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Structural development of high-pressure metamorphic rocks on Syros island (Cyclades, Greece)

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Abstract

New data obtained during re-mapping of the northern part of Syros (Cyclades, Greece) allow the refinement of existing concepts about the structural evolution of this island. A metabasite belt and its metasedimentary envelope near Kampos and San Michali were mapped at a 1:5000 scale, with special emphasis on the structural inventory. The HP/LT rocks ($P \sim 1.5$ GPa; $T \sim 500$ °C at ~50 Ma) exposed are intensely deformed by at least two isoclinal folding events. Relative age relations between deformation and peak metamorphism indicate that isoclinal folding took place before or during peak metamorphism. Later deformation stages include Miocene non-penetrative upright kink folding and crenulation, transpressional strike-slip faulting, and open upright cylindrical folding, followed by the development of steep normal faults. Taking into account recent zircon U–Pb geochronological constraints, lithostratigraphic observations, and published data on the Neogene structural evolution of the Aegean, we present a history for the rocks of Syros, beginning with the formation of the oceanic crust (represented by the metabasites) to the present.

Keywords: Cyclades; Syros Island; High-pressure metamorphism; Tectonic history; Ductile deformation; Brittle deformation

1. Introduction

The island of Syros is famous for its blueschist- and eclogite-facies rocks and is one of the best-preserved areas in Europe to study subduction zone processes. Syros consists of a series of stacked tectonic nappes that include a greenschist-facies basement, one or several metasedimentary marble–schist piles, and a metabasite unit (Fig. 1). The latter includes lithologies typical of a magmatically and hydrothermally active oceanic crust, such as Mn–Fe-rich hydrothermal metasediments and relict pillow lava structures. During the Eocene, all rocks were subjected to high-pressure blueschist–eclogite facies metamorphism that reached peak conditions around 1.5 GPa and 500 °C (e.g. Dixon, 1976; Ridley, 1982a; Okrusch and Bröcker, 1990). High-pressure metamorphism was followed in the Miocene by selective to pervasive greenschist-facies overprint.

The protolith age of the oceanic crust relics in northern Syros is constrained by U–Pb zircon geochronology from meta-igneous rocks to about 80–75 Ma (Keay, 1998; Tomaschek et al., 2003). Geochemical evidence suggests that the protoliths represent oceanic crust that formed in a back arc basin at a convergent plate margin of the Cretaceous Tethys (Kötz, 1989; Seck et al., 1996; Lagos et al., 2003). Note though that other authors interpret similar zircons as metamorphic (Bröcker and Enders, 1999, 2001). The high pressure peak metamorphism is dated at approximately 50 Ma, constrained by Ar–Ar dating of white mica, U–Pb ages on metamorphic zircon, and Lu–Hf dating of the peak paragenesis garnet–omphacite (Maluski et al., 1987; Baldwin, 1996; Lagos et al., 2003; Tomaschek et al., 2003). An age for the peak metamorphism of about 50 Ma seems to be commonplace throughout the Cyclades (e.g. Wijbrans et al., 1993; Ring and Layer, 2003), supporting the notion that the whole Cycladic Unit represents a coherent continental fragment that experienced an overall identical structural and metamorphic evolution.

A vital aspect in understanding the development of these rocks is their tectonic history. The structures on Syros are a result of polyphase ductile deformation, overprinted by

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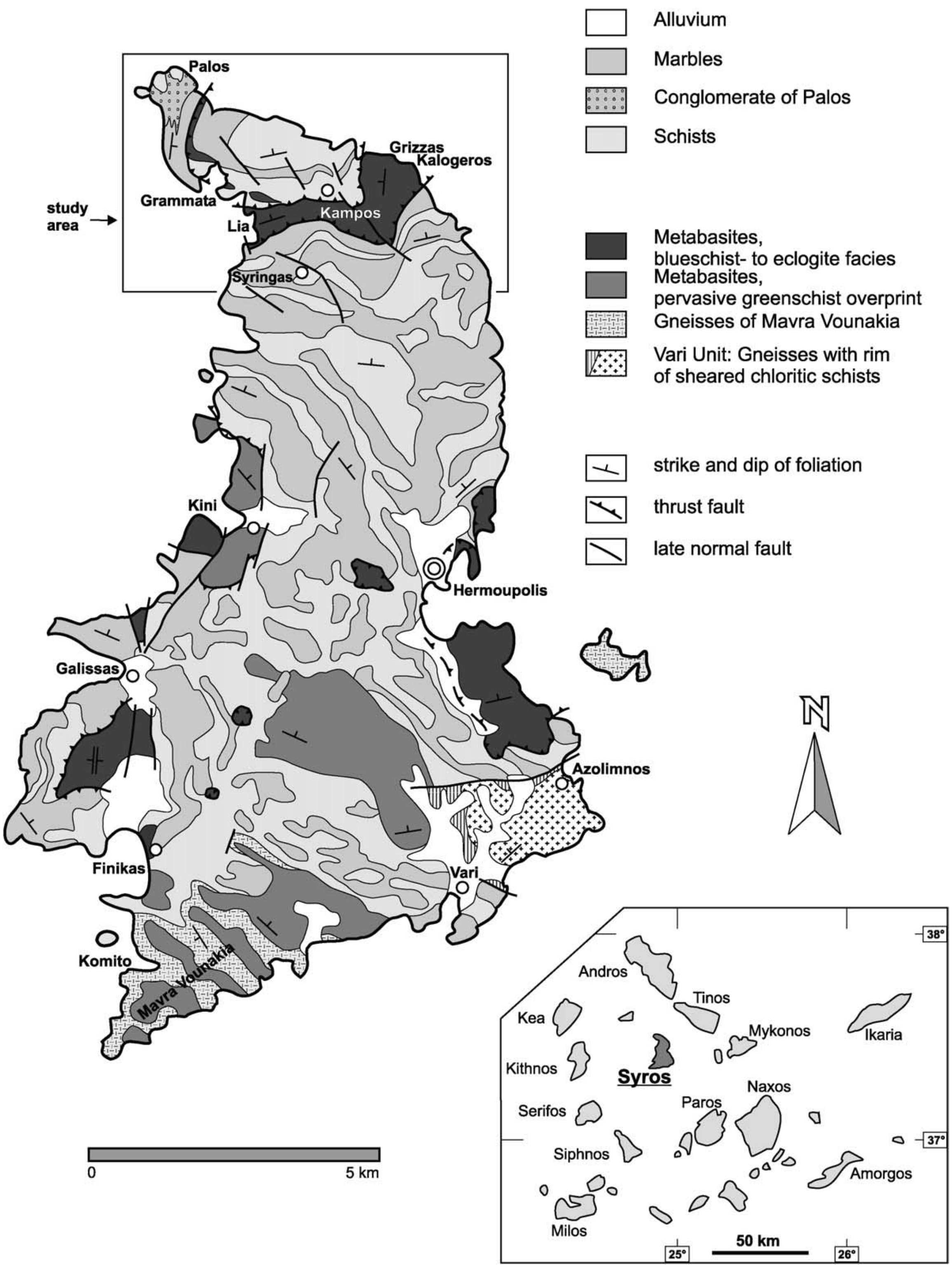

