Structural analysis of a complex nappe sequence and late-orogenic basins from the Aegean Island of Samos, Greece

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Abstract

The island of Samos in the Aegean Sea exposes high-pressure metamorphic rocks of the Cycladic blueschist unit which are sandwiched between the mildly blueschist-facies Kerktas nappe below and the overlying non-metamorphic Kallithea nappe. Structural and metamorphic analysis shows that deformation can generally be divided into four main stages: (1) Eocene and earliest Oligocene ESE–WNW-oriented nappe stacking (D1 and D2) associated with blueschist- and transitional blueschist–greenschist-facies metamorphism (M1 and M2). D2 caused emplacement of the blueschist unit onto the Kerktas nappe indicating that thrusting occurred during decompression. (2) A subsequent history of Oligocene and Miocene horizontal crustal extension (D3) before and after greenschist-facies metamorphism (M3). Ductile flow during D3 was characterized by a high degree of coaxial deformation but in general caused displacement of upper units towards the ENE. Nonetheless, the late-stage D3 emplacement of the Kallithea nappe between 9 and 10 Ma had a top-to-the-NW/NNW sense of shear. (3) A short period of brittle E–W crustal contraction (D4) occurred between <8.6 and ~9 Ma. (4) A phase of N–S-directed normal faulting (D5, <8.6 Ma to Recent). ESE–WNW-directed tectonic transport during D1 through D3 is in contrast to uniform NNE–SSW-directed tectonic transport in the adjacent Cyclades, Greece, as well as in the neighbouring Menderes Massif of western Turkey. Published paleomagnetic data reveal sinistral rotation between the Cyclades and western Turkey. We interpret this rotation as a consequence of differential extension between the severely extended Aegean and the moderately extended Menderes Massif during D3. The onset of D3 crustal extension is coeval with a marked change in the thermal structure. We propose that the thermal reorganization was associated with the retreat of the subduction zone towards the external Hellenides in the Early Oligocene and a subsequent increase in magmatic activity. © 1999 Elsevier Science Ltd. All rights reserved.

1. Introduction

Research over the last two decades revealed the importance of NNE–SSW-directed horizontal crustal extension in the Aegean (e.g. Le Pichon and Angelier, 1979; Lister et al., 1984; Avigad and Garfunkel, 1991) and the adjacent Menderes Massif of western Turkey (e.g. Hetzel et al., 1995). Eocene blueschist-facies rocks of the Cycladic blueschist unit are envisioned as having been dragged to the surface in the footwall of major crustal shear zones to form the Aegean metamorphic core complexes (Lister et al., 1984). However, no Tertiary high-pressure rocks have been reported so far from the base of the metamorphic core complex of the central Menderes Massif (Ring et al., 1999). Differential extension between the Aegean and the Menderes Massif is indicated by the fact that the Aegean, especially the Cretan Sea, is underlain by intensely attenuated continental crust whereas the Menderes Massif is not (Makris and Stobbe, 1984). The island of Samos is located between these differentially extending areas in the Aegean and the Menderes Massif and the structural evolution of Samos might have been controlled to some degree by differential extension. Paleomagnetic data from Lesbos Island and
the Izmir region (Kissel and Laj, 1988) (Fig. 1) shows an abrupt switch from slight clockwise rotation (68°) on Lesbos Island to pronounced anticlockwise rotation (0308°) in westernmost Turkey. The age of this rotation is not precisely known but Middle Miocene appears most reasonable (Kissel and Laj, 1988).

The pre-extension emplacement of the Cycladic blueschist unit is poorly understood due, at least in part, to the fact that the deep tectonic units onto which the blueschists have been emplaced are not exposed on most Aegean islands. In this regard the island of Samos (Figs. 1 and 2) in the easternmost Aegean Sea is of particular interest because it exposes the Kerketas nappe of the external Hellenides which occurs structurally below the Cycladic blueschist unit.

In this contribution, we analyse the structural evolution and its relationship to the metamorphic history of the nappe pile and its Neogene cover on Samos Island in order to decipher the roles of crustal contraction and crustal extension. The manifold problems in trying to relate a particular phase of deformation to crustal shortening or lengthening have been pointed out recently by a number of studies (Ring and Brandon, 1994; Wheeler and Butler, 1994; Ring, 1995). Then we discuss the tectonometamorphic history of Samos within the framework of Aegean tectonics.

2. Regional setting

The Cycladic Massif consists of a pile of nappes (Altherr and Seidel, 1977) (Fig. 3). In ascending order the major tectonic units are: (1) The Basal unit as part of the external Hellenides. (2) The Cycladic blueschist

