Briançonnais domain now incorporated into the present day cross section and future detachment horizons related to Alpine collision (thrusts); (2) a minimum estimate of crustal thickness arrived at by assigning an upper plate margin position to the Briançonnais domain during two subsequent events of passive margin formation (compare Figures 14-9a and 14-9b); (3) subduction of excess crustal material being replaced by Valais Bündnerschiefer in an accretionary wedge scenario.

dominantly impure carbonates) and the Valser Schuppen (highly variable lithologies, imbricated with the Adula basement). Steinmann (1994) reports ophiolite-bearing mélange zones, also containing slivers of continental basement, Permo-Triassic cover and in particular Liassic Gryphea-bearing limestones (Nabholz 1945), at the base of the Grava and Tomül units. The Aul marble unit and the Valser Schuppen also contain ophiolitic slices. Typical Bündnerschiefer lithologies, however, are largely missing within these two lowermost units. Instead the Aul marble represents an impure marble presumably representing a platform area N of the Bündnerschiefer.

It appears that the Areua-Bruschghorn-Martegnas mélange and Schams cover slices represent the topmost elements of a large accretionary wedge mainly consisting of the four Bündnerschiefer units mentioned above. It will be argued later that this accretionary wedge formed during the Tertiary. It roots in the Misox zone and on top of the Adula high-P unit. Interestingly, the age of the Bündnerschiefer in Grava- and Tomül units is predominantly Cretaceous (with possibly some Late Jurassic sediments at the base). Moreover, parts of these N-Penninic Bündnerschiefer appear to have been deposited onto oceanic rather than on continental crust (see Steinmann 1994).

In summary, the N-Penninic ophiolites, and at least parts of the Bündnerschiefer are likely to represent a domain of oceanic crust, as proposed by Frisch (1979), Rück (1990), Schmid et al. (1990) and Stampfli (1993), situated to the N of the Schams paleogeographical domain. The Areua-Bruschghorn-Martegnas mélange would have formed at the N boundary of the Briançonnais platform. This platform, represented by the Schams and the Falknis-Sulzfluh cover nappes and, additionally, the Tambo-Suretta basement nappes with their own Mesozoic cover, separates the Valais (N-Penninic Bündnerschiefer) oceanic domain from the S-Penninic or Piemont-Liguria ocean.

14.2.4 Paleogeographical setting in a larger context

In order to discuss basement-cover relationships in the Penninic nappes and in order to sketch large scale paleogeographical maps it is important to have some estimate on the original width of the Briançonnais platform. The foregoing discussions showed that the original width of the Falknis-Sulzfluh and Schams cover nappes amounts to about 70 km measured in a N-S-direction. Admittedly, this estimate is very crude. It is based on area balancing applied to a N-S oriented cross section (see chapter 22, plate 1), assuming plane strain deformation within the profile plane. While the thrusting direction cannot have been far off from N (see discussion of structural data in section 14.3.1.2), out of plane deformation during orogen-parallel extension is a more serious problem. The assumed thickness of the Mesozoic cover is an additional source of error.

While these cover nappes are completely detached from a basement which is no longer preserved in the present-day cross section, the situation is different for the southernmost part of the Brianconnais platform comprising basement flakes (Tambo and Suretta basement nappes). The affiliation of these basement nappes to the southernmost Briançonnais platform was deduced earlier from the nature of their autochthonous to parautochthonous cover (Suretta cover and Splügen zone). Taking again the area of both Suretta cover (12 km^2) and Splügen zone (16 km^2) in the present-day N–S-section and assuming an original thickness of 600m for the Suretta and Tambo cover, another 47 km have to be added to the N-S width of the Briançonnais platform. Hence the total original width of this platform is in excess of 100 km (Figure 14-8). While this figure may be considered an overestimate due to an underestimate of the original thickness of the cover, those parts of the Suretta cover intruded by the Bergell batholith and possible extensions of the Suretta cover into the southern steep belt were not included in the area balance and would add extra length to our estimate. Evidence for orogen parallel late stage stretching discussed later also results in an underestimate of the area occupied in the profile plane and, consequently, also the total N–S-width of the Briançonnais platform.

It is obvious that the present-day cross sectional area occupied by Tambo nappe (111 km^2) and Suretta nappe (145 km^2) can only represent a fraction of the continental crust formerly underlying the cover of the Briançonnais platform (Figure 14-8). Together with the original width of the Tambo-Suretta part of the Briançonnais platform an original depth of about 5.5 km can be calculated for the basal décollement of the Tambo and Suretta crustal flakes. The depth to detachment within the Tambo basement has been independently estimated to be around 8 km by Mayerat (1989), who restored the present day thickness to the pre-Alpine geometry considering the intensity of Alpine overprint. With the same procedure the base to detachment within the Suretta nappe was found to be at about 4 km beneath the Triassic quartzites. All these estimates are within 4–8 km and compare to the ones obtained in the Southern Alps (Schönborn 1992) and the Aar massif (chapter 13), i.e. areas lacking substantial ductile overprint modifying the original depth to detachment.

The numbers discussed so far have been used for the very simple paleogeographic profile depicted in Figure 14-8. This crude sketch primarily serves for illustrating a dramatic problem concerning basement-cover relationships during Alpine orogeny. Even given extreme amounts of crustal thinning predating convergence, the conclusion that much of the basement originally underlying the Briançonnais cover must have been eliminated by subduction is unavoidable. The structurally lower Adula nappe cannot be regarded to represent a crustal flake derived from the Briançonnais. It is separated from higher tectonic units by the Misox zone, representing the oceanic suture between the Briançonnais and the European foreland. The Adula nappe therefore must represent the southernmost edge of this European foreland.

It is interesting to note that Menard et al. (1991) came to the conclusion that subduction of a significant amount of "European" continental crust (including the Briançonnais platform) is not required. While these authors based their estimate on a bulk mass budget which includes the entire profile across the French-Italian Alps, our conclusion (which is valid for the cross section in Eastern Switzerland) is essentially based on the recognition that most of the basement nappes of the Penninic structural domain (with the exception of the Tambo-Suretta pair) are part of the European distal margin N of the Valais suture. Hence all the area occupied by the Adula and structurally lower nappes down to the Moho cannot represent continental crust belonging to the Briançonnais. As pointed out earlier, the Schams and Falknis-Sulzfluh cover nappes lack their basement (except for thin basement slivers) in the present-day cross section while the Tambo and Suretta basement slices are covered by their own sediments.

The paleogeographic reconstruction proposed by Schmid et al. (1990, their Figure 11), sketching the en-echelon arrangement of two oceanic domains (Valais and Piemont-Liguria) simultaneously forming to the N and S of the Briançonnais platform during the Middle Jurassic was arrived at in order to explain the existence of a continental fragment (Briançonnais) caught between two oceanic spreading centers but at the same time firmly attached to the European foreland in front of the Western Alps. Classically, the Valais zone is considered to blindly end within the Western Alps. Lemoine et al. (1986) kinematically linked the Valais zone to the Piemont spreading center. This "classical" reconstruction cannot explain the ongoing paleotectonic activity until "Mid"-Cretaceous times, as recorded by the Vizanbrekzien-Serie in the Schams nappes. Additionally, it does not explain the short-lived Late Jurassic rifting pulse recorded in the Falknis nappe. If both continental margins, N and S of the Briançonnais, are interpreted to have become passive

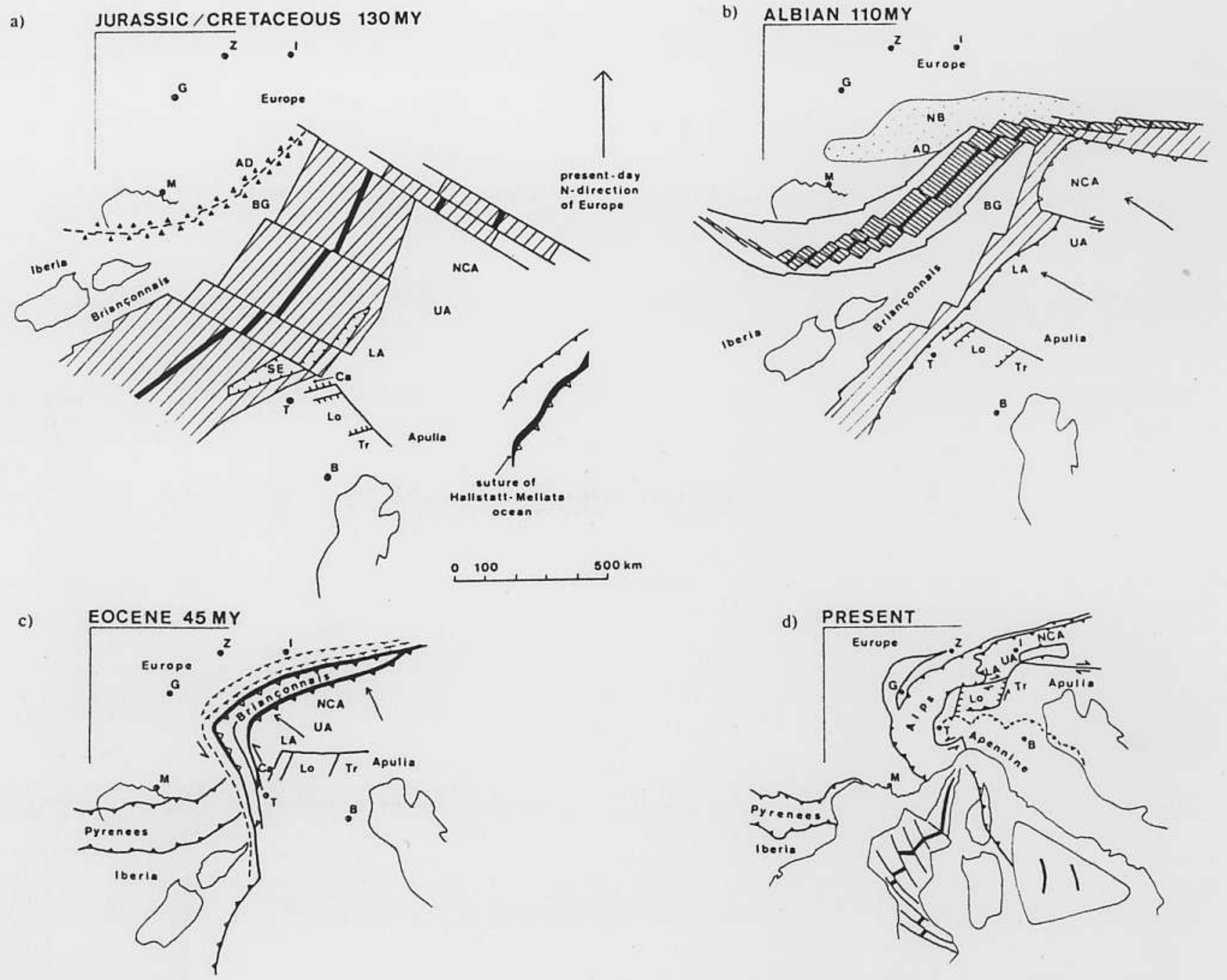


Figure 14-9

Palinspastic model of the Alps and surrounding areas. The model closely follows the restoration of Stampfli (1993) regarding the European and Iberia-Briançonnais plates and is identical with the restoration of Dercourt et al. (1986) regarding the movements of the Apulian plate relative to stable Europe (with a slight modification regarding the Albian stage). The Southern Alpine passive continental margin (Ca: Canavese; Lo: Lombardian basin; Tr: Trento platform) is assumed to be firmly attached to the Apulian block (after retrodeformation of Neogene S-directed thrusting in the Southern Alps).

Other geological units indicated are: AD: Adula nappe; NB: part of the N-Penninic Bündnerschiefer deposited on continental crust and Vocontian trough; Sch: Schams, Falknis, Sulzfluh and Tasna nappes; BR: Breccia nappe; SE: Sesia-Dentblanche and Margna-Sella extensional allochthon; LA: Lower Austroalpine nappes; UA: Upper Austroalpine nappes excluding the Northern Calcareous Alps; NCA: Northern Calcareous Alps.

Geographical reference: I: Innsbruck; Z: Zürich; G: Geneva; M: Marseille; T: Torino; B: Bologna.

a: Jurassic-Cretaceous boundary. Spreading in the Piemont-Liguria ocean linked to the Gibraltar transform system and rifting along the future break-up between Europe and the Iberia-Briançonnais block (stippled line, triangles refer to syn-rift breccia deposits). The Hallstatt-Meliata ocean already closed in the Late Jurassic (see discussion in Chapter 22).

b: Albian. Spreading in the Valais ocean linked to the Pyrenean fracture zone and active margin between the Piemont-Liguria ocean and Apulia, leading to subduction of the Sesia extensional allochthon.

c: Eocene. Closure of the Valais ocean, head-on collision in the Central and Eastern Alps, oblique collision associated with major sinistral strike-slip motion (Ricou and Siddans 1986) in the Western Alps separating the Briançonnais from Iberia (including Corsica-Sardinia) and Pyrenees.

d: Present day configuration, largely resulting from W-directed indentation of the „Insubric" plate during the Neogene (see Laubscher 1971, 1991).

during mid-Jurassic times, both margins ought to behave passively after the final break-up. Finally, the predominantly Cretaceous age of the N-Penninic Bündnerschiefer in the Grava and Tomül units (Steinmann 1994) argues for a later rifting and drifting event in the Valais trough.

An alternative palinspastic sketch is presented in Figure 14-9, largely based on a new model proposed by Stampfli (1993), a somewhat similar model having been proposed much earlier by Frisch (1979). Both these authors proposed a two-stage break-up. The Briançonnais first represented the passive continental margin of the Europe-Iberia block in respect to the Piemont-Liguria oceanic crust starting to form during Early to Middle Jurassic times (Figure 14-9a). Rifting along the future break up of the Valais ocean initiated later and during the Oxfordian (Figure 14-9a). Late Jurassic to "Mid"-Cretaceous drifting of the Briançonnais-Iberia block (Figure 14-9b) is responsible for the opening of the Valais ocean, partly accompanied by closure of the Piemont-Liguria ocean starting in the Cretaceous. According to this model, the Briançonnais platform, in particular its northern part represented by the Schams, Falknis, Sulzfluh and Tasna nappes suffers a second episode of rifting corresponding to large-scale sinistral transtension. A transition from rifting to drifting in the N-Penninic Valais domain during the Early Cretaceous has indeed been confirmed by recent field work in the N-Penninic Bündnerschiefer (Steinmann 1994) and at the northern margin of the Briançonnais (Tasna nappe, Florineth and Froitzheim 1994).

Because of dating problems it is hard to decide how much of the Vizanbrekzien-Serie could have formed during each rifting stage. The very long time interval covered by the Vizanbrekzien-Serie suggests that both rifting stages may be represented by these breccias. Tentatively local sinistral transpression indicated by the work of Rück (1990) in the Schams nappes could be attributed to the second rifting cycle whereas the extensional scenario in the Suretta cover discussed above might be related to the first rifting cycle. The main rifting episode recorded by the Falknis-Brekzie, however, is well dated (Latest Jurassic) and has to be related to the second rifting event associated with the opening of the Valais ocean.

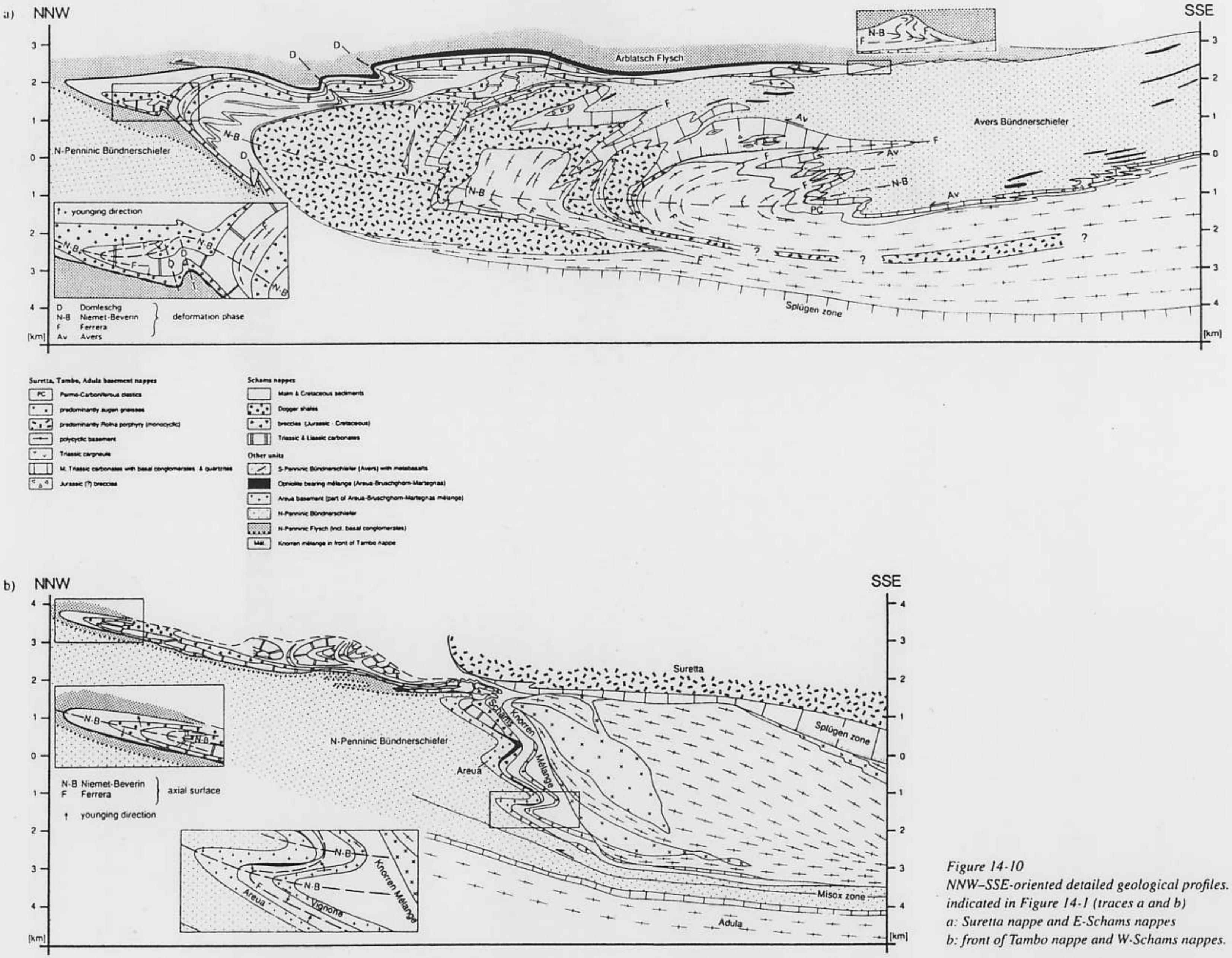


Figure 14-10
NNW–SSE-oriented detailed geological profiles. Profile traces are indicated in Figure 14-1 (traces a and b)
a: Suretta nappe and E-Schams nappes
b: front of Tambo nappe and W-Schams nappes.

