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Structural history of the Arosa Zone between Platta and Err Nappes east of Marmorera (Grisons): Multi-Phase deformation at the Penninic-Austroalpine Plate boundary

By Sören B. Dürr [1]

ABSTRACT

West of St. Moritz, the Arosa melange zone shows a lithofacial E–W separation, the eastern part being dominated by Austroalpine and the western part by South Penninic rocks.

Three alpine deformational phases D1–D3 are described. During D1 tectonic flow had a noncoaxial component and was top-to-the-west. Incremental strain analysis reveals an early D1 phase of NE–SW stretching and a later period of NW–SE to W–E stretching. D1 is attributed to Cretaceous westward overriding of Austroalpine onto Penninic units. During D2 tectonic transport was top-to-the-south, related to complex Tertiary backfolding. Evidence for a comparable deformation phase has so far been only reported from this Alpine suture further south. During D3 E–W extension with a top-to-the-east sense of shear prevailed, attributed to Late Tertiary uplift and extension in the Lepontine area.

Metamorphic conditions accompanying deformation decrease from D1 to D3.

ZUSAMMENFASSUNG

Westlich von St. Moritz zeigt die Arosa-Mélange-Zone eine lithofazielle E–W-Unterteilung. Während der östliche Teil von ostalpinen Gesteinen dominiert wird, überwiegen südpenninische Gesteine im westlichen Teil.

Drei alpine Deformationsphasen D1–D3 werden beschrieben. Während D1 hatte der tektonische Fluss eine nichtkoaxiale Komponente und war Top-nach-West. Eine Analyse der inkrementellen Deformation ermöglicht die Unterteilung von D1 in eine frühe Phase der NW–SE-Streckung und eine spätere Phase der NW–SE- bis E–W-Streckung. D1 wird dem westgerichteten kretazischen Überfahren von ostalpinen auf penninische Einheiten zugeschrieben. Der tektonische Transport während D2 war Top-nach-Süd. Dies wird mit komplexer tertiärer Rückfaltung in Verbindung gebracht. Diese Deformationsphase ist bisher nur von weiter südlich an der alpinen Sutur dokumentiert. Während D3 herrschte E–W-Extension mit einem Top-nach-E-gerichteten Schersinn vor. Die Ursache für D3 wird in spättertiärer Heraushebung und Extension der Lepontin-Gegend gesehen.

Introduction

The aim of this paper is to add data on the structural evolution of the Penninic/Austroalpine boundary in the eastern Central Alps, represented by the Arosa melange zone. The data result from remapping a part of the Arosa melange zone between the Penninic Platta nappe to the west and the Austroalpine Err nappe to the east (Fig. 1). A simplified geological map of the area is shown in Fig. 2.

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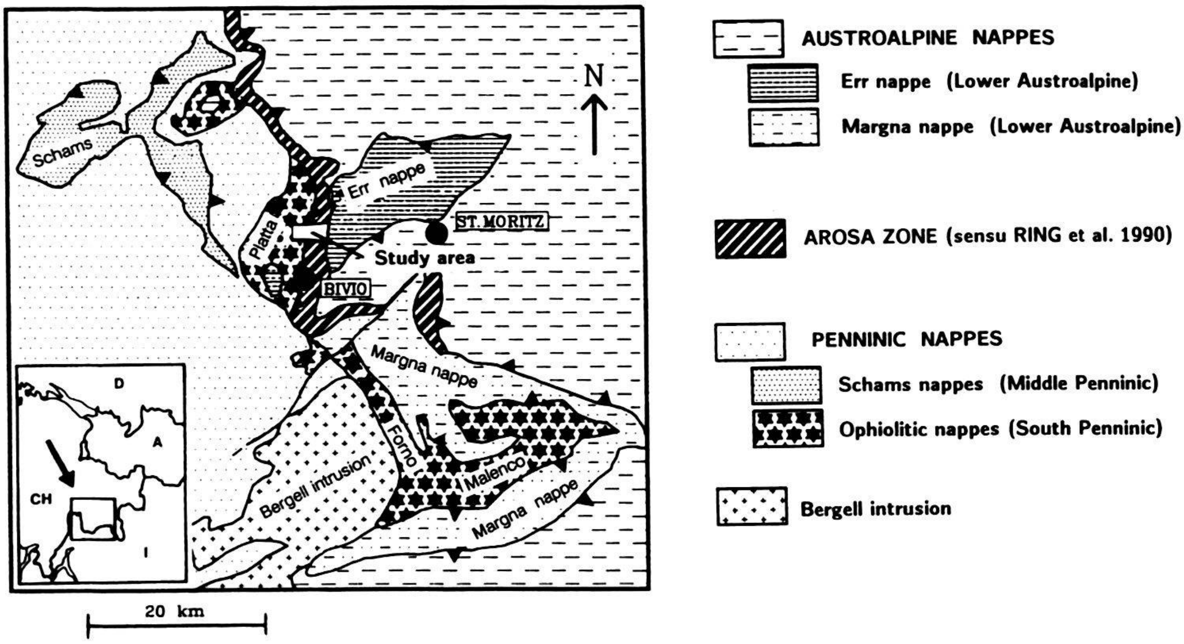

Fig. 1. Tectonic sketch map of SE Switzerland, showing the location of the study area and tectonic units mentioned in the text. Map modified after Cornelius (1932), Weissert & Bernoulli (1985), Gwinner (1978) and Liniger & Guntli (1988).

The Arosa melange zone is part of the main Alpine suture zone along which the South Penninic oceanic plate was subducted underneath the Austroalpine. The latter formed the northern continental margin of the Apulian plate (e.g. Frisch 1979).

The structure of the Arosa zone is of a melange-type (Ring et al. 1990). It is made up of an incompetent, sheared matrix of mostly South Penninic origin (serpentinite and shale), which surrounds rigid blocks of up to km-size that consist of both Austroalpine and Penninic rocks (mostly dolomite, basalt and weakly altered ultramafic rock). The melange formed by two processes (Ring et al. 1989): accretion of the upper portion of the subducting South Penninic oceanic crust and its overlying sediments and simultaneous overthrusting of the overriding plate (Austroalpine). The age of the melange formation is considered to be Cretaceous to Early Tertiary (Ring et al. 1990). The thickness of the roughly N–S trending Arosa zone generally ranges from a few tens to more than one thousand meters. It is underlain by ophiolitic and sedimentary Penninic units and overlain by Austroalpine nappes comprising basement as well as sedimentary and volcanic cover rock. In general, the metamorphic grade increases from north to south along the Arosa zone and adjacent units. Very low grade metamorphosed rocks are found in the Rätikon (Biehler 1990) and as far South as the Platta nappe (Dietrich 1970), medium to upper low grade metamorphism occurs south of Bivio (Trommsdorff & Dietrich 1980, Liniger & Guntli 1988).