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*Fig. 1. Generalized tectonic map of the Aegean and adjacent mainlands (greatly modified from Jacobshagen, 1986 and Le Pichon et al., 1995) showing major tectonic units and present-day Hellenic subduction zone north of the Libyan coast. The Pelagonian–Cycladic zones, which have been joined together here, and the Vardar–Izmir–Ankara zones form high-pressure belts of Cretaceous to Early Tertiary age which span a wide area from the Greek mainland across the northern Aegean Sea into Turkey. Miocene high-pressure rocks (vertically ruled pattern, Theye et al., 1992) occupy a more restricted area than the older high-pressure rocks and occur to the south/southeast of the Late Pliocene to Recent volcanic arc delineated by calc-alkaline volcanoes (asterisks). Inset: Miocene to Recent thrust fronts in the Mediterranean region and location of main map; arrows show the integrated relative motion of Africa with respect to Europe from 35 Ma to present in the eastern Mediterranean.*
Fig. 2. Simplified geologic and tectonic map of Samos Island showing major rock units, thrusts and faults (with bars and ticks on the hanging wall sides) and representative structural attitudes of rocks. The different nappes are described in Table 1. Lower series of graben fill comprise Basal Conglomerate, Pythagorion and Hora Formations; Upper series of graben fill include Mytilini and Kokkari Formations (see also Figs. 3 and 4 and Table 2). The D₁ Ampelos and Selçuk thrusts are the basal thrusts of the Ampelos and Selçuk nappes, respectively. The D₂ Pythagoras thrust put the Cycladic blueschist unit on top of the Kerkeras nappe (the Pythagoras thrust is named after ‘Pythagoras Cave’ c. 50 m above the thrust plane west of Marathokampos). The Kallithea detachment is a late-D₃ low-angle extensional fault. Widespread Middle and mainly Late Miocene volcanic and volcanoclastic rocks occur at the eastern and northeastern margin of the Karlovasi and Pyrgos graben. Subordinate volcanics also occur at the western side of the Mytilini basin. Numerous reverse (D₄) and normal (D₅) faults overprinted all earlier ductile contacts. We interpret the curved D₅ high-angle normal faults to have a listric geometry. Positions of cross-sections A–A' and B–B' (Fig. 4) and locations of maps shown in Figs. 13–16 are indicated.
unit comprising at least three separate members: (i) A Carboniferous basement nappe, overlain by (ii) a nappe made up of a probably post-Carboniferous shelf sequence (Altherr and Seidel, 1977). (iii) Above this shelf sequence occurs a melange-like nappe of ophiolitic rocks embedded in a serpentinitic and metapelitic matrix (Okrusch and Bröcker, 1990; Erdogan and Güngör, 1992; Candan et al., 1997). (3) The Upper unit consists of three subunits: (i) The Lycian nappes which are capped by Eocene to Oligocene clastic rocks (Okay, 1989; Collins and Robertson, 1998). (ii) The Cretaceous to Paleogene Vardar–Izmir–Ankara suture zone which fringes the Cycladic and the Menderes Massif to the north. (iii) The non-metamorphic composite Cycladic ophiolite nappe, which contains tectonically intercalated low-pressure metamorphic rocks of Cretaceous age (Reinecke, 1982). (4) Sedimentary basins filled with Miocene and younger sediments. These basins formed at $13 \pm 5$ Ma (Le Pichon and Angelier, 1979).

Both the Cycladic and the Menderes Massif have been suggested that extension started in the latest Oligocene (Seyitoglu et al., 1992; Gautier et al., 1993) or already in the Early Oligocene (Raouzaios et al., 1996). Crustal extension in the Cycladic Massif is generally considered to be the result of the retreat of the Hellenic slab (Lister et al., 1984; Buick, 1991) and occurred and still occurs in the fore-arc high (Crete; Fassoulas et al., 1994; Thomson et al., 1999) in the arc (on the island of Thera; Lister and Forster, 1996) and in the back-arc region (Cycladic Massif; Lister et al., 1984; Gautier et al., 1993).

3. Geology of Samos

Our mapping at the 1:10 000 scale over the last four years covered approximately 50% of Samos Island (Fig. 2) and augmented the work of Theodoropoulos (1979) and Papanikolaou (1979). The general structure is dominated by numerous late faults. A major, N-dipping normal fault occurs just north of the island and it tilted Samos Island to the south. Cross-sections...
Fig. 4. (a) and (b) Serial cross-sections A–A' and B–B' through Samos Island showing general architecture of the island (refer to Fig. 2 for transect positions). The trace of the main foliation illustrates the generally E-dipping structure (see also Fig. 5). (c–e) Summary plots of contraction and extension axes as deduced from fault-slip analysis of D3, D4 and D5 faults.
(Fig. 4) and orientation data (Fig. 5) show that the nappe pile dips to the east. This nappe pile is described in Table 1.

The final emplacement of the non-metamorphic Cycladic ophiolite nappe (Kallithea nappe in Samos) above the blueschist unit is generally regarded to be due to horizontal crustal extension (e.g. Böger, 1983; Lister et al., 1984). Böger (1983) suggests a generalized age of 8–10 Ma for nappe emplacement across the entire Aegean. West of Kallithea (Fig. 2), the Katavasis complex is in fault contact with the Kallithea nappe (Fig. 4). Mezger and Okrusch (1985) speculated that the Katavasis complex belongs to the Cretaceous high-temperature rocks, which form crystalline slices within the Cycladic ophiolite nappe. If so, the Katavasis complex would belong tectonically to the Kallithea nappe. We will return to this problem below.

Fig. 5. Orientational data (lower-hemisphere equal-area projections). (a) L_{str1/2}; (b) S_3; (c) L_{str3}; (d) L_{cren3}; (e) S_4; (f) L_{cren4} and F_4 axes.
The Karlovasi, Pyrgos and Mytilini graben are filled with Miocene/Pliocene fluviatile and lacustrine sediments (Table 2, Fig. 3). Above the Basal Conglomerate Formation follow the Pythagorion and Hora Formations. Both formations also laterally interfinger with each other. The sediments of the Hora Formation are thought to have formed in a deeper basin than the limestone of the Pythagorion Formation (Weidmann et al., 1984). A major angular unconformity occurs on top of the Hora Formation. Lacustrine sedimentation is succeeded by fluviatile conglomerate of the basal Mytilini Formation (Old Mill Beds sensu Weidmann et al., 1984). Weidmann (1984, fig. 6, p. 486) showed that in some places the unconformity occurs on top of the Old Mill Beds, whereas in other places it occurs below the Old Mill Beds. This difference might indicate that the unconformity did not occur at the same time in all parts of the basin or it