According to the reconstructions of Stampfli (1993) the Valais ocean was directly connected to the Pyrenean fracture zone (Figure 14-9b). Hence the Valais ocean does not end blindly within the European foreland according to this model. Instead, it is kinematically linked to the Pyrenean fracture zone through an oceanic domain formerly present in the Western Alps. In present day map view the Valais zone in the sense of a tectonic unit terminates in the Upper Val d'Isère. The reasons for this are still unclear. Sinistral strike-slip motion parallel to the N–S trending part of the W Alps (Late Cretaceous to Eocene "décrochement Briançon-Ligurie" of Stampfli, 1993, "Subbriançonnais strike slip fault" of Ricou and Siddans,1986) might have obscured this important suture. According to this reconstruction strike-slip motion will produce a relative displacement of the Adriatic wedge over some 500km towards the N (Figure 14-9c) contemporaneous with the closure of the Valais ocean in the Central and Western Alps during the Eocene. During the Neogene this strike-slip fault must have been overprinted by E–W-shortening in the Western Alps, related to the W-ward indentation of the Adriatic wedge (Laubscher 1971, 1991). Oceanisation in the western Mediterranean during the Neogene obscured the former connections of the Valais ocean with the Pyrenees (Figure 14-9d).

Clearly, the paleotectonic evolution depicted in Figure 14-9 remains speculative and, admittedly, it also meets with some difficulties. For example, the radiolarian cherts and pelagic limestones of the N-Penninic Martegnas mélange are identical to those found in the S-Penninic ocean. Hence these sediments must have been deposited contemporaneously in both oceanic domains, in contrast to Figure 14-9a and in agreement with the "classical" view saying that both oceans formed during the Middle Jurassic. It is also possible that the Briançonnais already formed a ribbon continent during the Mid-Jurassic break-up, a ribbon continent which was further removed from the European margin during the later opening of the Valais trough (Figure 14-9b) during the Early Cretaceous (see Steinmann 1994 for an extensive discussion)

Returning to the area of investigation and the basement-cover problem depicted in Figure 14-8, the two-stage rifting model proposed by Stampfli (1993) offers a viable model for eliminating much of the continental crust underlying the Briançonnais before the onset of convergence. Uniform-sense simple shear rifting (Wernicke 1985) is in fact able to eliminate all the lower and much of the upper crust of the passive margin referred to as the upper plate margin (see Stampfli et al. 1991). Upper plate margins are characterized by large-scale updoming related to initial uplift caused by the isostatic response to replacing subcontinental mantle by less dense asthenospheric mantle, while lower plate margins are characterized by strong initial subsidence due to extreme thinning of the crust. For Jurassic times such an asymmetry is in fact indicated by the strongly contrasting subsidence history between the northern passive margin (Briançonnais) and the southern passive margin (Austroalpine and Southern Alps) situated on opposite sides of the Piemont-Liguria ocean (Lemoine et al. 1987). This strongly supports an upper-plate margin setting of the Briançonnais, as proposed by Lemoine et al. (1987) and Stampfli (1993). Following Stampfli (1993) the Briançonnais again formed an upper-plate margin during the second rifting event near the Jurassic-Cretaceous boundary which led to the opening of the Valais ocean. Hence the uppermost crust of the Briançonnais microcontinent may be envisaged as being extremely thin and directly juxtaposed onto mantle rocks along low angle detachments, as depicted in Figure 14-8. It must be noted, however, that Florineth and Froitzheim (1994) postulate a lower plate scenario for the northern margin of the Briançonnais during the second rifting event.

However, even given the most "optimistic" scenario depicted in Figure 14-8 (an upper plate margin was chosen N and S of the Briançonnais) leading to the omission of large volumes of continental crust during rifting and drifting, substantial amounts of upper crust must have been subducted during the Eocene. All the basement formerly underlying the Schams and Falknis-Sulzfluh nappes must have been subducted and replaced by the cover slices of the Valais Bündnerschiefer, currently underlying these Briançonnais cover nappes. Décollement of these cover nappes and juxtaposition with the Valais Bündnerschiefer may be readily explained in a context of accretion of the extremely thin Briançonnais cover slices, followed by in-sequence accretion of the four Bündnerschiefer units described earlier onto an upper plate formed by the N-wards advancing orogenic lid of the Austroalpine nappes during the Eocene. This accretion would have occurred in conjunction with S-ward directed subduction of the distal European lower plate margin of the Valais ocean leading to Tertiary high pressure metamorphism in the Adula nappe (Becker 1992 and 1993, Gebauer et al. 1992, Gebauer 1996), in spite of the fact that dating of the Adula high pressure metamorphism (Cretaceous versus Tertiary) is still controversial (for a different view see Hunziker et al. 1989).

14.3 Alpine convergence: from early imbrication to exhumation

14.3.1 Structural analysis and nappe geometry

This chapter will discuss the deformation phases, proceeding from older to younger. Each subchapter first describes the large-scale tectonic structures (see profiles in Figure 14-10a and b), followed by a description of small scale structural features. The various deformation phases have been defined using the classical geometrical rules of superposition of structures as visible on the individual outcrop and/or in profile view within an area of limited extent (Schams nappes, frontal parts of the Suretta and Tambo nappes). In order to avoid confusions resulting from the different numbering system used by previous authors (D1, D2 etc.) each defomation phase is characterized by a local name referring to a specific locality where a particular phase is well developed. Although these phases have been originally defined on strictly geometrical arguments an attempt will also be made to discuss their kinematic significance within the study area (chapter 14.3.3.).

In a second step these deformation phases will be correlated with phases found in the surrounding areas, including a discussion of the metamorphic evolution and an attempt to date these phases (chapter 14.3.2.). This will finally allow for a discussion of the kinematic evolution along the entire NRP 20 East profile (chapter 14.3.3.). This last chapter will form an important basis for the discussion of an integrated cross section along this eastern transect, including the geophysical data (see chapter 22).

In a very general way the earliest Alpine deformation involved the detachment of sedimentary units from their pre-Triassic basement (Schams nappes) or from probably oceanic lithosphere (Avers Bündnerschiefer) as well as the imbrication of thin basement slivers. Only in the case of the Avers Bündnerschiefer is it possible to unequivocally define a separate deformation phase (the Avers phase) related to this detachment event, which is probably associated with accretionary wedge formation. The resulting thrust faults (Avers phase and/or early Ferrera phase) were subsequently folded, and the pervasive associated deformation resulted in a prominent first foliation and stretching lineation so characteristic for the Ferrera phase. The next following Niemet-Beverin phase implies post-nappe folding and locally produced a second foliation. This phase was in turn followed by local later overprints (Domleschg and Forcola phases) related to exhumation by thrusting and normal faulting.

14.3.1.1 The Avers phase: precursor of the Ferrera phase?

Evidence regarding Alpine pre-Ferrera-phase deformation stems from the paleogeographically most internal unit, the Suretta nappe and its contact with the overlying Avers Bündnerschiefer. Clearly, the interface between Suretta cover and Avers Bündnerschiefer is affected by Ferrera phase isoclinal folding and cleavage formation. This led many authors to conclude that the Avers Bündnerschiefer simply represent the post-Triassic cover of the Suretta nappe. However, as discussed previously, such a view is untenable. The so-called "Triassic" cover of the Suretta turned out to also encompass post-Triassic members and, additionally, the lithological composition of some components of the Avers Bündnerschiefer (radiolarian cherts, ophiolitic slivers) preclude deposition on the Briançonnais platform. Direct structural evidence for thrusting along this contact is given by the local occurrence of cargneule or other sedimentary or basement lenses at the base of the Avers Bündnerschiefer. The presence of numerous mafic boudins within the Avers Bündnerschiefer – some of them at the basal contact – suggests pre-Ferrera phase intense imbrication or mélange formation within the Avers Bündnerschiefer unit. The only mesoscopic evidence for such a pre-Ferrera event within the Avers Bündnerschiefer is reported by Hitz (1989) who found pre-Ferrera phase folds in prasinites, embedded in Avers Bündnerschiefer.

Milnes and Schmutz (1987) suggested S-directed movement of the Avers Bündnerschiefer over the Suretta cover during the Avers phase. Their argument for a tectonic contact is based on the seemingly chaotic structure of the contact zone ("torn-apart rock masses of mappable size") which in many places are now believed to represent sedimentary breccias and megabreccias. Evidence for S-directed thrusting was based on a different scenario of these authors regarding the reconstruction of the nappe pile formed during Avers and Ferrera phases (in contrast to our results regarding post-nappe refolding, they reconstructed the Ferrera phase folds as having been originally formed in a S-facing orientation).

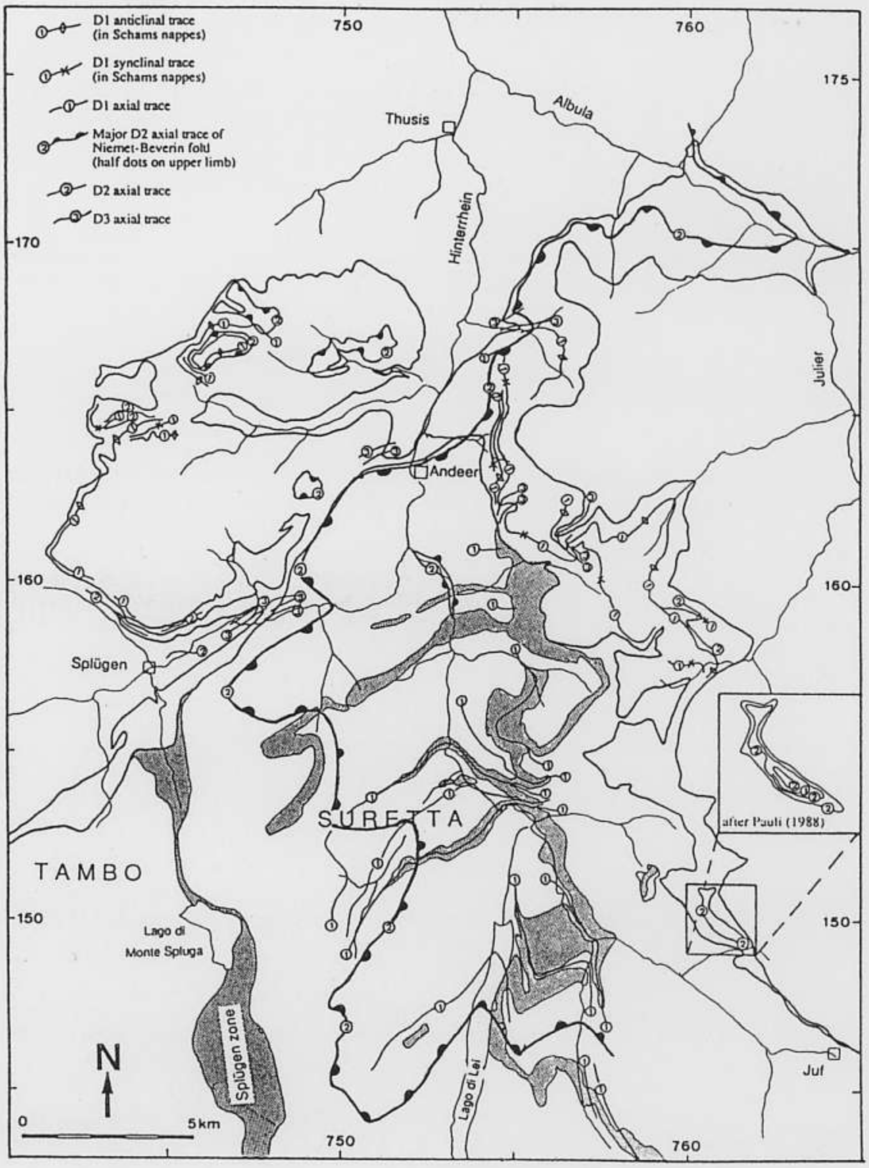


Figure 14-11
Axial plane of traces of the Ferrera (D1), Niemet-Beverin (D2) and Domleschg (D3) phase folds within the frontal Suretta nappe and the Schams nappes.

Pre-Ferrera phase foliations can be observed in pre-Triassic basement and Triassic cover rocks of the Suretta nappe and in the Avers Bündnerschiefer (Hitz 1989, Schreurs 1990, and own work). However, such older mesoscopic structures are extremely scarce and their tectonic significance remains unclear.

Ring (1992a, b) reports E–W lineations associated with high pressure metamorphism and top to the W shear senses of supposedly Cretaceous age in both Tambo and Suretta basement. If correct, these findings would indicate that the Briançonnais realm would have been deeply involved in the Cretaceous (Eoalpine) orogeny, a conclusion that would be completely incompatible with the palinspastic model previously derived for the rifting and drifting stage. These E–W lineations are conspicuously absent in the Adula basement nappe and in the case of the Suretta and Tambo nappes they turn out to be reoriented Ferrera phase lineations and/or lineations formed during the Niemet-Beverin post-nappe folding event (Baudin et al. 1993, Mayerat 1989, Schreurs 1990). Furthermore, shear sense indicators indicate top to the E movement where observable (Marquer 1991, Mayerat 1989 and own work). Regarding the age of high pressure or pressure dominated metamorphism radiometric evidence suggests considerably younger ages, a point that will be discussed in section 14.3.2.

In the Schams nappes, Carnian cargneule, representing the principal detachment horizon, occasionally occurs in the core of Ferrera phase folds. Strictly, this implies that Ferrera phase folding post-dates detachment. However, we interpret these structures as detachment folds, with detachment and folding in terms of a continuous process during the Ferrera phase deformation.

In the case of the Schams nappes located at the front of the Tambo nappe, however, isoclinal Ferrera folds also fold thrust faults delimiting the Areua and Vignone basement slivers (see Figure 14-10b, inset, and Mayerat 1989). Considering the large displacement that occurred along these detachment thrust faults, the large amplitude of the subsequent Ferrera phase fold, and the analogy to the basal thrust of the Avers Bündnerschiefer, it is more logical, in this case, to associate these thrust faults with the precursor, the Avers phase.

In summary, a distinct separate Avers phase can only be clearly defined in case of the detachment of the Avers Bündnerschiefer. There are considerable difficulties in assigning detachment events elsewhere to either the Avers phase or to an early stage of the Ferrera phase. Therefore we consider the Avers phase (where well defined) as a precursor of a continuous evolution from early detachment (Avers phase) to isoclinal folding and penetrative ductile deformation (Ferrera phase) as a result of steadily increasing burial and metamorphism during the Paleogene. According to our work there is no structural evidence for a separate Cretaceous (Eoalpine) tectonic event below the Platta and Austroalpine teconic units. In profile view the Avers Bündnerschiefer, unlike the Schams nappes and N-Penninic Bündnerschiefer and Flysch, are not wrapped around the front of the Suretta nappe. Instead they wedge out S of the Suretta front. Southwards they can be followed into the direct footwall of the Lizun and Forno ophiolitic units until they are cut by the Bergell intrusion (Liniger 1992). This suggests thrusting of the S-Penninic Avers Bündnerschiefer onto the most internal Briançonnais platform in a direction with a northerly component during the Avers phase.

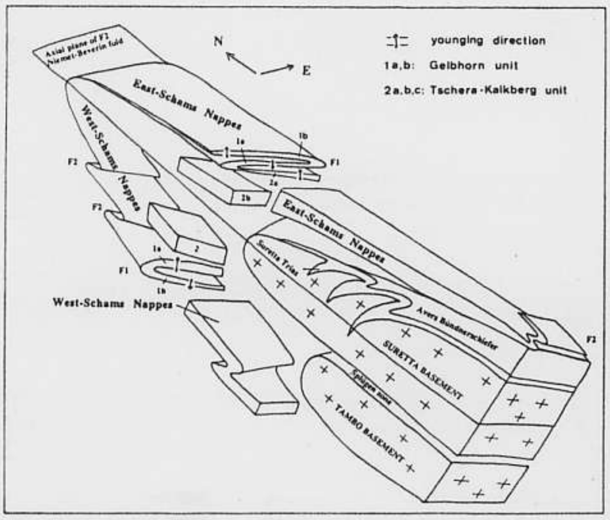


Figure 14-12
Schematic block diagram after Schreurs (1993) showing the relationships between the units and subunits of the Schams nappes on either side of the Niemet-Beverin axial trace.