Recent research shows Cretaceous top-NW- to top-SW-directed tectonic movement in the Austroalpine and along the Austroalpine/Penninic boundary (e.g. Biehler 1990, Laubscher & Bernoulli 1982, Liniger & Guntli 1988, Ratschbacher 1986, Ratschbacher & Neubauer 1989, Ring et al. 1989, Schmid & Haas 1989). In the lower plate (Penninic domain), Tertiary N-directed thrusting (Ring et al. 1989), followed by large-scale backfolding (Merle & Guillier 1989, Schmid et al. 1990), dominates the structural evolution.

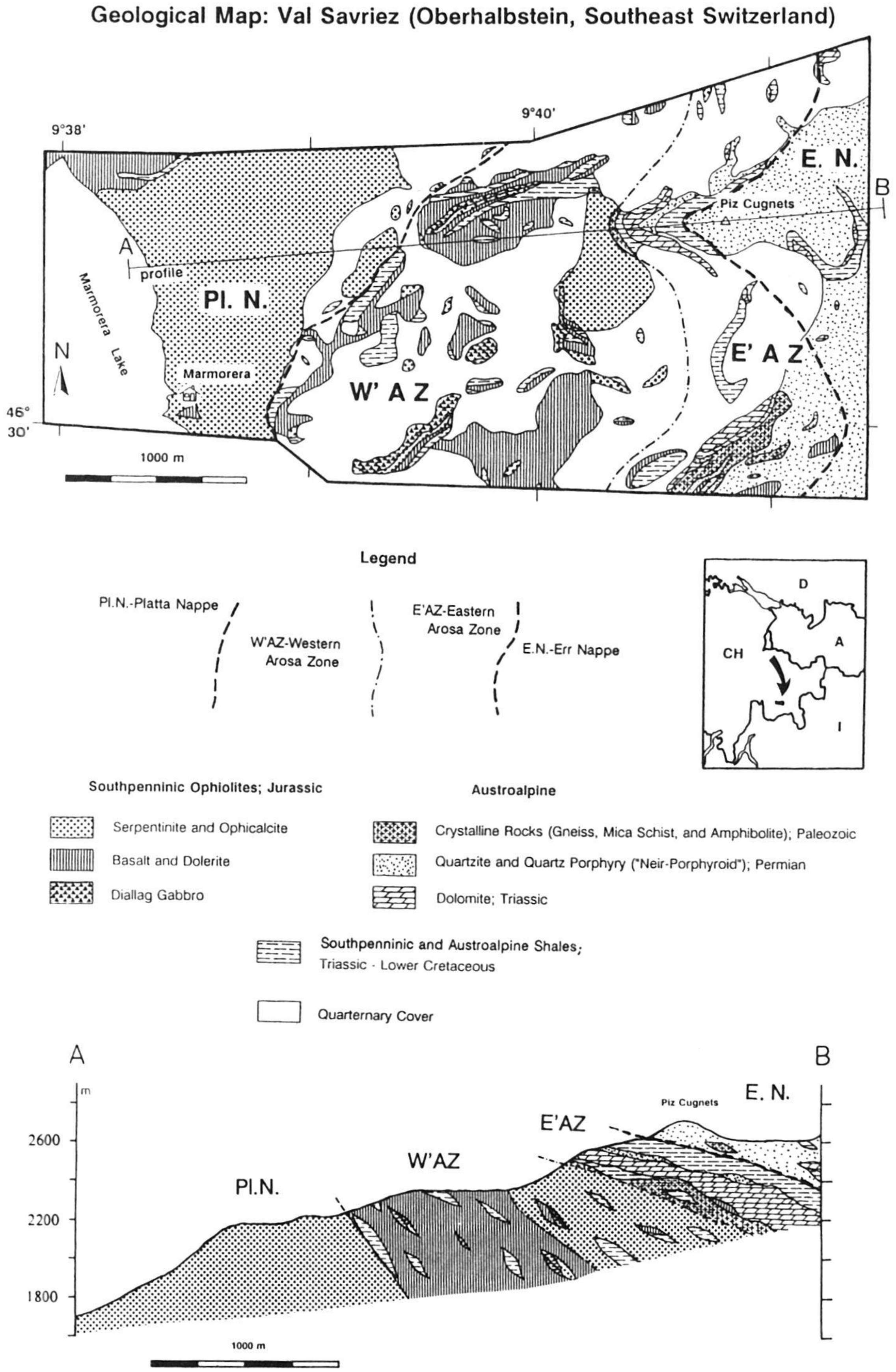

Fig. 2. Slightly simplified geological map and profile of the study area.

In the late Tertiary, an E–W extensional event with a top-E tectonic transport affects the Penninic and Austroalpine nappes of eastern Switzerland (Ring et al. 1991).

Geological setting

In the study area, the Arosa melange zone is situated between the Austroalpine Err nappe to the east and the ophiolitic South Penninic Platta nappe to the west (Fig. 1). The only so far published map of the area is by Cornelius (1932).

Here, the Arosa zone shows a lithofacial E–W separation (Fig. 2). The eastern part consists of little Austroalpine basement, of Permomesozoic Austroalpine sediments and volcanics, Mesozoic Penninic sediments, and rare South Penninic oceanic basement. The boundary to the Err nappe is here defined by the easternmost occurrences of ophiolitic rocks. The western part of the Arosa melange zone is made up of ophiolitic, South Penninic material, mostly basalt, ultramafic rock and gabbro, minor sediments of Penninic origin, and no distinct Austroalpine elements. The boundary to the Platta nappe is marked by a near cessation of melange structures and a marked decrease in the amount of sedimentary rocks (Fig. 2).

The Err nappe consists of Variscan basement rocks, Permotriassic clastic and volcanic rock ("Neir-Porphyroid"), Triassic dolomite, limestone and shale, Jurassic and Cretaceous limestone, shale, arenitic layers and radiolarian chert. The Platta nappe comprises Jurassic ophiolitic material (mostly serpentinite and basalt), Upper Jurassic radiolarian chert and shale, Upper Jurassic and Lower Cretaceous pelagic limestone as well as Cretaceous shale and arenites (Dietrich 1970), the latter partly as turbiditic layers (Weissert & Bernoulli 1985). Following Ring et al. (1990), I do not consider the Platta nappe and Arosa zone as structural equivalents as proposed by other authors (e.g. Weissert & Bernoulli 1985 or "Platta Zone" of Liniger & Nievergelt 1990), but as separate tectonic units since the Platta nappe lacks melange structures and, with few and partly ambiguous exceptions (see Dietrich 1970), also lacks elements other than such of Penninic origin.