Table 1
Geology of the nappes of Samos Island

<table>
<thead>
<tr>
<th>Nappes</th>
<th>Regional tectonic unit</th>
<th>Cretaceous and Tertiary metamorphism</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kallithea nappe</td>
<td>Composite Cycladic ophiolite nappe</td>
<td>Non-metamorphic</td>
<td>Chaotic brittle disruption; melange-like internal structure</td>
</tr>
<tr>
<td>(uppermost nappe)</td>
<td></td>
<td></td>
<td>At ~10 Ma intruded by up to ~10 m thick igneous dikes; dikes do not intrude directly overlying Kallithea nappe</td>
</tr>
<tr>
<td>Katavasis complex</td>
<td>Composite Cycladic ophiolite nappe</td>
<td>Upper amphibolite facies (no Tertiary blueschist-facies overprint)</td>
<td></td>
</tr>
<tr>
<td>Sızuk nappe</td>
<td>Cycladic blueschist unit (ophiolitic melange)</td>
<td>Blueschist facies followed by greenschist facies overprint</td>
<td>Resembles metabasite association of Syros Island</td>
</tr>
<tr>
<td>Ampelos nappe</td>
<td>Cycladic blueschist unit (post-Carboniferous cover nappe)</td>
<td>Blueschist facies followed by greenschist facies overprint</td>
<td>Two augengneiss slices in lower part yielded Pb/Pb zircon ages of 230–240 Ma</td>
</tr>
<tr>
<td>Agios Nikolaos</td>
<td>Cycladic blueschist unit (Carboniferous basement nappe)</td>
<td>Blueschist facies followed by greenschist facies overprint</td>
<td>Orthogneiss sample yielded Pb/Pb zircon age of 302 Ma; Carboniferous amphibolite-facies metamorphic relics &gt;1500 m thick sequence of monotonous dolomite marble</td>
</tr>
<tr>
<td>Kerketas nappe</td>
<td>Basal unit</td>
<td>Incipient high-pressure metamorphism followed by greenschist facies overprint</td>
<td></td>
</tr>
</tbody>
</table>

The Karlovasi, Pyrgos and Mytilini graben are filled with Miocene/Pliocene fluviatile and lacustrine sediments (Table 2, Fig. 3). Above the Basal Conglomerate Formation follow the Pythagorion and Hora Formations. Both formations also laterally interfinger with each other. The sediments of the Hora Formation are thought to have formed in a deeper basin than the limestone of the Pythagorion Formation (Weidmann et al., 1984). A major angular unconformity occurs on top of the Hora Formation. Lacustrine sedimentation is succeeded by fluviatile conglomerate of the basal Mytilini Formation (Old Mill Beds sensu Weidmann et al., 1984). Weidmann (1984, fig. 6, p. 486) showed that in some places the unconformity occurs on top of the Old Mill Beds, whereas in other places it occurs below the Old Mill Beds. This difference might indicate that the unconformity did not occur at the same time in all parts of the basin or it

Table 2
Neogene basins on Samos Island

<table>
<thead>
<tr>
<th>Formation</th>
<th>General lithology</th>
<th>Distinct sedimentologic features</th>
<th>Stratigraphic age</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kokkarion</td>
<td>Porous lacustrine limestone, tufaceous marl and silt</td>
<td>Erosional surface at top</td>
<td>Zanclean</td>
</tr>
<tr>
<td>Mytilini Formation</td>
<td>Conglomerate, sandstone, siltstone, tufaceous marl, paleosol, channel-fill deposits, chalky lacustrine limestone, tufaceous marl and sand at top</td>
<td>Contains pebbles of volcanic rocks, Cycladic blueschist unit and Kerketas nappe</td>
<td>Late Tortonian and Messinian</td>
</tr>
<tr>
<td>Hora Formation</td>
<td>Thick-bedded limestone, thin-bedded marl, tuff</td>
<td>&gt;400 m thick; marls contain abundant slump folds; deep lacustrine deposition; pronounced unconformity on top</td>
<td>Early Tortonian</td>
</tr>
<tr>
<td>Pythagorion Formation</td>
<td>Thick-bedded lacustrine limestone, locally silt, gravel and tuff</td>
<td>Desiccation cracks, wave ripples and bioturbation features attest to shallow lacustrine deposition</td>
<td>Early Tortonian</td>
</tr>
<tr>
<td>Basal Conglomerate</td>
<td>Poorly sorted angular cobble, conglomerate, rare thin tuff intercalations</td>
<td>Contains clasts of Agios Nikolaos, Ampelos and Selçuk nappes</td>
<td>Serravallian</td>
</tr>
</tbody>
</table>

For radiometric age data from tuff horizons see Weidmann et al. (1984) and Fig. 3.
might indicate that the Old Mill Beds are time-transgressive.

4. Metamorphic history

The $P$–$T$ conditions for high-pressure metamorphism (Table 3, Fig. 6) reveal mildly blueschist-facies conditions in the Kerketas nappe and a pronounced metamorphic break (up to 10 kbar) towards higher pressures and temperatures above the Kerketas nappe. Within the Cycladic blueschist unit, maximum metamorphic conditions decreased structurally upward. As will be shown below, maximum high-pressure assemblages in the Cycladic blueschist unit developed during the first deformational event (D1) and are therefore referred to as M1. The associated S1 foliation was porphyroblastically overgrown by glaucophane, chloritoid and kyanite during a static growth event (Fig. 7a). Maximum pressure in the Kerketas nappe occurred during the D2 deformation and therefore the mildly blueschist-facies event in the Kerketas nappe is regarded as M2 (Table 3). During D2, the M1 high-pressure assemblages in the Cycladic blueschist unit were replaced by M2 transitional blueschist–green schist-facies assemblages.

A subsequent Barrovian-type metamorphic overprint (M3) was characterized by the prograde formation of garnet and more rarely by biotite in metapelite of the Agios Nikolaos and Ampelos nappes (Chen, 1995). Chen et al. (1995) estimated about 6–7 kbar and 450–490°C for M3 with slightly higher temperatures in the western than in the eastern part of the island. The data show that M3 occurred during further decompression but increasing temperature (Table 3).

Age data for the Cycladic blueschist unit for a number of islands (e.g. Sifnos, Naxos, Ios, Syros, Tinos, Ikaria) across the entire Aegean are remarkably similar and are interpreted to date the peak of high-pressure metamorphism at $\geq 50$ Ma. Cooling below $\sim 350$–$450$°C (assumed phengite closure temperature) took place between 35 and 40 Ma and the Barrovian-type overprint occurred at about 18–25 Ma (Altherr et al., 1982; Wijbrans and McDougall, 1988; Wijbrans et al., 1990; Bröcker et al., 1993; Bröcker and Enders, 1999) (Fig. 6) concurrent with and followed by calc-alkaline magmatism. Unpublished K/Ar age data on white mica from Samos Island by H. Kreuzer are interpreted to agree with this age pattern (Chen et al., 1995).

