Geol Rundsch (1996) 85:832–851

ORIGINAL PAPER

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The transition from passive to active margin tectonics: a case study from the Zone of Samedan (eastern Switzerland)

Received: 22 January 1996 / Accepted: 19 August 1996

Abstract The Zone of Samedan is part of a fossil, early Mesozoic rift system originally situated in the distal, Lower Austro-Alpine domain of the Adriatic passive continental margin. An early Mesozoic configuration of asymmetrical rift basins bounded by relative structural highs compartmentalized Late Cretaceous active margin tectonics; Jurassic half-grabens were folded into arcuate synclines, whereas relative structural highs engendered thin, imbricated thrust sheets. West-directed thrusting and folding initiated at the surface and continued to depths favoring mylonitization under lower greenschist-facies conditions. At this time Liguria-Piemontese ophiolites were accreted to Lower Austro-Alpine units directly underlying the Zone of Samedan. Late Cretaceous orogenic collapse of the Adriatic active margin involved the reactivation of west-directed thrusts as low-angle, top-to-the-east, normal faults. These faults accommodated extensional uplift of Liguria-Piemontese ophiolites and Lower Austro-Alpine units beneath and within the Zone of Samedan. During Paleogene collision, some Late Cretaceous faults in the Zone of Samedan were reactivated under lower anchizonal conditions as north-directed thrusts. The latter stages of this early Tertiary thickening were transitional to brittle, high-angle normal faulting associated with top-to-the-east extension and spreading above the warm, uplifting Lepontine dome.

Key words Rifting · Basin inversion · Active margin · Passive margin · Austro-Alpine · Alpine tectonics · Tethys

Introduction

How did the ancient Tethyan passive margins evolve and what was their role in shaping the Alpine orogeny? In Argand's (1934) classical treatise on Alpine orogenesis, Tethys and its rifted continental margins were regarded as comprising a series of parallel, E- to W-trending oceanic troughs ("geosynclines") and swells ("geanticlines") which formed during N–S to NW–SE convergence of Europe and Africa in early Mesozoic time. Thus, basins and highs were viewed as cylindrical, contractional features which were the embryonic antecedents of the large, recumbent folds (i.e., nappes in the sense of Heim 1922) observed in some parts of the Alps presently.

The advent of the plate tectonic paradigm led to the presently accepted idea that oblique opening of Tethys in an E–W direction accommodated sinistral rotation of Africa with respect to Europe in early Mesozoic time (Dewey et al. 1973; Laubscher and Bernoulli 1977). This inspired Kelts (1981) to infer a transform geometry for the Tethyan margins, similar to that seen presently in the Gulf of California. Evidence for early Mesozoic transform margins in the Alps includes the regional distribution and petrology of Alpine ophiolite-bearing sequences (Lemoine et al. 1987), sedimentary features in these sequences (Weissert and Bernoulli 1985) and in Jurassic basinal sediments (Rück 1990), as well as the kinematic reconstruction of early Mesozoic extensional faults (Handy 1987; Froitzheim 1988; Froitzheim and Eberli 1990; Bertotti 1991) and basins (Eberli 1988; Castellarin and Picotti 1990; Channell et al. 1990; Steffen et al. 1993). Still, examples of pull-apart rift systems in the Alps are rare. A prime objective of this paper, therefore, is to present structural and stratigraphic evidence for the early Mesozoic geometry and evolution of part of such a rift system preserved in the Zone of Samedan, Lower Austro-Alpine units.

The Lower Austro-Alpine units at the transition from the eastern to western Alps (Fig. 1) are attractive for a detailed study of basin evolution for two main reasons: Firstly, they represent the northwesternmost margin of the Apulian microplate (Biju-Duval et al. 1976) or Adriatic promontory (Channell and Horvath 1976) of the African plate. This Early Mesozoic passive mar-

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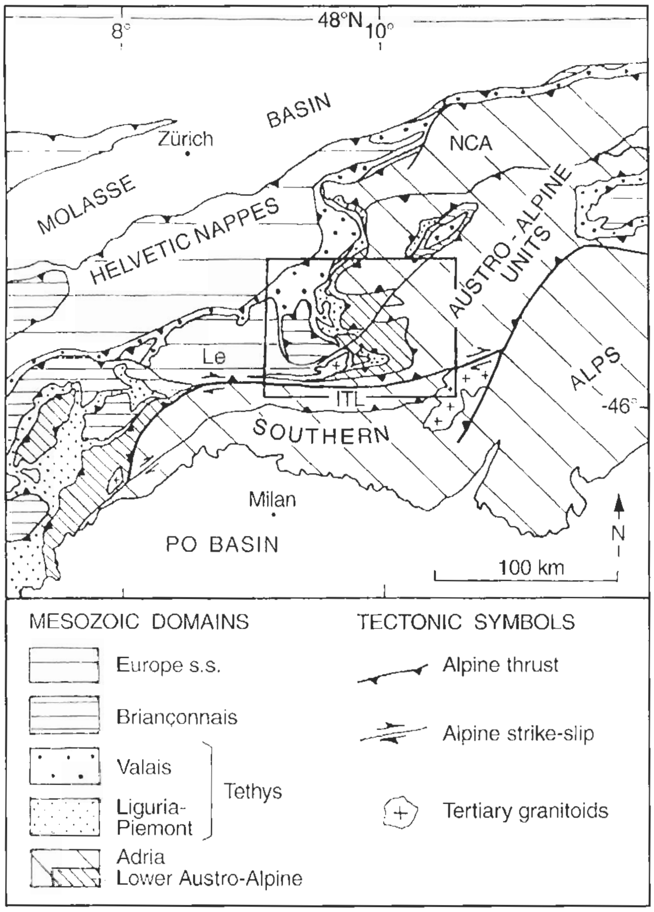


Fig. 1 Geological map of the boundary between the eastern and western Alps. The major Alpine tectonic units are marked according to their early Mesozoic provenance (see text). *Boxed area* shown in greater detail in Fig. 2. *Le* Lepontine dome within the lower Penninic nappes; *ITL* Iorio-Tonale part of the Insubric Line; *NCA* northern Calcereous Alps

gin (Trümpy 1975) became the site of subduction and accretion in Mid- to Late Cretaceous time (Winkler and Bernoulli 1986; Winkler 1988) and was thrust westward onto the Piemontese ophiolites (Ring et al. 1989). Secondly, the transition from passive to active margin tectonics is particularly well preserved in the Lower Austro-Alpine units because these units largely escaped the penetrative Tertiary deformational and metamorphic overprint which affected the underlying Penninic and Helvetic nappes of the European continental margin (Fig. 1). The area investigated below contains some excellent examples of structural basin inversion (as defined by Harding 1985) during these Alpine deformational events. Thus, a second major goal of this paper is to document the effect of inherited passive margin architecture on the evolution of part of the Adriatic active margin during the early stages of Alpine orogeny. It is shown below that this margin experienced a complicated history of repeated thickening and extension during Late Cretaceous and Tertiary time.

Following a brief exposé of the local geology, this paper is divided into two parts: The first part outlines the Alpine structure and kinematics in the Zone of Samedan and reviews structural and stratigraphic evidence for early Mesozoic rifting. In the second part, this information is used to constrain the Mesozoic geometry and evolution of basins in the study area. This local evolution is then discussed in the context of plate tectonic reconstructions for the Mesozoic Tethys. Although primarily devoted to Mesozoic tectonics, this paper also examines Tertiary structures insomuch as these are necessary to reconstitute the earlier history.

Local geology and terminology

The Zone of Samedan (Figs. 2 and 3) is a wedge of Lower Austro-Alpine, Mesozoic sediments between two Lower Austro-Alpine basement nappes: the Err basement unit below and the Julier-Bernina basement unit above (Cornelius 1932, 1950; Finger 1978; Staub 1948). Together, these Lower Austro-Alpine rocks overlie the Liguria-Piemontese Platta ophiolites, the Valais Arblatsch flysch and equivalent units to the north, and middle Penninic (Briançonnais) basement and cover rocks (Fig. 2). This entire nappe pile is transected by the Engadine line, a Late Oligocene to Miocene fault which accommodated sinistral strike-slip (Trümpy 1977) and/or differential uplift (Schmid and Froitzheim 1993). Tertiary doming and erosion of the Lepontine area in the western Alps (Hurford et al. 1989; Milnes 1978) are responsible for the exposure in map view of progressively deeper structural levels from east to west (Figs. 1 and 2).

Various movements affected the contacts between major tectonic units structurally underlying the Zone of Samedan: The Adriatic–Piemontese suture represented by the contact between the Lower Austro-Alpine units and Platta-Malenco ophiolites (Fig. 2) accommodated Late Cretaceous, west- to southwest-directed thrusting (Liniger 1992; Liniger and Nievergelt 1996), whereas the base of the Piemontese Platta ophiolites was a Paleogene, north-directed thrust. This thrust transported the Austro-Alpine margin and Piemontese ophiolites a minimum distance of 50–75 km onto Paleocene to Eocene flysch of Valais affinity covering the European margin (e.g., Fig. 3 in Milnes 1978; Ring et al. 1989). In Oligocene time, north-directed thrusting yielded to top-to-the-east, mylonitic extension (the Turba mylonite zone, marked T in Fig. 2) associated with incipient Tertiary doming to the west (Nievergelt et al., 1996). Although all of these events affected the Zone of Samedan, pre-Tertiary sructures are reasonably well preserved because the Austro-Alpine units formed part of the cold, rigid orogenic lid of the Tertiary Alpine orogen (Froitzheim et al. 1994; Laubscher 1983; Liniger and Nievergelt 1990).

A short digression is needed here to elucidate some Alpine terminology: Traditionally, the modifiers Aus-

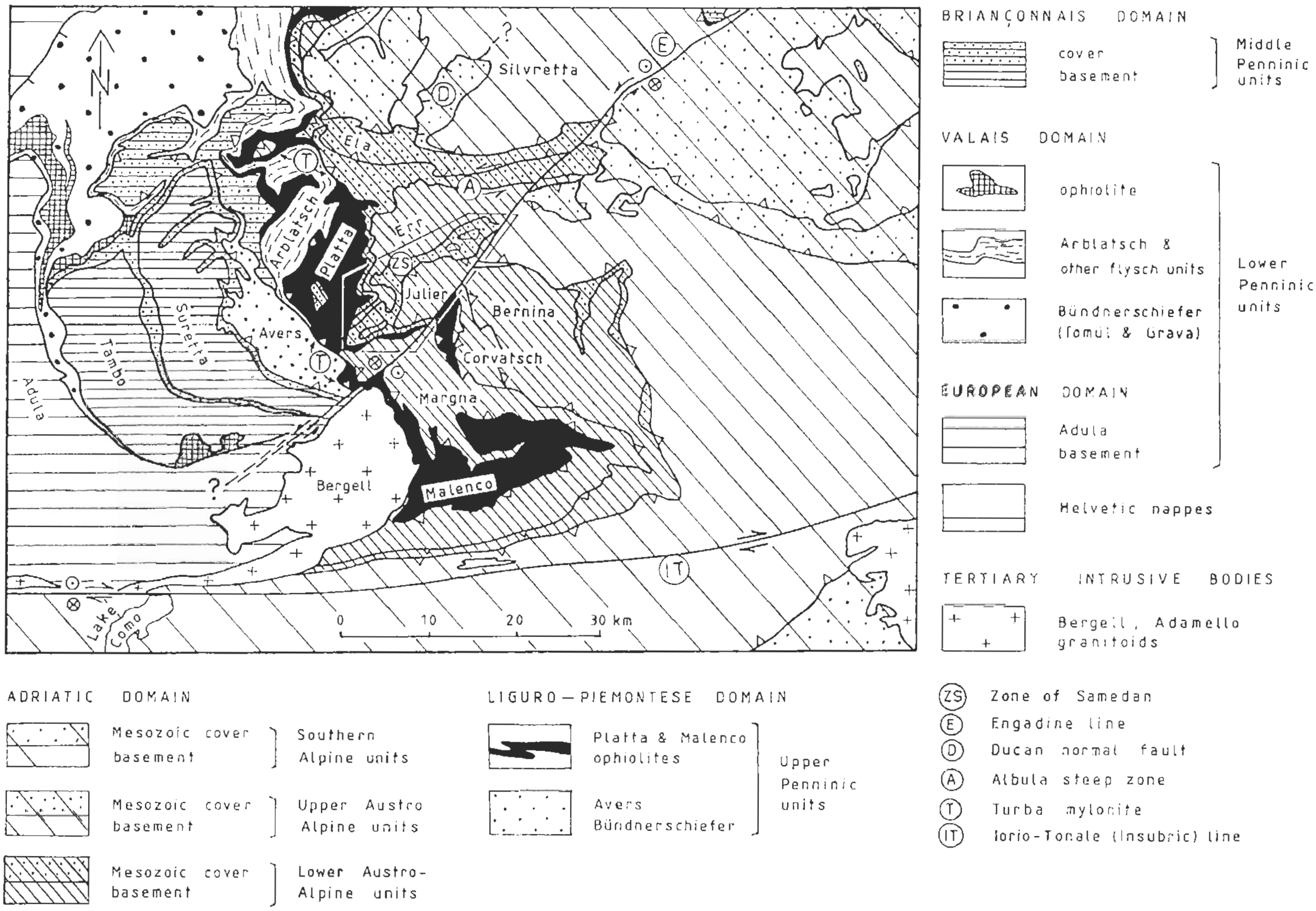


Fig. 2 Geological map of the boundary between Austro-Alpine and Penninic units in the eastern part of the Central Alps. The *outlined area* is depicted in greater detail in Figs. 3 and 6. Map based on Spicher (1980) with modifications from Schmid and Froitzheim (1993) and Schreurs (1993)