Ductile rock flow affected, to varying degrees, all rocks except for dolomite and weakly altered ultramafic rock. Rock slices of different lithologies and meter- to kilometer-sizes were welded together to form a chaotic picture (cf. Fig. 2). Although the three-dimensional geometry of single rock slabs remains unclear, many appear to dip east (just west of Piz Cugnets) or south (central and SW part of the study area).

Metamorphism in the study area is transitional between very low grade and low grade. This transition is defined by the mineral assemblage albite, chlorite, actinolite, epidote, pumpellyite, calcite and sphene in metabasaltic rocks (Dietrich et al. 1974, Trommsdorf & Dietrich 1980).

Structural history

Three major deformation phases D1 to D3 are distinguished. The distinction was made by cross-cutting relationships of deformation structures: during D1 the main foliation developed, which is folded by D2 folds. Finally, D3 structures cut and therefore postdate D1 and D2 structures. Additionally, a decrease in metamorphic conditions from D1 to D3 helps to distinguish the deformation phases. This is visible in thin

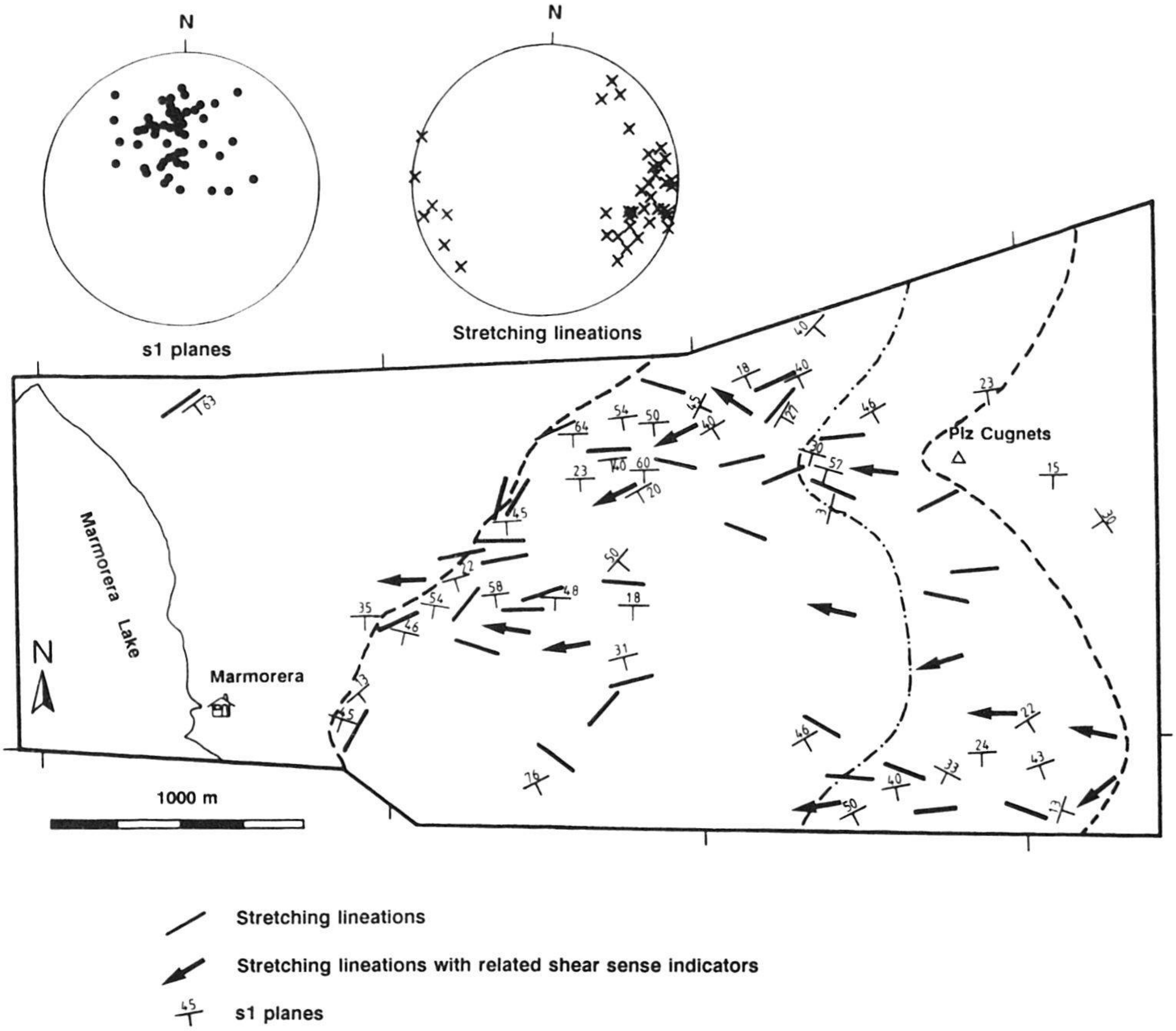

Fig. 3. D 1. Structural map of the study area and stereographic projections (lower hemisphere) of S 1 planes and stretching lineations. Pointing directions of arrows (shear sense indicators) give directions of shear.

sections from the way different minerals behaved under stress (e.g. White 1976). For example, calcite recrystallizes at lower temperatures than quartz. Also, the grade of ductility decreases from D 1 to D 3. While during D 1 incompetent rocks like shale or highly altered ultramafics were deformed in a ductile manner, D 2 structures show deformation of these rocks at the ductile/brittle transition. During D 3, all rocks in the study area were deformed brittly.

First deformation (D 1)

The main D 1 deformation caused isoclinal folding (F 1) of bedding planes and the formation of an axial plane foliation (S 1) in incompetent rocks (Figs. 3, 4 a).

The foliation is penetrative, the folds only rarely visible. S 1 surfaces are closely spaced and defined by fine-grained phyllosilicates and opaque minerals. A NW–SE to WSW–ENE oriented stretching lineation is frequently developed on S 1 planes. It is