5. Deformation history

Detailed structural studies were conducted at the tectonic contacts between the various tectonostratigraphic units and within the Agios Nikolaos, Ampelos and Kallithea nappes and the Neogene graben sediments. Based mainly on microstructural relations of mineral parageneses and overprinting criteria, we recognized sets of structures formed during five major deformational events (D1–D5). In general, D3 is the dominant ductile deformation. Earlier structures and fabrics have been substantially modified and transposed and therefore only a little information about the

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**Table 3**

<table>
<thead>
<tr>
<th>Nappe</th>
<th>Metamorphic event</th>
<th>Lithology and mineral assemblagesa</th>
<th>$P$–$T$ conditions</th>
</tr>
</thead>
<tbody>
<tr>
<td>Selcuk nappe</td>
<td>M1</td>
<td>Metabasite: opx + zo + ep1b and gl1 + ep2 + phen1 + ab1 + chl1 + sp1 + rt + hem + cc1</td>
<td>400–470°C, 10–14 kbar</td>
</tr>
<tr>
<td>Ampelos nappe</td>
<td>M1</td>
<td>Metabasite: gl1 + ep2 + phen1 + chl1 + ab1 + sp1 + rt + hem + cc1</td>
<td>420–490°C, 11–13 kbar</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Metapelit: qtz1 + gl1 + phen1 + par1 + chl1 + ctd1 + ky1 + ab1 + car + tur1c</td>
<td>~ 500°C, ~15 kbar</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Metapelite: bt + phen2 + chl2 + ab2 + ep3</td>
<td>450–490°C, 6–7 kbar</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Metapelite: gl1 + ep1 + grt1 + phen1 + par1 + chl1 + ab1 + clz + sp1 + rt + hem + cc1</td>
<td>500–550°C, &gt;12 kbar</td>
</tr>
<tr>
<td>Agios Nikolaos nappe</td>
<td>M1</td>
<td>Metapelite: qtz1 + gl1 + ep1 + grt1 + phen1 + par1 + chl1 + ab1 + clz + sp1 + rt + hem + cc1</td>
<td>520–530°C, ~19 kbar</td>
</tr>
<tr>
<td></td>
<td>M2</td>
<td>Metapelite: qtz2 + phen2 + par2 + chl2 + ep3 + ab2</td>
<td>380–420°C, 7–8 kbar</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Metabasite: calc. amph + chl2 + par2 + pl2f</td>
<td>420–450°C, 9.5–10 kbar</td>
</tr>
<tr>
<td>Kerketas nappe</td>
<td>M1</td>
<td>Metapelite: grt1 + bt + phen1 + chl1 + ab1 + ep1d</td>
<td>450–490°C, 6–7 kbar</td>
</tr>
<tr>
<td></td>
<td>M2</td>
<td>Impure marble: qtz + phen + par + chl + tlc + dol + cc</td>
<td>350–400°C, 7–10 kbar</td>
</tr>
<tr>
<td></td>
<td>M3</td>
<td>Metabauxite: crn + hem + ilm + spl + rtg</td>
<td>~ 450°C, 4–6 kbar</td>
</tr>
</tbody>
</table>

a Mineral abbreviations after Kretz (1983).


c Okrusch et al. (1984).

d Chen (1995).

e Chen et al. (1995).

f Will et al. (1998).

g Mposkos (1978).
kinematics of D₁ and D₂ exists. We first establish typical structural characteristics, deformation/metamorphism and overprinting relationships, and then illustrate key geologic aspects from the nappe contacts and the Pyrgos graben.

5.1. D₁/D₂ deformations

Deformation/metamorphism relationships indicate that the first two generations of ductile structures (D₁ and D₂) were related to blueschist- and transitional blueschist–greenschist-facies metamorphism (Fig. 7). In a few outcrops in the Agios Nikolaos and Ampelos nappes either overprinting relationships between D₁ and D₂ were observed or growth of blueschist-facies porphyroblasts occurred between D₁ and D₂ (Fig. 7a) which allowed us to distinguish between D₁ and D₂ structures. In most cases, however, both sets of structures are hard to distinguish and are therefore described together.

Much of the metapelite, quartzite, marble and metabasite of the Agios Nikolaos and Ampelos nappes and, in part, also the gabbro of the Selçuk nappe have a regionally shallow east-dipping S₁ foliation which is...
Fig. 7. Photomicrographs of D1/2 structures in XZ sections (X Y Z principal finite-strain axes). (a) Porphyroblastic glaucophane overgrowing S1 statically. (b) Slightly rotated syn-D1 garnet with strain shadow containing glaucophane (arrow). (c) Rotated D2 albite prophyroblasts; calcic amphibole and sphene, both of which replace glaucophane, make up external and in part also internal foliation; sense of shear is top-to-the-SE. (d) Asymmetrically sheared chloritoid crystal in quartzite of Ampelos nappe directly above Pythagoras thrust; asymmetry indicates top-to-the-E sense of shear.
strongly transposed. $S_1$ is defined by glaucophane, phengite, paragonite, quartz, chlorite, epidote, albite, chloritoid, kyanite, zoisite, sphene and opaques. Omphacite and garnet also grew during $D_1$ (Fig. 7b).

In a few cases it can be demonstrated in thin sections that $S_1$ is axial planar to rootless, highly sheared, small-scale folds ($F_1$). Static growth of glaucophane, chloritoid and kyanite followed $S_1$ and $F_1$.