Fig. 1. Simplified geological map of Syros island and location of study area (modified after Hecht (1984)).

several brittle deformation phases. Ridley (1982a,b, 1984) identified one ductile isoclinal folding event related to crustal shortening on the prograde P–T path. He also described several minor deformation events that accompanied exhumation and emplacement of the rocks of Syros to their present-day position. Further research in the Aegean concentrated mainly on general and special aspects of the exhumation history of high-pressure rocks (Avigad et al., 1992; Gautier and Brun, 1994; Gautier et al., 1999; Jolivet and Patriat, 1999). In this context, the timing

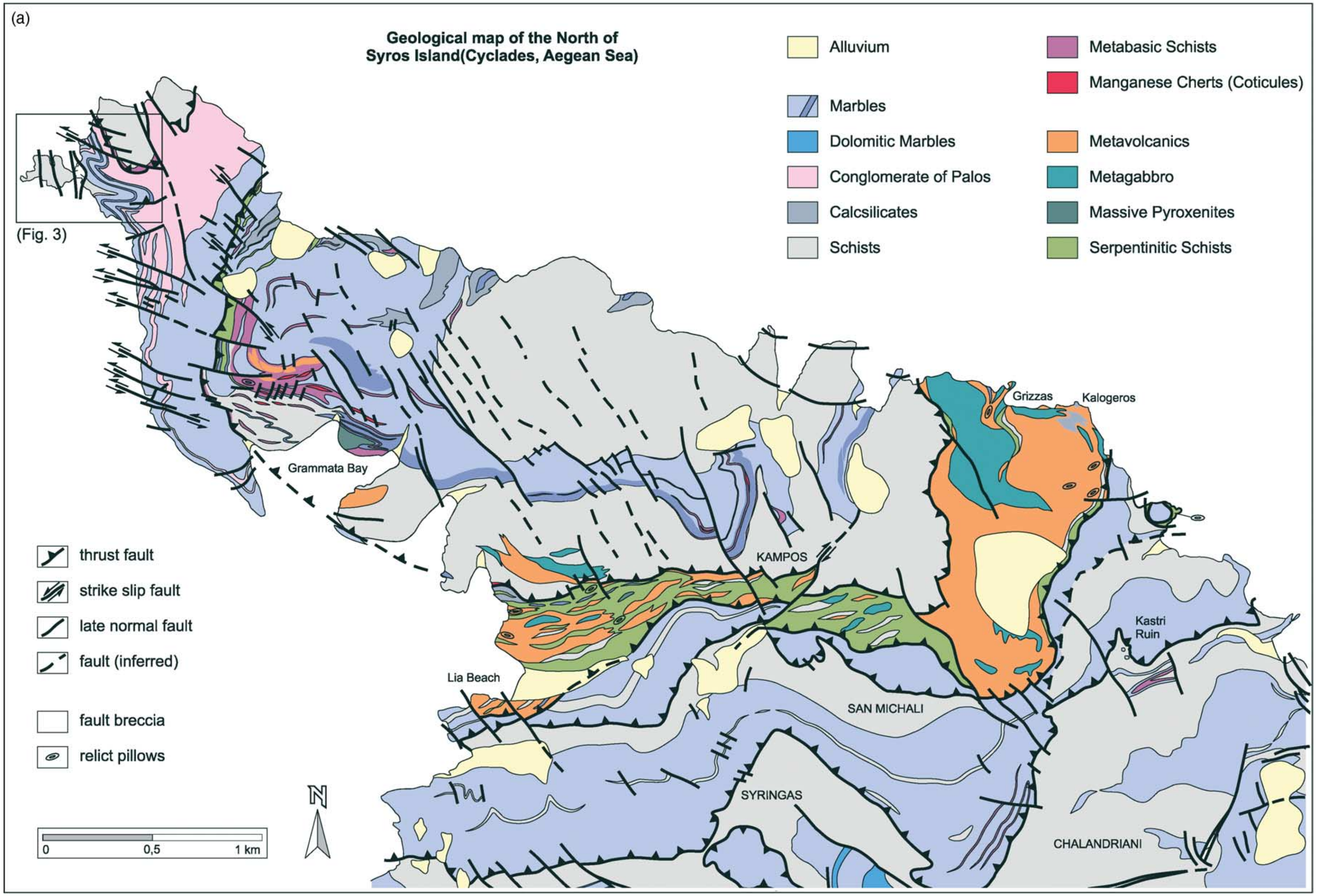

Fig. 2. (a) Geological map of the Kampos metabasite belt (northern Syros) and its metasedimentary envelope based on 1:5000 maps by the authors. (b) Structural map of the same area. Prominent large-scale isoclinal F2 folds are noted, as well as examples for young brittle upright cylindrical folds.

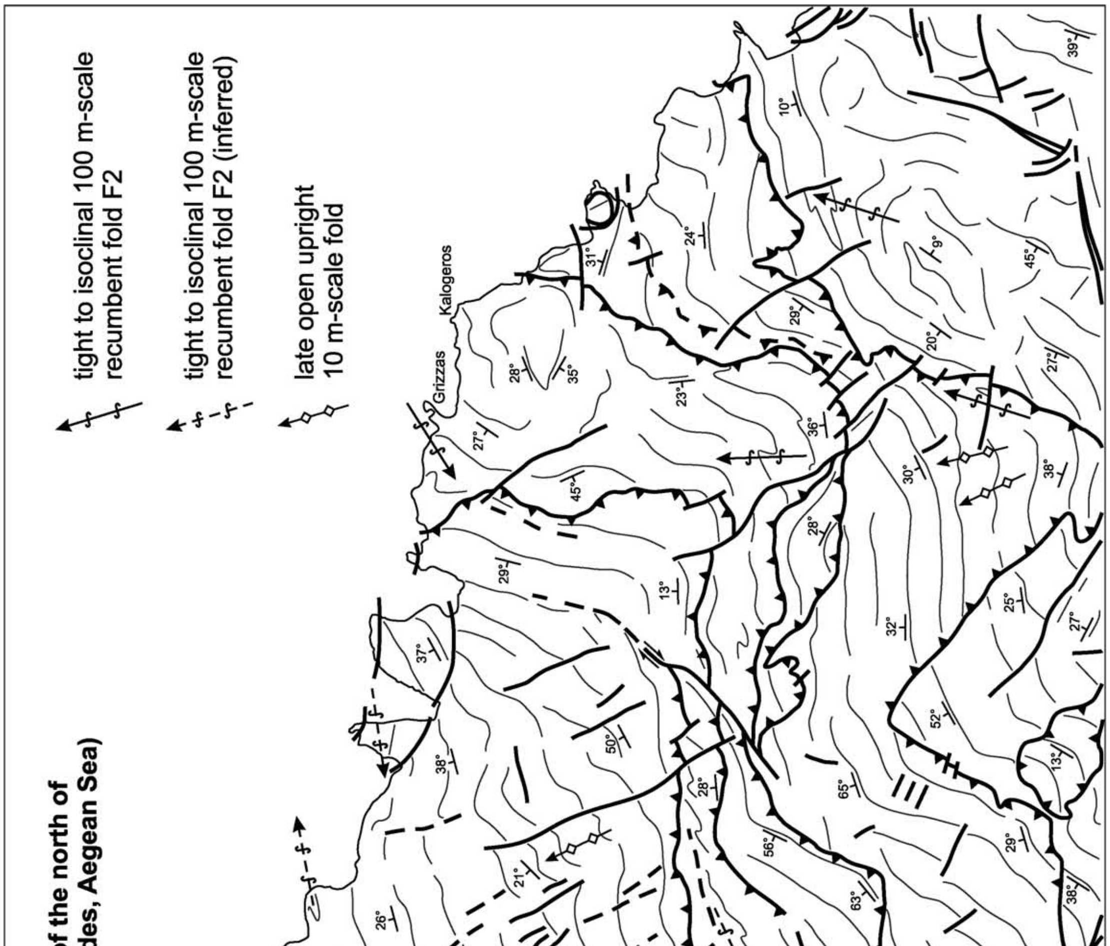
(b)
Structural map of the north of
Syros Island (Cyclades, Aegean Sea)
tight to isoclinal 100 m-scale
recumbent fold F2
tight to isoclinal 100 m-scale
recumbent fold F2 (inferred)
late open upright
10 m-scale fold
trend of foliation
thrust fault
strike slip fault
late normal fault
fault (inferred)
strike and dip of foliation
N
0
0,5
1 km
Grizzas
Kalogeros
Grammata Bay
Lia Beach