14.3.1.2 The Ferrera phase: nappe imbrication and isoclinal folding

For a better understanding of Ferrera large scale structures one major result regarding the subsequent post-nappe folding phase (Niemet-Beverin phase) has to be anticipated: The Schams cover nappes and originally N-facing Ferrera phase folds in the Suretta cover have been refolded around the hinge of the Niemet-Beverin fold (see Figures 14-2, 14-10 and 14-11). The stacking order of the different Schams units (Figure 14-12) produced during the Ferrera phase is upright below the axial plane of this megafold (W-Schams) and inverted above the axial plane (E-Schams). The pattern of younging directions is more complicated but also is reversed across the axial plane of the Niemet-Beverin fold (see younging directions indicated for subunits 1a and 1b in Figure 14-12). This implies that the solution "infra" (as defined by Trümpy 1980), i.e. "rooting" of the Schams nappes below the Tambo-Suretta pair, is correct. An extensive discussion of numerous arguments in favour of such a reconstruction is found in Schmid et al. (1990) and Schreurs (1990, 1993) and will not be repeated here.

Discussions concerning mutual relationships between thrusting and nappe formation (discontinuous deformation on a large scale) on the one hand, and folding associated with pervasive straining, leading to a first generation of folds, planar and linear fabrics on the other hand, are very typical for the Penninic structural domain. Some authors envisage a foreland-type thrust-and-fold belt geometry being overwhelmed by subsequent penetrative deformation. This clearly would imply a two-stage scenario. Others envisage continuous progressive deformation with overlapping events of detachment and penetrative strain.

Ferrera phase deformation in the Schams cover nappes (Schreurs 1990) strongly argues for the second view. Ferrera phase thrusting of the Schams cover nappes over the N-Penninic Bündnerschiefer was accommodated by a mélange zone composed of gneissic material, Schams cover sediments and ophiolites: the Areua-Bruschghorn and Martegnas mélange zones (Figure 14-10). However, before juxtaposition with this mélange zone the Schams cover nappes must have been detached from their substratum first (see reconstruction of principal décollement horizons in Figure 14-3). In case of the Gelbhorn unit 1b this décollement horizon is made up of Upper Triassic cargneules which are frequently but not always found in the core of isoclinal Ferrera phase folds. In most places (an exception discussed later is found in front of the Tambo nappe) these folds do not affect the underlying mélange zone and could well be interpreted as detachment folds which developed coeval with thrusting over the mélange zone.

The tectonic contacts between the Schams subunits (in a paleogeographical sense) also argue for a continuous scenario: The limits between individual subunits may be defined by still visible Ferrera phase isoclinal fold hinges in one place (between units 1a/1b and 1a/2a, see Figure 14-12) or marked by thrusts in other places (occasionally between 1a/1b and 1a/2a, always at contacts of subunits within the Tschera-Kalkberg unit). Furthermore, Ferrera phase foliations (S1), axial planar to D1 isoclinal folds, are parallel to D1 thrust contacts which often but not always remain unfolded by D1 folds.

Similar observations are reported from the thin basement and cover slices in front of the Tambo nappe (Figure14-10b) by Mayerat (1989). There, a very thin (occasionally a few m) basement slice (Areua gneiss, covered by a thin veneer of Permo-Triassic clastics), which can be traced E-wards into the Areua-Bruschhorn mélange, can be continuously followed over a distance of more than 10km. This slice was thrusted onto N-Penninic Bündnerschiefer and is in turn covered by allochthonous Schams cover slices. Another mélange zone (Knorren zone, predominantly made up of Schams cover rocks) defines a tectonic contact in respect to the front of the Tambo nappe. Mayerat (1989) argued that this Knorren mélange extends a short distance into the Splügen zone and hypothesized that the Schams nappes could have derived from the Tambo nappe. As reported earlier, we disagree with this interpretation because recent field work enables us to trace Schams elements southward and into the Misox zone beyond the village of Mesocco. In front of the Tambo nappe a large scale isoclinal Ferrera fold locally affects the thrust contact between Areua gneiss and Schams slices and is probably responsible for the westward termination of the Schams nappes (Mayerat 1989). This is in contrast to most of the other Ferrera phase thrusts in the Schams nappes which, as mentioned earlier, remain unfolded during the Ferrera phase. As in the Schams nappes further to the N and E, S1 is parallel to these thrust contacts and the axial surface of the isoclinal fold.

Décollement horizons and, occasionally, the position of (Ferrera) fold hinges are controlled by the paleogeography (see Figure 14-3). A spectacular example of such a control is the position of fold hinges at the transition from breccias to basin sediments between subunits 1a and 1b (note that the younging direction is systematically inversed across the boundary between these subunits, in the W-Schams as well as in the E-Schams, Figure 14-12). Another example of the influence of the paleogeographic geometry concerns the detachment of basement slivers such as the Taspinit, Nolla and Areua slivers. It can be shown that the Taspinit basement sliver is directly covered by the Vizanbrekzien-Serie (Rück 1990), the potential décollement horizons in the Triassic having been eliminated by pre-depositional erosion. These basement slices probably represent "decapitated" horsts (Mayerat 1989, Schmid et al. 1990), extremely thinned out by subsequent straining.

In the Suretta nappe, Ferrera phase thrusting and folding occurred coevally (Pfiffner et al. 1990, Schreurs 1990). A major thrust fault occurs in the frontal part of the Suretta nappe (Figure 14-10a). This thrust emplaces an upper basement digitation (with polycyclic basement at its base) onto the Mesozoic cover of a lower basement digitation (largely made up of Rofna porphyry in its frontal part). Both imbricates are in an upright position below the axial plane of the Niemet-Beverin fold (Figure 14-10a). The pre-Niemet-Beverin phase geometry suggests that the thrust – although following a flat within Triassic evaporites – cuts up section towards the N, possibly indicating a northerly transport direction during the Ferrera phase. The central part of the Suretta nappe contains a number of tight, more or less symmetric folds affecting the Mesozoic cover of the Suretta nappe and the Avers Bündnerschiefer, cored by polycyclic basement (immediately SSE of the central part of Figure 14-10a). The contrast in style of this nappe-internal deformation may well reflect the mechanical stratigraphy involved: the mechanically stiffer monocyclic Rofna porphyry reacted by brittle thrust faulting, whereas the polycyclic basement containing more schistose lithologies deformed by ductile folding. It is important to note that the pervasive Ferrera foliation in basement and cover is parallel to both the thrust fault and the axial surfaces of the folds, and that the thrusts remain unaffected by Ferrera phase folding (of course except for the basal thrust of the Avers Bündnerschiefer). This suggests that the two structures (thrusts and folds) are more or less coeval. These large scale Ferrera phase folds, as well as subsequent Niemet-Beverin folds, involve the previously emplaced Avers Bündnerschiefer but not the E-Schams nappes. Milnes & Schmutz (1978) concluded from this, that the Schams nappes had been emplaced onto the Suretta nappe and Avers Bündnerschiefer at some later, post-Ferrera stage.

In the Tambo nappe the autochthonous cover is completely missing at the front of the nappe and at the base, whereas in the Suretta nappe relics of a thin autochthonous cover (basal quartzites) are locally preserved at the front of the nappe. For both, Tambo and Suretta nappe the contact with the Schams slices is tectonic everywhere. Hence, on a very large scale these basement nappes represent right way up thrust sheets which, on a smaller scale, are internally sliced by thrusts and penetratively deformed by folding and pervasive strain. The basal thrusts of both basement nappes, however, are often overprinted by post-Ferrera movements discussed later.

The Ferrera phase nappe stack consists of (from bottom to top): (1) N-Penninic Bündnerschiefer and flysch; (2) Areua-Bruschghorn-Martegnas mélange; (3) Schams units with subunits 1b, 1a, 2, occasionally linked by isoclinal folds; (4) mélange zones such as the Knorren zone; (5) Tambo nappe including Splügen zone; (6) Suretta nappe including its Mesozoic cover and previously emplaced Avers Bündnerschiefer. Note that these elements are of

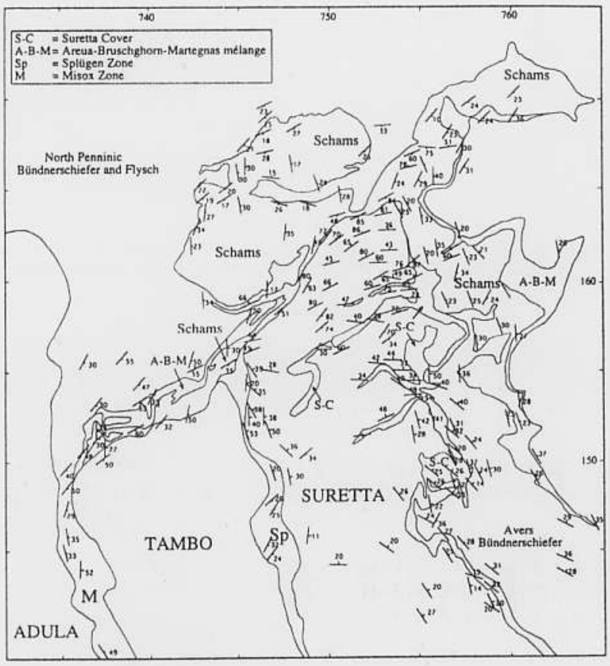


Figure 14-13
Strike and dip of Ferrera phase foliations.

extremely heterogeneous nature and include thick basement thrust sheets, detached and isoclinally folded cover sheets, extremely thin basement slices and mélanges containing ophiolitic elements. The pre-Triassic basement of the Schams cover nappe is missing apart from a few basement slivers.

Due to the regional axial plunge, Ferrera phase foliations mostly dip towards the E. Deflections from this direction are clearly visible in Figure 14-13 at the front of the Suretta and Schams nappes positioned in the hinge zone of the later Niemet-Beverin fold. Foliation, schistosity or cleavage is penetrative in all lithological units except in dolomites and in parts of Rofna porphyry and Truzzo granite, the latter deforming by networks of shear zones (Marquer 1991). A strain analysis regarding Ferrera phase straining was carried out in the Vizanbrekzien-Serie (Schreurs 1990) and the Tambo basement (Mayerat 1989). Axial ratios indicate plane strain deformation producing a thickness reduction of around 50%, assuming constant volume. This is a more or less representative strain measurement regarding basement and limestone lithologies but certainly an underestimate regarding less competent lithologies (basin sediments of Schams subunit 1b, post-Mid-Cretaceous sediments).

Surprisingly, the Ferrera phase stretching lineations exhibit a reasonably well developed preferred orientation trending NNW–SSE in the Schams nappes, despite of intense post-Ferrera folding (Figure 14-14a). Rare occurrences of reliable shear indicators and intense post-Ferrera phase folding prevent an unambiguous determination of the overall sense of movement related to this stretching lineation.

In contrast, the stretching lineations in the Schams units in front of the Tambo nappe and within the Tambo nappe (Figure 14-14a,b) are very variably oriented (Mayerat 1989, Baudin et al. 1993 and new work by Pfiffner and Schmid). Examples of variably oriented stretching lineations from the Misox zone and the top of the Suretta nappe are given in Figure 14-15 (c and d). The stretching lineations from these areas are seen to form a girdle along a great circle whose pole coincides with the average foliation poles (S1 or S2). This scatter is due to two effects which cannot always be separated: (1) reorientation of earlier formed Ferrera stretching lineations L1 during near-isoclinal Niemet-Beverin F2 folding due to non-parallelism between L1 and F2, and, (2) formation of a new stretching lineation L2 during the Niemet-Beverin phase. Mayerat (1989) finds both stretching lineations L1 and L2 straddling a complete great circle on a stereoplot (see Figure 14-15b). At the same time, a clear preference of E–W orientations is found. The great circle distribution argues for reorientation of L1. However, there is also clear evidence for finite E–W stretching from L2 stretching lineations found within axial planar S2 schistosities (Mayerat 1989, Baudin et al. 1993 and own work), supported by strain determinations (Mayerat 1989). Therefore it is impossible to separate L1 and L2 at many localities, especially when S2 completely transposes S1. Mayerat (1989) assigned a third deformation phase to be responsible for what we (and Baudin et al. 1993) refer to as F2 (Niemet-Beverin phase) E–W stretching, but this will be discussed later. At this point it is important to re-emphasize that E–W stretching very clearly postdates NNW–SSE stretching during the Ferrera phase, contrary to the findings of Ring (1992 a and b) who claims E–W-stretching to be followed by NNW–SSE stretching. At some locations it can be clearly seen that the Ferrera phase stretching lineation is folded around Niemet-Beverin folds reorienting the stretching lineations from a more N–S orientation on the lower limb to a more easterly dip on the upper limb.

NNW–SSE trending Ferrera phase lineations not affected by D2 with unambiguous sense of shear are found in the contact zone to the Splügen zone in the mylonitized top of the Tambo basement (at Splügen-Pass: Mayerat 1989, Schreurs 1990) and at the base of the Tambo basement and within the Misox zone (Schreurs 1990). A kinematic analysis of shear zone networks within

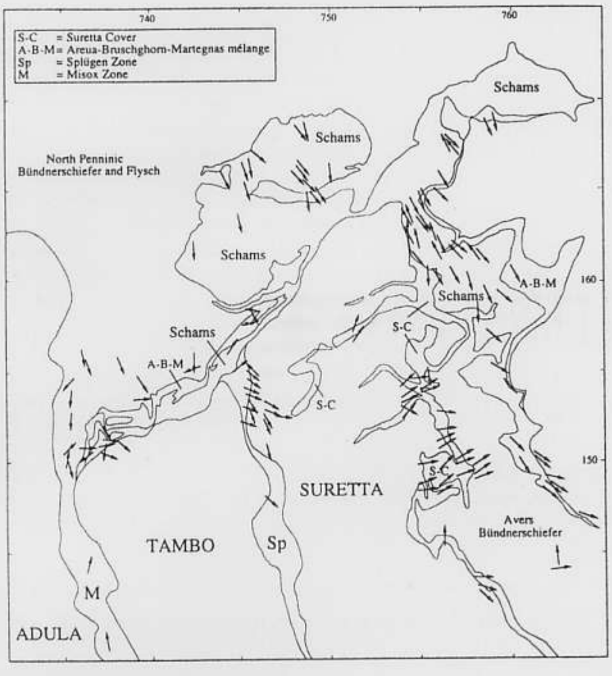


Figure 14-14a
Ferrera phase stretching lineations in the Schams nappes.

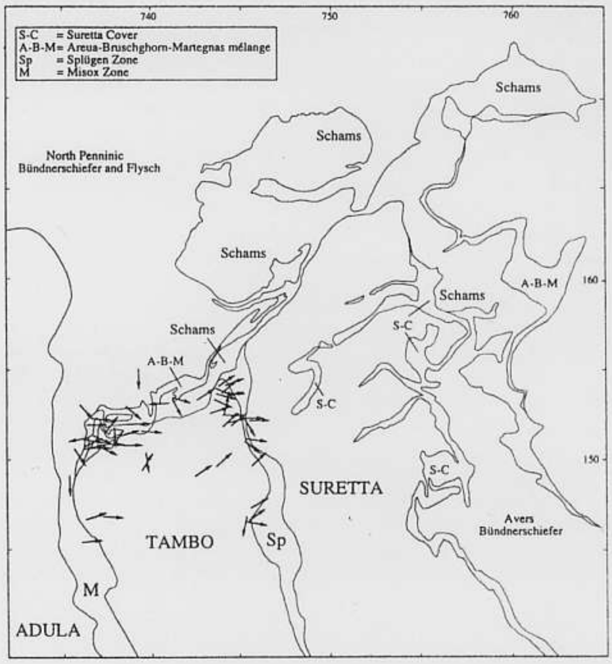


Figure 14-14b
Ferrera and/or Niemet-Beverin phase stretching lineations (where assignment to one or the other deformation phase is impossible) in Suretta and Tambo nappes and in the Schams units in front of the Tambo nappe.

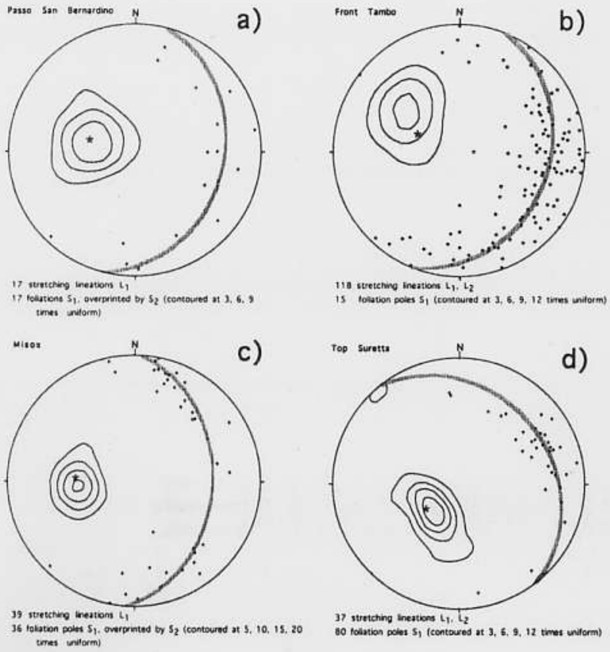


Figure 14-15
Stereograms showing foliation poles (contoured) and stretching lineations (points). Stretching lineations scatter on a great circle (best fit shaded) which is identical with the average orientation of the foliation (pole to best fit great circle for lineations (star) lies within maximum of foliation poles); a: Schams units near Passo San Bernardino; b: front of the Tambo nappe (from Mayerat Demarne 1994); c: Bündnerschiefer of the Misox zone; d: top of the Suretta nappe.