$D_2$ structures are associated with the breakdown of glaucophane to barroisite, actinolite, white mica, albite and chlorite. Quartz veins folded during $D_2$ are commonly rootless with strongly attenuated limbs and thickened hinges. The quartz veins and $S_1$ were isoclinally folded about moderately ESE-plunging $F_2$ axes and in the hinges of $F_2$ folds a new $S_2$ foliation developed subparallel to $S_1$. In these zones, the orientation of $S_1$ is subparallel to the orientation of $S_1$ in zones where $D_2$ is weak or absent. Therefore, $S_2$ is regionally subparallel to $S_1$. Both foliations are commonly indistinguishable and then termed $S_{1/2}$; $S_{1/2}$ is commonly subparallel to the nappe contacts.

In some outcrops within the blueschist unit, glaucophane, barroisite, albite, white mica, chloritoid and kyanite are strongly lineated in $S_{1/2}$ and this lineation has a WNW–ESE orientation ($L_{str1/2}$, Fig. 5a). These mineral lineations are parallel to aligned quartz–albite aggregates and strain shadows around $D_1$ garnet and $D_{1/2}$ albite (Figs. 7b and c). $Rt/ϕ$ analysis of quartz and albite grains in the quartz–albite aggregates shows that the maximum elongation direction of the quartz and albite grains parallels the alignment of the quartz–albite aggregates, suggesting that at least the quartz–albite alignment represents a true stretching lineation.

In zones of strong $D_{1/2}$ deformation, $Rt/ϕ$ work on dynamically recrystallized quartz grains yielded strain ratios of $X/Z ≥ 4.5$ ($X ≥ Y ≥ Z$, principal strain axes). These are minimum estimates because quartz is recrystallized.

Kinematic indicators associated with $S_{1/2}$ are asymmetric strain shadows around garnet containing barroisite, chloritoid and glaucophane (Fig. 7b), rotated glaucophane, chloritoid and albite (Fig. 7c), asymmetries of chloritoid crystals (Fig. 7d), and asymmetries of $D_{1/2}$ isoclinal folds. The kinematic indicators do not supply a consistent sense of shear (Fig. 8).

The lack of a consistent sense of shear might be due to a general non-coaxial deformation. To verify this, the degree of non-coaxiality for $D_{1/2}$ was estimated using the approach of Wallis et al. (1993). In a general shear-flow regime, particles with an aspect ratio above a critical value will rotate until they reach a stable orientation whereas particles below this critical value will rotate freely (Jeffery, 1922; Ghosh and Ramberg, 1976). According to Passchier (1987), the value of this critical aspect ratio ($R_c$) is a function of the degree of non-coaxiality ($W_m$) only, and is expressed by:

$$W_m = R_C^2 - 1/R_C^2 + 1.$$  

Critical prerequisites for applying this method are reasonably homogeneous deformation at the mesoscopic scale and a relatively large number of particles (Passchier, 1987). Five samples of chloritoid–kyanite schist from the Ampelos nappe containing abundant syn-$D_1$ chloritoid porphyroblasts and another three samples of Ampelos metabasite containing abundant
syn-D$_2$ albite were selected, and the rotation angles of the porphyroblasts were measured with respect to the mesoscopic S$_{1/2}$ foliation. The results for one representative sample is shown in Fig. 9(a). The values for R$_C$ range from 2.9 to 3.6; the mean of R$_C$ for the five samples containing D$_1$ chloritoid crystals is 3.3 (corresponding to $W_m = 0.83$), and the mean of the three samples containing D$_2$ albite is R$_C = 3.2$ which corresponds to $W_m = 0.82$.

5.2. Main D$_3$ deformation

D$_3$ structures were associated with greenschist-facies mineral assemblages. In some outcrops, we observed that S$_{1/2}$ has been thrown into a set of isoclinal, commonly rootless F$_3$ folds. F$_3$ folds have similar morphological characteristics to F$_{1/2}$ folds. The S$_3$ foliation is defined by oriented growth of chlorite, albite, quartz, white mica and, in part, biotite and actinolite or a differentiated layering of quartz and sheet silicates on a 0.3–1.5 mm scale. S$_3$ transposed S$_{1/2}$ and became the dominant regional foliation (Fig. 5b). An associated stretching lineation ($L_{str3}$) characterized by the preferred orientation of mica and chlorite, aligned quartz aggregates and quartz fibres in strain shadows developed. R$_{f/\phi}$ analysis of quartz grains in the quartz aggregates and in phyllosilicate-bearing quartzite shows that the maximum elongation direction of the quartz grains parallels the alignment of the quartz aggregates and the preferred orientation of mica and chlorite. L$_{str3}$ has a variable orientation across the island (Figs. 5c and 8). In addition to L$_{str3}$, a crenulation lineation ($L_{cren3}$ in Fig. 5d), is defined by the intersection of S$_{1/2}$ and S$_3$. Throughout the region of strong D$_3$ shear, where D$_3$ tectonites are mylonitic and R$_{f/\phi}$ work on dynamically recrystallized quartz grains yielded strain ratios of $X/Z \geq 5$, the angle between L$_{str3}$ and L$_{cren3}$ and F$_3$ axes does not exceed 10°. In intervening low-strain areas, the angle between L$_{str3}$ and L$_{cren3}$ and F$_3$ is of the order of 45–60° (in low-strain areas, strain ratios as obtained from R$_{f/\phi}$ analysis of dynamically recrystallized quartz were $X/Z \leq 3$).

A detailed description of consistent top-to-the-E/ENE D$_3$ sense-of-shear indicators directly above the Selçuk and Pythagoras thrusts (see also Fig. 10a) will be given below. Away from these nappe contacts, especially in the central part of the island, the shear senses are less consistent (Figs. 8 and 11), although we infer a general top-to-the-E/ENE sense of shear from the data.