Fig. 2 (b).

and the process responsible for exhumation of the Cycladic blueschists are of special interest. A study presented by Trotet et al. (2001) favours the process of low angle detachment faulting, and places a major normal fault contact along the footwall of the Kampos metabasite belt in the north of Syros. Rosenbaum et al. (2002) suggested that the dominant ductile fabrics on Syros developed during exhumation as a result of coaxial stretching.

This paper discusses the structural evolution in the light of these recent publications, focussing on the mechanics and timing of ductile deformation. We present field observations made during lithological and structural mapping of the northern part of the island, i.e. the Kampos Metabasite Belt and its adjacent envelope of metasedimenary rocks, at a scale of 1:5000 (Fig. 2a,b). Data used for this paper include field measurements of fold axes and interference relationships of superposed folds, foliation measurements, fold geometry, fault orientations, and general relative age relationships.

As a result of this study, we identify two discrete prograde isoclinal folding events followed by several semi-brittle to brittle deformation events related to the retrograde exhumation path. All identified ductile deformation phases occurred on the prograde branch of the P–T path, here defined as that part of the P–T path with simultaneous pressure and temperature increase. We show that exhumation of the rocks of northern Syros was most probably static since at least 1 GPa. In addition, we correlate the different deformation events with geochronological data and provide a model for the structural evolution of Syros Island from the Upper Cretaceous to the present day.

2. Structural evolution

The rocks of Syros are intensely deformed and affected by at least four folding events. The structural development is characterized by a transition from ductile to brittle tectonism through time. The ductile deformation events (polyphase tight to isoclinal folding and thrusting) took place on the prograde branch of the P–T path. They caused crustal thickening during the Mesohellenic phase of the Alpidic convergence of the Hellenic Orogen. On the retrograde P–T path, a transition from semi-brittle to brittle deformation can be observed. Semi-brittle deformation is represented by kink folds and development of a crenulation cleavage and brittle deformation phases include transpressional structures and late normal faults.

2.1. First deformation D1: isoclinal folding

The first recognizable deformation phase, i.e. D1, is characterized by relict rootless isoclinal and intrafolial fold hinges. These folds are rarely of more than several decimetres in scale. D1 deformed primary sedimentary fabrics s0, e.g. Mn-rich hydrothermal cherts, and early

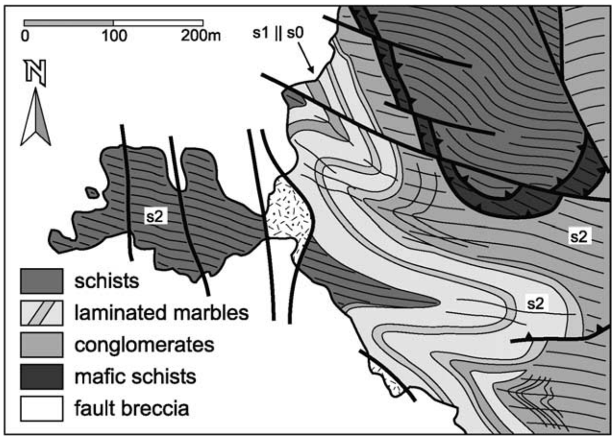

Fig. 3. Inset map showing the different orientations of the dominant foliations s1 and s2, respectively, in marbles and schists on Palos Peninsula (see Fig. 2a). The dominant foliation, s1, in the marble unit is folded by an F2 isoclinal fold. Note the schists and conglomerates, which developed a penetrative foliation s2 parallel to the fold axial plane of the large-scale F2 fold. The marbles developed a widely-spaced cleavage parallel to the regional s2 only in the highly strained fold hinges.

metamorphic quartz segregations. Intense deformation in the course of later events, especially during the ductile D2 event, has obscured or entirely destroyed most of the F1 folds, as well as an s1 foliation. Some lithologies, though, have preserved an s1 fabric, mainly caused by competence contrasts. These include the metavolcanics and marbles, as can be seen on Palos Peninsula (Fig. 3). Due to the scarcity and poor preservation of F1 folds, there is no means to quantify a transport direction related to D1.

2.2. Second deformation D2: tight to isoclinal folding and thrusting

2.2.1. Tight to isoclinal folding

Structures formed by this deformation event are the most prominent In almost all lithologies except the marbles and some metavolcanics, the s2 foliation related to D2 is the penetrative fabric. All map-scale, several 100-m-scale folds (Fig. 2b) are D2 structures. The F2 folds re-fold s1 and are intrafolial with respect to the s2 foliation. On Palos Peninsula, for example, a marble unit is folded isoclinally at a several-hundred-metre scale by an F2 fold. Development of a macroscopically visible s2 schistosity is confined to the hinge zone of this fold structure (Fig. 3).

Fig. 4a shows refolded layers of manganese cherts in an outcrop north of Lia Beach. The chert layers are composed of quartz and spessartine and result from metamorphic recrystallisation of formerly hydrothermal Mn deposits in mafic tuffites, deposited on a Late Cretaceous ocean floor. In this outcrop, the axes of the two fold generations F1 and F2 are not parallel to each other. A stereographic projection of structural data from this outcrop shows the geometric relationships between the two fold stages. Following the method of Ramsay and Huber (1987), the direction of