tro-Alpine, Penninic, and Helvetic have referred both to Alpine structures and to early Mesozoic paleogeographic provenance. This dualistic usage is implicitly genetic and stems from Argand's dated notion that the present structure of the Alps derives from cylindrical folding of early Mesozoic troughs and swells. To remain clear without inventing a new nomenclature, the old adjectives above are simply used together with the terms "domain" for paleogeographic areas having a common early Mesozoic facies and "unit" for rocks presently nested together within the Alpine orogen. This distinction is important because, as discussed below in the final section, some units comprise originally separate early Mesozoic paleogeographic domains which were juxtaposed later during Alpine strike-slip movement (e.g., Kelts 1981; Trümpy 1982, 1992).

Structure of the Zone of Samedan

Alpine structure and kinematics

The name Zone of Samedan betrays earlier interpretations of the Mesozoic sediments between the Err and Julier basement units as the remains of a single basin which was folded and imbricated during Paleogene north-directed thrusting (Cornelius 1932, 1950; Staub 1948). Our studies have shown, however, that the Zone actually comprises several basins and relative highs preserved within two structurally and stratigraphically distinct domains north and south of the Julier pass road (see map in Fig. 3). These domains individuated during early Mesozoic rifting and later during two phases of Late Cretaceous Alpine deformation, herein termed D1 and D2.

The part of the Zone of Samedan north of the Julier pass road contains primarily two imbricated F1 synclines (sections B–B′ and C–C′ in Fig. 4), whereas the part south of the road comprises thin thrust sheets emplaced on top of an isoclinal F1 anticline which is refolded by two large, tight F2 folds (section D-D′ in Fig. 4). The difference in large-scale structure between these two segments is associated with facies and thickness variations of their constituent sediments. The F1

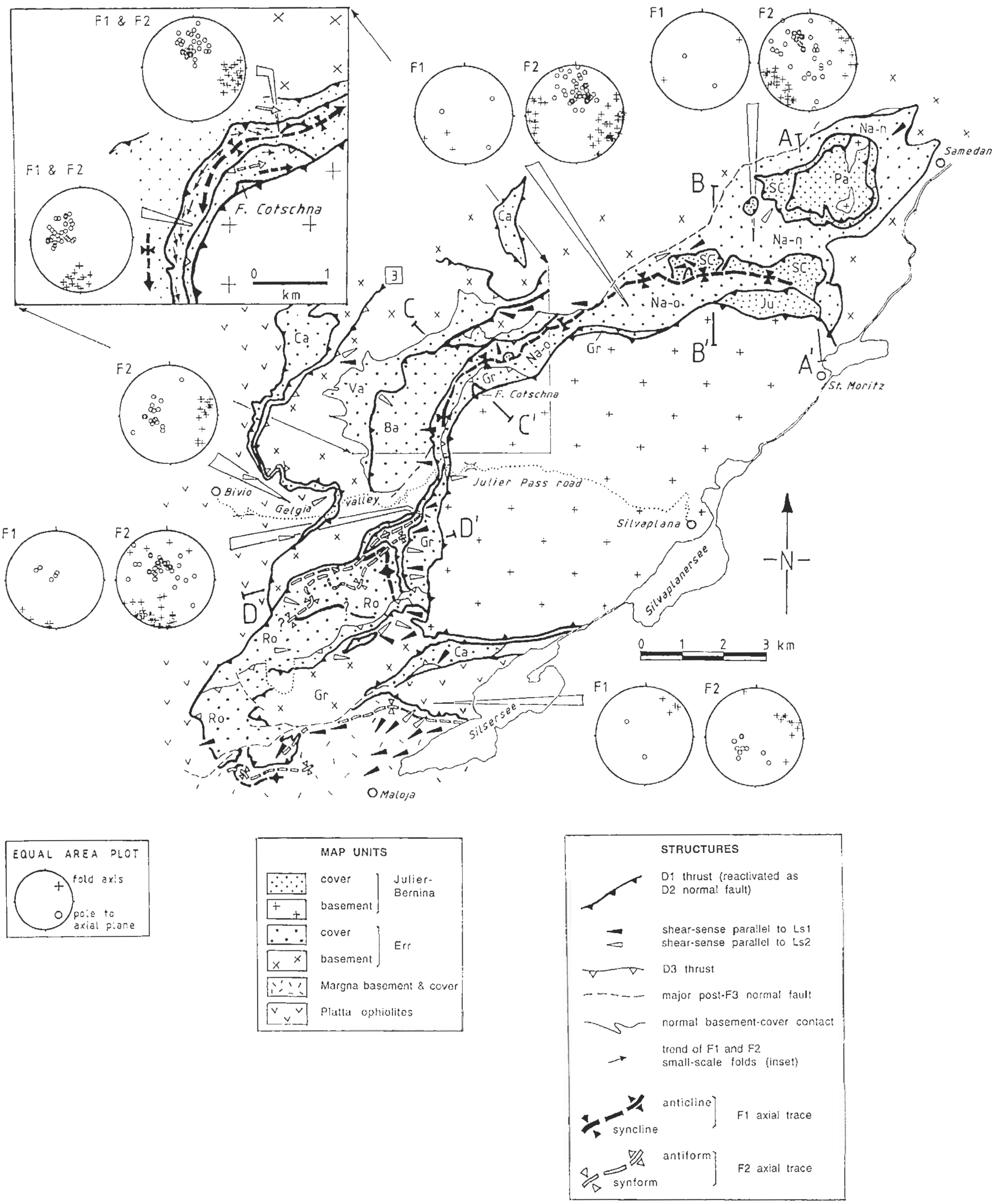


Fig. 3 Map of D1 and D2 structures in the Zone of Samedan and adjacent units. Julier-Bernina cover units: *Ju* Julier; *SC* Schlattain-Clavadatsch; *Pa* Padella; Err cover units: *Na-n* normal Nair unit; *Na-o* overturned Nair unit; *Ba* Bardella; *Va* Vairana; *Gr* Grevasalvas; *Ro* Roccabella; *Ca* Carungas. Lines marked *A–A'* to *D–D'* indicate traces of four cross sections in Fig. 4. *Inset* shows change in F1 and F2 fold orientations in the Fuorcla Cotschna area (see text). Fault labelled with square 3 explained in text

synclines north of the pass road fold the Mesozoic cover of the Err basement, including Triassic pre-rift platform carbonates and two thick, Middle Jurassic syn-rift basinal formations: the Bardella and Saluver formations (Finger 1978; Furrer et al. 1985). The core of the longer, structurally higher syncline in Fig. 3 contains tectonic klippen with similarly aged platform and basin sediments of the overlying Bernina unit (section A–A′ in Fig. 4). In comparison, the Grevasalvas thrust sheets in the southern Zone of Samedan (units labelled Gr in Fig. 3) contain complete, but relatively thin, Mesozoic sequences which also include slivers of Err basement. The doubly folded F1 anticline in the south comprises mostly late Middle Jurassic to Early Cretaceous radiolarian cherts, pelagic limestones, and marly shales. The anomalous thickness of these post-rift sediments in this area is attributable to a combination of early D1 thrusting and subsequent F1 and F2 folding (see doubly folded, early D1 thrust labelled 1 in section D–D′ of Fig. 4).

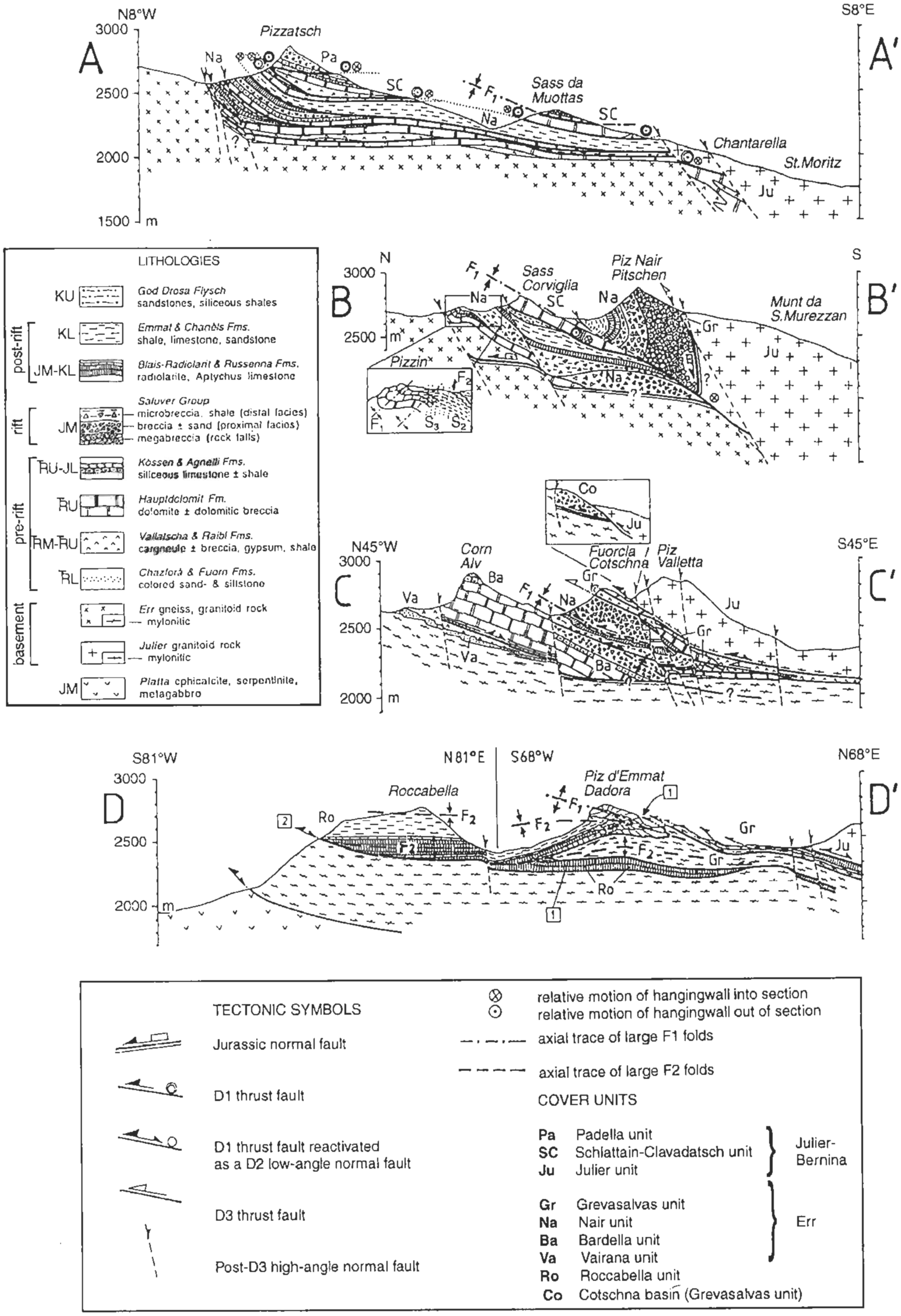


Fig. 4. Cross sections through the Zone of Samedan. Traces of these cross sections shown in Fig. 3. Sections *A–A′* to *C–C′* are oriented perpendicular to the inferred axis of the Middle Jurassic Nair basin (see text and Fig. 8). Sections *A-A′* and *B-B′* are oriented normal to the D1 and D2 movement directions, whereas sections *C-C′* and *D-D′* are, respectively, oblique and parallel to these directions. See text for description of insets to *B-B′* and *C-C′*. Sections *C-C′* and *D-D′* simplified from Fig. 8 of Handy et al. (1993). Age abbreviations next to lithological descriptions: *KU* Upper Cretaceous; *KL* Lower Cretaceous; *JM* Middle Jurassic; *JL* Lower Jurassic; *TRU* Upper Triassic; *TRM* Middle Triassic; *TRL* Lower Triassic. Abbreviations for cover units are identical to those in Fig. 3. See text for explanation of faults labelled with squares *1* and *2*