Microstructures show that Fe–Mg silicates were substantially altered to chlorite during D$_3$ (Fig. 10b). In zones of strong D$_3$ deformation, glaucophane-bearing metabasite has been completely reconstituted to chlorite–albite schist. Most of the albite grew during D$_3$ as indicated by rotated D$_3$ microstructures within albite. Garnet also grew during the early stages of D$_3$. Microprobe work on phengite and chlorite in D$_3$ strain shadows shows that D$_3$ structures started to develop at different stages. Early D$_3$ structures are frequent in the Agios Nikolaos nappe and the lower parts of the Ampelos nappe and developed before the climax of M$_3$, i.e. before the growth of biotite and M$_3$ garnet. Phengite barometry (Massone, 1995) suggests that the early D$_3$ structures started to develop at about 7 kbar with phengites in the Agios Nikolaos nappe having slightly but consistently higher Si values per formula unit than those in the Ampelos nappe.

However, in most of the Ampelos nappe, especially in its upper parts, but also in the other nappes of the Cycladic blueschist unit, D$_3$ structures developed during and closely after the peak of M$_3$ metamorphism. These late D$_3$ structures are more frequent than...
Fig. 10. Photomicrographs of $D_1$ structures. (a) Rotated marble clast in calcareous phyllite above the Pythagoras thrust on the eastern side of the Kerktas Massif indicating top-to-the-E sense of shear. (b) Asymmetric strain shadow around garnet; note that garnet breaks down to chlorite; sense of shear is top-to-the-E. (c) Riedel structures at the base of the Kallithea nappe; sense of shear is top-to-the-NW. (d) Outcrop photograph of cataclastic fault zone at the top of the Katavasis complex north of Kallithea where Kerktas nappe (upper half of photograph) occurs between Katavasis complex and Kallithea nappe.
Fig. 11. Quartz c-axis fabrics from D₃ tectonites from various parts of Samos Island (for localities of samples refer to Fig. 8); principal strain axes X, Y and Z and the S₃ foliation have the same position for all samples and are only indicated for sample 13.7 (upper left-hand side). The angle β between the central part of the fabric skeleton and the S₃ foliation, the sample numbers and the number of measured grains is also shown. Most of the fabrics have a kinked or roughly orthorhombic topology suggesting deformation with 0 < Wₘ < 1. The fabrics do not have a consistent sense of asymmetry although we interpret the majority of the fabrics to indicate a top-to-the-E sense of shear.
early D3 structures and are characterized by asymmetric strain shadows around M3 garnet (Fig. 10b). Biotite in these strain shadows was almost completely replaced by chlorite. Some of the late-D3 ductile structures progressively grade structurally upwards into D3 normal faults (Fig. 12).

In order to quantify the degree of non-coaxiality for the D3 deformation, six samples of chlorite–albite schist from within the Ampelos nappe containing abundant syn-D3 albite porphyroblasts were selected, and the rotation angles of the porphyroblasts were measured with respect to the mesoscopic S3 foliation. The result for one of the samples is shown in Fig. 9(b). The values for $R_C$ range from 1.9 to 2.4; the mean of $R_C$ for the six samples is 2.2 which is equivalent to $W_m = 0.71$.

5.3. $D_4$ and $D_5$ deformations

Ductile deformation in the Cycladic blueschist unit during D4 was very limited and not accompanied by extensive recrystallization and grain growth. Fabrics associated with D4 are therefore not pervasively developed. The most prominent D4 structures are folds at the 10–>100 m scale that have been observed in the blueschist unit and in the cover. S3 has been thrown into a set of mostly tight, commonly west-vergent F4 folds. F4 folds in the deeper parts of the blueschist unit are tight and associated with a generally NE-dipping crenulation cleavage (S4, Fig. 5e). Here, the older S3 foliation is crenulated and in part kinked, and irregular or stylolitic solution-transfer seams marked by concentrations of opaques, sericite and chlorite form a spaced S4 cleavage. The seams wrap around and thus postdate M3 albite and garnet porphyroblasts. The intersection of S4 with S2 created a pronounced intersection lineation ($L_{crena}$) that defines a strong NNE-trending maximum (Fig. 5f). Higher in the sequence, F4 folds are open to tight with E-dipping to subvertical axial planes and NW/N-trending axes (Fig. 5f). Other prominent D4 structures are discrete reverse faults, the planes of which dip mainly to the east, but also to the west (Figs. 13–16). A prominent WSW-dipping D4 reverse fault is exposed at the contact between the Ampelos nappe and the Mytilini graben at the north coast (Boronkay and Doutsos, 1994) and places marble onto Early Tortonian marls of the Hora Formation. D4 structures only occur in Miocene sediments below the unconformity at the base of the Mytilini Formation, supplying an age constraint for D4.

D5 is an entirely brittle event and includes a set of E–W- to NW-striking faults which have a normal sense of displacement. The map pattern (Fig. 2) demonstrates that D5 faults crosscut all other structures and occur in all exposed rock units in Samos. The map furthermore shows that early D5 faults have curved outcrop traces and are cut by steeper normal faults that have relatively straight outcrop patterns. NW-striking D5 faults appear to be the latest set of D5 faults. Both sets of D5 normal faults are due to N/NE-oriented extension (Figs. 4, 15 and 16).

5.4. Ampelos thrust

The Ampelos thrust separates the Ampelos and Agios Nikolaos nappes. The Ampelos thrust was severely affected by later deformations and is poorly
exposed. Garnet-mica schist directly underneath the Ampelos thrust north of Agios Nikolaos preserved D1 structures. D1 shear-sense indicators are internal foliations within rotated M1 garnet and asymmetric strain shadows around M1 garnet which contain glaucophane (Fig. 7b). The rocks of the Ampelos nappe above the Ampelos thrust were strongly sheared during D3 and therefore D1 fabrics are rare. In general, D1 kinematic indicators do not supply a consistent sense of shear in the Agios Nikolaos area.