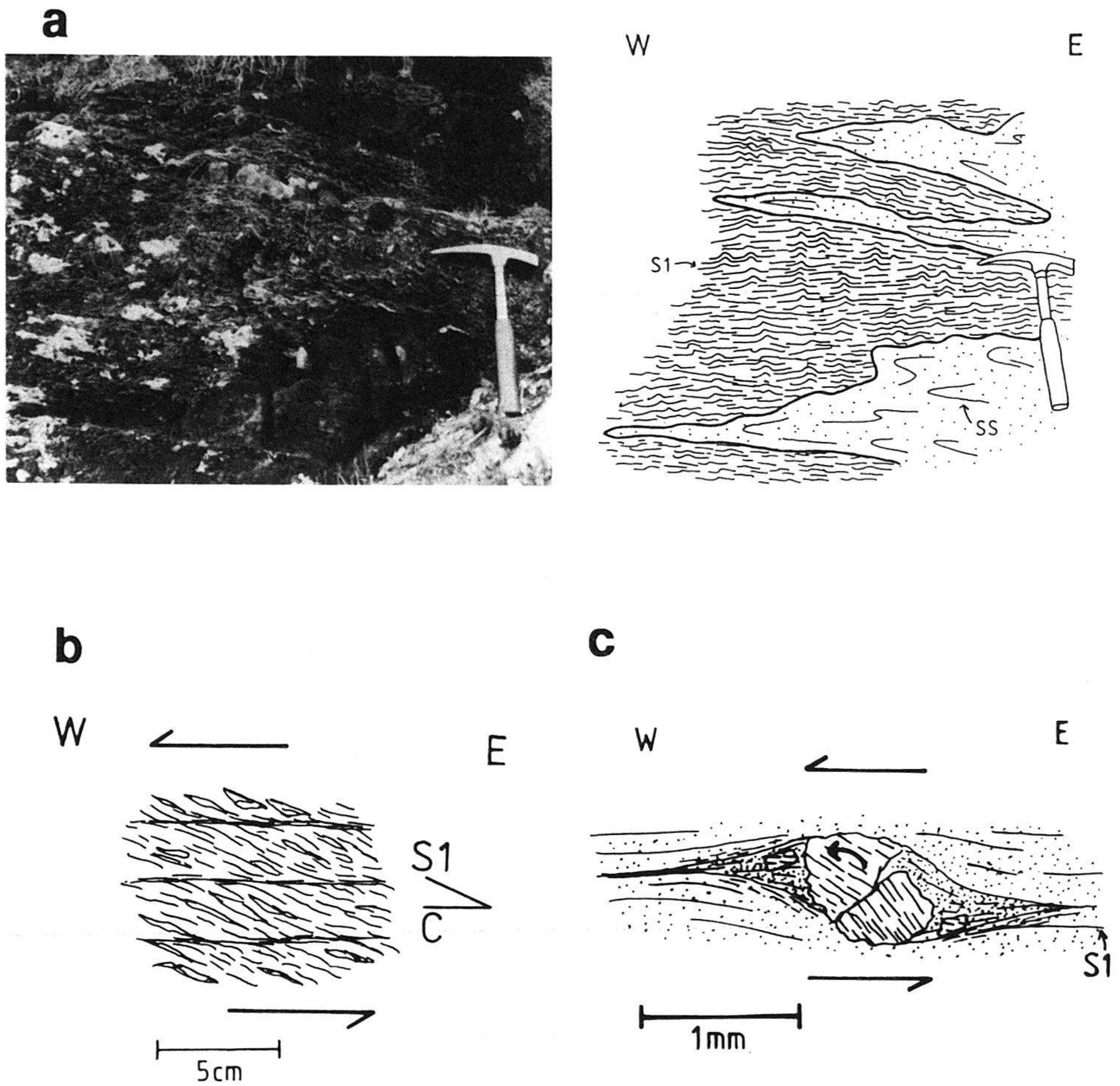

Fig. 4. D 1 deformational structures. **a**: Isoclinally folded (F 1) bedding plane between carbonate (stippled) and radiolaritic shale. Carbonate shows folded internal bedding (SS; thin sandlayers). More incompetent shale shows axial plane S 1 cleavage, crenulated by D 2. 500 m ESE Piz Cugnets. **b**: S–C structure in Ophicalcite. 1 km NE Marmorera. **c**: Asymmetric pressure shadows (mainly quartz and sericite) around a broken and partly rotated Amphibole in porphyroid. Near Piz Cugnets.

formed by elongated minerals, fossils (radiolarians?) and varioles, and by pressure shadows around mineral clasts. S–C structures, extensional crenulation cleavage, en-echelon-veins and asymmetrical pressure shadows show a top-to-west sense of shear during D 1 (Figs. 3, 4 b, c). Competent rocks show roughly N–S striking extension veins. The thickness of these does not exceed a few cm. The veins are filled by NE–SW to NW–SE oriented calcite, quartz or serpentine fibers.

Mineral textures are in accordance with the thermal regime during peak metamorphic conditions: feldspar and dolomite crystals do not show intracrystalline deformation but are deformed in a brittle manner. Calcite shows complete recrystallization and also static grain growth. Quartz exhibits sutured grain boundaries, dynamic as well as static recrystallization, and rare static grain growth.

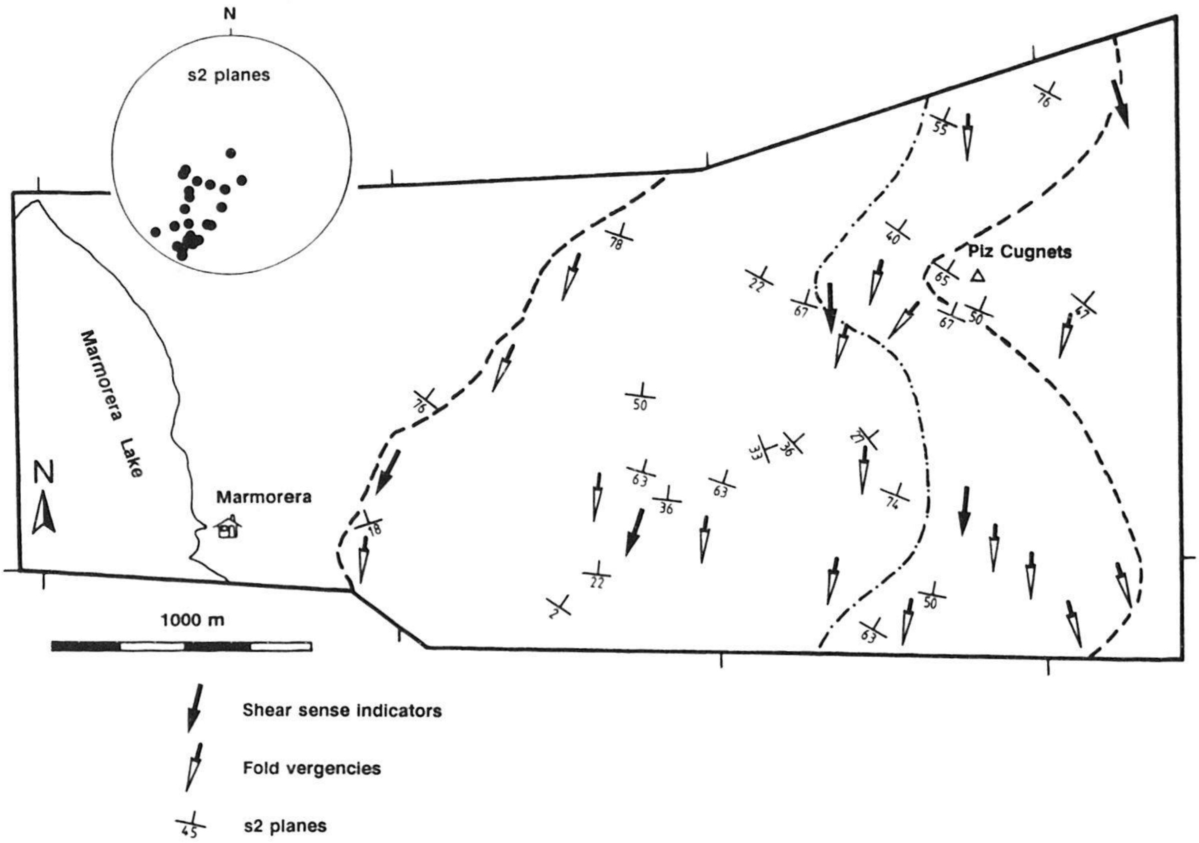

Fig. 5. D 2. Structural map of the study area, stereographic projection (lower hemisphere) of S 2 planes. Pointing directions of filled and open arrows give directions of shear and fold vergencies, respectively.