The structural map in Fig. 3 shows that the orientation of small-scale D1 and D2 structures varies with location in the Zone of Samedan. In the Err cover units, F1 axes and axial planes are sharply around the bend in the northern part of the Zone of Samedan near Fuorcla Cotschna (inset to Fig. 3). The F1 axial planes lie parallel to a penetrative S1 mylonitic foliation formed under lower greenschist-facies conditions (Handy et al. 1993). Within the thin thrust sheets in the southern Zone of Samedan as well as in the Margna basement unit, F1 axes plunge gently to the northeast or southwest (Fig. 3), parallel to a well-developed mineral stretching lineation, L1. The sense of shear parallel to L1 in these areas is consistent with top-to-the-west to -southwest thrusting during D1 (Fig. 3). Similar shear senses are observed in the northern Zone of Samedan, although top-to-the-WNW to -NW shear senses are found there locally as well (e.g., near Fuorcla Cotschna in Fig. 3). Unfortunately, L1 stretching lineations are often impossible to distinguish from L2 lineations in the area just east of the town of Samedan.

D2 deformation also occurred under lower greenschist-facies conditions (Handy et al. 1993), but is more heterogeneous on the outcrop scale than D1. It is marked by tight F2 folds and by an S2 axial plane schistosity which in the Zone of Samedan generally dips more gently than S1. Like S1, S2 follows the bend in the northern Zone of Samedan (Fig. 3). S2 becomes pervasively mylonitic and parallel to S1 along nappe contacts. Kinematic indicators in D2 mylonites are often ambiguous (e.g., conjugate shear bands, dual-sense rotation of clasts), but the drag direction of S2 surfaces along the more steeply inclined D1 nappe contacts indicates top-down-to-the-east to -ENE sense of shear (Fig. 3). Such nappe contacts are often marked by tectonic and stratigraphic repetitions which have been tectonically thinned or even partly faulted out, suggesting that D1 thrusts were reactivated as D2 low-angle normal faults (examples cited below).

Within the Zone of Samedan, noncoaxial D2 extensional shearing localized between the Err cover and the Julier-Bernina units, along the base of the Err sediments, and in the lower part of the Err basement. Deformation of the Julier basement lid was primarily cataclastic, except at its base where lower greenschist-facies D2 mylonite and cataclasite coexist (Handy et al. 1993). In the northern Zone of Samedan, Upper Triassic monomict carbonate breccias with a fine-grained, mylonitic matrix (presumably a member of the Kössen formation) in the normal limb of the upper F1 syncline contain east-plunging D1 or D2 stretching lineations defined by elongate breccious components (Fig. 3). The stratigraphic thickness in both the normal limb of this syncline and the overturned limb of the underlying F1 syncline is reduced and locally appears to have been faulted out during east-directed D2 normal faulting (sections B–B′, C–C′ in Fig. 4). D2 extensional faulting is also manifest in the southern Zone of Samedan, where pre- and syn-rift sediments are missing and post-rift sediments directly overlie Err basement rock (section D–D′ in Fig. 4). D2 mylonitization is pervasive and mylonitic along and within the base of the Err basement unit. Between extensionally reactivated D1 thrusts, F2 folds verge uniformly antiform-to-the west (section D–D′; Fig. 4) and opposite shear senses appear in their attenuated upper and lower limbs (Handy et al. 1993). This observation, combined with the general coaxiality of F1 and F2 folds throughout the area (equal area plots in Fig. 3), suggests that F2 folds are similar folds which flattened the previously steepened S1 schistosity, as proposed by Froitzheim (1992).

Beneath the Zone of Samedan, a broad zone of noncoaxial mylonitic D2 shear affects the entire Platta ophiolite unit and the uppermost part of the underlying Margna basement and cover units (Fig. 3). Extensional uplift of the Platta and Margna units along this shear zone is probably responsible for the preservation of a sharp increase in Late Cretaceous metamorphic grade from upper anchizone in the Err units of the Zone of Samedan to epizone in the underlying Platta and Margna units (Klahr 1994, see his Fig. 7.4).

The physical conditions of metamorphism in the Zone of Samedan during D1 and D2 are not well constrained. A temperature range of 250–400°C is estimated for D1 from the observation of dynamically recrystallized quartz and brittle feldspar clasts within mylonitic granitoids. This syn-tectonic temperature estimate is relatively crude because it is based on the microscopic identification of deformation mechanisms in fault rocks and the extrapolation of experimental flow laws for these mechanisms to geological strain rates (e.g., Carter and Tsenn 1987).

Post-D2 deformation locally folds and imbricates D1 and D2 structures, but does not significantly alter the structure of the Late Cretaceous nappe pile in the Zone of Samedan. A D3 thrust in the southern Zone of Samedan ramps northwards along the base of the structurally highest Grevasalvas imbricate, transporting this unit and the overlying Julier basement lid a maximum distance of 2 km to the north (Fig. 5). This thrust can be traced as far as Fuorcla Cotschna (location in Fig. 3), where Err basement rocks are cataclastically infolded beneath Middle Triassic corneole along the D3 thrust plane (section C–C′ in Fig. 4). Herwegh (1992) observes that this D3 thrust truncates, and therefore syn- to post-dates F3 folds in the southern Zone of Samedan (see photo in his Fig. 5). In the absence of D3 stretching lineations, the transport direction is assumed to be perpendicular to the ENE- to WSW-trending F3 axial planes, whereas the net displacement of 2 km is the distance parallel to the transport direction between displaced lithological contacts in the hanging- and footwalls of the thrust (see map reconstruction in Fig. 6 of Herwegh 1992, profile B–B′ in Mattenberger 1994). Another possible D3 thrust with cataclastic fault rocks accommodates an unknown amount of displacement along the Err-Platta contact and appears to ramp northward within the Err basement unit (fault labelled

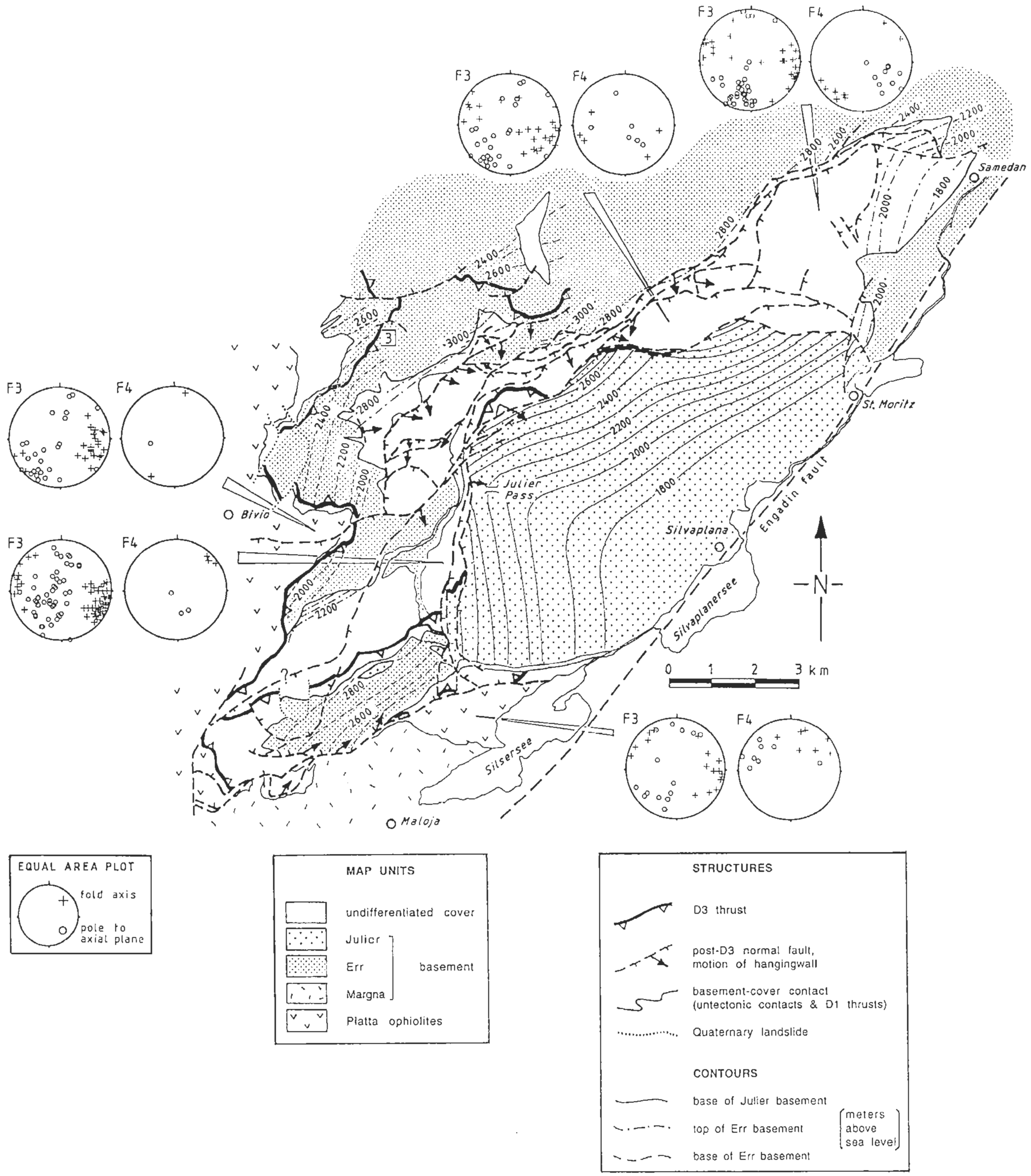


Fig. 5 Map of D3 and later structures in the Zone of Samedan and adjacent units; basement contours discussed in text. Fault labelled with square *3* explained in text

3 in Figs. 3 and 5). Yet other D3 thrusts are suspected (e.g., at the base of the Julier basement lid, see Finger 1978), but are either unmappable or unverifiable due to the complicated pre-D3 history and the obliterating effects of later normal faulting.

Unlike the F1 and F2 folds, F3 folds maintain the same orientation and style throughout the area, and do not follow the bend in the northern Zone of Samedan (equal area plots in Fig. 5). The ESE–WNW trend of the F3 axial planes is consistent with NNE- to SSW-directed shortening during D3. The strong flexural slip component of F3 folding in most lithologies (Figs. 11a and b in Handy et al. 1993), the cataclasitic behavior of quartz-bearing rock along D3 thrusts, and the restriction of crystal plastic deformation to fine-grained carbonate rock all point to low-temperature, lower anchizonal conditions during D3.