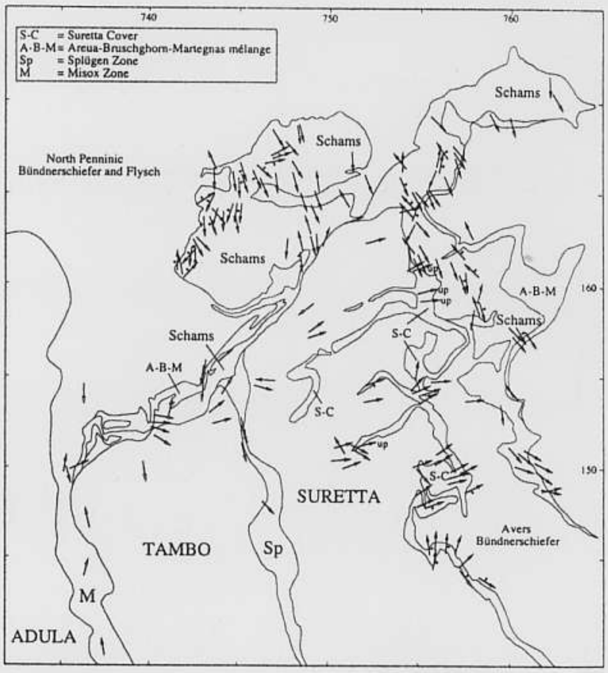


Figure 14-16
Orientation of subhorizontal Ferrera phase folds (arrows) and structural facing direction (indicated by short tick with dot; up = upward directed facing direction)

the Truzzo granite (Marquer 1991) documents top NNW sense of shear within the Tambo nappe.

Stretching lineations in the Suretta nappe, discernible as stretched deformed pebbles, are invariably parallel to Ferrera phase fold axes and intersection lineations (compare Figure 14-14a with Figure 14-16). Strictly, they only represent finite elongation during the Ferrera phase and they do not necessarily have a kinematic signification. In fact, strain measurements indicate a strong flattening component and microstructural analysis suggests near-coaxial deformation (Schreurs 1990). Also, the lineations do not turn into a preferred orientation near the basal thrusts of the nappes. For these reasons they have no kinematic significance in terms of a transport direction which can only be assumed to be subparrallel to the stretching lineation in the case of simple shearing.

Arguments for top to the NNW movement regarding the Suretta nappe and the Schams nappes entirely rely on analyses of fold facing. Regarding the Schams nappes the axes of perfectly isoclinal F1 folds lie fairly close to the orientation of L1 (Figures 14-14a,14-16), except for competent horizons. The analysis of structural facing (younging direction projected normal to the fold axis and lying within the axial plane, Figure 14-16) yielded the following results (Schreurs 1990, 1993): (1) Facing azimuths systematically depart in one direction away from the stretching lineation in the W-Schams, suggesting unidirectional rotation of fold axes towards the stretching lineations resulting in W to SW facing. This excludes true sheath fold formation and argues for rotation of fold axes due to systematic non-parallelism between early formed fold axes of buckle folds with the shearing plane, progressively rotating these fold axes during ongoing shearing deformation. (2) Facing directions in the E-Schams have locally been reoriented by Niemet-Beverin phase folding but NE-facing predominates. (3) The facing direction of non- or little rotated fold axes in the W-Schams clearly indicate top to the NNW transport.

F1 fold axes in the Tambo basement and Splügen zone are curviplanar and variable in orientation (Baudin et al. 1993). This is in contrast to the relatively constant ENE trending fold axis orientation found within the frontal part of the Suretta nappe (Milnes and Schmutz 1978, Schreurs 1990) for both large and small scale isoclinal folds affecting the Suretta cover (Figure 14-16). Since these folds have been coaxially folded around a large scale F2 fold (Niemet-Beverin fold) their present day facing (up or SSW, Figure 14-16) restores into NNW-facing after retrodeformation, again compatible with NNW-directed transport during D1. The stretching lineation, if in fact acquired during D1, suggests finite elongation perpendicular to the transport direction. However, a D2 stretch accidentally parallel to earlier formed F1 fold axes cannot be ruled out.

Orientations of fold axes in the Suretta cover dramatically change in the area of Val Madris going south (see southernmost data in Figure 14-16; Hitz 1989, Schreurs 1990). The large scale Ferrera phase fold axes prefer a NW–SE trend but small scale intersection lineations and fold axes straddle a complete great circle in S1 (Hitz 1989). Facing is variable but predominantly NE directed, resulting in apparent N-facing when projected into a N–S-section (Figure 14-10a). Stretching lineations, where observable in the form of pebble stretches, are mostly NW–SE to NNW–SSE oriented and close to large scale F1 folds. This change in orientation of D1 structures occurs across the trace of the Niemet-Beverin fold surface trace which leaves the Suretta basement across Lago di Lei, running into the Suretta cover and Avers Bündnerschiefer (southernmost part of Figure 14-11), contrary to the findings of Milnes and Schmutz (1978). Hence, this southerly area being situated in the lower limb of the Niemet-Beverin fold may have largely preserved the NNW–SSE-orientation of the Ferrera phase lineation which, in this case, would be parallel to the supposed transport direction (contrary to the stretching lineations near the front of the Suretta nappe which are perpendicular to the transport direction). Locally, however, Niemet-Beverin phase folds are also observed in Val Madris. Therefore a reorientation of Ferrera phase folds by later straining during the Niemet-Beverin phase cannot be excluded.

In summary, the Ferrera phase represents at the same time (i) the major event of nappe imbrication and (ii) the main phase of ductile penetrative deformation in the area. Penetrative deformation follows earlier detachments during the Avers phase and/or during the early stages of the Ferrera phase. The orientation pattern of stretching lineations is very complex. This is partly due to later deformation during the Niemet-Beverin phase, partly to the fact that not all stretching linations can be inferred to have formed subparallel to the transport direction. Arguments for approximately NNW directed movement during this phase can only be locally found (base of Tambo and Suretta nappes) or be inferred from an analysis of fold facing (Schams nappes). It cannot be excluded, however, that this NNW transport direction may have been reoriented to some extent by later deformation phases. In fact, neighbouring areas with less intense post-Ferrera phase overprint exhibit slightly different movement directions. Froitzheim et al. (1994) inferred a N to NNE directed transport direction for Tertiary thrusting in the Austroalpine nappes (their Blaisun phase), probably coeval with the Ferrera phase as defined for the Penninic units. Therefore the NNW transport direction inferred from our study area may be of rather limited precision and in fact may have been deflected from strictly N to NNW due to later strains.

14.3.1.3 The Niemet-Beverin phase: nappe refolding and vertical shortening

On a large scale the D2 Niemet-Beverin fold clearly overprints Ferrera phase thrust contacts and isoclinal folds (Figures 14-10,14-11). Large scale folding inverts the nappe pile presently found in the upper limb of the Niemet-Beverin fold axial trace consisting of (from top to bottom in its present position): (1) N-Penninic Bündnerschiefer and flysch (mainly Arblatsch flysch); (2) ophiolitic mélange zones (Martegnas); (3) part of the Schams nappes (E Schams); (4) the top of the frontal Suretta nappe with its spectacular backfolds. The axial trace of the Niemet megafold discovered by Milnes and Schmutz (1978) can be followed into the frontal Beverin fold (Figures 14-10a and b) in the Schams nappes (Schreurs 1990, 1993) and as far N as the Stätzerhorn (Figure 14-2, above Rothenbrunnen) into a S-facing isoclinal fold (Jäckli 1941). Inversion of Ferrera phase imbricate thrust sheets occurs over a distance as long as 40km (measured in a N–S section from Rothenbrunnen to the S end of the Schams nappes and the Arblatsch flysch near Alp Sovrana; see Figure 14-2). Large scale Z-shaped parasitic folds (viewed towards the E, Figure 14-10b) are observed to refold the imbricates at the front of the Tambo and Suretta nappes (W-Schams, Schreurs 1990, Mayerat 1989). Parasitic folds are also observed above the front of the Suretta nappe, where they affect the E-Schams nappes, the overlying Martegnas slice and the Arblatsch flysch (Schreurs 1990). These Niemet-Beverin phase folds are not visible in Figure 14-10a since their fold axes run parallel to the plane of the section. The Z-shaped folds in the front of the Tambo nappe may be treated as parasitic folds in respect to the Niemet-Beverin megafold. Most of the "backfolds" found within the northern part of the Suretta nappe (new work by Pfiffner, DallaTorre 1991 and Christen 1993), however, represent formerly N-facing Ferrera-phase megafolds reoriented by the Niemet-Beverin phase. The large backfold at Piz Grisch is a notable exception and has formed (or was at least amplified) during the Niemet-Beverin phase. On the other hand, the Z-shaped folds found above the trace of the Niemet-Beverin megafold (Figure 14-10a, above the frontal part of the Suretta nappe) formed during the Domleschg phase. On a mesoscopic scale the Niemet-Beverin cleavage, S2, can be continuously traced through the Niemet-Beverin fold axial trace from lower limb to upper limb. Interestingly, the basal "backthrust" of the E-Schams nappes over the Suretta cover and Avers Bündnerschiefer (Figure 14-10a, central part) is seemingly planar (where not affected by D3 folding discussed later) and suggests that post-nappe (re)folding is associated with some amount of "backthrusting" in the upper limb of the Niemet Beverin fold. We will discuss later that "backthrusting" and megafold formation probably both result from N to NW-directed differential movement of the Suretta and Tambo nappes in respect to the orogenic lid.

Pauli (1988) showed that the southern termination of the E Schams nappes (Figure 14-10a, inset near southern end of the profile) is caused by isoclinal Niemet-Beverin folding associated with a S-closing (in N–S profile view) megafold (Wissberg fold). This confirms the early views of Haug (1925) and Streiff (1962) and shows that the Middle Penninic Falknis-Sulzfluh nappes (northern part of Figures 14-1, 14-2) in principle represent the N continuation of the E-Schams nappes in the hangingwall of the Arblatsch flysch. A mega-scale Z-shaped fold affecting all the Briançonnais units of Graubünden appears in a large scale cross section (cross-hatched in Figure 14-2): The Falknis-Sulzfluh nappes form the upper right-way-up limb, the E-Schams nappes the overturned short middle limb, and the W-Schams nappes the lower right-way-up limb in respect to Niemet-Beverin phase mega folding.

How can such mega-scale folding lead to the overturning of a 40km long (measured in N–S-direction) "short" middle limb possibly have formed? The only feasible model we know of is that proposed by Merle and Guillier (1989), extensively discussed in Schmid et al. (1990). In brief: Post-collisional shortening in the root zone near the Insubric line during the Early Oligocene led to the vertical extrusion of the southern Penninic units in the Bergell area. Due to the presence of a rigid orogenic lid vertical flow within the ductile Penninic units was deflected into a near-horizontal direction and towards the N at shallower levels immediately below the lid. This led to a horizontal differential movement towards the N of the Tambo-Suretta pair both in respect to the overlying units (Austroalpine) and the underlying units (Adula nappe). It is this differential movement which is responsible for the apparent back-folds in the Suretta cover which simply represent originally N-facing Ferrera phase folds reoriented by the Niemet-Beverin phase strain. However, the second and structurally higher (in respect to the Niemet-Beverin-fold) megafold (Wissberg fold) is severely overprinted by a later, ESE directed normal fault, the Turba mylonite zone (Liniger 1992, Nievergelt et al. 1996) running along the base of the Platta nappe and cutting down section from N to S (from top Arblatsch flysch to top Avers Bündnerschiefer, Figure 14-2). Hence, omission by out of section normal faulting is responsible for the complete lack of Briançonnais tectonic units in the upper limb of the Wissberg fold along our N–S profile which runs approximately parallel to the

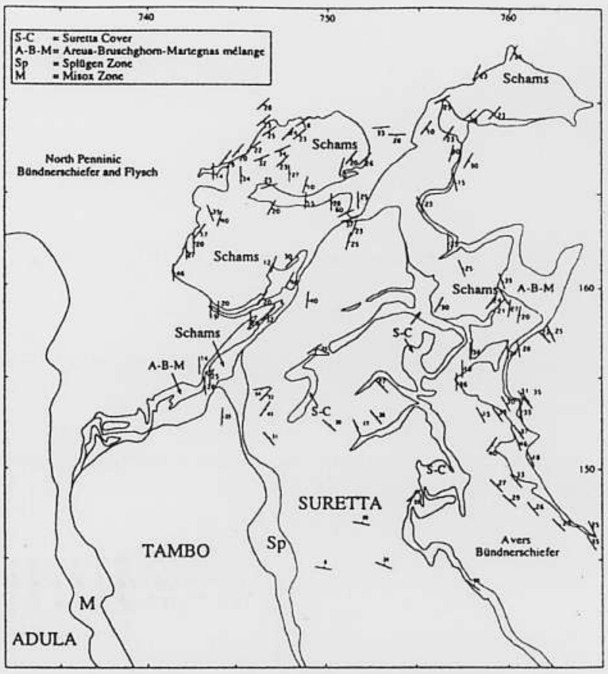


Figure 14-17
Strike and dip of the Niemet-Beverin phase foliation.

Turba normal fault (Figure 14-2). Between the Wissberg fold and the S termination of the Falknis nappe near Lenzerheide, only a very thin lens of Briançonnais units is found near Tiefencastel and in the hangingwall of the Arblatsch flysch (Streiff 1962).

In the surroundings of the frontal Tambo nappe, Mayerat (1989) attributed a local E-dipping lineation ("L3") with top to the E shear sense to an E–W extension. Marquer (1991) and Baudin et al. (1993) also described top to the E shearing in the Tambo nappe associated with a stretching lineation. These authors, however, correlated this E–W-extension with their second phase (the Niemet-Beverin phase). Thus, it appears that Niemet-Beverin folding is coeval with E–W extension. This E–W stretch, which we attribute to the Niemet-Beverin phase following Marquer (1991) and Baudin et al. (1993), leaves no trace in terms of large scale structures in a N–S profile except for the above mentioned omission at the Turba normal fault. Large scale parasitic folding in front of the Tambo nappe could be the combined effect of megafold formation largely related to N-directed differential movement of the Suretta-Tambo pair accompanied and/or immediately followed by E–W extension. Interestingly, no E–W extension was recorded in the Schams nappes further to the N where such Z-shaped parasitic folds also occur in the lower limb of the Niemet-Beverin axial trace.

Strains associated with the Niemet-Beverin phase vary in space. In the Mesozoic cover of the frontal Schams nappes (Schreurs 1990) the associated foliation is usually only well developed as a discrete or zonal crenulation cleavage in pelitic rocks. Limestones, even when isoclinally folded, only rarely exhibit a weak planar fabric. Penetrative fabrics in pelitic rocks are restricted to the Schams slices in front of the Tambo nappe and to the southern part of the E-Schams nappes, where strains are markedly higher. In spite of the generally weak axial planar fabrics, Niemet-Beverin folds are often near isoclinal on all scales in the cover.

In the basement strain intensity is in general weaker, but may be high locally. In the frontal Suretta nappe the early Ferrera foliation is gently wrapped around the very broad hinge of the Niemet-Beverin megafold, contrasting with the isoclinal megafold at Piz Beverin in the Schams nappes (along the same axial trace, Figure 14-10). On a small scale strain within the Rofna porphyry, as well as the Tambo nappe (according to Baudin et al. 1993) is pervasive but heterogeneous. The axial traces of the Niemet-Beverin large scale parasitic folds developed in the cover slices in front of the Tambo nappe cannot be traced into the Tambo basement (Mayerat 1989). This disharmonic structure again suggests a considerable contrast in mechanical behaviour between basement and cover during the Niemet-Beverin phase.

Axial planes of Niemet-Beverin folds and foliations are generally E-dipping (Figure 14-17), subparallel to the Ferrera foliation and they appear subhorizontal in a N–S cross section. Fold axes orientations are highly variable (Figure 14-18) although an ENE–WSW trend predominates in many places. An E to ENE strike characterizes the hinge line of megafolds such as defined by the front of the Suretta nappe and the basal D1 thrust of the Schams nappes. These hinge lines can be followed over a considerable distance along strike.