Second deformation (D 2)

The most prominent D 2-structures are crenulations and up to m-sized S- to SSW-vergent folds (F 2) in incompetent rocks (Figs. 5, 6 a, c). The folds are classified as open parallel to similar folds (class 1 C; Ramsay 1967, Ramsay & Huber 1987), the interlimb angles not falling short of 70°. The fold axes strike E–W to ESE–WNW and the axial planes dip N to NNE with 30–60°, indicating top-south shear.

A widely spaced axial plane cleavage (S 2) dips around NNE with an average of 45° (Figs. 5, 6 a). Other shear sense indicators that can be assigned to D 2 are rare and restricted to localities where shear zones developed due to contrasting lithologies (Fig. 6 b). They include extensional crenulation cleavages, S–C structures and rotated clasts (Figs. 5, 6 b). Shear senses determined at such locations must, nevertheless, be viewed with caution, since shear senses in shear zones around rigid blocks may differ from bulk shear sense.

The southward dip of the S 1 planes is attributed to a south-directed D 2 rotation. Northward backrotation of S 1 planes around an average F 2 axis (110° strike) results in a shallow pre-D 2 dip of S 1 planes towards east. Very rarely D 2 extension veins filled by N–S oriented fibrous quartz, calcite and serpentine are developed in competent rocks. In the same outcrop, D 2 veins are narrower than D 1 veins. D 2 deformation was less intense than D 1: D 2 structures are only locally developed and mostly affect very incom-

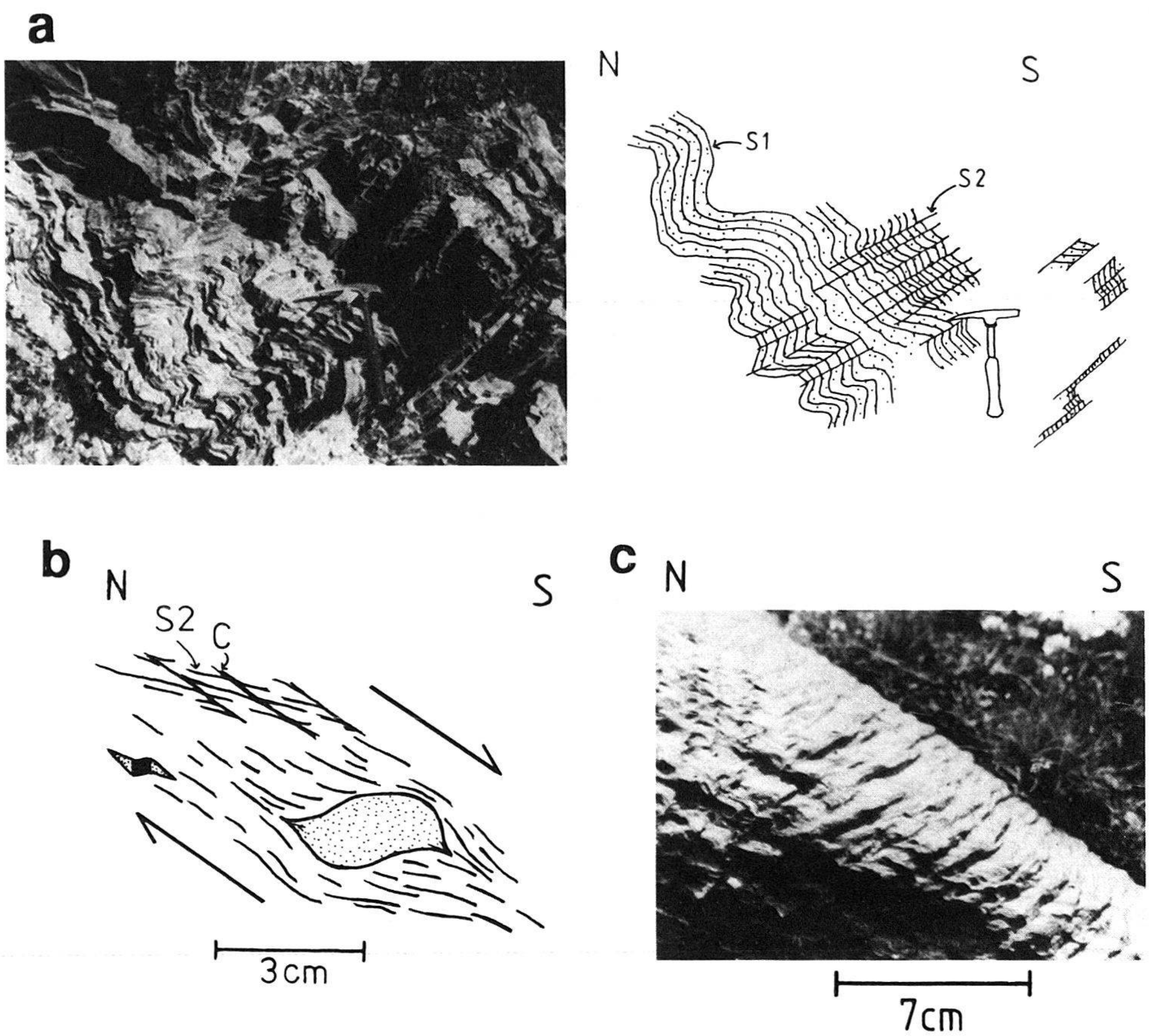

Fig. 6. D 2 deformational structures. **a**: F 2 folds in calcareous shale. S 2 planes begin to develop at fold hinges. 800 m SSW Piz Cugnets. **b**: D 2 shear sense indicators in strongly sheared shale under rigid dolomite block: S–C structure, asymmetric calcite pressure shadows around pyrite crystal, calcite Sigma clast. 500 m W Piz Cugnets. **c**: Crenulated clay-rich radiolaritic shale. 1 km NNE of Marmorera.

petent rocks such as calcareous and mica-rich shales. S 2 is not a penetrative foliation. Additionally, the steeper angle of dip of S 2 planes as opposed to the shallow pre-D 2 dip of S 1 planes indicates a lower strain for D 2 (Ramsay & Graham 1970). Temperatures during D 2 were lower than during D 1. While calcite deformed ductilely, quartz behaved in a rigid manner.