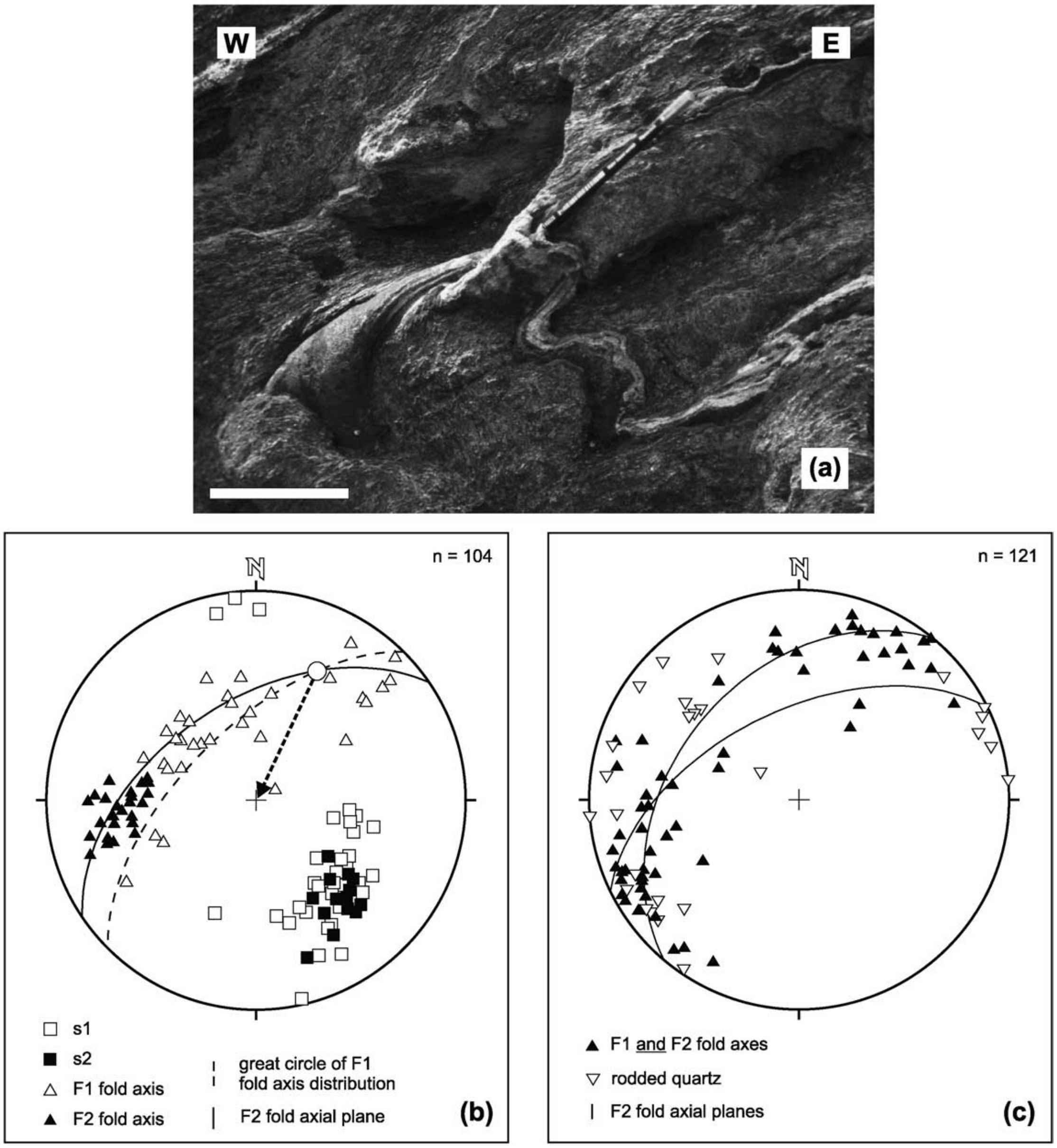

Fig. 4. Ductile refolding structures. (a) Thin layers of manganese chert interbedded in a matrix of glaucophane-bearing mica schist (mafic tuffite) north of Lia Beach. An isoclinal F1 fold hinge (left) is refolded by an F2 fold (axis parallel to pencil). Size of scale bar ~10 cm. (b) Stereographic projection of the structural inventory measured at this location. Note the distinct great circle distribution of F1 fold axes, indicating a strong shear component for the F2 folding event. The arrow shows direction of tectonic transport. (c) Stereographic projection of linear elements measured in the area between Lia Beach and Kampos (for location see map Fig. 2a). Similarity between the distributions in Fig. 5b and c emphasizes that outcrop scale deformation pattern is applicable to the regional scale. For further details see text.

relative tectonic transport for the D2 deformation event can be established to be roughly top-to-the-SSW (see Fig. 4b).

The existence of two sets of non-axial-parallel folds in certain outcrops was recognized earlier, but interpreted as non-representative exceptions from the rule (Ridley, 1982a). Both isoclinal F1 and F2 folds on Syros were regarded as a product of one single, continuous deformation event. We tested this by plotting measured fold axes and rodded quartz from the metabasite belt between Kampos and Lia Beach (Fig. 4c). Most of the measured folds in the area chosen are preserved only as isolated fold hinges, so that a distinction between F1 and F2 folds was not possible. Nevertheless, the plot shows within error the same distribution as the plot from the quartz–spessartine chert outcrop where superposed folding is evident (Fig. 4b). We therefore suggest that the deformation pattern observed in the manganese chert outcrop is representative for the entire area of northern Syros, and we assume that two generations of non-axial-parallel folding events can be distinguished.

2.2.2. Thrusting

Prominent structural features to the north of Syros are

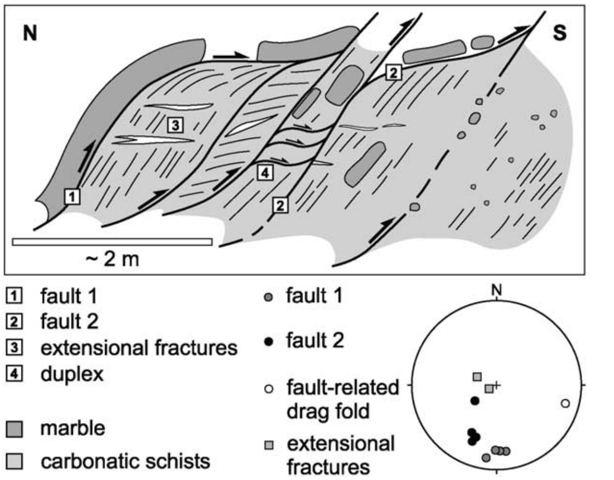

Fig. 5. Sketch of a brittle stacking structure at a marble–schist contact between San Michali and Kampos. Duplex structures, bending of fault planes, fracture geometry and fault related drag folds indicate top-to-the-S thrust sense of movement.

thrusts. The most distinctive is located at the footwall of the Kampos metabasite belt. This contact can be traced across northern Syros, running from Palos Peninsula, Lia Beach, Kampos and north of San Michali to the coast south of Kalogeros Point (Fig. 2a). It is regarded as a thrust because it juxtaposes mafic and ultramafic rocks against pure marbles, an association that is not likely to be primary. Furthermore, the orientation of this contact is broadly parallel to the penetrative foliation s2 and the axial planes of F2 folds. Note though that in places, the Kampos thrust, and the other mapped thrust planes, cross-cut s2 at a small angle (Fig. 2b) and show signs of brittle overprint. This overprint is caused by a poorly defined later stage brittle thrust sense reactivation (see Fig. 5). Metamorphic grade across the Kampos contact shows no sharp boundary; retrograde greenschist overprint south of Kampos is not penetrative but rather gradual/selective with, in places, well-preserved blueschist facies lithologies (but see Trotet et al., 2001). Furthermore, the structural inventory north and south of the metabasite belt, as well as within the belt itself, is generally identical.