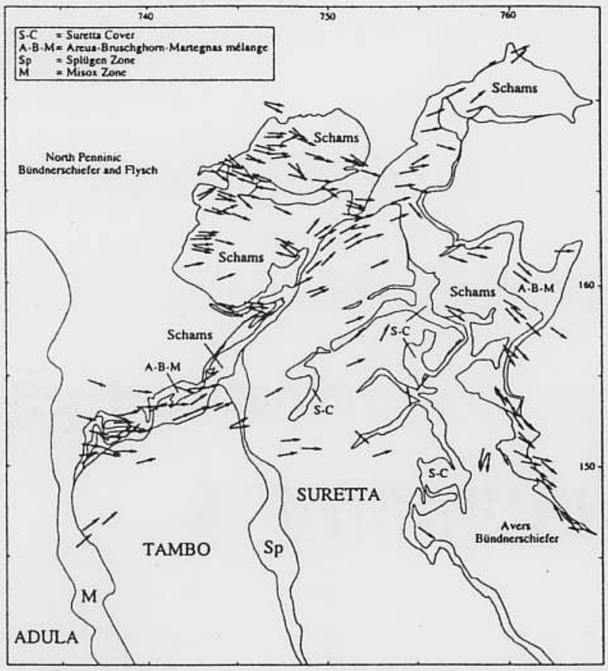


Figure 14-18
Niemet-Beverin fold axes (subhorizontal)

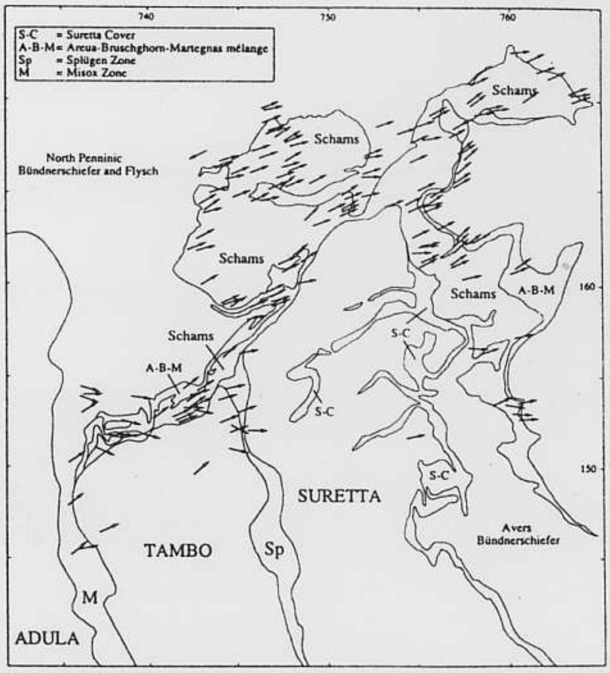


Figure 14-19
Domleschg phase fold axes (subhorizontal)

A systematic departure in orientation is observed in Figure 14-18 regarding the E Schams nappes where fold axes continuously swing into a NW–SE orientation towards the S. The rotation of these fold axes in the cover nappes away from the Niemet-Beverin hinge zone seems to indicate that the front of the Suretta nappe acted as a relatively rigid indenter producing a zone of high shearing strain on top of the Suretta nappe. This shearing strain, increasing southwards and associated with top to the SSE shear sense indicators, rotated the fold axes into a direction parallel to the local Niemet-Beverin transport direction. A SSE directed transport direction was also constructed by Schreurs (1990), based on deformed Ferrera phase lineation patterns around Niemet-Beverin folds, occupying a great circle on a stereogram.
In general, Niemet-Beverin phase stretching lineations and sense of shear indicators are only developed locally and are of variable orientation. Stretching lineations are absent in the frontal Schams nappes. In the E-Schams nappes shear bands indicating a top to the SE movement are widespread and also affect the overlying Arblatsch flysch, and the underlying Avers Bündnerschiefer (Schreurs 1990, Pauli 1988). Approaching the Turba mylonite in the hangingwall of the Arblatsch flysch the transport direction changes into the more easterly direction characteristic of the Turba mylonite normal fault formed during the closing stages of the Niemet-Beverin phase (Liniger 1992).

In summary, the Niemet-Beverin phase is associated with large scale post-nappe megafolding visible in a N–S cross section (Niemet-Beverin and Wissberg folds). Vertical shortening during the final stages of the Niemet-Beverin phase substantially thinned the pre-existing nappe pile. During the closing stages this vertical shortening was taken up by E–W extension, restricted to certain areas only (Turba mylonite zone, Schams nappes in front of the Tambo nappe and Tambo basement).

14.3.1.4 The Domleschg and Forcola phases: overprint related to final exhumation

The entire region is overprinted by a regionally consistent NW to NNW fold vergence with steeply inclined SE-dipping axial planes and NE to ENE plunging fold axes (Figure 14-19). This deformation also affects the N-Penninic Bündnerschiefer and flysch where it has been described as the Domleschg phase by Pfiffner (1977). This phase, although locally well developed as an intense crenulation, does not significantly alter the large scale geometry. Closely spaced large scale antiform-synform pairs produce a staircase geometry in the E-Schams nappes above the Niemet-Beverin axial trace (Figure 14-10a), affecting underlying units (backthrust of the Schams nappes over the Avers Bündnerschiefer) and, in particular, overlying units (up to the Austroalpine). Such large scale folds also concentrate near the front of the Suretta nappe, where they affect the axial trace of the Niemet-Beverin fold (Figure 14-10a). Staircase folds are again widespread in the Tambo basement (Baudin et al. 1993) where they increase in intensity towards the S and are, at least partially, responsible for the steeper northern dip of the Tambo nappe towards the S, observable in a N–S section (Figure 14-2).
The orientation of small scale fold axes (F3 of Schreurs 1990, 1993 and Schmid et al. 1990, F4 of Mayerat 1989) is amazingly constant, plunging NE to ENE (Figure 14-19), vergence being invariable towards the NW. Domleschg phase crenulations are easily distinguished from flat-lying Niemet-Beverin folds, because the former generally exhibit a more steeply inclined SE dipping axial plane. NE to ENE trend and plunge of small and large scale Domleschg folds is subparallel to most of the older large scale structures characteristic for the entire area situated E of the Ticino culmination of the Lepontine dome. This axial plunge is variable, as revealed by large scale contouring of Penninic basement nappes (Pfiffner et al. 1990 and Chapter 9, Steck et al., Chapter 12) and tectonic units in the hangingwall of these basement nappes, and locally varies between 10° and 35° to the ENE. On the other hand, the plunge of the Domleschg phase folds is only gentle to the ENE or near-zero (Pfiffner 1977, Schreurs 1990, Mayerat 1989, Baudin et al. 1993). This suggests that the large scale axial plunge between 10° and 35° is partly due to E-side-down offsets on later normal faults formed during the Forcola phase. SSE-dipping lineations, related to the Domleschg phase are only found in the southern Tambo nappe (Baudin et al 1993).
Normal faulting during the Forcola phase clearly overprints Domleschg phase folding and is only well developed in the S and in lower structural levels (Tambo nappe and base Suretta nappe, cf. Baudin et al. 1993). These authors describe NNW–SSE-striking steeply inclined E-dipping normal faults with a systematic downthrow towards the ENE, documented by stretching lineations. The largest of these faults, the Forcola normal fault, has not yet been investigated systematically and is situated at the base of the Tambo nappe. It is held responsible for the wedging out of the Misox zone towards the SE in map view (Figure 14-1) and can be followed as a retrograde greenschist facies mylonite into the Valle della Mera near Chiavenna (A. Berger, pers. comm.). Schreurs (1990) describes very low grade quartz mylonites with top to the E shear sense at the base of the Tambo nappe in Val Mesocco, overprinting higher grade top to the N or NNE mylonites related to earlier phases. However, the Forcola normal fault dies out northwards and cannot be traced further within the Bündnerschiefer. This retrograde and often brittle, steeply inclined normal faulting represents a later stage of E–W-extension in respect to the Turba low angle normal fault, as will be discussed later. In the upper and southern part of the Suretta nappe, E–W extension is documented by (brittle) shear bands, kink bands and joints and is associated with a top to the east shear sense (Hitz 1989).

In summary, the Domleschg phase moderately shortens the entire nappe pile in a SSE–NNW direction without producing significant décollement or large scale nappe folding. The systematic axial plunge to the ENE suggests that uplift of the Lepontine dome at its eastern margin postdates the Domleschg phase. Axial plunge is most likely related to final uplift and tilting of the nappe units, related to E–W extension manifested by the Forcola phase normal faults.

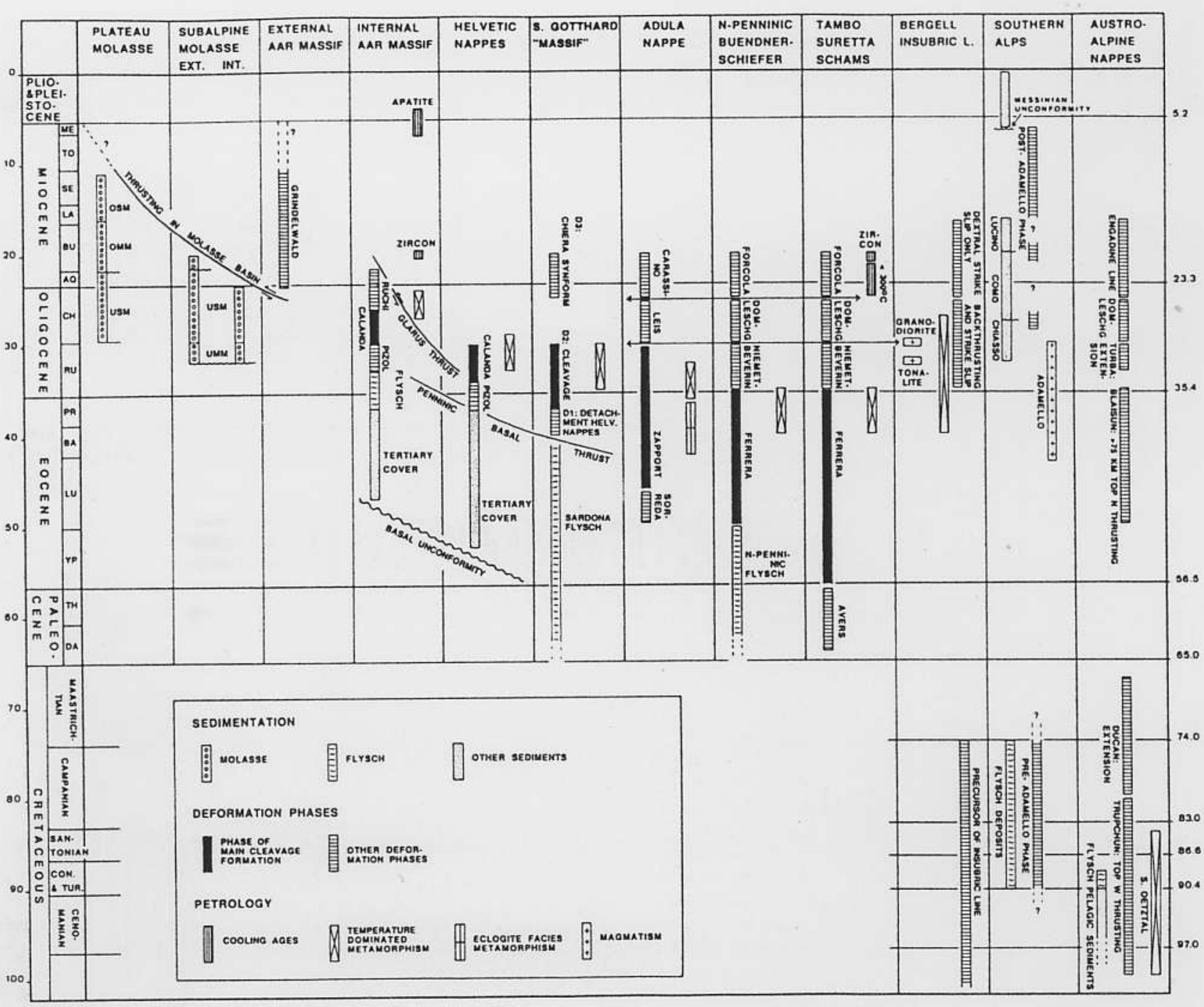


Figure 14-20

Correlation table attempting to date deformation phases and stages of the metamorphic evolution along the Eastern Traverse. Time scale according to Harland et al. (1989). See text for further explanations. Information concerning the Bergell area, the Insubric line, the Austroalpine nappes and the Southern Alps is added for convenience, but will be discussed in Chapter 22.

14.3.2 Regional correlation, metamorphism and dating of deformation phases

So far deformation phases were primarily based on structural superpositions analysed on all scales but within an area of limited extent. The tectonic and kinematic significance of such phases can only be further assessed after an attempt is made (1) to correlate these phases with those found in neighbouring areas, (2) to evaluate metamorphic conditions during deformation, and, (3) to date these phases with geochronological methods. As long as deformation phases are entirely defined by locally observed structural superpositions they simply represent a classical methodological tool. The correlation table given in Figure 14-20 summarizes the interpretation given in this chapter.

14.3.2.1 Regional correlation of deformation phases

In the units below the Tambo-Suretta pair, four deformation phases have been recognized by Löw (1987) for the frontal sector and by Meyre and Puschnig (1993) for the middle sector of the Adula nappe. The first phase (Sorreda phase of Löw 1987) predates eclogite facies overprint and is related to intense slicing of the quartzo-feldspatic basement with Mesozoic slivers. The Zapport phase, as defined by Löw (1987) represents the principle phase of penetrative cleavage formation in the Adula nappe. This longlasting phase sets in under eclogite facies conditions and spans a wide interval of pressures during near-isothermal decompression. Two stages of this Zapport phase have recently been distinguished by Meyre and Puschnig (1993) and Partzsch et al. (1994). An earlier deformation stage D2a, related to eclogite facies conditions, produced an early penetrative foliation and subsequent isoclinal folding of this foliation. Stage D2a structures are preserved in mafic boudins only. Outside these boudins the dominating penetrative Zapport phase foliation corresponds to stage D2b. This stage, prevalent outside the mafic boudins, is clearly post-eclogitic and associated with an alteration of the rims of the eclogite facies boudins under amphibolite facies conditions.

The post-eclogitic D2b stage foliation (formed during the closing stages of the Zapport phase in the sense of Löw, 1987) carries a N–S oriented stretching lineation associated with top to the N shearing according to numerous sense of shear criteria (Meyre and Puschnig 1993, Partzsch et al. 1994). Recent field work (M. Frey, J.Partzsch, S.Schmid, work in progress) shows that this main schistosity related to stage D2b can be followed without a structural break across the entire Misox zone (N-Penninic Bündnerschiefer). As described below, this Misox zone is characterized by a composite Ferrera and Niemet-Beverin phase foliation. From these field relations we conclude that the closing stages of the Zapport phase in the Adula nappe (stage D2b) are contemporaneous with the Niemet-Beverin phase of the higher tectonic units.

The subsequent Leis phase in the Adula nappe (Löw 1987), however, cannot be correlated with the Niemet-Beverin phase of our working area. This Leis phase corresponds to a large scale antiform in front of the Adula nappe, the Lunschania antiform (Löw 1987, Probst 1980) found in the N-Penninic Bündnerschiefer. According to Steinmann (1994) this antiform is equivalent to the Domleschg phase as defined by Pfiffner (1977). The correlation between Leis and Domleschg phase (Figure 14-20) confirms that the Niemet-Beverin phase has to be roughly contemporaneous with the closing stages of the Zapport phase in the Adula nappe. In the Bündnerschiefer of the Misox zone near the base of the Tambo nappe, Niemet-Beverin phase deformation is extremely intense and completely transposes the Ferrera phase first penetrative cleavage. There, Ferrera phase lineations, strongly deformed during the Niemet-Beverin phase, straddle an entire great circle due to "in-plane" folding (see Figure 14-15c). This indicates very strong syn-Niemet-Beverin strains across the Misox zone, which represents a telescoped section across all the four major units of the N-Penninic Bündnerschiefer, with relics of the Schams nappes in its hangingwall.

The Carassimo phase (Löw 1987) is related to the Alpettas synform north of the Adula front, overprinting the Lunschania antiform and, therefore, also post-dating the Domleschg phase. No equivalent of the Carassimo phase (and similar late synforms so characteristic for the formation of the Northern Steep Belt of the Penninic structural zone such as the Chiera synform, see Milnes 1974, 1976) seem to have overprinted our working area situated further to the E.

Moving further down the nappe pile, the N-Penninic Bündnerschiefer are observed to be thrust onto the largely allochthonous "cover" of the Gotthard "massif" (Etter 1987) along what we refer to as the Penninic Basal Thrust, to be correlated with the "Frontal Penninic Thrust" of the Western Alps (Bayer et al. 1987). According to Steinmann (1994), this thrusting post-dates the main phase of cleavage formation in the N-Penninic Bündnerschiefer attributed to the Ferrera phase. However, we do not regard his structural arguments to be conclusive in this respect. In analogy to the findings in the Suretta nappe and the Schams units we regard detatchment and penetrative cleavage formation (Ferrera phase in the N-Penninic Bündnerschiefer) as a continuous process, detachment being followed by ductile penetrative overprint and not vice-versa as postulated by Steinmann (1994). Therefore we prefer to attribute detachment along the Penninic Basal Thrust to the Ferrera phase. Steinmann (1994) observes Domleschg phase overprinting of the Penninic Basal Thrust. This overprint is also seen in the large scale profile (Lunschania antiform in Figure 14-2). Intense Domleschg phase folding (equivalent to the Leis phase) also affects the Penninic Basal Thrust.

The main phase of cleavage deformation in the Gotthard "massif" and its autochthonous cover remnants (Etter 1987, his D2) and the Helvetic nappes (Calanda phase of Pfiffner 1977 and 1986), however, post-date the main cleavage formation in the N-Penninic Bündnerschiefer, attributed to the Ferrera phase (Dreibündenstein phase of Pfiffner 1977) and the emplacement of the latter onto the Helvetic nappes (Pfiffner 1977 and 1986). A correlation of the Domleschg phase with movement along the Glarus thrust in the Helvetic realm has been suggested by Pfiffner (1977, 1986).