Third deformation (D 3)

During D 3 E–W extension was dominant. Deformation structures are E-dipping dip-slip normal faults ranging from microscopic to m-size (Fig. 7). The maximum amount of offset observed is about 0.5 m. The sense of movement on the fault surfaces is top-to-E. In two cases, extension gashes of up to 10 cm width were observed in dolomite and basalt. The gashes are vertical, not filled, and strike N–S, showing E–W

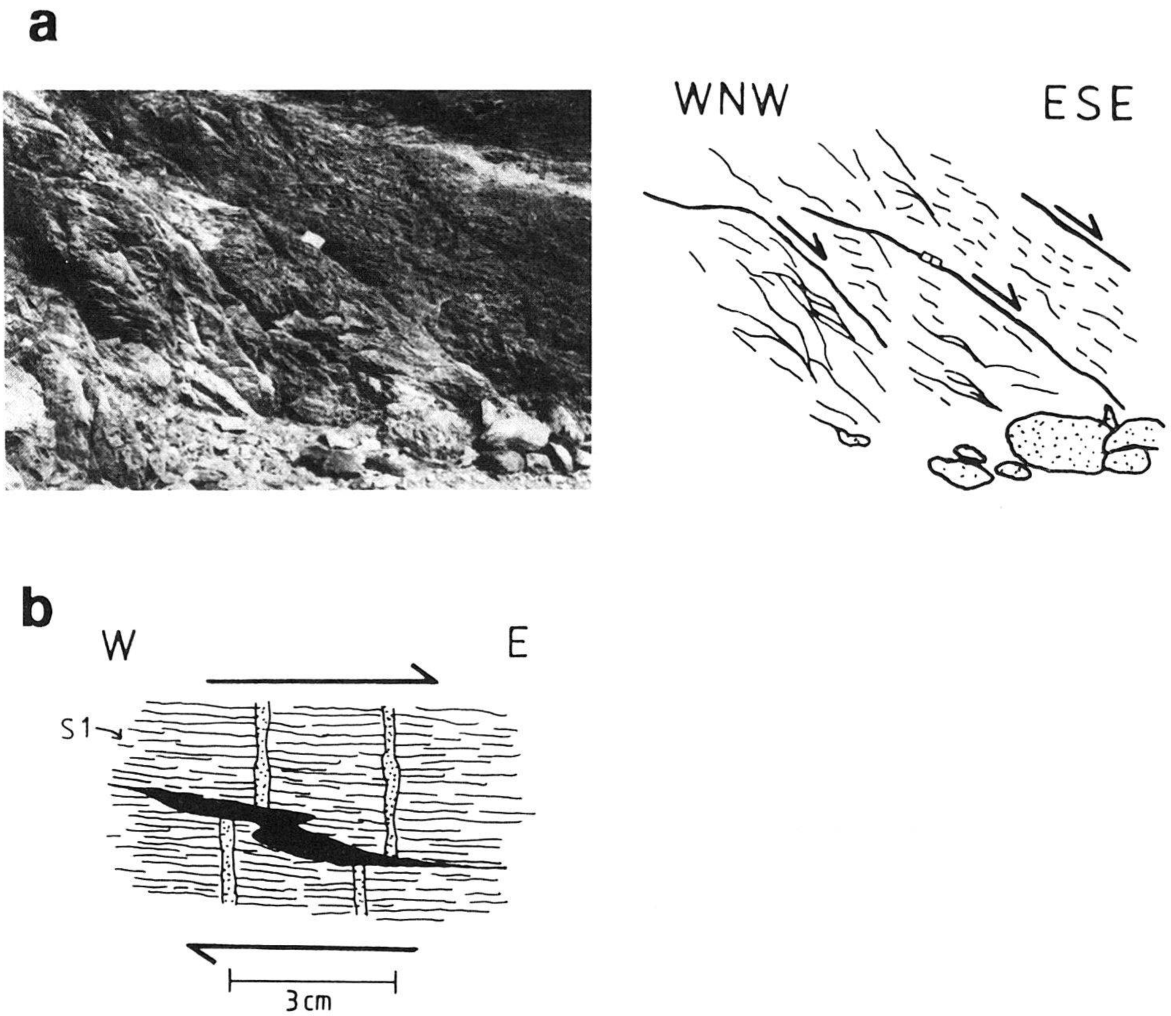

Fig. 7. D 3 deformational structures. **a**: D 3 normal faults in strongly sheared serpentinite. Shear sense is determined by shear fibre steps on major fault surfaces (marked by thick lines). Compass in picture centre for scale. Lakeside at Marmorera. **b**: D 3 extension vein (unhealed) in calcareous shale offsetting calcite filled D 1 veins. 1 km WNW Piz Cugnets.

extension. Temperatures during D 3 were considerably lower than during D 2. D 3 structures show only brittle fracture and no F 3 folds were found.

Incremental and Finite strain

Incremental strain paths are deduced from veins filled with fibrous quartz, calcite (Fig. 8 b) and serpentine as well as from fibrous pressure shadows around mineral clasts. Extensional veins were mostly found in competent rocks such as dolomite and weakly altered ultramafic rock. Age relations were determined by cross-cutting relationships of veins or from the growth sense of curved fibers in veins or fibrous pressure shadows (after Ramsay & Huber 1983). Three major extension periods were thus determined (Fig. 8 a). During the first, roughly NE–SW extension prevailed. NW–SE to W–E extension defines the second period. The amount of extension was larger during the second period; this is obvious both from cases where fibres in pressure shadows were measured and