F4 folds are generally open (Fig. 11c in Handy et al. 1993) and accommodated minor NW- to SE-directed shortening perpendicular to their axial surfaces (Fig. 5). They are observed only locally on the outcrop scale and effect variations in the orientation of F3 axial surfaces. F5 kink folds with N–S trending axes indicate a late phase of minor E–W shortening.

Brittle normal faulting modified the essentially Late Cretaceous structure of the Zone of Samedan. N–S and NE–SW to ENE–WSW trending, moderately to steeply dipping (40–70°), normal faults displace D1 and D3 thrust contacts (Fig. 5). The N–S faults dip consistently to the east, whereas the NE–SW to ENE- to WSW-trending faults have conjugate dip azimuths: The large, NNW-dipping Lunghin fault just north of the Silsersee dips towards a series of SSE-dipping faults to the north (sections A–A′ and B–B′ in Fig. 4, narrowly spaced contours at top of Err basement in Fig. 5). This conjugate, asymmetrical fault system delimits a large, ENE- to WSW-trending graben which contains the structurally highest units of the area (Julier basement unit, Grevasalvas unit) and is juxtaposed with lower units both to the north (Err basement and cover units) and south (Platta and Margna units). Thus far, the NE–SW to ENE- toWSW-trending fault set has not been recognized regionally (Froitzheim et al. 1994) and appears restricted to the Zone of Samedan. Locally, the NE- to SW-trending faults offset, and therefore appear to post date the N- to S-trending normal faults (e.g., Julier basement lid 1 km north of the Julier pass; Fig. 5). However, these cross-cutting relationships only reflect the latest movements, so the nucleation of these two fault systems may have been broadly coeval. The presence of young, oblique-dip slickenfibers on some steep, normal fault surfaces (arrows on dashed fault traces in Fig. 5) suggests that they may have served as restraining and releasing bends during Miocene strike-slip movement along the nearby Engadine fault in the upper Engadine valley.

Table 1 summarizes the Alpine structural events in the Zone of Samedan and the criteria used to constrain their ages. These age criteria are discussed extensively in Handy et al. (1993) and therefore are treated only cursorily below. Two episodes of crustal thickening and extension are distinguished: (a) Late Cretaceous, west- to southwest-directed thrusting and nappe stacking (D1), followed by E–W extension and differential uplift (D2); and (b) middle to late Paleogene, north-directed thrusting and folding (D3) transitional to both E–W and N–S to NE- to SW-directed extension (D4). The same sequence of events here is also recognized regionally in the Lower Austro-Alpine units, as summarized in Froitzheim et al. (1994). D3 deformation in the Zone of Samedan is interpreted to have been broadly coeval with the aforementioned northward thrusting of the previously sutured Platta and Austro-Alpine units onto Paleocene to Eocene flysch of Valais affinity (Arblatsch flysch in Fig. 2). D4 high-angle, normal faulting in the Zone of Samedan manifests a change from this Paleogene, N–S shortening to lateral, inhomogeneous spreading above the Turba low-angle normal fault at the base of the Platta ophiolites to the west (Fig. 2; Nievergelt et al., in press). The regional significance of F4 and F5 folding is more difficult to ascertain; F4 may reflect NW- to SE-directed shortening during Insubric backfolding along the Iorio–Tonale Line (Fig. 1; Schmid et al. 1989) and/or local shortening during intrusion of the Bergell granitoids to the south.

Early Mesozoic rift features

Early Mesozoic rift tectonics are manifest as stratigraphic variations preserved within D1 nappes and as relic extensional structures. Representative stratigraphic columns for the D1 cover nappes in the Zone of Samedan appear in Fig. 6.

Middle and Upper Triassic evaporites (Vallatscha and Raibl formations) and Upper Triassic platform carbonates (Hauptdolomit formation) in tectonic units north of the Julier pass road are up to 400 m thick, whereas the same rocks in the Grevasalvas units attain a primary stratigraphic thickness of only 15–20 m. Likewise, Lower Jurassic limestone of the Agnelli formation occurs in Err cover units north of the pass road, but not to the south. These stratigraphic variations among tectonic units probably reflect tectonically induced differences in early Mesozoic sedimentation, rather than Alpine tectonic omissions, because D2 extensional shearing and D4 normal faulting affected primarily the contacts, rather than the bulk, of the D1 nappes. Differential subsidence therefore began in the Middle to Late Triassic, before the onset of rifting and rapid subsidence in Early to Middle Jurassic time (e.g., Trümpy 1975).

Most vestiges of early Mesozoic extensional tectonics are preserved within the folded and imbricated remains of two thick, Middle Jurassic basinal sequences in the northern Zone of Samedan (Fig. 3): the Bardella formation to the west and the structurally higher Saluver formation to the north and east (Finger 1978; Furrer et al. 1985). These two formations occupy F1 synclines within, respectively, the Bardella and Nair cover units in Fig. 6. A third Middle Jurassic basinal sequence, the Padella formation (Eberli 1988; Furrer et al. 1985), forms part of a thrust sheet (the Padella unit marked Pa in Fig. 6) within the northeasternmost Bernina klippe overlying the Err cover units. In all of these formations, inverse-to-normally graded, poly- and monomict brec-

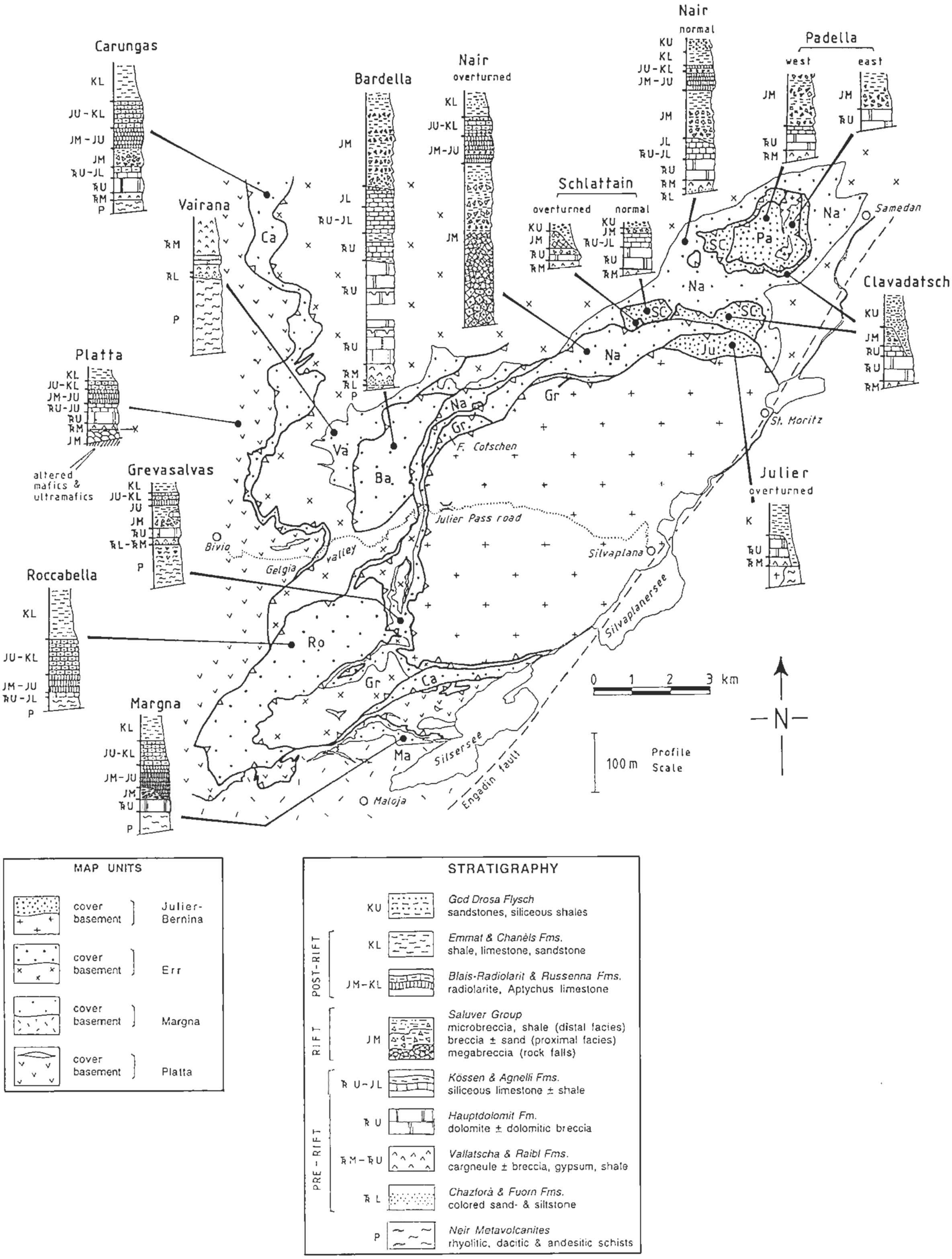


Fig. 6 Stratigraphy of cover units in the Zone of Samedan and vicinity. Sediment thicknesses from own observations and from Dietrich (1970), Finger (1978), and Liniger and Guntli (1988). Age abbreviations next to stratigraphic columns: *KU* Upper Cretaceous; *KL* Lower Cretaceous; *JU* Upper Jurassic; *JM* Middle Jurassic; *JL* Lower Jurassic; *TRU* Upper Triassic; *TRM* Middle Triassic; *TRL* Lower Triassic; *P* Permian

cias alternate with sandstones which grade upsection into fine-grained argillaceous shales (Finger 1978). This overall fining up sequence is typical for syn- to post-rift turbiditic megacycles developed within extensional basins of rifted continental margins (e.g., Williams 1993; Prosser 1993).

Table 1 Alpine history of the Zone of Samedan and adjacent units above the Turba mylonite zone. Numbers of references in alphabetical order: 1. Dietrich [1970], 2. Dürr [1992], 3. Finger [1978], 4. Froitzheim [1988], 5. Froitzheim [1992], 6. Froitzheim and Eberli [1970], 7. Froitzheim et al. [1994], 8. Furrer et al. [1985], 9. Handy et al. [1993], 10. Handy et al. [1996], 11. Liniger [1992], 12. Liniger and Guntli [1988], 13. Liniger and Nievergelt [1990], 14. Nievergelt et al. [1996], 15. Peretti [1985], 16. Rösli [1946], 17. Schmid et al. [1989], 18. Schmid et al. [1990], 19. Schmid and Froitzheim [1993], 20. Trommsdorff and Nievergelt [1983], 21. Trümpy [1977], 22. Ziegler [1956]. For purposes of regional correlation, phases D1–D4 are listed with the regional names suggested by Froitzheim et al. [1994]

Regional event	Local phase	Structures and sediments	Conditions in fault rocks	Kinematics in fault rocks	Deformation age (Ma)	Age criteria
		Pre-F1 slump folds and veins in Aptian-Albian marls and sandstones			108–100	Youngest sediments in Platta, Carungas, and Roccabella units are Aptian-Albian age [1]
Subduction and collision: Nappe stacking and accretion	D1 "Trupchun"	E-dipping pre-F1 thrusts [3, 7, 9]	Cataclastic	Top-to-W thrusting	100–88	Cenomanian-Turonian God-Drosa flysch beneath pre-F1 thrusts [3, 8, 16]
		S1, F1, Ls1, E-dipping thrusts [2, 4, 5–7, 9, 11, 13]	Greenschist facies, mylonitic	Top-to-W, -SW, locally also -WNW to -NW shear	89–76	K-Ar white mica forming S1 in axial planes of F1 Emmat fold [10]
Collapse: Extensional Uplift	D2 "Ducan-Ela"	S2, F2, Ls2, D1 thrusts reactivated as normal faults [5, 7, 9, 11]	Greenschist facies, mylonitic	Top-to-E shear and flattening	80–69	K-Ar white mica and Ar/Ar amphibole in axial planes of F2 folds [10]
Collision: Nappe stacking and thickening	D3 "Blaisun"	S3, F3, S-dipping thrusts [2, 7, 9, 11–13]	Anchizone, cataclastic in quartzose rocks, mylonitic in calcite rocks	Top-to-N to -NNE thrusting	50–30	F3 cut by Bergell granitoids [14, 20], F3 deforms Paleocene-Eocene Arblatsch flysch [7, 18, 22]
Doming: Spreading (Turba extension)	D4 "Turba"	E-dipping normal faults [7, 9, 11, 13, 14]	Cataclastic	Top-to-E extension	35–30	Turba mylonite zone syn- to post-dates Schams D2 [7, 11, 14]
		N- and S-dipping normal faults [9]	Cataclastic	N-S extension	≤35–30	S-dipping normal faults cut E-dipping faults [9]
Oblique convergence		F4, F5 [7, 9, 11, 13]	Unmetamorphic, cataclastic	NW-SE shortening, E-W shortening	(30?) 25–12	Cooling ages along the Iorio-Tonale line [17]
Lateral extrusion?		NE- to SW-trending slickensides on normal fault surfaces [9]	Cataclastic	Sinistral strike-slip and vertical block rotation [19]	25–12	Engadin line cuts Bergell granitoids [19, 21]

In contrast to the deep basinal formations described above, the basinal sequence in the thrust sheet forming the base of the Bernina klippe (i.e., the Schlattain and Clavadatsch units marked SC in Fig. 6) is very shallow and therefore is interpreted to represent part of a relative structural high in Middle Jurassic time (Handy et al. 1993, and below). Similarly, the thin Grevasalvas units contain only reduced Mesozoic sediment thicknesses (Fig. 6) and the deformed remains of one or more small, relatively shallow Middle Jurassic basins (Handy et al. 1993). The Grevasalvas units therefore represent a relative structural high which was imbricated during west-directed D1 thrusting and again during north-directed D3 thrusting (see reconstruction in next section).