We regard this thrusting as a final stage of D2 horizontal shortening, which took place whenever shortening exceeded a degree of what could be accommodated by the F2 folding. The above-mentioned observations on metamorphic grade and structural features indicate that, after the cessation of prograde crustal shortening, no major displacements along the Kampos contact took place.

2.3. Third deformation D3: kink folding and crenulation

D3 produced small (millimetre to decametre scale) upright to slightly vergent kink folds referred to as F3, and a sub-vertical crenulation cleavage s3 (Fig. 6a–c). These structures are restricted to small isolated zones, often no more than outcrop size. From the orientation of the crenulation cleavage and the axes of the kink folds, we derive an east–west horizontal shortening for this event (Fig. 6d).

Cross-cutting relationships (Fig. 7) show that F3 folds postdate the tight to isoclinal F2 folds. In thin section, micas deformed by D3 are bent without recrystallisation in the fold hinges (Fig. 8), indicating that D3 took place under semi-brittle conditions. We place D3 at a late stage of the retrograde greenschist facies overprint, probably at temperatures of about 300 °C.

2.4. Later stage postmetamorphic deformation events

The rocks of Syros are affected by a number of late brittle deformation phases post-dating greenschist-facies overprint. Quite often, brittle deformation has considerable influence on the geometry of the whole succession. A common characteristic for all events described below is that they lead to brittle behaviour in marbles. Assuming a geologically realistic strain rate of $10^{-15}\ s^{-1}$ (Heard and Raleigh, 1972), calcite reacts brittly below 250 °C, indicating that these deformation events either formed part of a late stage of the metamorphic evolution or reflect final emplacement of the rocks to their present-day position. The brittle structures identified are (1) brittle upright folds (Fig. 9) and left-lateral strike slip faults (Figs. 2 and 10) as well as (2) late normal faults (Figs. 2 and 11). Constraints on their relative and absolute age, if available, are discussed below.

2.4.1. Brittle upright folding and transpressional strike-slip faults

The map view (Fig. 2a and b) shows that the north of Syros is bent on a kilometre-scale, forming a northerly dipping synform and leading to a change in strike of the s2 foliation from E–W in the central area around Kampos, to N–S on Palos Peninsula and between San Michali and Grizzas Bay (cf. Hecht, 1984). Frequently, we observe outcrop-scale brittle shortening structures, such as upright folds in marbles (Fig. 9a). These structures are open folds with interlimb angles between 120 and 80° and are upright to slightly vergent with north dipping axes. Such folded marble layers were measured systematically to test a possible geometric coherence with the above-mentioned large-scale structure. A stereographic projection of measured and constructed fold axes (Fig. 9b) shows a general northerly dip direction. Therefore, we regard these folds as parasitic structures of the map-scale synform. Obviously, this folding event is more important for the structural evolution of Syros than previously thought. The arcuate lineation trends described by Ridley (1982b) for the north of Syros are here assumed to be a product of the large-scale upright folding. A distinct east–west directed contraction seems to be most likely.

Related to this event is a closely spaced system of brittle left-lateral strike-slip faults on Palos Peninsula (Fig. 2a and b).

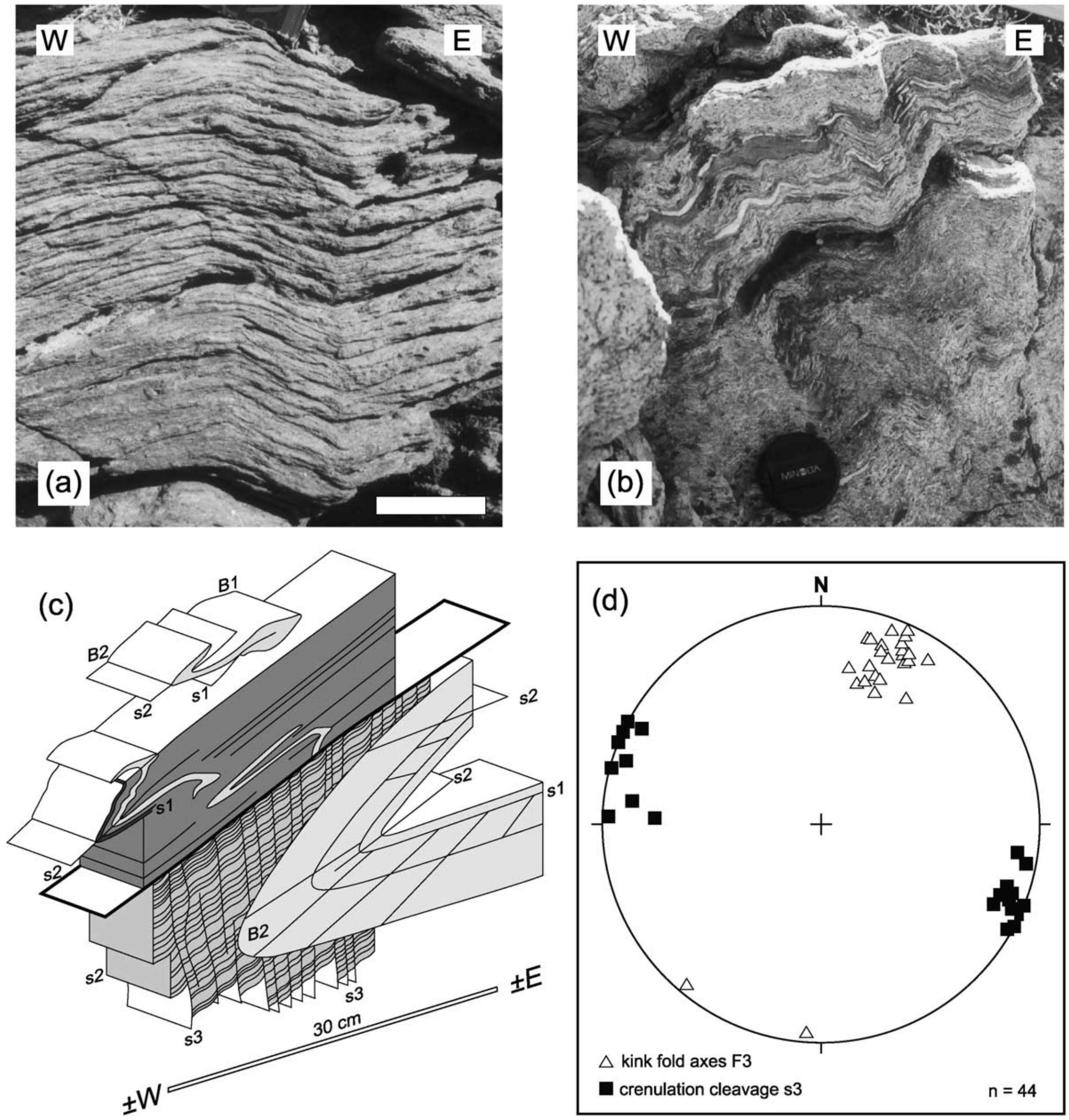

Fig. 6. Structures formed by D3. (a) Kinking of blueschists northwest of Kampos; size of scale bar 10 cm. (b) Deformed albitized layers in banded schists north of Plati Vouni, size of camera lid 5 cm. (c) Sketch showing a vertical crenulation cleavage s3 cross-cutting the penetrative metamorphic s2 foliation and an F2 fold. Note that the bright-grey, more competent lithology is not affected by the crenulation. (d) Stereographic projection of F3 fold axes and s3 cleavage planes. Horizontal shortening was roughly E–W directed.