Correlations with deformation phases found in the units above the Tambo-Suretta pair have been provided by Liniger (1992). During his D1 phase (equivalent to our Ferrera phase) a nappe pile, formed during Cretaceous orogeny, was passively transported over viscously deforming underlying units (orogenic lid in the sense of Laubscher 1983, Merle & Guillier 1989, Schmid et al. 1990). Eoalpine deformation is considered to represent a separate orogenic event ("Cretaceous orogeny") because an important extensional phase (Froitzheim 1992, Froitzheim et al. 1994) separates Cretaceous WNW-directed convergence from the Tertiary (Meso- and Neoalpine) orogeny characterized by NNE to N directed convergence (the Blaisun phase of Froitzheim et al. 1994, see Figure 14-20). In regard to the dating and kinematics of Cretaceous deformation the reader is referred to data and discussions provided by Liniger (1992), Froitzheim (1992), Thöni (1986), Schmid & Haas (1989), Deutsch (1983) and Froitzheim et al. (1994). Cretaceous deformation will also be briefly discussed in chapter 22 of this volume. Tertiary age structures have been observed in the orogenic lid by Liniger (1992) Froitzheim (1992) and Froitzheim et al. (1994), but their importance is minor N of the Engadine line and they certainly are not related to the formation of the main schistosity in this orogenic lid, which is of Cretaceous age.

This upper nappe pile, already formed during the Cretaceous, represented an orogenic lid during the Tertiary. It consists of (from top to bottom) the Austroalpine nappes, the Arosa-Platta ophiolite units, the Margna nappe and the Lizun-Forno-Malenco ophiolitic units. This suturing of the remnants of the Piemont-Liguria oceanic domain (ophiolitic units) with the Apulian plate (Austroalpine) during Cretaceous orogeny predates all deformation phases observed in our working area. The Avers Bündnerschiefer represent the only derivatives of the Piemont-Liguria ocean which are in a structural position below the orogenic lid and which have consequently been penetratively deformed by Tertiary age deformation. Unfortunately, no structural correlation of the Avers phase with phases found in the orogenic lid can be made. If we propose that this Avers phase is related to an early phase of Tertiary convergence, this is based on their present-day structural position in the footwall of the orogenic lid, and additionally, on the large scale configuration in a N–S profile discussed earlier, suggesting roughly N-directed transport of the Avers Bündnerschiefer over the Suretta cover.

The basal thrust of the orogenic lid is severely overprinted by normal faulting during the closing stages of the Niemet Beverin phase in the Turba mylonite zone. This Turba mylonite zone forms an extremely valuable time marker: Liniger (1992) was able to follow this mylonite zone into the contact aureole of the Bergell intrusion where it is cut by the well dated Bergell granodiorite. A subsequent event of NNW–SSE compression (D3 of Liniger 1992) folds the Turba mylonite zone and underlying Avers Bündnerschiefer. This D3 of Liniger (1992) is associated with differential uplift of the Bergell intrusion and the southern part of the Suretta nappe. It can easily be correlated with D3 structures of Baudin et al. (1993) and, consequently, with our Domleschg phase.

14.3.2.2 Metamorphism

The discussion will be restricted to the Mesozoic cover and the monocyclic basement in order to avoid problems with pre-alpine metamorphism. The Adula nappe, where an Alpine age of high pressure metamorphism has been demonstrated by Löw (1987), forms an exception. Temperatures related to metamorphism will be discussed first. Problems arise with regard to pressures discussed thereafter, because different interpretations are possible. Classically, high pressure metamorphism is regarded as Cretaceous in age (Eoalpine), followed by Tertiary Barrowian type metamorphism (Meso- to Neoalpine, "Lepontine metamorphism"). In view of the previous discussion on Cretaceous versus Tertiary orogeny, this would imply a polyphase metamorphic history. In contrast to this, a continuous p-T-loop was first proposed by Löw (1987). A Tertiary age for such a p-T-loop has been proposed by Gebauer et al. (1992) and Gebauer (1996) while Schmid et al. (1990) cast doubts about a Cretaceous age in the Adula nappe purely based on the inferred paleogeographic position of this basement nappe. In the absence of undisputable radiometric ages we will assess these alternatives based on data concerning interactions between metamorphism and deformation.

In the Schams nappes the occurrence of white mica, stilpnomelane, chlorite and epidote indicates lower greenschist or blueschist facies conditions (discussion and references in Schreurs, 1990). These conditions prevailed during the Ferrera phase but persisted during the Niemet-Beverin phase (Schreurs 1990). Mayerat (1989) reports syn- and post-Ferrera phase growth of amphibole, followed by retrograde conditions, but still within greenschist (or blueschist) facies conditions during the Niemet-Beverin phase. This evidence agrees well with the style of deformation observed during both deformation phases and demonstrates that both basement and cover may undergo ductile deformation at relatively low temperatures. Final exhumation clearly postdates these two phases of deformation.

A large part of the Suretta nappe and the Avers Bündnerschiefer, but only the NE edge of the Tambo nappe, are situated N of the stilpnomelane-out isograd (Frey et al. 1983). New occurrences of stilpnomelane in the Schams nappes have been mapped by Schreurs (1990). This indicates temperatures below about 450°C (experimental data of Nitsch, 1970, valid for a pressure of 4 kb). Temperatures are better constrained in the Misox zone with regard to "Lepontine metamorphism" where Teutsch (1982) indicates 500–550°C around Mesocco and 600–660°C near the Forcola pass. In this context "Lepontine metamorphism" is not used in the sense of a metamorphic event. Instead, this term refers to a final equilibration under moderate pressures (around 6 kb in this case) leading to the famous pattern of isograds in the Lepontine dome. Teutsch (1982) reports growth of staurolite during and after his second deformational phase which corresponds to either the Ferrera or the Niemet-Beverin phase. In the light of the pressure estimates by Baudin and Marquer (1993) from the Tambo nappe discussed later, equilibration during the Niemet-Beverin phase is more likely. This agrees well with the classical picture of NW–SE-running isograds in the E part of the Lepontine dome, suggesting that final equilibrium related to Lepontine metamorphism overprints all major tectonic contacts formed during the Ferrera and Niemet-Beverin phases. The strike of these isograds indicates final exhumation related to differential uplift and erosion of the southern Penninic region during the Domleschg phase (N–S gradient), in combination with the axial plunge to the E, associated with E–W extension during the Forcola phase (E–W gradient).

Metamorphism related to the Lepontine isograds postdates D2 in the cover of the Gotthard massif (Etter 1987). Since we found this D2 to be roughly contemporaneous with our Niemet-Beverin phase, peak temperatures must have been reached at some later stage in this area.

Mafic eclogites of the Adula nappe (Heinrich 1983) indicate temperatures around 450–550°C in the extreme north, increasing to 550° to 650°C at Trescolmen in a middle sector E of Mesocco (Trescolmen). These temperatures agree with those obtained outside mafic boudins by Heinrich (1983) and Löw (1987). Hence eclogite facies related temperatures (and pressures) are not restricted to mafic lithologies and presumably prevailed all across the Adula nappe. Interestingly, these temperature estimates are above those reported for Lepontine equilibration in case of the northern Adula nappe. This implies that the Lepontine equilibration, as for example indicated by the staurolite "isograd" (Frey et al. 1980), established primarily within Mesozoic cover rocks, does certainly not apply to the northern Adula nappe situated a long way outside this "isograd", indicating temperatures of around 500°C. In fact, Klein (1976) reports staurolite north of this "isograd" and within the northern Adula nappe, while Koch (1982) inferred temperatures in excess of 500° C for the middle part of the Adula nappe. This is important since it indicates that the rocks in the northern Adula nappe must have cooled considerably before they were equilibrated during the Lepontine stage. Löw (1987) reports that temperatures and pressures compatible with Lepontine zonation were reached only after his Zapport and Leis phases, while Meyre & Puschnig (1992) find that Lepontine p-T conditions where established at the end of the Zapport phase (during their stage D2b) but before the Leis phase.

This confirms that Lepontine isograds postdate the major nappe contacts, formed during the early stages of the Ferrera phase. However, some difficulties still arise with regard to the timing of the higher or similar temperatures reached during the eclogitic stage. Strictly speaking, only D2b of Meyre & Puschnig (1993) can directly be correlated with the Tertiary aged Ferrera and Niemet-Beverin phases, the eclogitic event D2a being only detectable within mafic boudins. However, D2a and D2b probably have to be viewed as two stages during a continuous tectonic evolution associated with near-isothermal decompression (Löw 1987) and top to the N senses of shear documented for the later stages (D2b). This evolution is viewed to be contemporaneous with the Ferrera and Niemet-Beverin phases. The two stages D2a and D2b do in fact correspond to the Zapport phase of Löw (1987) who found this Zapport phase to straddle a range of decreasing pressures, starting at peak pressures. If this eclogitic stage is of Tertiary age indeed, as suggested by the correlation of D2a and D2b with the Ferrera and Niemet-Beverin phases, respectively, a high pressure overprint in the tectonic units overlying the Adula nappe would be expected. The following discussion on pressures will show that this is the case indeed.

The peak pressures reported for the Adula nappe increase from 10–13 kb (northern Adula nappe) to 15–22 kb (Trescolmen) according to Heinrich (1983, 1986). Eclogites are also found at the base of the Misox zone, but no Alpine age eclogites are known from the upper portions of the Misox zone or higher tectonic units. This suggests a drop in pressure and/or temperature across the Misox zone towards the E.

In basement slices of the Misox zone (Gadriolzug) Teutsch (1982) reports that phengitic white mica related to the Ferrera phase foliation are preserved in the N. His Si-contents reach values of around 3.5 (based on 11 oxygens per formula unit), indicating pressures of around 10 kb. This is compatible with a more recent study by Baudin & Marquer (1993) on the monocyclic Tambo basement. These authors analysed white mica in different microstructural sites corresponding to their deformation phases. Pressures in the order of 10–13 kb during the Ferrera phase are followed by pressures of around 5-10 kb recorded during the Niemet-Beverin phase. In spite of considerable difficulties concerning the applicability of the geobarometer provided by Massone and Schreyer (1987), pressures in excess of those reported for the Lepontine stage nearby (6 kb near Mesocco according to Teutsch, 1982) must have been reached. The results of Baudin & Marquer (1993) also suggest, at least qualitatively, a substantial pressure drop during the Niemet-Beverin phase. Temperatures dropped only insignificantly. This is analogous to the near-isothermal decompression reported for the Adula nappe by Löw (1987) and Meyre & Puschnig (1993).

Data on a possible pressure dominated overprint of higher tectonic units are scarce and no evidence for such an event is reported yet from the Schams nappes. Ring (1992 a) reports pressures in excess of 10 kb, also based on the Si-content of white mica, from the Suretta nappe and the Avers Bündnerschiefer. Oberhänsli (1986), who studied blue amphiboles in the Avers Bündnerschiefer, concludes that glaucophane and crossite indicate "pressure-accentuated" greenschist facies overprint but gives no estimate of pressure. Goffé & Oberhänsli (1992) found pseudomorphs in the Avers Bündnerschiefer which resemble carpholite, which indicates minimum pressures of around 7 kb.

Regarding the N-Penninic Bündnerschiefer, Goffé & Oberhänsli (1992) found carpholite preserved in the Lugnez valley, near the Basal Penninic Thrust. They estimate pressures to be in excess of 7 kb. There, chloritoid formed as a break-down product of carpholite when temperatures rose to about 350°C. Since carpholite pseudomorphs are also found near Thusis, these authors suggest that high-p low-T relics in the Bündnerschiefer are only preserved at the periphery of the Lepontine isograds. In the light of the common deformation history this strongly suggests that the Schams nappes were also affected by such a high-p event before the temperature rose to somewhere above 350°C but below about 450°C, as indicated by the stability of stilpnomelane.

In conclusion it appears that the entire studied area has been affected by high-pressure metamorphism as originally suggested by Frey et al. (1983) based on the occurrence of 3T white mica polymorphs. The decrease in pressure across the Misox zone is substantial but cannot be quantitatively assessed yet. The relations between mineral growth and deformation phases carefully established by a number of workers (e.g. Baudin & Marquer 1993, Löw 1987, Meyre & Puschnig 1993) exclude an Eoalpine age for this high pressure overprint which prevailed during the initial stages of the Ferrera phase. Near-isothermal decompression is indicated during the late stages of the Zapport phase in the Adula nappe (D2 of Meyre & Puschnig, 1993) and during the Niemet-Beverin phase in the Tambo nappe. In the N-Penninic Bündnerschiefer and Schams cover slices decompression from more moderate pressures must have been associated with increasing temperatures during the final stages of the Ferrera phase. The conclusion regarding a single p-T evolution of Tertiary age will have to be confronted with radiometric data discussed in the following section.

14.3.2.3 Dating of deformation and metamorphism

Deformation phases which can be correlated between different areas do not need to be strictly synchronous. If an attempt is made to date a particular deformation phase at a certain localitiy in the following discussion, summarized in Figure 14-20, such dating is only strictly valid at this particular locality. The same, of course, applies to the dating of metamorphic "events", or better, certain stages of a metamorphic evolution. Only in the absence of clear indications for migrating deformation phases do we assume contemporaneity (for example in case of the Domleschg phase, see Figur14-20). In other cases (for example the Calanda phase, see Figure 14-20) the available data point to the heterochroneity of the same deformation phase (in a strictly structural sense) in different tectonic positions.

Solid stratigraphical constraints for the earliest onset of deformation in parts of the Valais domain are only available for the Arblatsch and Prättigau flysch (Early Eocene, about 56–50 Ma) and the Falknis nappe (Paleocene). Sedimentation in the remainder of the Briançonnais and Valais domains may theoretically have stopped at some earlier stage because no fossils younger than Late Cretaceous in age have been found. Hence these constraints do not exclude the onset of collision at some earlier stage regarding the southern Briançonnais realm (Avers phase) nor do they exclude tectonic and metamorphic events in the basement while sedimentation continues in the cover. However, for the Tomül unit of the N-Penninic Bündnerschiefer the Ferrera phase is constrained to be younger than Early Eocene.

Regarding the more external Helvetic units (Glarus flysch), sedimentation continued well into the Early Oligocene below the Glarus thrust (Herb 1988, Pfiffner 1986), constraining the local maximum age of the Pizol and Calanda phase in the Infrahelvetic complex to substantially postdate the Eocene-Oligocene boundary (about 35 Ma). In the Helvetic nappes, however, sedimentation stopped earlier (Herb 1988, Lihou 1995) and synchronously with the emplacement of exotic cover sheets of S-Helvetic origin (Pizol phase) near the Eocene-Oligocene boundary. These data confirm the classical view of orogenic foreland propagation from S to N. Both the Pizol and the Calanda phases are not contemporaneous in the Helvetic nappes and the units below the Glarus thrust, as depicted in Figure 14-20.

Radiometric dating of the intrusion of the Bergell granodiorite at 30.13 ± 0.17 Ma (von Blanckenburg 1992) provides an absolutely reliable and extremely useful time marker. As discussed earlier, this intrusion constrains the end of the Niemet-Beverin phase to predate 30 Ma.

Radiometric dating of metamorphism is less straightforward and is a misnomer in that metamorphism also has to be viewed as an evolution in time and space which, strictly, has no fixed age. Additionally, numerous problems are inherent to the method used, such as to know wether isotopic compositions record cooling, formation or mixed ages.

Evidence, often quoted in favour of Eoalpine or Cretaceous metamorphism below the orogenic lid, stems from the Suretta nappe (Hanson et al. 1969, Steinitz and Jäger 1981). A Rb-Sr whole rock isochron age of 118 Ma was obtained for the northern Suretta nappe. This age, however, is in contrast to Tertiary phengite mineral ages, and, according to Hurford et al. (1989), the possibility of an artifact due to partial rejuvenation of pre-Alpine ages has to be considered. The second piece of evidence for Eoalpine ages comes from eclogites of the Adula nappe where Hunziker et al. (1989) quote a time bracket of 76–180 Ma based on work by Muralt (1986). This contrasts with Tertiary U-Pb zircon ages reported from the Cima Lunga unit, a western continuation of the Adula nappe (Gebauer et al. 1992, Gebauer 1996). Recent Sm-Nd mineral data by Becker (1992, 1993) are controversial too, in that they indicate both Tertiary (Cima Lunga) and Cretaceous (Adula) ages within the same tectonic unit. All this shows that radiometric evidence for the age of high-pressure metamorphism in the Adula nappe is still contradictory. Based on stratigraphic and structural arguments we prefer a Tertiary age for high pressure metamorphism. In accordance with the Tertiary ages given by Becker (1993) and Gebauer (1996) we place the age of this high pressure event in the Adula nappe somewhere between 42 and 36 Ma (Figure 14-20).