There is strong evidence from the facies distribution and bedding angles in the aforementioned Middle Jurassic formations that the rift basins in the Zone of Samedan were asymmetrical half-grabens: Bedding in the syn-rift Bardella formation dips gently to the east and lenses out to the west (Fig. 7a). Intra-formational bedding discordances form acute angles which open to the

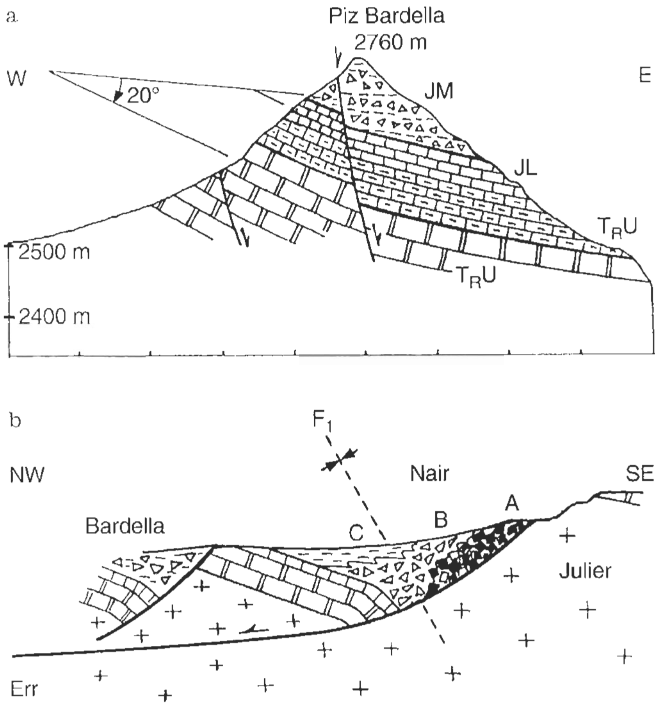


Fig. 7a, b Early Mesozoic extensional features in the Zone of Samedan. **a** Cross section through Piz Bardella. *Arrow* showing angular discordance between Middle Jurassic syn-rift clastics (*JM*), Lower Jurassic–Upper Triassic limestone (*JL-TRU*), and Upper Triassic dolomite (*TRU*). Lithological symbols as in Fig. 6. Steep normal faults are mid- to late Tertiary features (see text for interpretation). **b** Reconstructed cross section through the Early Mesozoic Nair basin containing members A, B, and C of the Middle Jurassic Saluver formation. The section is oriented perpendicular to the inferred strike of the basin axis and shows the trace of the later F1 synclinal axial plane

east (Fig. 5A in Handy et al. 1993). The Bardella formation is therefore inferred to have been deposited onto an extensional fault block which rotated above an originally west-dipping, syn-sedimentary growth fault. The 20° angle between bedding dips at the bottom of the Bardella formation and the top of the underlying pre-rift sediments indicates that this putative extensional fault block rotated at least 20° about a north- to south-trending axis in Jurassic time (Fig. 7a). The orientation of this rotation axis corresponds approximately with the current north–south strike of bedding in the Bardella formation and is therefore inferred to represent the local trend of the basin axis.

A fault preserved in the Grevasalvas units beneath the Julier thrust at Fuorcla Cotschna (location in Fig. 6) juxtaposes steeply southeast-dipping syn-rift Dogger clastics directly above gently southeast-dipping basement schist (inset with unit labelled Co in section C–C′ of Fig. 4; see field photos in Figs. 5B and C in Handy et al. 1993, and Figs. 3-8 to 3-10 in Regli 1992). The Dogger rift breccia contains components of reworked cataclasites identical to those found along the fault plane (photomicrograph in Fig. 6D of Handy et al. 1993). Faulting was therefore coeval with clastic sedimentation in Dogger time. The sense of shear in the cataclasites along the southeast-dipping fault plane is top-to-the-northwest (shear criteria in Fig. 6C of Handy et al. 1993). This sense of shear, together with the omission of Triassic stratigraphy along the fault plane, suggests that the fault was originally a west- to northwest-dipping normal fault which later attained a southeastern dip during D1 and/or D3 Alpine thrusting. Froitzheim (1988) and Froitzheim and Eberli (1990) describe similar structures in parts of the Err and Ortler units further to the north and east.

The depth and asymmetry of a former rift basin containing the Saluver formation is reflected in this formation's differentiation into three members (Fig. 6; Finger 1978): A component-supported megabreccia with house-size blocks of Julier basement rock (member A) grades to coarse, polymict breccias and arkosic sandstones (member B) in formerly deep, proximal parts of the basin, whereas fine-grained sandstones and dark marly shales (member C) were deposited in the originally shallower, distal parts of the basin. Presently, members A and B of the Saluver formation occupy the overturned limb of an imbricated F1 syncline, whereas member C forms the normal limb of this fold (sections B–B′ and C–C′ in Fig. 4). Triassic strata are entirely missing in the overturned limb, but are usually complete (although tectonically thinned) in the normal limb. Unfolding this F1 syncline about its subhorizontal axis yields an elongate, asymmetrical basin with a more proximal facies in the southeast than in the northwest (Fig. 7b; see also map reconstructions in the next section). In unfolding the syncline, the F1 fold axis is assumed to be parallel to the axis of the former rift basin as indicated by the present strike of the contact between members B and C in the Saluver formation. The limbs of the syncline are rotated by amounts required to bring the bedding of the stratigraphically highest syn-rift clastics into a subhorizontal orientation. The restored cross-section of the Nair basin in Fig. 7b offers an explanation for the missing Triassic strata in the overturned limb of the F1 syncline: These pre-rift strata are interpreted to have been omitted within the deeper part of the Nair basin, where a putative, west- to northwest-dipping, Middle Jurassic extensional fault accommodated top-to-the west downthrow and rotation of pre- and syn-rift Saluver sediments in its hangingwall. This normal fault or its reactivated Alpine successor may form the present basement-sediment contact beneath the Zone of Samedan (double-lined fault marked with "?" in section B–B′ of Fig. 4).

Middle Jurassic clastics in the western part of the Padella unit (Fig. 6) become thicker and more breccious from west to east, where they are juxtaposed with Upper Triassic platform carbonates along a steeply west-dipping normal fault (photo in Fig. 8 of Froitzheim and Eberli 1990). This suggests that the Padella basin was an asymmetrical half-graben which formed above a west-dipping master normal fault (Froitzheim and Eberli 1990). For the purposes of the paleogeo-

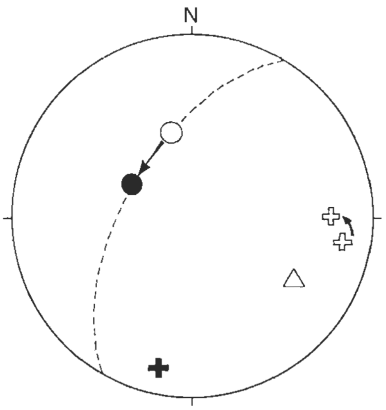


Fig. 8 Equal area projection showing back rotation of average S1 and S2 foliation poles (*circles*) and F1 and F2 fold axes (*crosses*) about a rotation axis (*triangle*) defined by contrasting orientation of basement contours (Fig. 6) and foliation about F. Cotschna (location in Fig. 3). The back rotation (*arrows*) brings the average S1 and S2 poles from NE of F. Cotschna (*open circle*) into coincidence with those to the SW (*filled circle*). Note, however, that the fold axes in these two locations (*open* and *filled crosses*) are not colinear, even after back rotation (see text for explanation)

graphic reconstruction discussed in the next section, the former rift basins containing the Padella, Saluver, and Bardella formations are henceforth referred to, respectively, as the Padella, Nair, and Bardella basins.

Early to Middle Jurassic basinal clastics also occur within the Margna unit structurally underlying the Zone of Samedan (Liniger 1992, Liniger and Guntli 1988). Unfortunately, these sediments were isoclinally folded and severely boudinaged during D1 and D2 mylonitic shearing, hindering realistic reconstructions of the original basin geometry in the Margna unit.

Middle Jurassic to Early Cretaceous radiolarian cherts and pelagic limestones mark the end of rifting in all Lower Austro-Alpine units. Breccious layers within Middle to Upper Jurassic radiolarian chert rock of the Saluver formation (Finger 1978; Staub 1948) indicate that submarine basin topography remained pronounced, even after the onset of post-rift subsidence in late Middle Jurassic time. These clastic radiolarian cherts grade upwards into thin, alternating layers of fine, sandy limestone and marl (Aptian-Albian; Dietrich 1970) before yielding to early Late Cretaceous flysch (Cenomanian–Turonian; Rösli 1945) beneath D1 thrust sheets forming the Bernina klippe in the northern Zone of Samedan. The local transgression of this flysch directly onto pre- and syn-rift sediments of these thrust sheets (see stratigraphic columns for the Schlattain and Clavadatsch units in Fig. 6) suggests that at least part of the Bernina domain was exposed to erosion prior to or during early D1 thrusting.

In conclusion, the structural and stratigraphic relations in the Zone of Samedan indicate that asymmetrical basins developed above west- to northwest-dipping extensional faults at the westernmost edge of the Adriatic continental margin. This accords well with the observations of Bernoulli et al. (1990), Eberli (1988), Froitzheim (1988), and Froitzheim and Eberli (1990) for other parts of the Lower Austro-Alpine margin exposed further to the north (Fig. 2).

Mesozoic paleogeography and evolution of the Zone of Samedan

Passive margin tectonics

Any attempt to reconstruct the early Mesozoic paleogeography in the Zone of Samedan must begin by clarifying the relationship of the bend in the northern Zone of Samedan to the change in orientation of D1 and D2 structures around Fuorcla Cotschna (Fig. 3). Do these features stem from early Mesozoic extension, Alpine tectonics, or both?

The bend in the Zone of Samedan is a large, ESE-plunging open fold whose limbs are defined by the contrasting orientations of S1 and S2 about Fuorcla Cotschna (equal area plots in inset to Fig. 3) and by a change in strike of the Julier and Err basement contours (Fig. 5). This fold probably pre-dates D2 and D3 deformation because it is unlikely to have formed during subhorizontal flattening associated with D2 extensional shearing, and because its interlimb angle is much greater (150°) than that of the typically close to tight F3 folds. Thus, bending of the Zone of Samedan probably occurred prior to or during D1. Unfolding the S1 and S2 planes about the open fold axis on an equal area net does not restore the F1 and F2 fold axes northeast and southwest of Fuorcla Cotschna to a coaxial orientation (Fig. 8). The change in orientation of D1 and D2 structures about Fuorcla Cotschna therefore pre- to syndates open folding and may reflect primary variations in the orientation of bedding within the basinal sediments with respect to the D1 shearing plane.