Within the Agios Nikolaos nappe a discrete, late D3 shear zone is exposed (Fig. 12a). The granitic gneiss depicts a moderately to steeply E-dipping mylonitic S3 foliation. On S3 an ESE-plunging Lstr3 (Fig. 12b) associated with top-ESE shear-sense indicators developed. The mylonitic foliation is characterized by chlorite and quartz ribbons and grades structurally upwards into a cataclastic foliation in which feldspar and quartz are intensely fractured. The fractures are commonly associated with red–brown iron oxides. Fracturing is accompanied by growth of chlorite, in part as fibrous overgrowth. The ductile and brittle–ductile structures are cut by faults whose kinematics is similar to that of the ductile structures (Fig. 12). The granitic gneiss is interlayered with mica schist that grades over a short distance into a chlorite breccia. The chlorite breccia is cut by moderately dipping faults which are in turn cut by steeply dipping faults. The latter two sets of faults dip either to the SE or NW, and their kinematics is again similar to the ductile fabrics in the granitic gneiss (Fig. 12).
5.5. Selçuk thrust

The Selçuk thrust is best exposed west of the village of Mavradzei (Fig. 13). Here, the Selçuk nappe is represented by strongly sheared flaser gabbro, which contains relic magmatic diopside. The diopside is marginally replaced by rare glaucophane, which in turn is severely transformed into blue–green calcic amphibole. The latter defines, together with albite, clinozoisite, epidote and quartz, an S2 foliation. Associated kinematic indicators are characterized by the replacement of glaucophane to barroisite and supply a top-to-the-W sense of shear. However, in a less well exposed part of the Selçuk thrust southwest of Pagondas, rare kinematic indicators supply a top-to-the-E sense of shear.

The relic S2 foliation is overprinted by a mylonitic ESE-dipping S3 foliation (Fig. 13c). In rare cases, the
new S₃ foliation cuts S₂ at a higher angle and is characterized by the replacement of barroisite by actinolite. The S₃ foliation contains a down-dip stretching lineation (Lstr₃) defined by elongated quartz and actinolite aggregates. Associated kinematic indicators are asymmetric strain shadows around amphibole porphyroblasts, which supply a top-to-the-ESE sense of shear.

Commonly, the steeply dipping foliation planes contain shiny, striated surfaces. The first set of brittle structures produced moderately SE-dipping striations that occur on reactivated S₃ planes and on newly formed outcrop-scale faults (Fig. 13d). Riedel planes of up to one metre in size indicate a dextral, down-dip sense of shear. In thin section, a weak P foliation (sensu Rutter et al., 1986) defined by solution-transfer seams and chlorite can be observed within the composite Riedel structures.

A younger set of crosscutting, subvertical striations also occurs on S₃ planes and on small-scale faults that cut across D₃ faults. Small-scale Riedel structures yielded a reverse sense of shear due to E–W contraction (D₄) (Fig. 13e). D₄ faults are cut by E–W-trending D₅ normal faults (Fig. 13a).

5.6. Pythagoras thrust

The Pythagoras thrust separates the Kerketas nappe from the Ampelos nappe. Impure dolomitic marble of the Kerketas nappe immediately below the Pythagoras thrust at the eastern side of the Kerketas Massif in part preserved D₂ structures. These are a foliation made up by phengite, chlorite, talc, quartz, calcite and opaques and a WNW-trending stretching lineation expressed by phengite alignment. The penetrative foliation and chloritoid alignment in Ampelos quartzite immediately above the Pythagoras thrust has the same orientation (Fig. 14). Rₓ/φ analysis of quartz grains in the quartzite shows that the maximum elongation direction of the quartz grains parallels the phengite and chloritoid alignment, suggesting that this lineation is a true stretching lineation. The chloritoid preserved
asymmetric internal fabrics suggesting a top-to-the-E sense of shear during D2 (Fig. 7d).

Penetrative ductile D3 structures in the Ampelos nappe above the Pythagoras thrust are an E-dipping S3 and an E-plunging Lstr3. Lstr3 immediately above the nappe contact is associated with abundant and consistent top-to-the-E kinematic indicators (Figs. 8, 10a and 11).

The D3 structures are in turn strongly overprinted by D4 faults. Faulting caused pronounced cataclasis of the Kerketas dolomite, which resulted in a structureless, fine-grained microstructure in the dolomite adjacent to the fault. Fault-slip analysis indicates that faulting is due to E–W contraction (Fig. 14c). D4 reverse faults are then cut by W/NW-trending D5 normal faults.

5.7. Kallithea detachment

A structural analysis in the Kallithea nappe is hampered by severe and in part chaotic brittle disruption of all lithological units and poor outcrop conditions. The partly serpentinitized peridotite and the ophiolitic mélange at the base of the Kallithea nappe are cataclastically sheared resulting in a crude, spaced P foliation. The P foliation is defined by the preferred orientation of tabular fragments or compositional layering in clayey and serpentinitic matrix and inclined.
The youngest set of structures (D5) are approximately WNW-striking faults. The cross-section (Fig. 16b) reveals a half-graben geometry of the Pyrgos graben with a major graben-bounding fault at the northeastern side and onlapping geometries at the southwestern side of the graben. Fault-slip analysis reveals that D5 resulted from ENE/NE horizontal extension followed by WNW-oriented horizontal contraction during D4, which in turn is succeeded by NE horizontal extension during D3 (Fig. 16c–e). Crosscutting striations on D3 faults show that during D4 and D5 some D3 faults were reactivated.

6. Tectonic interpretation

6.1. D1 and D2 crustal contraction

Although the kinematic data for D1 and D2 do not supply a consistent sense of shear, they do suggest WSW–ESE tectonic transport. These structures probably formed in the Eocene and earliest Oligocene (Fig. 6) and therefore predated the Miocene ~30° anticlockwise rotation described by Kissel and Laj (1988). If the effects of this rotation were removed, the data for D1 and D2 would indicate NW–SE tectonic transport during and shortly after high-pressure metamorphism.

Kinematic indicators for the early orogenic evolution in the Aegean are scarce. Vandenberg and Lister (1996) reported NW-trending glaucophane lineations from Ios Island. The only shear-sense data for deformation associated with blueschist-facies metamorphism we are aware of are those of Ridley (1984) from Syros Island which indicate top-to-the-SE tectonic transport. Eocene to earliest Oligocene top-to-the-SE tectonic transport towards the foreland is in accord with Late Cretaceous to Eocene top-to-the-SE tectonic transport of the overlying Lycian nappes (Collins and Robertson, 1998), general southward progradation of orogenesis in the Aegean and top-to-the-S tectonic transport during Early Miocene high-pressure metamorphism in Crete (e.g. Fassoulas et al., 1994).