Radiometric evidence for Tertiary metamorphism in the Suretta, Tambo and Schams nappes has been compiled by Schreurs (1990, 1993). For the Ferrera phase cleavage in the Schams nappes K-Ar ages between 45 and 30 Ma are obtained for the <2 mm fraction (analyses by S.Huon and J. Hunziker reported in Schreurs 1990, 1993). An age bracket of 41–38.4 Ma is indicated for phengites in the Suretta cover by K-Ar dating (Steinitz and Jäger 1981,

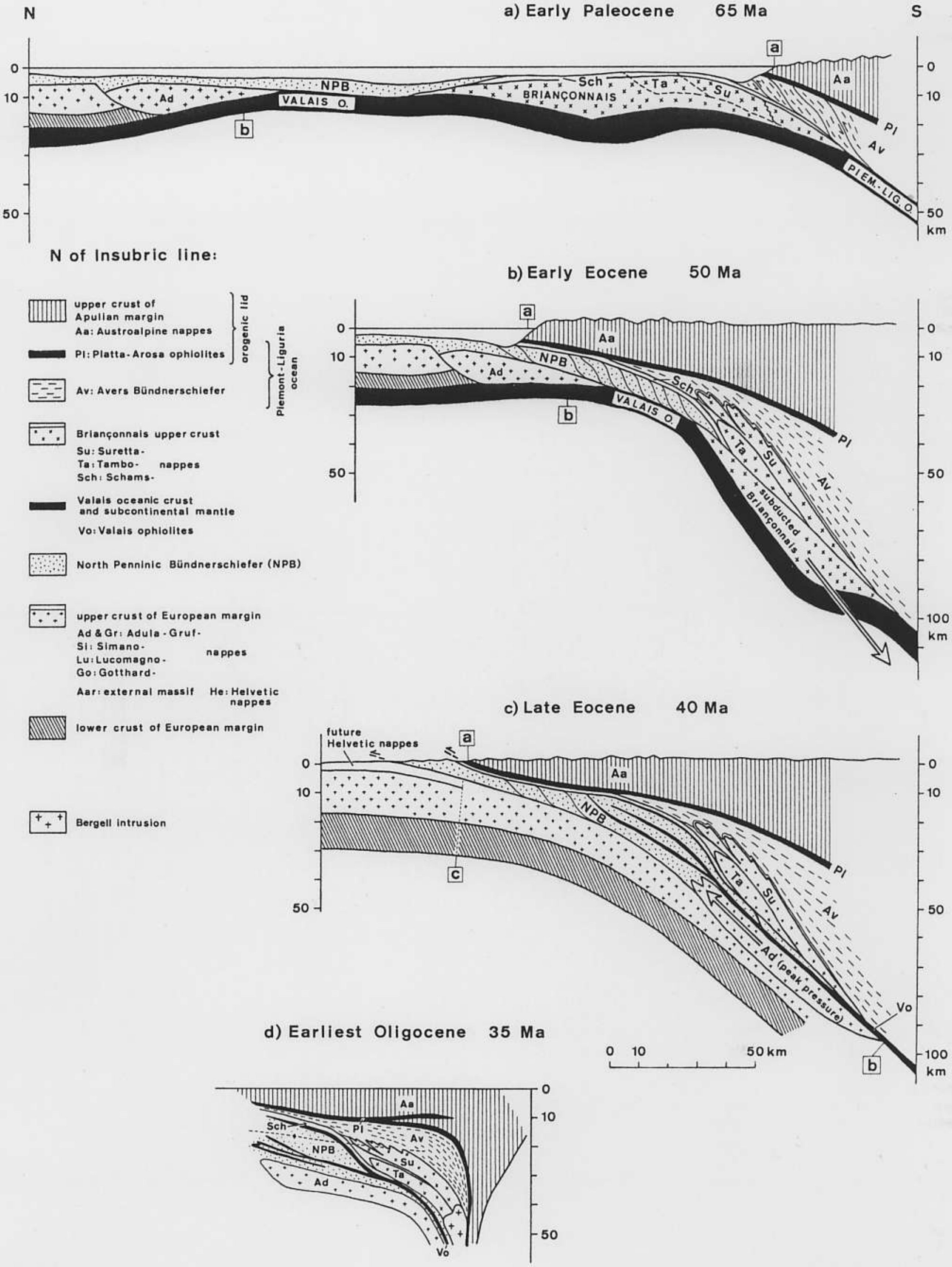
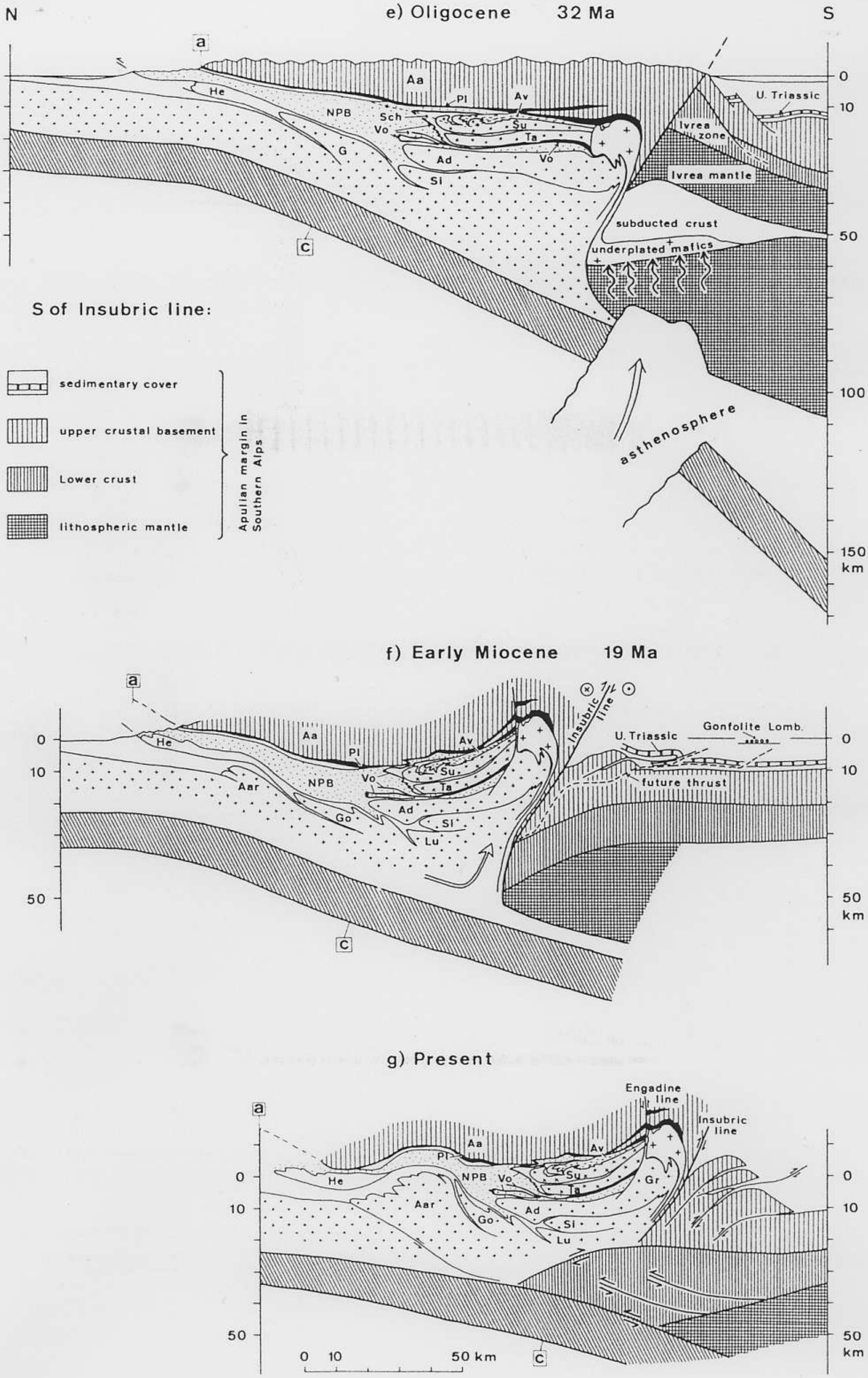


Figure 14-21
Scaled and area-balanced sketches of the kinematic evolution from early convergence and subduction (stages a and b) to collision (stage c) and postcollisional shortening (stages d to g).



N
e) Oligocene 32 Ma
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50
0
10
50
100
150
km
Aa
Pl
Av
He
NPB
Sch
Su
Vo
Ta
G
Ad
Vo
Si
C
U. Triassic
Ivrea zone
Ivrea mantle
subducted crust
underplated mafics
asthenosphere
S of Insubric line:
sedimentary cover
upper crustal basement
Lower crust
lithospheric mantle
Apulian margin in Southern Alps
f) Early Miocene 19 Ma
a
Insubric line
U. Triassic
Gonfolite Lomb.
future thrust
Aa
Pl
Av
He
Aar
NPB
Vo
Su
Ta
Ad
Go
Si
Lu
C
0
10
50
0
10
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km
g) Present
Engadine line
Insubric line
a
Aa
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Av
He
NPB
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Hurford et al. 1989). Similar phengite ages (36.6–37.6 Ma) are reported for the Rofna porphyry by Baltzer (1989) but phengite ages may be older in other places within the Suretta basement (up to 49.6 Ma, Hurford et al. 1989). Amphibole ages are still older (maximum age of 55 Ma, Hurford et al. 1989). All minerals dated are related to the main schistosity formed during the Ferrera phase and do not date the Niemet-Beverin phase, as erroneously reported by Steinitz and Jäger (1986). The dated amphiboles are aligned within the Ferrera stretching lineation, but also broken and stretched, possibly by late Niemet-Beverin E–W extension. Given the possibility of excess or inherited argon, Hurford et al. (1989) consider the age of 49–55 Ma as being maximum formation ages.

In view of the large spread of K-Ar mineral ages from the Suretta and Schams units (35–55 Ma, excluding Rb-Sr ages which are in part younger) an exact age for the Ferrera phase deformation and associated metamorphism cannot be deduced. Hurford et al. (1989) conclude that the age bracket of 35–40 Ma, encompassing concordant Rb-Sr and K-Ar ages, is the most realistic estimate for the age of metamorphism associated with the Ferrera phase (Figure 14-20). However, we provided evidence that the pressure peak probably predates the temperature peak in this area. Hence we consider the 35–40 Ma bracket to date the temperature peak postdating the onset of early pressure dominated stages of the Ferrera phase. Taking the spread of K-Ar phengite ages to reflect a spread in their age of formation, 50 Ma (Early Eocene) seems to be a good estimate for the beginning of the high-P metamorphic stage related to the Ferrera phase. Hence, the Ferrera phase deformation in the Tambo, Suretta and Schams area is inferred to have started before 50 Ma, somewhere around the Paleocene-Eocene boundary.

Hurford et al. (1989) use the 35–40 Ma bracket as the most likely age interval for evaluating the cooling history of the Suretta nappe, starting between 350° and 450°C. Fission track zircon ages for the Suretta nappe (20–21 Ma, Hurford et al. 1989), together with a temperature estimate of 400°C at 37.5 Ma, result in a slow average cooling rate (around 10°C/ Ma) during the Oligocene. This agrees well with the conclusion that temperatures did not decrease significantly during the Niemet-Beverin phase, ending in Mid-Oligocene times. Apatite fission track ages between 10–16 Ma indicate a more rapid cooling rate during Early Miocene times. Since accelerated cooling fits nicely into the context of the Forcola phase, related to final exhumation, this phase, and consequently also the Domleschg phase, are likely to predate this 10–16 Ma age interval (Figure 14-20). However, the Suretta cooling data do not have enough resolution to date the onset of rapid cooling related to final exhumation more precisely.

An attempt to date the Domleschg and Forcola phases more precisely has to use data from the more internal Penninic zone around the Bergell intrusion. The southern Tambo nappe cooled to below 300° C during Early Miocene times (biotite Rb-Sr and K-Ar ages between 21 and 25 Ma, KAW 105 and KAW 281 of Jäger et al. 1967, and Purdy and Jäger 1976). The greenschist facies mylonites related to the Forcola line found near Chiavenna (recent field work by A. Berger, Basel) can therefore not have formed substantially later than this time interval. Younger biotite cooling ages in the southern Misox valley (around 18–19 Ma, Purdy and Jäger, 1976, Wagner et al. 1977) show that the axial plunge of the eastern Lepontine area already existed at around 18–19 Ma. An age interval somewhere between 21 and 25 Ma (Oligocene-Miocene boundary) is therefore realistic for the Forcola phase (Figure 14-20).

The Domleschg phase, related to northward tilting of the southern Tambo nappe, is intimately related to uplift of the Bergell region. According to Giger (1991) and Giger and Hurford (1989), cooling and exhumation of the Bergell region immediately postdates the Bergell intrusion (30–32 Ma, von Blankenburg 1992). Work in progress (Rosenberg et al. 1994 & 1995) even suggests that backthrusting along the Insubric line already initiated during the intrusion of the Bergell tonalite (32 Ma, Blanckenburg 1992). It appears that the Domleschg phase initiated immediately after the final stages of the Niemet-Beverin phase. Liniger (1992) in fact argues that normal faulting along the Turba mylonite may even be contemporaneous with his D3 (correlated with the Domleschg phase). A scenario for contemporaneous E–W extension and N–S compression was recently proposed along the Simplon line by Mancktelow (1992) and, for Early Miocene times, for the region between Engadine line and Tonale line (Schmid & Froitzheim 1993). The Glarus thrust in the adjacent Helvetic zone, whose final stages can be correlated to the Domleschg phase, leads to a locally inverse metamorphic zonation and can be dated as Late Oligocene to Early Miocene (Groshong et al. 1984). In conclusion then, the Domleschg phase is inferred to have been active within the 30–25 Ma time interval (Figure 14-20).

Concerning deformation phases in the Helvetic units radiometric dating has to rely on illite-muscovite data discussed in Hunziker et al. (1986) and on stratigraphic evidence (Pfiffner 1986). Hunziker et al. (1986) conclude that Calanda phase deformation within the Helvetic nappes above the Glarus thrust occurred at around 35–30 Ma ago, a time interval also given in the reconstruction of Pfiffner (1986). Figure 14-20, however, depicts a somewhat younger time interval for Calanda phase deformation and associated metamorphism in the Helvetic nappes because, as discussed earlier, sedimentation did occur up to the Priabonian (i.e. in the Blattengrat unit of S-Helvetic origin, Lihou 1995). The Calanda phase within the Infrahelvetic units below the Glarus thrust is still younger (post 30 Ma, see Figure 14-20, "Internal Aar massif"). Final movements along the Glarus thrust related to the Ruchi phase (Milnes & Pfiffner 1977, Schmid 1975) offset the metamorphic zonation indicated by illite crystallinity data (Frey 1988 and Groshong et al. 1984). Pfiffner (1986) suggests that movement on the Glarus thrust causing the inverse metamorphic zonation stopped at around 20 Ma. From this time onwards further shortening within the Aar massif is related to thrusting in the Molasse basin (Grindelwald phase of Figure 14-20).

Geochronological data from areas further to the N along the NRP 20 have recently been provided by Michalski & Soom (1990). They are discussed in Chapter 13.1 of this volume in more detail. Zircon fission track ages around 20–21 Ma from the eastern Aar massif (Vättis window), the Ilanz area and the N-Penninic Bündnerschiefer near Rothenbrunnen do not significantly differ from those reported for the Suretta nappe. At first sight this implies a near-horizontal isotherm at 20 Ma ago in relation to the present-day position of these units. However, the assumption of horizontal isotherms during Miocene times is unjustified and N-dipping isotherms are inferred for the Helvetic nappes (Frey et al. 1973) and the Infrahelvetic complex, including the Aar massif (Groshong et al. 1984 and references therein). As discussed in Chapter 13.1, the Aar massif zircon fission track ages of around 20 Ma are related to the initiation of uplift of this external massif associated with the initial stages of shortening within the Aar massif (Grindelwald phase). Uplift, documented by apatite fission track ages, continued until the Late Miocene in the Central Aar massif and was associated with thrusting within the Subalpine Molasse Figure 14-20).

The most important conclusions on dating deformation phases and metamorphic stages are summarized in Figure 14-20, together with events within the Austroalpine and South-Alpine units which are discussed in Chapter 22 of this volume. The most important features regarding the Penninic and Helvetic units emerging from Figure 14-20 are the following:

1. Youngest sedimentation ages suggest that the onset of tectonic activity migrates towards the northern foreland over a time span of about 40 Ma.
2. The main phase of deformation related to nappe imbrication, associated with the development of a first penetrative cleavage also migrates towards the foreland. The same applies to the activity of the major thrusts: The Penninic Basal Thrust is followed by the Glarus Thrust and finally by thrusting within the Subalpine Molasse.
3. Thrusting along the Glarus thrust is preceded by early décollement of exotic strip sheets of S-Helvetic and Penninic origin during the Pizol phase. This Pizol phase is not associated with cleavage formation but is, in case of the Helvetic nappes, contemporaneous with the closing stages of Ferrera phase deformation and thrusting along the Penninic Basal Thrust. Similarly, the Ferrera phase deformation and associated thrusting follows earlier phases of detachment and imbrication within more internal units (Sorreda phase of the Adula nappe, Avers phase).
4. Metamorphic isograds drawn across the Penninic-Helvetic tectonic boundary (between cover of the Gotthard massiv and N-Penninic Bündnerschiefer) must be heterochronous. Peak temperatures were reached in the Suretta and Tambo nappes at around 40–35 Ma. This predates the peak temperatures inferred for the Gotthard massif and the Helvetic nappes at 35–30 Ma.
5. Lepontine equilibration at moderate pressures in the Adula, Tambo and Suretta nappes immediately follows and is intimately related to a high pressure event of Tertiary age. This is not the case with respect to the Gotthard region, where indications of an earlier pressure dominated stage are missing.
6. Cooling due to exhumation started relatively early (around 35 Ma) in the Eastern Lepontine units. This exhumation is locally associated with E–W extension (Niemet-Beverin and Forcola phases), but contemporaneous with ongoing N–S to NW–SE compression in the external zones, and differential uplift of the Bergell region related to backthrusting along the Insubric line.

14.3.3 Kinematic evolution

The following synthesis of the kinematic evolution also makes use of data provided in other chapters of this volume in order to arrive at a more synthetic view. This particularly concerns the geophysical infomation on the present day deep structure of the Alps contained in Figure 14-21g (compare Chapters 9 and 22) and geological data on the northern (see Chapter 13.1) as well as the southern (see Chapter 15) foreland of the Alps. On the other hand, this following discussion forms an important basis for Chapter 22, where some of the features discussed here will be further discussed with the help of an integrated present day cross section of the eastern transect, forming the base of the sketch presented in Figure 14-21g. Some overlap between the present discussion and Chapter 22 could not be avoided and we recommend the reader to also consult parts of Chapter 22 while reading this discussion of the kinematic evolution (in particular the integrated cross section, Plate 22-1).