Tectonic stress could be identified as roughly east–west directed, as derived from geometric relationships between the main strike-slip fault plane and associated Riedel shears (Fig. 10).

2.4.2. Late normal faults

The last recorded deformation event is characterised by the development of steep normal faults cutting all other structures to the north of Syros. Recorded vertical displacements along the faults range between 10 and 100 m. The most prominent strike direction of steep normal faults is NW–SE to NNW–SSE (see maps in Fig. 2a and b). This tendency is also observed in a statistical analysis of measured small-scale normal faults (Fig. 11). In the northern part of Syros, a mainly NE–SW directed brittle extensional strain prevails.

3. Discussion

The following section integrates the deformation succession into a reasonable P–T–t path. The various events are presented in a chronological order corresponding to our field observations. Integrated into this is a discussion of the mechanical behaviour of the rocks during exhumation. We will show that the isoclinal fold stages D1 and D2 are very

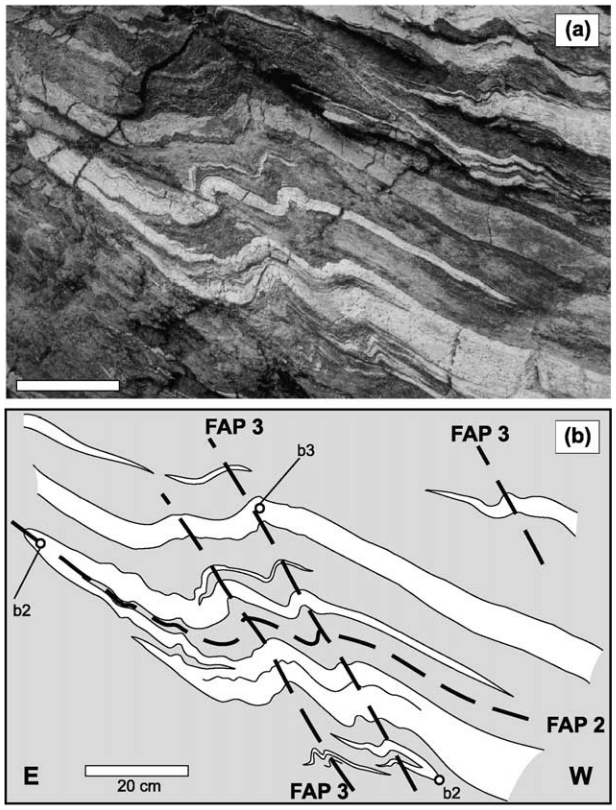

Fig. 7. Interference between isoclinal fold stage D2 and crenulation stage D3, south of Chalandriani. (a) An intrafolial isoclinal F2 fold (white layers) is refolded by F3 and truncated by a slightly vergent s3 cleavage. Size of scale bar 20 cm. (b) Graphic interpretation of (a); the isoclinal fold axial plane FAP2 is refolded by D3 fold structures (fold axial planes FAP3 are noted). Crenulation cleavage s3 and kink folding postdate the ductile D2 structures.

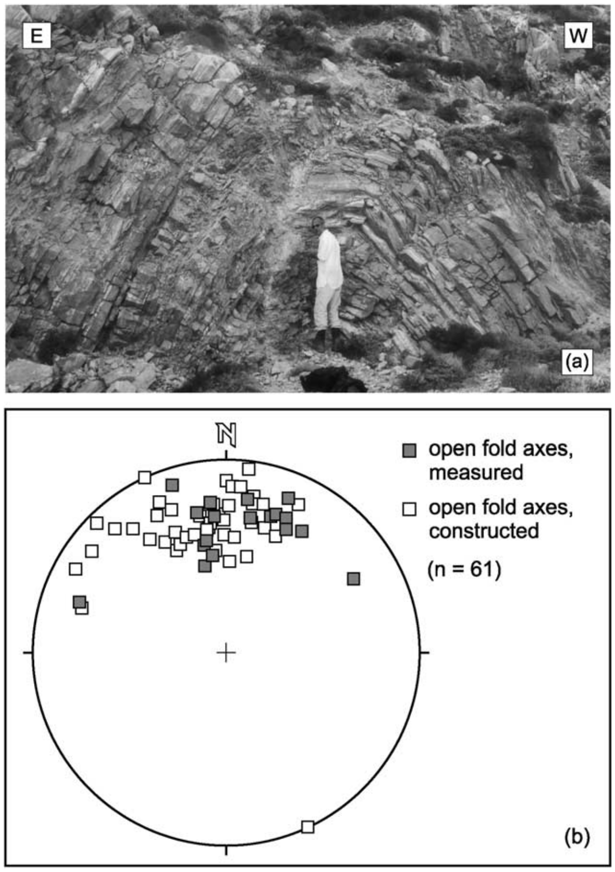

Fig. 9. Late upright folds. (a) Brittle cylindrical folds in marbles south of Chalandriani. (b) Stereographic projection of measured and constructed late open folds in northern Syros. General slight to moderate dip of fold axes in a northerly direction corresponds to the geometry of the large-scale synform seen on the maps in Fig. 2a and b. For further explanations see text.

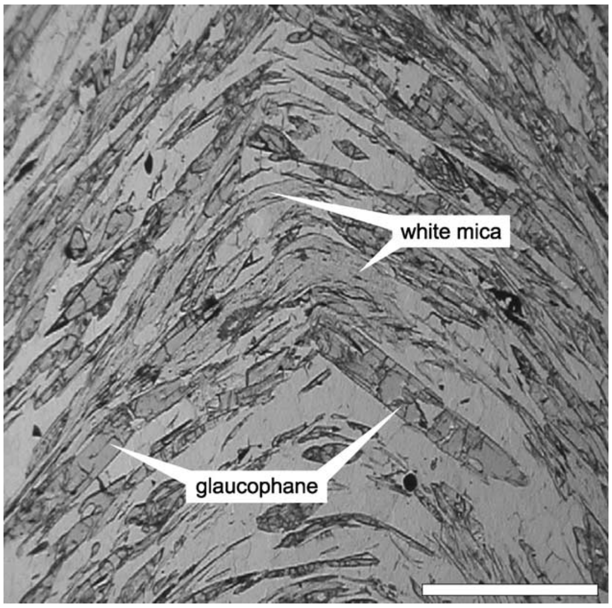

Fig. 8. Thin section of D3 structures perpendicular to F3 kink fold axes. Note the broken glaucophane needles and the bent mica aggregates that show no signs of recrystallisation in the hinges. Size of scale bar 1 mm.