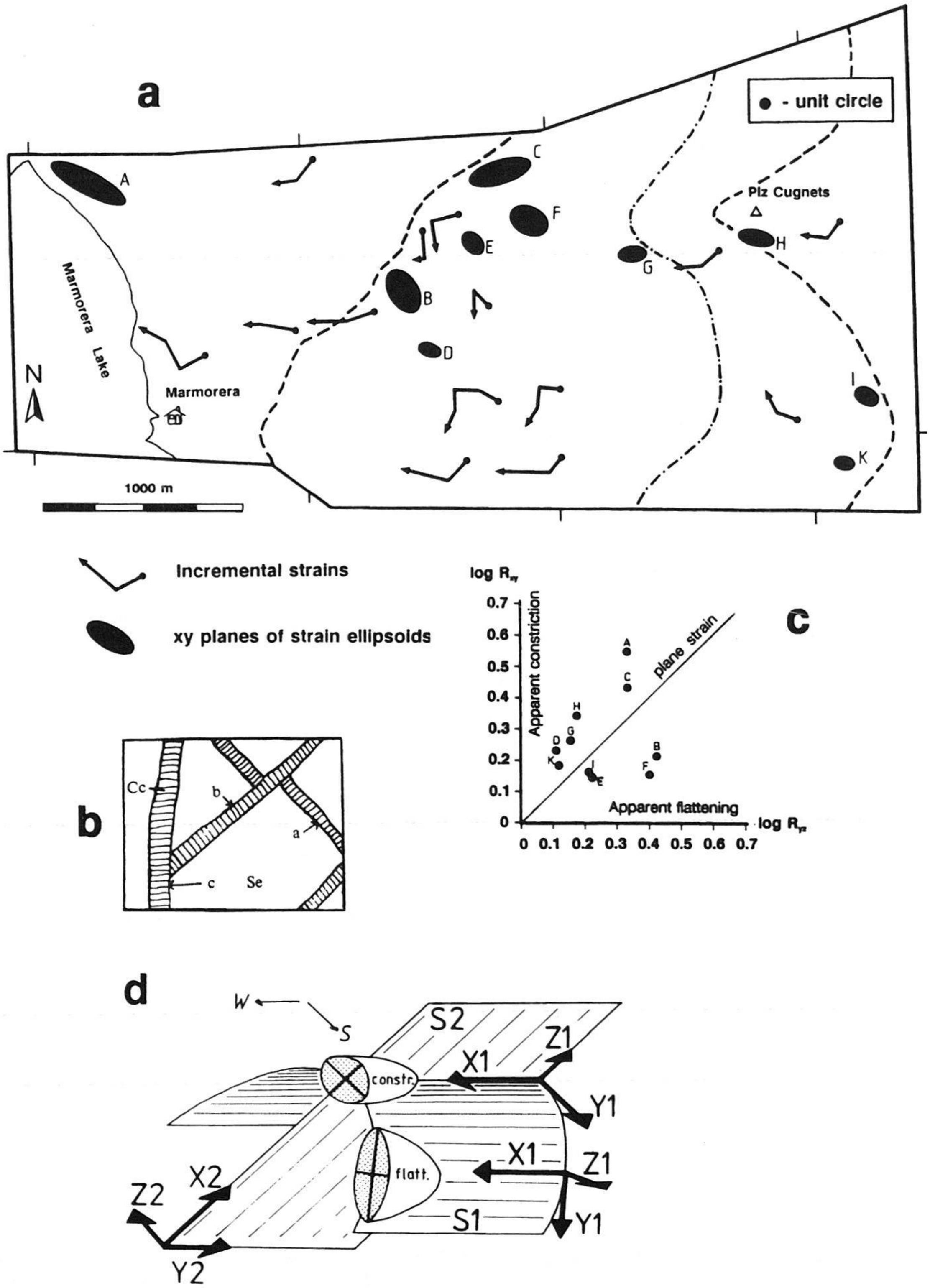

Fig. 8. **a**: Structural map showing finite strain data (XY projection of strain ellipsoids, normalized to unit circle = Z) and incremental strain arrows. Arrow heads point toward younger increments. **b**: Example of progressive sequence of carbonate (Cc) vein formation (arrow N Marmorera) in serpentinite (Se). a, b, c: different generations of fibres; a is oldest, c is youngest. **c**: Flinn diagram shows 2 groups of strain ellipsoids: flattened and slightly constrictional. **d**: Interpretation of both flattened and constrictional strain geometries by D1 – D2 superposition. X1, Y1, Z1 and X2, Y2, Z2 are the principal strain axes during D1 and D2, respectively (X > Y > Z); foliations are the XY planes. D1 strain > D2 strain. In F2 fold hinges, S1 and S2 are perpendicularly oriented and the superposition causes constrictional finite strain geometry; on F2 limbs, S1 and S2 are subparallel and flattened strain geometry results.

where different generations of veins crosscut. N–S extension characterizes the third period and only occurred in the central part of the study area (Fig. 8a).

Due to largely varying vein widths no absolute extension values were calculated. According to the extension orientations and their succession, I assign the first and second increments to D1, the third one to D2.

Finite strain ellipsoids (Fig. 8a) were determined using deformed varioles and pillow outlines in basalt, deformed quartz phenocrysts in porphyroid, and dark and light coloured mineral aggregates in doleritic basalt as strain markers. For all markers, the R_f-Phi method (Dunnet 1969) was applied.

Except for B, E and F on Fig. 8a, the ellipsoid X-axes are oriented E–W to ESE–WNW. This roughly agrees with the first and second strain increments (D1 stretching direction). Ellipsoids B, E and F show a NW–SE orientation of their X-axes.

The Flinn diagram (Fig. 8c; Flinn 1962) depicts the geometry of the ellipsoids. Except for B, E, F and I, the ellipsoids show a slightly constrictional geometry. For B, E, F, and to a lesser degree I, the geometry is flattened.

With a model proposed by Ratschbacher & Oertel (1987) and Ring et al. (1989), both geometries can be interpreted as results of overprinting of a weaker D2 strain upon stronger D1 strain (Fig. 8d). F2 folding of S1 planes will result in perpendicular orientation of S1 and S2 planes at F2 hinges, and subparallel orientation on F2 limbs. Taking the foliations as XY planes with X1 and X2 being the stretching directions (nearly orthogonal), a constrictional finite strain geometry will result at F2 hinges, where the D2 stretching direction is parallel to the D1 direction of least extension. On F2 limbs a flattened geometry results, because S1 and S2 are subparallel and so are the directions of least extension for both deformations. According to this interpretation, the flattened strain geometries require a coaxial component for D2, because superposition of purely non-coaxial strains with a single shear plane results in total plane strain.

Discussion and Conclusions

The structural history determined is not in accordance with the structural history recorded from the Arosa zone to the north (e.g. Biehler 1990, Ring et al. 1988, 1989). During the major, penetrative deformational phase D1, tectonic flow had a noncoaxial component, deduced from E–W stretching and shear sense indicators showing top-to-the-west tectonic movement. Due to large similarities in structures and kinematics, D1 is interpreted here to be the equivalent of the major deformation reported from the northern Arosa zone (Ring et al. 1989, 1990) and the western part of the eastern Alps (Ratschbacher 1986). This deformation, dated around 110–60 Ma by various means (Deutsch 1983, Lüdin 1987, Phillip 1982, Ring et al. 1988, 1989, Weissert & Bernoulli 1985, Winkler 1987), has been attributed to westward overriding of Austroalpine onto Penninic units, obliquely to the Austroalpine continental margin (Ratschbacher 1986, Ring et al. 1988).