The present shape of the Zone of Samedan is therefore attributed to differential rotations of the northeastern and/or southwestern parts of the Nair rift basin about vertical axes during west- to southwest-directed D1 thrusting, possibly combined with arcuation of this basin during early Mesozoic rifting. Unfortunately, the available structural evidence is insufficient to constrain the relative contributions of rift arcuation and D1 rotation in forming the present bend in the Zone of Samedan. The nature of this arcuation and rotation is discussed again below in relation to the paleogeographic reconstruction in Fig. 9 and the present map pattern in Fig. 3.

The map in Fig. 9 was constructed by retro-deforming the present nappe stack first in a N–S direction along D3 thrusts described above, then in an east–west direction approximately parallel to the trend of L1 and L2 stretching lineations along D1 nappe contacts. It should be emphasized that the reconstruction in Fig. 9 is only qualitative for two main reasons: Firstly, displa-

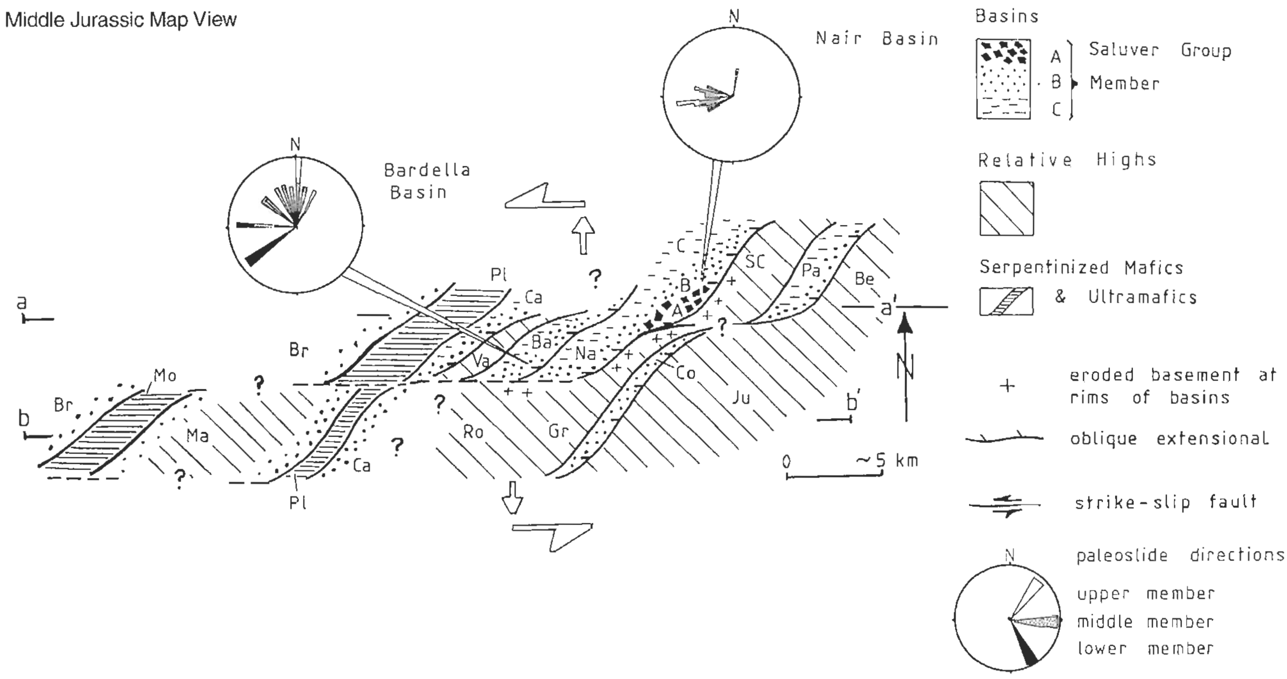


Fig. 9 Map of part of the Lower Austro-Alpine domain at the end of Middle Jurassic rifting (modified from Fig. 12A in Handy et al. 1993). *Letters* occupy original position of cover units within the Zone of Samedan. *Blank areas* and *question marks* indicate areas for which structural and stratigraphic information is missing or equivocal. *Rose diagrams* show back-rotated, paleo-transport directions of clastic basinal sediments from the Bardella formation (data from Finger 1978) and from member B of the Saluver formation (own data). The circles' radii indicate 30 times a uniform distribution of data points. Lines *a–a'* and *b–b'* are traces of restored cross sections in Fig. 10. Structural elements along the rifted margins: *Be* Bernina high; *Ju* Julier high; *Pa* Padella basin; *SC* Schlattain-Clavadatsch high; *Na* Nair basin; *Ba* Bardella basin; *Va* Vairana unit; *Ca* Carungas basin; *Co* Cotschna basin; *Gr* Grevasalvas high; *Ro* Roccabella high; *Ma* Margna extensional allochthon; Liguro-Piemontese domain: *Pl* Platta ophiolites; *Mo* Forno-Malenco-Lizun ophiolites; Briançonnais domain: *Br*

cements along D1 thrusts and D2 extensional faults are unknown, precluding any reliable quantification of Late Cretaceous, east–west shortening. Likewise, north-directed D3 movements are only poorly constrained, although it was concluded above that north–south shortening is minor compared with D1 shortening. Secondly, D2 extensional shearing and D4 normal faulting eliminated an unknown amount of early Mesozoic stratigraphy along some D1 nappe contacts. For example, most of the overturned limb of the F1 syncline containing the proximal facies of the syn-rift Bardella formation (section C–C′ in Fig. 4) was faulted out.

The Mesozoic domain containing the Zone of Samedan is depicted in Fig. 9 to have comprised deep basins in the west and north (Bardella, Nair, and Padella basins) bounded by relative structural highs to the east and south (Julier, Bernina, and Grevasalvas domains). The Margna basement nappe, currently embedded within Platta and Malenco ophiolites (Fig. 2), is interpreted to represent an extensional allochthon (or "microcontinent" of Trümpy 1975) of the Adriatic passive margin (Fig. 9; see also Liniger 1992). Piemontese ophiolites of the Platta and Malenco units comprise mostly altered mafic and serpentinized ultramafic rocks and therefore are considered to be remnants of a lithospheric ocean exhumed during the late stages of Jurassic rifting (Lemoine et al. 1987; Trommsdorff et al. 1993). The sigmoidal shape of the Nair basin in Fig. 9 reflects structural arguments above that the present bend in the Zone of Samedan may have originated partly during early Mesozoic rifting. Basin arcuation like this can arise during strike-slip or oblique transverse shearing (Hempton and Neher 1986; Schreurs 1994) and indicates an overall regime of sinistral strike-slip motion in map view.

The southward continuation of the Nair and Bardella basins in Fig. 9 is not clear (see blank area to the west of the Grevasalvas domain in Fig. 9) due to the absence of deep basinal formations south of the Julier pass road in Fig. 6. Either these rift basins were faulted out during D2 extension (e.g., along the fault labelled 2 in section D–D′ of Fig. 4) or they never existed in the units currently exposed south of this road. The former case implies a significant southward increase across the Gelgia valley in the amount of D2 displacement along the Err basement-cover contact, whereas the latter possibility implies the former existence of an east- to west-trending fault separating the Bardella and Nair basins in the north from the Grevasalvas domain to the south. Handy et al. (1993) tentatively proposed such a fault, namely, an early Mesozoic transfer fault whose trace in the Gelgia Valley was supposedly obliterated by D3 and D4 deformation.

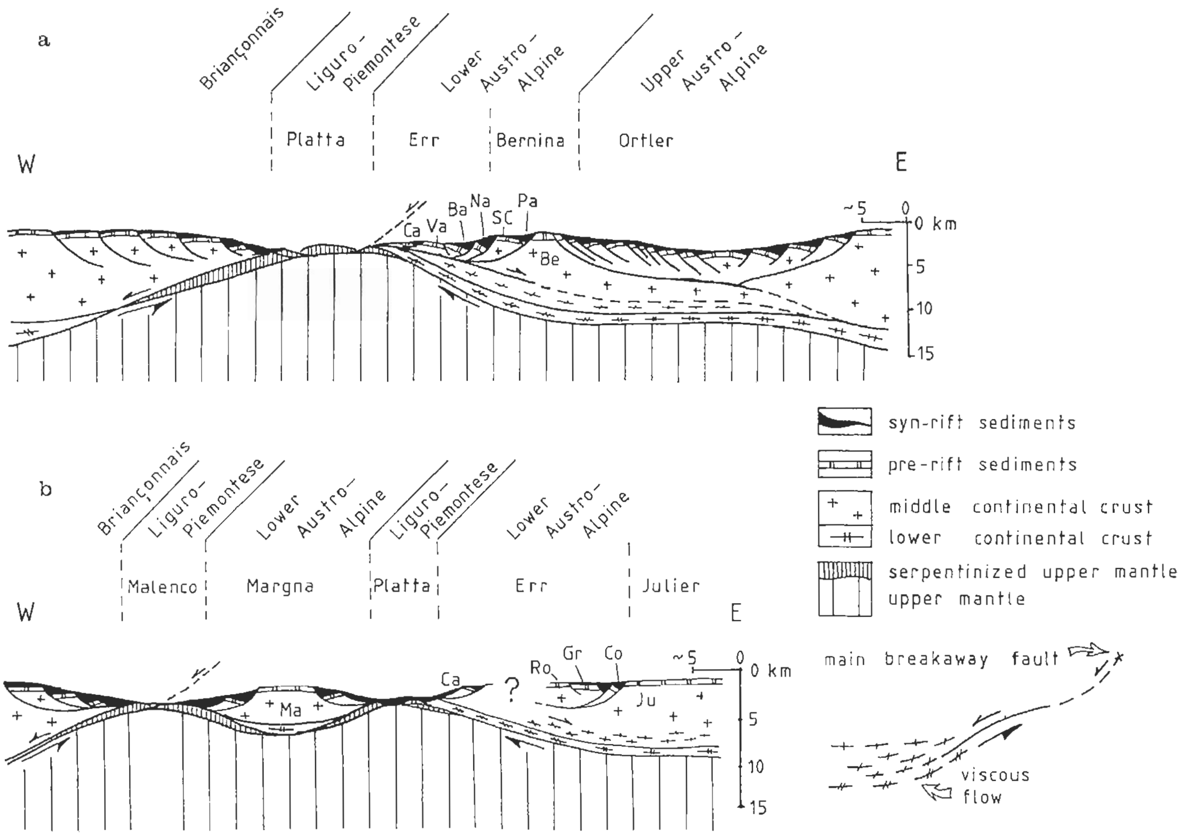


Fig. 10a, b Restored Eþ cross sections across the Adriatic passive margin in the Austro-Alpine domain. Labels as in Fig. 9. Vertical and horizontal scales are only approximate. Trace of sections shown in Fig. 9. **a** Section *a–a′* includes part of the rifted Adriatic margin presently exposed in units north of the Gelgia Valley. The part of the cross section west of the Bernina high is not shown in Fig. 9. Upper crustal structure west of the Bernina high is adapted from Bernoulli et al. (1990), Eberli (1988), and Froitzheim and Eberli (1990). **b** Section *b–b′* includes part of the rifted Adriatic margin presently exposed in units south of the Gelgia Valley

Unfortunately, there is no direct evidence for or against any of these explanations; the amount of D2 extension cannot be quantified and no diagnostic structures of strike-slip faulting (e.g., steep foliations) are observed in the Gelgia Valley. However, indirect evidence for the transform geometry in Fig. 10 comes from changing sediment transport directions determined from slide marks at the base of turbiditic sandstone layers within polymict breccias and sandstones of the Bardella formation (rose diagrams in Fig. 9): Southwestward to westward sliding directions in the stratigraphically lowest beds grade to northward sliding directions in the highest beds. This change in sediment transport direction coincides with a decrease in the proportion of basement components and an increase in the proportion of carbonate components in the stratigraphically highest breccias (Finger 1978). These features indicate that basin subsidence and tectonic relief initiated against normal faults to the east, but became progressively greater along a relative structural high which must have bounded the basin to the south. It is reasonable to conclude that the Grevasalvas units represent the imbricated remains of this former structural high, as drawn in Fig. 9. Based on similar reasoning, the back-rotated sediment transport directions in the northern part of the Saluver formation indicate the former existence of a relative structural high to the east of the Nair basin, depicted in Fig. 9.