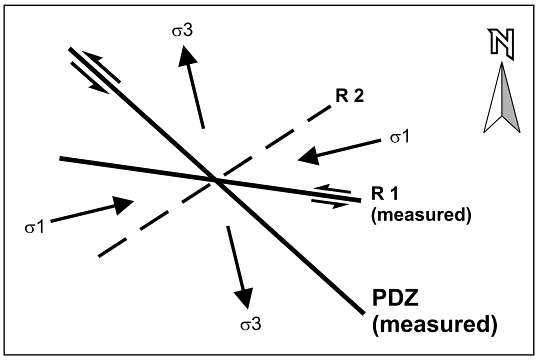

Fig. 10. Stress field of the sinistral strike-slip faults on Palos Peninsula (see Fig. 2a). Orientation of main fault (principal displacement zone—PDZ) and associated synthetic R1 Riedel shear faults measured in a brecciated fault zone on Palos Peninsula. Antithetic R2 Riedel shear faults are subdued here, their orientation being derived from the dextral strike-slip fault cutting through the Metabasite Belt S Kampos (see map Fig. 2a). Geometric relationships, according to Ramsay and Huber (1987), indicate E–W directed transpressive stress ($\sigma 1$).

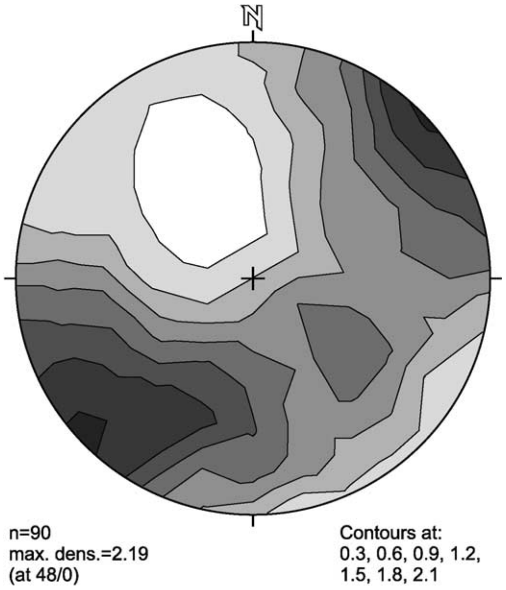

Fig. 11. Stereographic density diagram of small-scale steep normal faults measured in northern Syros. The maximum at a NE–SW dip direction of fault planes corresponds to the prominent orientation of map-scale normal faults (see also Fig. 2a and b), indicating a strong component of NE–SW directed extension affecting Syros in its younger structural history.

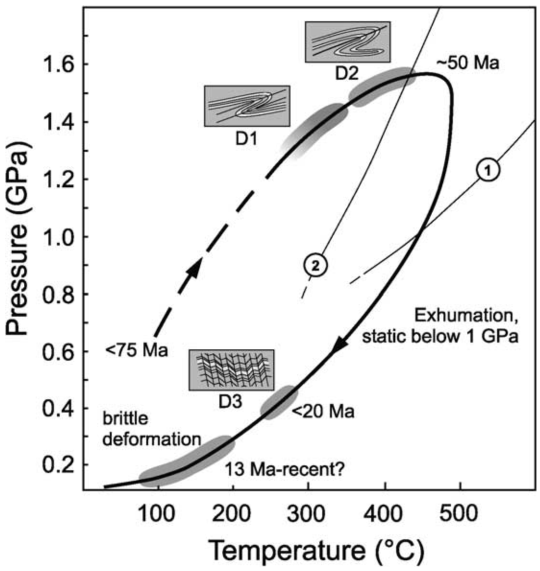

Fig. 13. Schematic P–T–t diagram showing the timing of the various deformation events based on observations described in this paper. The P–T path is clockwise, as expected in an orogenic belt: rapid burial during crustal stacking up to peak metamorphism, followed by near isothermal uplift at the early stage of exhumation, causing a Barrowian-type greenschist overprint. Mineral equilibria included in the diagram are as follows: (1) aragonite = calcite (Carlsen, 1983); (2) 52 lawsonite + 5 glaucophane = 26 clinozoisite + 3 chlorite + 10 paragonite + 27 quartz + 74 H_2O (Evans, 1990).

likely restricted to the prograde branch of the P–T path. We also demonstrate that exhumation of the rocks was most probably static, with very little or no internal deformation. Furthermore, we show that there are doubts concerning the interpretation of the Kampos contact as a major detachment fault, as proposed by Trotet et al. (2001).

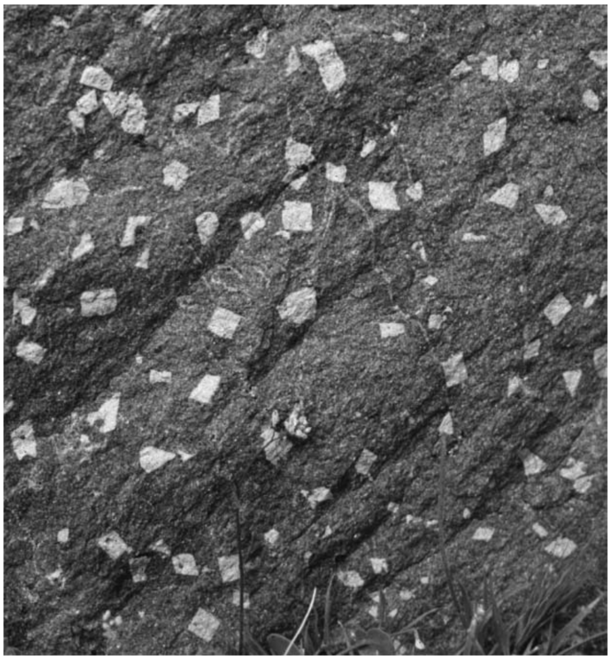

Fig. 12. Static peak minerals. Euhedral lawsonite pseudomorphs overgrowing the penetrative foliation s2 in blueschists, here running from lower left to upper right. Note that the pseudomorphs follow no preferred orientation, indicating that lawsonite replacement by pseudomorphic phases (clinozoisite, chlorite, phengite, quartz) postdated the development of s2 and was not dependent on pre-existing lithological anisotropies. Size of pseudomorphs 10–20 mm.

3.1. Timing of ductile deformations D1 and D2 relative to metamorphism

To constrain the timing of D1 and D2 penetrative deformation relative to metamorphism, deformation states of lawsonite pseudomorphs were used. Delicate fabrics, such as pseudomorphs of clinozoisite/chlorite/paragonite/-quartz after lawsonite, are often preserved undeformed (Fig. 12). The breakdown of lawsonite in the presence of glaucophane is a prograde reaction (Fig. 13), so that D2 must have been largely completed before the rocks passed from the glaucophane–lawsonite into the glaucophane–clinozoisite subfacies (cf. Evans, 1990) before peak metamorphic conditions were reached. Since U–Pb zircon ages of 75–80 Ma from the metabasite belt are interpreted as primary magmatic (Keay, 1998; Tomaschek et al., 2003), and considering the age of the high pressure metamorphism (Maluski et al., 1987; Baldwin, 1996; Tomaschek et al., 2003) it can be assumed that the deformation phases D1 and D2 took place within a time range between formation of the oceanic crust at 75 Ma and peak metamorphism at 50 Ma.

3.2. Strong ductile deformation during exhumation?

The process responsible for exhumation of the rocks on Syros is still not altogether clear (for a detailed review of exhumation mechanisms see Ring et al. (1999a)). Since a large amount of erosion can be excluded because there are insufficient amounts of post-Eocene molasse sediments in the Cyclades (Lister et al., 1984; Avigad and Garfunkel, 1991), there are two general concepts to explain exhumation from 60 km to the surface: (1) crustal thinning by coaxial stretching (Rosenbaum et al., 2002) and (2) crustal thinning by low angle detachment faulting (e.g. Trotet et al., 2001).