D2 was not a penetrative deformation. It is most pronounced in the central part of the study area, where it was responsible for N–S extension and had the most pronounced influence on finite strain geometry in the study area (cf Fig. 8a). The northward dip of F2 axial planes and shear sense indicators suggest a noncoaxial component for D2 with a top-to-the-south tectonic displacement. A comparable deformation phase has so far only been reported from further south at the Penninic-Austroalpine boundary in the central Alps (Liniger & Guntli 1988, Liniger & Nievergelt 1990). From the Arosa zone and underlying Penninic units further north, an increasing prevailance of second-phase north-directed over first-phase west-directed tectonic displacement towards deeper parts of the rock pile (i.e. towards the west) is reported (Ring et al. 1988, 1989). This cannot be observed in the study area, but just north of it (on Alp Flix, Borchert 1991), and also further south in the Arosa zone (near Bivio, Dörre 1989, Knaus 1991).

Schmid et al. (1990) report southward tectonic motion from a stack of Penninic units including the eastern part of the sedimentary Schams nappes. They attribute this deformation to refolding of viscous Penninic units under a rigid orogenic lid of previously emplaced Austroalpine units but including also the Platta nappe. This refolding happened when additional, postcollisional shortening occured between the Penninic realm and the Apulian plate during Late Oligocene to Neogene time, resulting in a huge north-closing recumbent fold with opposite senses of shear in the normal and overturned limbs, with the upper, inverted limb revealing top-to-south shear (see Schmid et al. 1990, for more details). A base of this model is a silicone putty experiment (Merle & Guillier 1989; cf. Fig. 9). Positioning the study area in one of the zones of southward shear which penetrate the rigid orogenic lid of Schmid et al. (1990) (Fig. 9), explains the observed

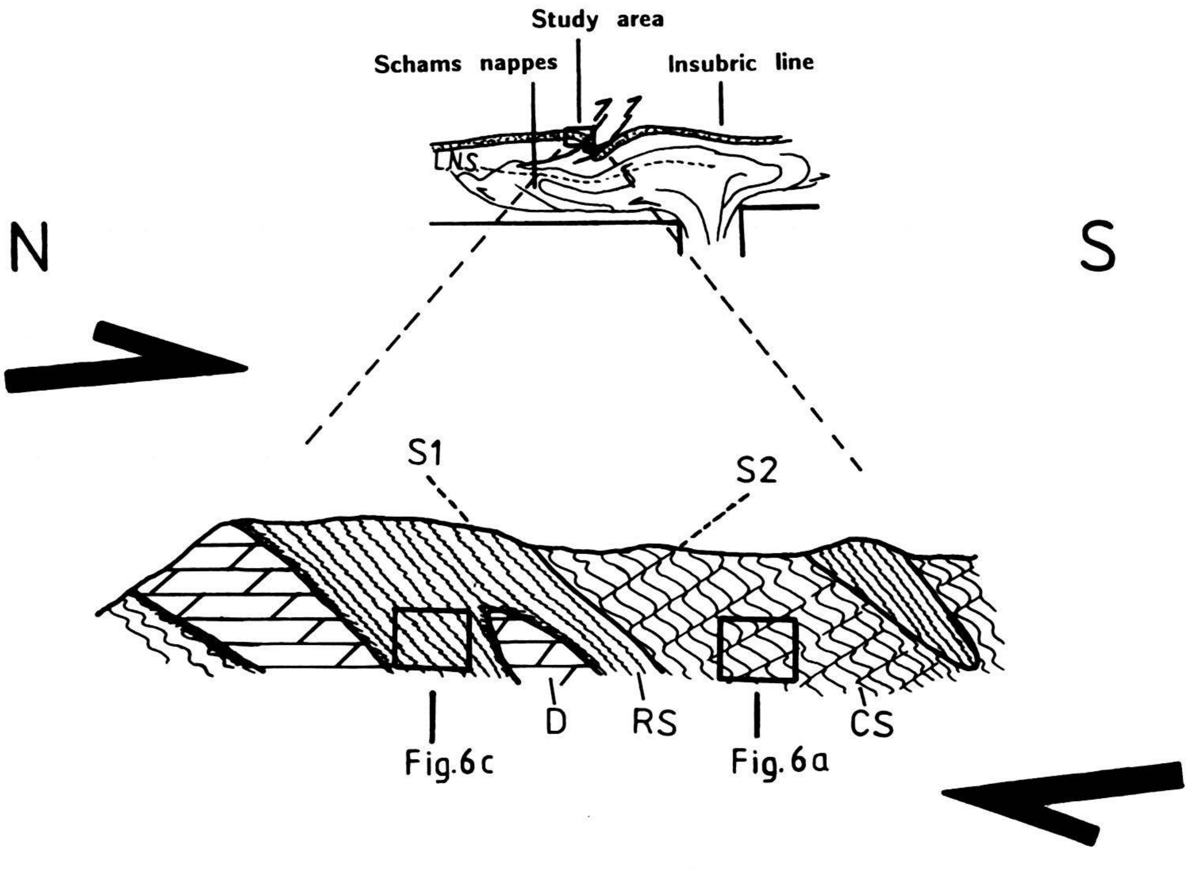

Fig. 9. Model explaining formation of D2. Sketch at top: silicone putty experiment from Merle & Guillier (1989). Study area is positioned at shear zone cutting through "orogenic lid" (stippled). L.N.S. = Line of No Shear. Lower sketch: Schematic N–S cross section through the study area, D = dolomite, RS = radiolaritic shale, CS = calcareous shale. S-ward thrust causes folding and crenulation of incompetent rock (shale); widely spaced cleavage S2 may form.

scattered occurrence of D 2 structures in and near the Arosa zone. The rigid orogenic lid is then represented by the Platta nappe. South-vergent folds with north-dipping axial planes are reported from and near the Austroalpine Margna nappe (Liniger & Guntli 1988, Liniger & Nievergelt 1990). This may mark another place of penetration of a shear zone with southward displacement. These authors argue in favour of a pre-Bergell age (60–30 Ma) for this folding phase by correlating it with folds in the Monte-del-Forno complex that are cut by the Bergell intrusion (Peretti 1985). The folds reported by Peretti reveal south-dipping axial planes; therefore this correlation and the following age for the S-vergent folds in the Margna nappe seem doubtful.

During the third deformational phase D3, structures caused by E–W extension developed with a top-to-east tectonic movement. Equivalent structures have also been found in the northern Arosa zone, the Platta nappe and lowermost parts of the Austroalpine thrust system (Ring et al. 1991). Ring et al. (1991) also report related compressional structures (folds) and conclude a gravitational origin for the coeval development of extensional and compressional deformation features. They relate this deformation to Late Tertiary uplift and tangential stretch in the Lepontine area, according to a model of Merle et al. (1989).

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