In keeping with this transform geometry, the deep structure of the passive margin probably varied along strike, as shown in Fig. 10 for the two cross-sectional traces in Fig. 9. No lower continental crustal rocks are exposed in the Err, Julier, and Bernina basement units, so the deep structure in Fig. 10 is inferred solely from the upper crustal basin geometry reconstructed in Fig. 9. In the north (Fig. 10a) the Bernina outer high separated east-dipping normal faults in more proximal parts to the east from west-dipping normal faults in the distal part of the margin studied here (Eberli 1988; Froitzheim and Eberli 1990). To the south (Fig. 10b) the Margna extensional allochthon is interpreted as a former outer high which separated from the rest of the Austro-Alpine margin during the latter stages of rifting. These relations suggest that total extension in the south probably exceeded that in the north.

Active margin tectonics

The map in Fig. 11 shows how Late Cretaceous D1 deformation may have modified the early Mesozoic configuration of the Zone of Samedan and adjacent units. In this kinematic reconstruction, west- to north- and southwest-directed D1 thrusting and folding is superposed on the generally NE- to SW-trending structural grain inherited from early Mesozoic oblique rifting. The resulting structural complexity as well as the position of the Lower Austro-Alpine units in the apex region of the converging Adriatic, European, and northern Calcereous microplates in Late Cretaceous time renders any reconstruction of the kinematic framework during D1 speculative (see next section). The wide range of D1 movement directions observed in the Zone of Samedan is compatible with a progressive change in the principal shortening direction from ENE–WSW to NW–SE during Late Cretaceous time, as inferred from plate kinematic considerations reviewed below. When combined with the early Mesozoic transform geometry proposed above, such a changing kinematic framework for D1 also provides a complementary mechanism to early Mesozoic arcuation to explain pre-D2 bending of the Zone of Samedan. Seen in paleogeographic map view (Fig. 10), early D1 west- to southwest-directed thrusting of the Julier high folded and imbricated the Nair and Bardella basins to the east; northwest-directed movement of this high during the late stages of D1 rotated the northeastern and southwestern parts of the Nair basin, respectively, clockwise and counterclockwise about vertical axes. Such indentation and rotation would have accentuated any early Mesozoic bend in the Nair basin. The open bends in the Err and Julier basement contours observed in present-day map view (Fig. 5) are possible relics of such late D1 differential rotation. They may have resulted from clockwise rotation of the northern part of the Err basement unit during west- to northwest-directed thrusting on one or more south-dipping oblique, lateral ramps located within the Err basement unit (e.g., fault labelled 3 in Fig. 3). Handy et al.'s (1993) previous view that the open bend in the Julier basement contours is due to hangingwall drag along steep, N–S and ENE–WSW trending, D4 normal faults bordering the Julier basement lid is now considered unlikely because this same bend also affects the Err basement and therefore appears to be more than just a manifestation of local drag.

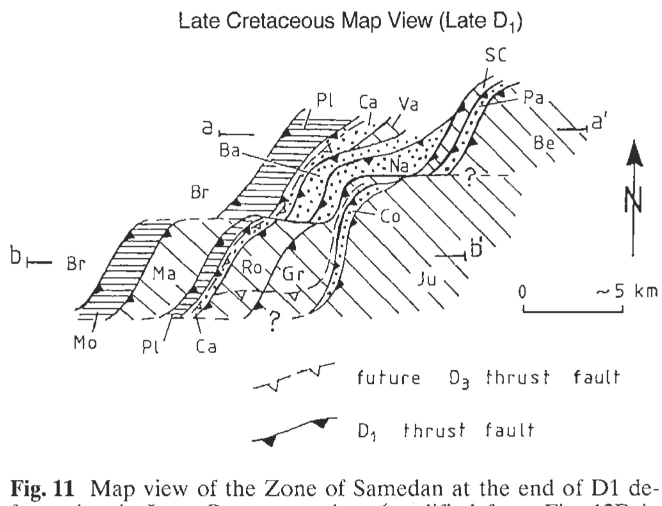


Fig. 11 Map view of the Zone of Samedan at the end of D1 deformation in Late Cretaceous time (modified from Fig. 12B in Handy et al. 1993). Lines *a–a′* and *b–b′* are traces of the cross sections in Figs. 12 and 13. Labels as in Fig. 9

The Late Cretaceous evolution of the Adriatic active margin in the vicinity of the Zone of Samedan is shown in a series of cross-sections in Figs. 12 and 13 whose map traces are labelled a–a′ and b–b′ in Fig. 11. The boxed insets in Figs. 12 and 13 specifically outline the contrasting evolution of structures, respectively, in the northern and southern parts of the Zone of Samedan.

In Fig. 12a imbricates of the early Mesozoic, Julier-Bernina high are shown to have been thrust westward onto the Nair basin during the earliest stage of D1, prior to their incorporation into the core of the F1 syncline in the Nair basin. The source area of these thrust sheets is located in the Bernina nappe, approximately 15 km east–southeast of the Zone of Samedan (Finger 1978). D1 thrusting of these units is inferred to have initiated at or near the surface, as indicated by the fact that the Bernina klippe override Late Cretaceous flysch containing olistoliths derived from the hangingwall (see photos in Figs. 2 and 3 of Finger 1978). Large F1 synclines developed in the Nair and Bardella rift basins, whereas anticlines and thrusts are inferred to have formed in the adjacent basin rims and highs. Shearing continued to depths compatible with high P/T greenschist-facies conditions (Fig. 12b; Handy et al., 1996) and led to the development of the S1 mylonitic foliation observed presently in the axial planes of F1 folds. Late D1 thrusting then separated the normal and overturned limbs of the large, arcuate F1 fold in the Nair basin into two thrust sheets, as seen presently (Fig. 3).

In Fig. 13a the doubly refolded F1 anticline in the southern part of the Zone of Samedan (Fig. 4; recall section D–D′) is interpreted to have originated above the thrust which carried Grevasalvas basement rock onto proximal post-rift sediments. With subduction to greater depths, this anticline became isoclinal and mylonitic in the footwall of the overriding Grevasalvas thrust sheets to the east (Fig. 13b).

D1 thrusts then served as loci for D2 low-angle normal faults (Figs. 12c and 13c). As discussed previously, the two large F2 folds which deform the F1 anticline in the southern Zone of Samedan (section D–D′ in Fig. 4) are interpreted to have formed in a flattening environment between such reactivated shear zones (Fig. 13c). Smaller-scale F2 folds also deform the large F1 synclines in the northern Zone of Samedan shown in Fig. 12c. As Froitzheim (1992) notes, F2 folding under coaxial conditions is only possible if the S1 schistosity attained a steep dip prior to F2 folding. Syn- to post-mylonitic steepening of the D1 structures is attributed to a change in the plate kinematic boundary conditions

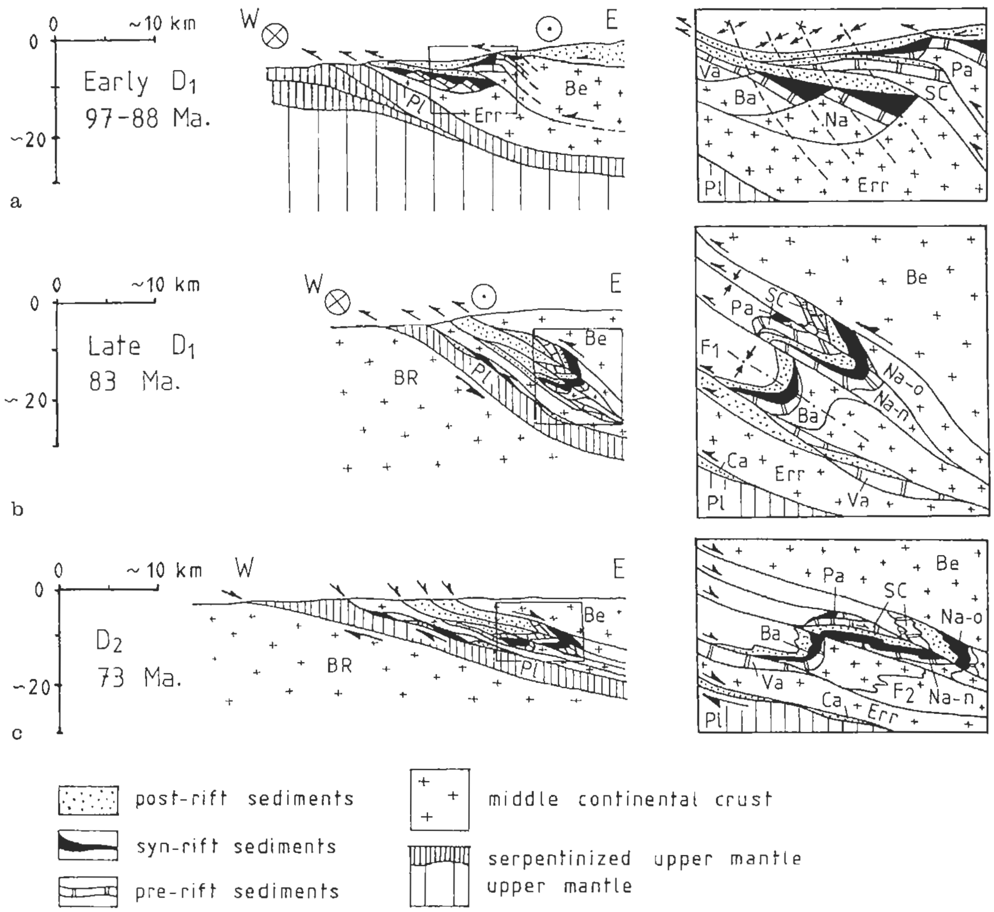


Fig. 12a–c Evolutionary cross sections across part of the Adriatic active margin presently exposed in units north of the Gelgia Valley (*a–a'* in Fig. 11). *Insets* show detail of *boxed areas*. Labels as in Fig. 9

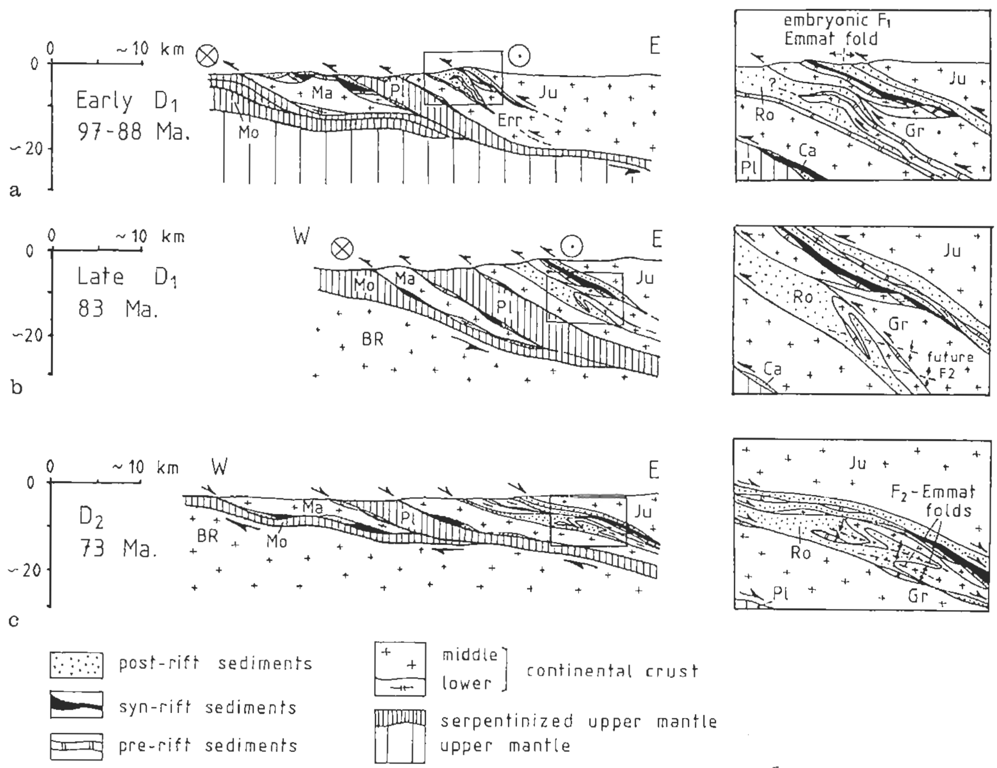


Fig. 13a–c Evolutionary cross sections across part of the Adriatic active margin presently exposed in units south of the Gelgia Valley (*b–b'* in Fig. 11). *Insets* show detail of *boxed areas*. Labels as in Fig. 9

at the end of D1 in Late Cretaceous time (see below, and in Froitzheim et al. 1994; Handy et al. 1993).