All over Syros, boudinage is a frequent phenomenon. Rosenbaum et al. (2002) attributed these structures to regional coaxial flattening during deep-level ductile stretching. This concept implies that the rocks were exhumed as a whole, without losing coherence by crustal-scale detachment faults. Note though that boudinage, the main argument for coaxial horizontal stretching/vertical flattening processes in Rosenbaum et al. (2002), can also develop on long limbs of isoclinal folds during crustal shortening, i.e. is not necessarily an indicator for *regional* extensional environments. Furthermore, the concept of regional coaxial stretching cannot be supported here because it would require a large amount of penetrative shearing. Our observations are in contrast to the assumption that the rocks of Syros were affected by internal deformation during crustal thinning, for the following reasons:

1. Calcite paramorphs after aragonite: Everywhere on Syros columnar calcite needles can be observed in marble layers (recently described by Brady et al. (2003)), oriented at a fairly constant angle of about 60° to the dominant foliation. These calcite needles result from a topotactic relation of retrograde calcite with prograde aragonite that developed during exhumation when the marbles re-entered calcite stability. Any internal deformation by means of the process envisioned by Rosenbaum et al. (2002) would have destroyed such delicate fabrics, especially in highly ductile lithologies like marbles. The fact that these calcite needles are so abundant and well-preserved all over Syros leads us to conclude that no penetrative internal deformation took place at least since the aragonite–calcite phase transition, i.e. ~1 GPa (cf. Carlsen, 1983).
2. Lawsonite pseudomorphs: As mentioned above, lawsonite pseudomorphs, where preserved, are often internally undeformed and overgrow the penetrative foliation (Fig. 12). Like the calcite fabric, such pseudomorphs would not have survived any penetrative deformation event that affected the rocks after the breakdown of lawsonite in the presence of glaucophane to its pseudomorphic phases, further supporting that ductile deformation was reserved entirely to the prograde branch of the P–T path.

Both examples indicate that a major part of the retrograde exhumation P–T path elapsed under static conditions. Therefore we support the general concept that exhumation principally occurred along discrete low angle simple shear detachment faults, as was postulated for many other Cycladic islands (e.g. Lister et al., 1984; Avigad and Garfunkel, 1991). The model provides a reasonable explanation for the existence of domains that apparently suffered no internal stretching during exhumation.

While acknowledging the importance of low-angle detachment faults for exhumation, we would not place one along the Kampos metabasite belt, contrary to assertions by Trotet et al. (2001). Trotet et al. (2001) speculated that the Kampos contact juxtaposed greenschist-facies rocks in the south against blueschist–eclogite-facies rocks in the north. We found no evidence for a sharp break in metamorphic grade across the Kampos contact. The generally identical structural inventory north and south of the Kampos contact (see Section 2.2.1) casts further doubts on the assertion that the Kampos contact as a major extensional fault structure constituting a tectonic offset of several dozen kilometres.

3.3. Timing of kink folding and crenulation D3

As described above, D3 accompanies late-stage retrograde greenschist overprint, rendering D3 a compressive sub-event in the otherwise mainly static exhumation. Since the age of the greenschist overprint is constrained with white mica geochronology from 25 to 16 Ma (Bröcker et al., 1993; Wijbrans et al., 1993), a rough and tentative age of <20 Ma for D3 can be suggested.

3.4. Timing of brittle deformation events

The ductile–brittle transition in the Cycladic area is dated using cross-cutting relations of brittle structural features with datable sediments (Boronkay and Doutsos (1994) for Serifos, Tinos and Naxos) as well as apatite fission-track ages (Lee and Lister (1992) for Mykonos). Accordingly, the ductile–brittle transition took place between 13 and 8 Ma. There is no reason to assume that the history of Syros was remarkably different to that of other Cycladic islands. By inference, the brittle structures described here are assumed to be younger than 13 Ma. Note, however, that more accurate dating is difficult, because Syros lacks datable sediments. Therefore, the following considerations remain speculative:

- the strike-slip fault system observed on Palos Peninsula and the open upright folding might be consistent with a transpressional deformation phase described by Boronkay and Doutsos (1994) on Samos, Paros and Naxos, terminating about 7 Ma ago. Ring et al. (1999b) were able to constrain a shortening event on Samos to a time interval between 8.6 and 9 Ma. The transpressional stress field found by Boronkay and Doutsos and Ring et al. was identified as roughly E–W directed, in fair agreement with the stress direction observed at the open upright fold structures and the strike-slip faults on Palos Peninsula;
- the steep normal faults cut all other structures and

therefore mark the youngest deformation event. They may be related to a change in the regional stress regime caused by the beginning of lateral extrusion of the Turkish–Aegean microplate along the North Anatolian Fault Zone. This prominent right-lateral strike-slip system is considered to have been active since the Early Pliocene at about 5 Ma (Gautier et al., 1999; Kurt et al., 2000). Furthermore, data on present-day crustal motion in the Aegean show a SW-directed extensional strain of about 20 mm/a (Kahle et al., 1998; Cocard et al., 1999). The strike direction of the steep normal faults to the north of Syros fits into this regional extensional stress field and suggests normal faulting to be active to the present day.

4. Conclusions

Based on the evidence discussed above we propose the following scenario for the geological and structural history of the rocks of Syros, as summarized in Fig. 13:

1. Slices of an Upper Cretaceous oceanic back arc succession with a sedimentary cover became involved in orogenic crustal stacking processes and were buried to a depth of 50–60 km. During burial, ductile deformation fabrics developed, now preserved as two generations of isoclinal folds and thrust shear zones (D1 and D2). The fact that two, non-axial parallel fold sets exist indicates that during burial, a change in stress direction took place. A plausible cause could be a change in general plate convergence vectors during the Eocene orogenic cycle in the Hellenides. The second deformation, producing the dominant fabrics on Syros, is characterized by a top-to-the-SSW thrust-sense movement. At the time of peak metamorphism (~50 Ma), ductile deformation was largely completed, i.e. the succession of Syros was structurally isolated.
2. Exhumation of the rocks occurred along localized detachment faults, leaving most of the succession on Syros unaffected by internal extensional deformation. Static exhumation was interrupted by a phase of contraction, probably under semi-brittle conditions, i.e. at a late stage of greenschist overprint. This contractional phase left behind upright to slightly vergent kink folds and in places a subvertical crenulation cleavage (D3).
3. The brittle deformation phases described are assumed to be younger than 13 Ma and continue to the present day. These deformation phases are seen in direct context with the rapidly changing stress field affecting the Aegean back arc region in Neogene times, including the migration of the Hellenic subduction zone, the beginning of lateral escape of the Turkish–Aegean microplate, and block rotations caused by the south- to southwestward extrusion of Aegean lithosphere towards the Hellenic trench. Postmetamorphic deformation events affecting the rocks of Syros in places have a strong influence on the orientation of prograde fabrics, and have to be kept in mind when evaluating the significance of structures related to the metamorphic history.

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