Any attempt to sketch different stages of the kinematic evolution, such as the one proposed in Figure 14-21, is of course highly speculative. Nevertheless, construction of these sketches respects constraints, including all the uncertainties, which evolve from the previous discussions. In fact many constraints turned out to be more severe than anticipated during preparation of the sketches. On the other hand retrodeformation needs to be performed more rigorously in the future. Figure 14-21 only represents a first attempt. The most important guidelines and constraints need to be briefly outlined: The stages a,b and c in Figure 14-21 are constructed on the basis of the palinspastic restoration as given in Figures 14-8 and 14-9 and propose stages of convergence leading to final collision in the Late Eocene arrived at by forward modelling. Stages d, e, f and g in Figure 14-21, on the other hand, represent stages obtained by backwards modelling based on the present day profile along the NRP 20 Eeastern Traverse (Figure 14-21g) integrating geological and geophysical evidence further discussed in chapter 22. These last stages represent the result of successive retrodeformation of post-collisional deformation stages. All profiles are scaled and area-balanced in a semi-quantitative way. The area occupied by the Avers Bündnerschiefer has been enlarged for the stages represented in Figures 14-21a,b,c,d in order to take account for out of profile movements associated with E-W extension related to the Turba mylonite zone. The depth of several tectonic units (e.g. Aar massif, Suretta nappe, Bergell intrusion) respects data on the P-T-t evolution discussed previously, and, in the case of the Bergell intrusion, data found in Rosenberg et al. (1994 & 1995) Reusser (1987), von Blanckenburg (1992) and in Giger & Hurford (1989). The position of the Helvetic foreland in respect to the Insubric line in Figures 14-21e,f is based on the curvimetric retrodeformation of the present-day N–S-extension of the orogenic lid (Austroalpine units), which was only insignificantly strained during Tertiary deformation.

14.3.3.1 The Avers phase (Early Paleocene, Figure 14-21a)

The exact timing of the Avers phase is ill-constrained. The Early Paleocene was chosen for dating the sketch of Figure 14-21a because the Avers phase must substantially predate the final closure of the Valais ocean during Early to Middle Eocene times. Given the total width of the Briançonnais domain (about 115 km, Figure 14-8) and a minimum width of the oceanic part of the N-Penninic domain (only 50 km of oceanic lithosphere were chosen in Figure 14-21a), any time later than 65 Ma for the onset of subduction of the Briançonnais domain would lead to unrealistically high convergence rates when compared to large scale plate reconstructions (e.g. Dewey et al. 1989).

The parts of the Avers Bündnerschiefer still preserved in a present-day cross section are interpreted to represent the frontal shallow parts of an accretionary wedge which is going to be underplated by the Briançonnais units during the Avers phase. The upper plate is represented by the Austroalpine units and the previously sutured Platta and Lizun-Forno-Malenco ophiolites. The future decollement horizons within the Briançonnais platform, related to the initial stages of the Ferrera phase, are sketched according to Figure 14-8.

14.3.3.2 Subduction of the Briançonnais unit (Early Eocene, Figure 14-21b)

This sketch represents early stages of the Ferrera phase deformation affecting the Tambo, Suretta and Schams nappes, as well as parts of the N-Penninic Bündnerschiefer. The part of the N-Penninic realm which is characterized by oceanic lithosphere has already been closed. However, marine sedimentation continues in those parts of the N-Penninic Valais domain (Arblatsch and Prättigau flysch), which are underlain by continental lithosphere. Alternatively, the Eocene flysches may be assumed to have been deposited on parts of the N-Penninic realm characterized by oceanic lithosphere. In this case a wider N-Penninic oceanic lithosphere (more than the 50 km assumed) would not have closed yet at the stage depicted in Figure 14-21b. However, in this alternative case the convergence rate between the stages depicted in Figure 14-21 b and c would have to be very high (more than 1.5 cm per year).

Peak pressures in the Suretta-Tambo pair are assumed to have been reached by Early Eocene times in order to allow enough time for the subsequent subduction of the Adula unit, placed at the southern distal continental margin of Europe still during the early stages of the Ferrera phase, i.e. during the Early Eocene. Even so, the rate of convergence during Eocene times must have been relatively high and in the order of 1.5 cm per year in order to allow subduction of some 150 km of lithosphere necessary to arrive from the stage depicted in Figure 14-21b to that of Figure 14-21c (compare the relative positions of the 2 reference points marked "a" and "b" in the profiles).

As proposed in Figure 14-8, large parts of the continental crust of the Briançonnais will be subducted during this stage without subsequent exhumation. Southern parts of the N-Penninic Bündnerschiefer, the Schams nappes and the Tambo-Suretta pair have been accreted to the orogenic lid. Hence nappe formation during this and subsequent stages of the Ferrera phase is interpreted to be related to accretionary processes at great depth. This would explain the contemporaneity of decollement and isoclinal folding so typical for the Ferrera phase.

14.3.3.3 Subduction of the Adula nappe (Late Eocene, Figure 14-21c)

Compared to the Briançonnais units, deformation and high pressure metamorphism must be younger in age in the case of the Adula nappe if the palinspastic scenario depicted in Figure 14-21a and Figure 14-9, postulating a N-Penninic ocean between Briançonnais and Adula unit, is correct. Heterochroneity in regard to the onset of Ferrera phase and Zapport phase deformation and high pressure metamorphism in regard to Briançonnais and Adula units (Figure 14-20) is primarily a corrolary of the paleogeographic and tectonic reconstruction and is still ill-constrained by the geochronological constraints discussed earlier.

Figure 14-21c depicts the Adula nappe at its position corresponding to the peak pressures as determined for its northernmost part (11.5 kb) and at Alpi Arami in the southernmost part (27 kb in the Cima Lunga unit, corresponding to the southernmost Adula nappe) as given by Heinrich (1986). The total length of the Adula nappe would be about 85 km at the particular ill-constrained subduction angle chosen in Figure 14-21c. This figure is not unrealistically high when compared to the present day N-S extension of the Adula nappe: 45km in map view, about 60 km in profile view due to backfolding and steepening into the southern steep belt (see profile in Figure 14-2).

Unfortunately, there still is no adequate model available in order to explain the subsequent early phases of exhumation which bring the Adula unit back to more moderate pressures of typically around 6–8 kb, prevailing during equilibration at the temperature dominated so-called Lepontine stage reached at around 35 Ma (see Figure 14-20 and the next stage represented in Figure 14-21d). This exhumation is very rapid since it occurs over a time span of only 5 Ma. This pressure drop corresponds to uplift rates between 3 and 14 mm per year for a pressure drop from 11.5 and 27 kb, respectively.

A corner flow model such as proposed by Cowan & Silling (1978) and Cloos (1982) would be unlikely to have preserved the very substantial and systematic pressure gradient as presently observed from N to S within the eclogites of the Adula nappe. The widely accepted extensional model proposed by Platt (1986) is only applicable for parts of the later stages of exhumation, related to the Niemet-Beverin and Forcola phases, when E–W extension does occur. The rise of a delaminated rock pile as a buoyant sheet along the shear zone formed by the proces of subduction, representing a zone of greatest weakness, was recently proposed by von Blanckenburg & Davies (1995) for the case of the Central Alps. This model meets with difficulties because the Adula nappe is not surrounded by high-density mantle rocks according to the reconstruction given in Figure 14-21c. A mechanism of forced extrusion parallel to the subduction shear zone (arrow in Figure 14-21c) seems to be the most likely mechanism. Such a mechanism was recently proposed for the Western Alps (Michard et al. 1993). This mechanism is analogous to the subsequent differential movement of the Tambo-Suretta pair, as postulated during the subsequent Niemet-Beverin phase by Merle & Guillier (1989) and Schmid et al. (1990).

Unfortunately, such a model of forced extrusion is not supported by the field data. Kinematic indicators at the base and top of the Adula nappe suggest top to the N movement, while top to the S movement sense would be expected at the top of the Adula nappe according to this model. However, the top to the N movements reported to have been active during the late stages of the Zapport phase (D2b of Meyre & Puschnig 1993, Partzsch et al. 1994) in the structurally higher parts of the Adula nappe are likely to have overprinted earlier top to the S movements. These late stages of the Zapport phase are considered to be contemporaneous with the Niemet-Beverin phase (Figure 14-20), when top N thrusting of the Tambo-Suretta pair over the previously exhumed Adula nappe did occur (between the two stages represented in Figures 14-21d and 14-21e, as discussed below).

14.3.3.4 The final nappe stack (Earliest Oligocene, Figure 14-21d)

The scenario depicted in Figure 14-21d is arrived at by retrodeformation of the Niemet-Beverin phase as proposed by Schreurs (1990, 1993). The marked southerly dip of the Penninic nappes depicted in Figure 14-21c must have been preserved until the Eocene-Oligocene boundary in order to allow for backfolding of the units in the upper limb of the Niemet-Beverin axial plane. Top to the N to NW shearing of the Tambo-Suretta pair in respect to the underlying Adula unit (Schmid et al. 1990) during the subsequent Niemet-Beverin phase results in a more southerly position of the Tambo-Suretta front in respect to the front of the previously exhumed Adula nappe in Figure 14-21d when compared to Figure 14-21c.

14.3.3.5 Nappe refolding and vertical thinning (Oligocene, Figure 14-22e)

The kinematic model based on the work of Merle & Guilier (1989) regarding megafold formation associated with the inversion of a Ferrera phase nappe stack over a N–S distance of some 40 km in the hangingwall of the Niemet-Beverin axial trace has been extensively discussed earlier and in Schmid et al. 1990 and Schreurs (1993). This model implies relatively slow differential movement of those parts of the deforming Penninic units which were in close contact to the Austroalpine orogenic lid during overall NW to NNW transport of the Tambo-Suretta pair. Hence backfolding in the Suretta nappe is only apparent and not related to backthrusting of the orogenic lid in respect to some fixed point of reference in the lower crust. Differential NW to NNW-directed flow of the Tambo-Suretta pair in respect to the underlying Adula nappe occurred contemporaneously. The intense Niemet-Beverin phase overprint in the Misox zone and top to the N shearing during the closing stages of the Zapport phase in the Adula nappe are related to this differential movement of the Tambo-Suretta pair.
However, the relative velocity profile proposed by Schmid et al. (1990) has to be modified for the effects of simultaneous vertical shortening during the closing stages of the Niemet-Beverin phase. As pointed out correctly by Schreurs (1993) the plane strain heterogeneous simple shear model proposed by Schmid et al. (1990) cannot adequately explain shortening of the Schams units in front of the Tambo nappe. Simultaneous vertical shortening, however, may easily place the originally S-dipping Schams nappes into the shortening field of an oblate strain ellipsoid.
On the large scale depicted in Figure 14-21e, one of the major effects of the Niemet-Beverin phase concerns the relative uplift of the southern Penninic units which leads to the subhorizontal attitude of the Ferrera phase nappe stack in Figure 14-21e. In Figure 14-21e the Lombardic Southern Alps of the present-day profile (Figure 14-21g) are replaced by the Ivrea cross section after Zingg et al. (1990) in order to account for retrodeformation of some 50 km dextral strike slip motion along the Insubric line after Early Oligocene times (Schmid et al. 1989). The southern steep belt (and consequently a precursor of the present-day Insubric line) must have already existed during the intrusion of the Bergell pluton, as indicated by intrusive relationships found in the lower Val Masino (Rosenberg et al. 1994 & 1995). Hence this tilting of the Penninic units into a subhorizontal attitude is intimately related to early stages of movement along a precursor of the Insubric line and final emplacement of the Bergell intrusion.
The final stages of the Niemet-Beverin phase overlap with the final emplacement of the Bergell tonalite, but predate the intrusion of the Bergell granodiorite. The ascent of the Bergell pluton along a pre-existing Insubric line is facilitated by the overall relative uplift of the entire southern Penninic zone with respect to the Southern Alps along a pre-existing southern steep belt. Following von Blanckenburg & Davies (1995) we relate the ascent of the Bergell batholith to slab breakoff which results as a consequence of continental collision. It has to be emphasized, however, that final exhumation of the Bergell intrusion from the considerable depth of intrusion indicated in Figure 14-21e, associated with backthrusting along the present-day Insubric mylonite belt (Rosenberg et al. 1994), post-dates the Niemet-Beverin phase. Only the ascent of magma and the final emplacement of the Bergell pluton (except for parts of the granodiorite which intrudes the Turba mylonite zone at the NE corner of the Bergell intrusion) are associated with the Niemet-Beverin phase.
The modelling work of Merle & Guillier (1989) proposes that upwards directed viscous flow in the southern steep belt results in N-directed viscous horizontal flow of the Tambo-Suretta pair and the Schams nappes indicated by nappe refolding. In principle we still regard this model to adequately describe flow during the Niemet-Beverin phase. However, part of this horizontal flow may also be due to vertical shortening. Consequently, upwards directed flow in the southern steep belt may be less dramatic than suggested by these authors.
In the northern part of the section, palinspastic reconstructions (Pfiffner 1986, 1992) suggest that the Helvetic nappes must be considered as being detached from their crystalline substratum (the Lucomagno-Leventina and Gotthard nappes) at this time. The latter was internally shortened, leading to the detachment of the allochthonous Gotthard "massif". Some 200 km of European crust must have been subducted below the higher Penninic units and the orogenic lid by this time. This distance is measured between points "b" and "c" in Figure 14-21 b and c, a distance which needs to be subducted in order to have the Adula nappe at its depth required for eclogite facies metamorphism. Together with the previously subducted parts of the Briançonnais domain this length is compatible with an absolute minimal length of 200 km inferred to have been subducted until present into the lithosperic mantle beneath the Southern Alps by Pfiffner (1992) and Stampfli (1993).

14.3.3.6 Final exhumation of the Southern Penninic (Early Miocene, Figure 14-21f)

Comparing Figures 14-21e and 14-21f illustrates the effects of the Late Oligocene Domleschg phase because the effects of the Forcola phase (installment of an axial plunge and E-W extension) are not visible in a N-S section. Insubric backthrusting along the Insubric mylonite belt sets in immediately after the Niemet-Beverin phase and results in amazingly rapid exhumation and erosion of the deep-seated parts of the Bergell intrusion (Giger & Hurford 1989). In fact, exhumation of the southern Penninic region may be seen as a continuous process between the stages represented in Figure 14-21d and 14-21f. Tilting of the southern Penninic nappes continues and results in the presently observed northerly dip.
Oblique block rotation along the Engadine line (Schmid & Froitzheim 1993) results in differential uplift of the Bergell region also in respect to Penninic and Austroalpine units N of the Engadine line. The activity of the Engadine line certainly postdates the Late Oligocene Domleschg phase but is likely to be Early Miocene in age and contemporaneous with dextral strike slip movement along the Insubric line (see discussion in Schmid & Froitzheim 1993).
N–S shortening during the Domleschg phase is rather moderate across the Penninic and Austroalpine units. Major components of N–S shortening are now taken up by the Helvetic foreland where the Glarus thrust accommodates some 25 km displacement between the stages represented in Figures 14-21e and 14-21f (Pfiffner 1986). Thus, as pointed out by Pfiffner (1985), a major part of the deformation in the Helvetic foreland is contemporaneous with backthrusting along the Insubric line. In the footwall of the moving Helvetic nappes the upper crust is involved in shortening, as evident from the initiation of bulging in the Aar massif. This deformation eventually leads to a broad antiformal structure depicted by the Austroalpine orogenic lid. Effects of Late Oligocene to Early Miocene shortening in the Southern Alps, in contrast, are rather moderate or even absent according to Schönborn (1992).

14.3.3.7 Lower crustal wedging (Middle and Late Miocene, Figure 14-22g)

It is interesting to note that this spectacular event, profoundly shaping the deep structure of the Alps as revealed by geophysical work, left little trace of internal deformation in the transect of the Alps of eastern Switzerland. Roughly speaking, the Central Alps float as an undeformed area above S and N dipping deformation zones confined to the northern and southern Alpine foreland, similar to a pop-up structure. However, the Central Alps were subjected to differential uplift in the order of 10 km during this time interval.
Shortening within the Aar massif in the eastern transect amounts to about 27 km, but is higher in the transect of central Switzerland. This uplift and shortening progressed from east to west as discussed in Chapter 13.1. During the Miocene time interval considered here, shortening concentrated mainly to the west, i.e. in the central and western transects. Miocene shortening in the Southern Alps is more substantial and amounts to 56 km in respect to the post-Adamello phase which does not set in before Middle Miocene (Burdigalian) times according to Schönborn (1992).
Contrary to early interpretations of lower crustal wedging (e.g. Bernoulli et al. 1990) this Miocene phase clearly postdates Insubric backthrusting and strike slip motions, as already proposed by Laubscher (1990) and confirmed by Schönborn (1992). Hence, it is not admissible to kinematically link the lower crustal wedge with backthrusting along the Insubric line. Instead, the Insubric line is cut by this lower crustal wedge beneath the Central Alps.

On the other hand, balancing of shortening in the Southern Alps demands excess volume of crustal material below the Central Alps (Pfiffner 1992, Schönborn 1992). This volume cannot be provided by the crustal material below the N-dipping Insubric line alone but has to include the area occupied by the lower crustal wedge. Hence lower crustal wedging and shortening in the Southern Alps are kinematically linked. As a consequence the Insubric line is part of the hangingwall and allochthonous in respect to the major thrusts formed during Miocene-Pliocene shortening in the Southern Alps (Lombardic phase of Laubscher 1990). The exact shape of this wedge, as well as its composition (exclusively lower crustal material vs. "mélange") remains unknown. Figure 14-22g merely presents a possible geometry which is area balanced against Figure 14-22f in terms of Southern Alpine upper and lower crustal material.

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