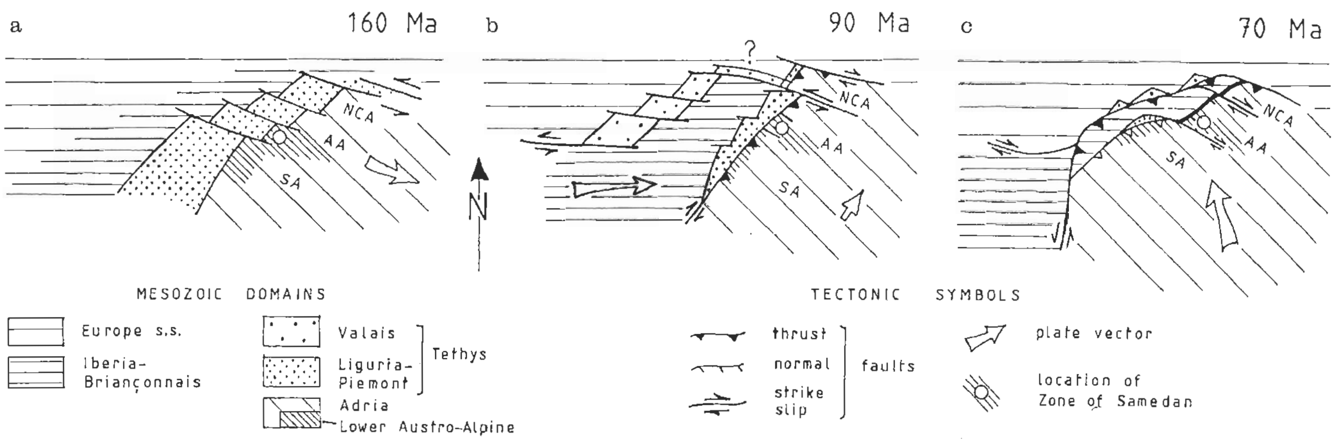
D2 low-angle normal faults in the Zone of Samedan uplifted the Err, Platta, and Margna units with respect to the Julier and Bernina units (Figs. 12c and 13c). The amount of this D2 extensional uplift is poorly constrained; Si-in-phengite geobarometry on chlorite- and biotite-phengite-K-feldspar-quartz-bearing assemblages oriented parallel to S1, S2, and S3 in cover rocks of the Margna (Liniger 1992) and Err basement units (Tietz 1993) bracket the D2-induced pressure drop at between 400 and 600 MPa (Handy et al. 1996). If one assumes an average mid-crustal density of 2.7 g/cm^3, this pressure drop corresponds to 11–16 km of vertical uplift in the footwall of the Julier and Bernina units. Taken together with the average 20° dip of the D2-reactivated D1 nappe contacts, such a pressure drop indicates a total top-to-the-east, dip-slip displacement of approximately 32–58 km! Detailed structural and microfabric studies indicate that most of this uplift was accommodated within the Err and Platta units (Handy et al. 1996). D2 extension started between 80 and 70 Ma (Table 1; K–Ar white mica ages given by Handy et al. 1996), but probably did not continue much beyond 60–65 Ma, the youngest age of Muttekopf-Gosau clastic marine sediments deposited in syn-orogenic, piggyback extensional basins of the Inntal nappe in the northern Calcereous Alps (NCA in Fig. 1; Eisbacher et al. 1991, and references therein).

Relationship of local structures to Mesozoic tectonics along the Adriatic continental margins

The pronounced relief, asymmetrical structure, and elongate, arcuate shape of the reconstructed rift basins in the Zone of Samedan are all diagnostic features of an oblique, pull-apart environment (Christie-Blick and Biddle 1985; Harding et al. 1985). The inferred arcuation and former NE–SW trend of these basins imply a sinistral sense of shear in map view, consistent with large-scale tectonic reconstructions (Fig. 14a) calling for sinistral transform opening of the Liguria-Piemontese Tethys in Early to Middle Jurassic time (Laubscher and Bernoulli 1977; Lemoine et al. 1987; Ricou et al. 1986; Weissert and Bernoulli 1985).

Southwest- to west-directed, early D1 shortening in the Zone of Samedan from 110 to 80 Ma (Table 1) is compatible with regional geological arguments reviewed in Stampfli (1993) for oblique dextral convergence along the NNE- to SSW-trending active margin between Adria and the Briançonnais domain. This margin is represented in the area by the contact of the Err and Platta units (Fig. 2). Such a ENE–WSW shortening direction during the early stages of D1 is also consistent with regional paleomagnetic evidence for sinistral oblique convergence between Adria and Europe from 110 to 80 Ma (Savostin et al. 1986) as well as with Froitzheim et al.'s (1994) local observation that D1, "Trupchun phase" westward thrusting of the Lower Austro-Alpine Err and Ela units (north of the Zone of Samedan in Fig. 2) was concomitant with sinistral displacement along a broad, east–west striking zone of mylonitic strike-slip shear (the "Albula steep zone" marked A in Fig. 2). Thus, early D1 deformation in the Lower Austro-Alpine units is attributed to a shear couple resulting from E–W to NE- to SW-directed, oblique dextral convergence between Adria and the Briançonnais domain to the west and NE- to SW-directed, oblique sinistral convergence between Europe and Adria to the north. This situation is depicted in Fig. 14b, where the Lower Austro-Alpine domain, including the Zone of Samedan, is depicted to have undergone NE–SW shortening in the apex of the Adriatic microplate,

Fig. 14a–c Map showing Mesozoic evolution of Alpine Tethyan margins. **a** Passive margin formation associated with transform opening of the Liguria-Piemontese Ocean (modified from Weissert and Bernoulli 1985); **b** active margin tectonics associated with subduction of the Liguria-Piemontese Ocean and spreading of the Valais Ocean (modified from Frisch 1979, and Stampfli 1993); **c** extensional uplift of the accreted and subducted Austro-Alpine passive margin south and east of the subducting Valais Ocean (modified from Ricou and Siddans 1986). Movement vector of microplates relative to stable Europe from Savostin et al. (1986; see text)



adjacent to the northern Calcereous Alps microplate and Iberian–Briançonnais microplate.

Several plate kinematic factors acting individually or in concert may have induced late D1, NW-directed thrusting, steepened the D1 structures, and ultimately triggered D2 extension in the Zone of Samedan and adjacent Lower Austro-Alpine units along the western margin of Adria. Changes in the direction and rate of convergence between any of the microplates involved in the Late Cretaceous Alpine orogeny (Adria, Europe, and Briançonnais in Fig. 14c) could have induced gravitational instability of the thickened Adriatic lithosphere (Ratschbacher et al. 1989). In particular, the paleomagnetic data reviewed in Savostin et al. (1986) indicate a switch from sinistral to dextral convergence between Europe and Adria approximately 80 Ma ago. This plate kinematic switch probably increased the component of transverse, N–S shortening in the Lower Austro-Alpine domain during the late stages of D1 and is therefore expected to have steepened the F1 and S1 structures in the Lower Austro-Alpine units. The change in convergence direction also coincided with a decrease in the convergence rate from 1.5 to 1 cm/year (Fig. 15a in Savostin et al. 1986), making Platt's (1986) model of an extending orogenic wedge one attractive mechanism to explain D2 extensional uplift within the Late Cretaceous Alpine orogen (Froitzheim et al. 1994; Handy et al. 1993). An alternative explanation for D2 extension is that Late Cretaceous to Early Tertiary southeastward subduction of the Valais oceanic domain (Fig. 14c) involved hinge roll-back towards the northwest, extending and uplifting the overriding Penninic (Briançonnais), Liguria-Piemontese and Austro-Alpine units in the east.

Conclusions

Although the Zone of Samedan makes up only a small part of the former Adriatic continental margin, it contains at least two features which may typify other ancient passive margins preserved within orogens: Firstly, early Mesozoic rifting introduced a complex structural and stratigraphic mosaic which compartmentalized subsequent active margin tectonics. Deep, asymmetrical half-grabens became the sites of synclines, whereas relative structural highs were imbricated to form stacks of thin thrust sheets containing slivers of basement rock. The axial planes of the synclines dip towards the deepest parts of the former rift basins, such that distal and proximal facies of the syn-rift basinal formations are preserved, respectively, in the normal and overturned limbs of the synclines. During the latter stages of shortening, these fold limbs became individual thrust sheets.

Secondly, thrust surfaces which were active during plate convergence were reworked during later episodes of extension and shortening. In particular, late orogenic collapse of the Adriatic active margin was accommodated primarily along a network of continentward-dipping low-angle normal faults which are reactivated thrusts. As shown above, such extensional faults uplifted distal parts of the Adriatic margin, as well as fragments of the Liguria-Piemontese oceanic crust which were accreted during plate convergence. In light of the evidence for Late Cretaceous extensional uplift in the Zone of Samedan, Hsü and Briegel's (1991) hypothesis that the Liguria-Piemontese ophiolites and Margna basement unit were brought to the surface as a result of convergent offscraping within a subduction zone mélange (Hsü and Briegel 1991, pp 139–140) becomes very unlikely (see also discussions by Platt 1992 and Hsü 1992). It is emphasized that D2 extensional shearing did not raise the Platta and Margna rocks all the way to the surface. Instead, it was the combination of Late Cretaceous east–west extensional uplift (D2), early Tertiary northward thrusting along the base of the Platta ophiolites (D3), and early to mid-Tertiary normal faulting (D4) above the uplifting Lepontine dome which ultimately led to their exhumation. The latter two events represent the collisional stage of the Alpine orogeny, during which the Adriatic margin formed a relatively cold and rigid orogenic lid overriding the warmer stack of Penninic nappes derived from the European margin (e.g., Laubscher 1983). Thus, the passive margin features in the Zone of Samedan survived a complex history of repeated crustal thickening and extensional collapse, first within an active margin and then again in the hangingwall of a major thrust within a collisional orogen.

Acknowledgements This paper synthesizes mapping conducted between 1990 and 1993 by the author and five diploma students: Marco Herwegh, Christian Regli, Reto Tietz, Christoph Mattenberger, and Urs Klahr. We enjoyed discussions and field visits with Niko Froitzheim, Ivan Mercolli, Peter Nievergelt, Tjerk Peters, and Adrian Pfiffner. A shorter version of this paper was accepted for publication in "Tectonics" at the end of 1993, but was withdrawn from that journal due to the unfounded request of its former European editor to halve the length of the manuscript. Further delay resulted from M. R. Handy's move to the University of Giessen in 1994. Nevertheless, the paper has benefitted, first from the critique of "Tectonics" reviewers Geoff Milnes, Uwe Ring, and Stefan Schmid, and then from the comments of "Rundschau" reviewers Niko Froitzheim and Lothar Ratschbacher. Finally, I thank Rudolf Trümpy for his perceptive remarks, based as they are on years of experience in the Alps in general and in the Austro-Alpides in particular. The financial support of the Schweizerische Nationalfonds in the form of project grant 2971-0.88 and Profil-2 grant 21-30598.91 is gratefully acknowledged.

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