We have not been able to relate the mesoscopic D1 and D2 structures unequivocally to nappes contacts. Deformation/metamorphism relationships (Fig. 6) indicate that the nappes are the kerketas, Agios Nikolaos, Ampelos and Selçuk nappes were largely assembled prior to D3/M4 and therefore formed during D3 and D4. The P–T estimates show that the degree of M3 metamorphism within the Cycladic blueschist unit decreases from the Agios Nikolaos nappe upward towards the Selçuk nappe. Therefore, and because D1 and M4 were followed by a stage of porphyroblastic glaucophane growth, we suggest that initial stacking of the Agios Nikolaos, Ampelos and Selçuk nappes took place during underthrusting and underplating.
that caused deep burial and maximum metamorphic pressure.

During D$_2$ the Cycladic blueschist unit was emplaced onto the Kerketas nappe along the Pythagoras (out-of-sequence) thrust. Deformation/metamorphism relationships show that D$_2$ structures formed during decompression indicating considerable exhumation of the Cycladic blueschist unit during D$_2$ (Fig. 6, see Avigad et al., 1997 for a similar case from Evia Island). These constraints also demand that the Pythagoras thrust must have moved upward relative to the Eocene/Oligocene Earth’s surface in the direction of tectonic transport and therefore resulted from horizontal crustal contraction (cf. Wheeler and Butler, 1994). Thrusting of the Cycladic blueschist unit onto a cold foreland unit, i.e. the Kerketas nappe, caused cooling of the blueschists (Fig. 6). Lister and Raouzaios (1996) described a similar case of nappe stacking during initial decompression and rapid cooling of the Cycladic blueschist unit from the island of Sifnos.

The pressure drop during D$_2$ thrusting requires the removal of ~30–40 km of overburden above the Agios Nikolaos nappe. There is no evidence for Eocene to earliest Oligocene normal faults in the Aegean. The Lycian nappes, which overly the Cycladic blueschist unit in western Turkey show no evidence for normal faulting during this period of time (Collins and Robertson, 1998). Using a value of $W_m = 0.83$ and $R_{XYZ} \geq 4.5$ yields a vertical shortening of more than 30% in zones of strong D$_{1,2}$ deformation (note that this is a minimum estimate because recrystallized quartz grains were used for strain quantification). Ring (1998) showed that a value of >30% vertical shortening in the hanging wall of the Pythagoras thrust associated with an erosion rate <1 mm/a is sufficient to account for the removal of ~30–40 km of metamorphic section within and above the Cycladic blueschist unit. (Note that in order to properly estimate the contribution that >30% vertical shortening of the Cycladic blueschist unit made to the total exhumation of this unit, both the vertical rate at which the blueschist moved upwards through the overburden and the rate of thinning of the remaining overburden at each step along the exhumation path has to be considered, i.e. the velocity-gradient field along the exhumation path has to be specified for both the blueschist and its cover (Feehan and Brandon, 1999; Ring and Brandon, 1999)).

6.2. D$_3$ crustal extension

The overall geometric relationship between D$_3$ structures which overprinted the E-dipping Pythagoras and Selçuk thrusts, the consistent D$_3$ top-to-the-E kinematic indicators immediately above both thrust planes and the general attitude of bedding in the graben sediments in the vicinity of the thrusts (which provides a relation between the fault plane and the Earth’s surface in the Miocene), suggests that top-to-the-E shear during D$_3$ descended relative to the Earth’s surface in the direction of tectonic transport and therefore was caused by horizontal crustal extension. Crustal extension during D$_3$ is corroborated by the progressive development from ductile to brittle conditions of D$_3$ structures in the footwall of the D$_3$ shear zone south of Agios Nikolaos. There, the parallelism between mylonitic L$_{str3}$ and the extension direction as deduced from fault-slip analysis suggests kinematic compatibility during progressive D$_3$ exhumation. The continuous evolution of these D$_3$ structures from ductile to brittle conditions resembles the structural evolution in the footwall of extensional faults in the Basin-and-Range province and other well-known extensional settings (e.g. Lister and Davis, 1989). Likewise, the identical kinematics of ductile D$_3$ structures and their brittle overprint at the Selçuk thrust west of Mavradzei implies kinematic compatibility during progressive exhumation. The D$_3$ half-graben geometry in the Pyrgos graben supplies further evidence for crustal extension during D$_3$.

Despite the consistent D$_3$ kinematic indicators at the Pythagoras and Selçuk thrusts, D$_3$ structures within the Ampelos nappe are somewhat difficult to interpret. The overall lack of a strong preferred orientation of L$_{str3}$ across the island and the variable sense of shear may suggest a relatively strong component of coaxial flattening during D$_3$. $W_m$ of about 0.71 proves pronounced deviations from simple-shear deformation and together with $R_{XYZ} \geq 5$ gives a minimum vertical shortening of more than 50% in zones of strong D$_3$ shear. We suggest that D$_3$ crustal extension was characterized by pronounced vertical shortening in a complexly ductilely flowing crust. Shear strains apparently localized heterogeneously around pre-existing structures and contacts in the extending rock mass (cf. Vandenberg and Lister, 1996).

The scarcity of overprinting relationships in the Kallithea region hinders an exact interpretation of the relationship between the Kallithea nappe and the Katavasis complex. Two general scenarios are discussed. (1) In the cross-section in Fig. 17(a), the D$_3$ extensional Kallithea detachment is cut by a knife-sharp D$_4$ thrust that eventually brought the Kallithea nappe on top of the Katavasis complex. In this interpretation, the Katavasis complex and the Kallithea nappe would have occupied very different paleo-geographic positions—the Kallithea nappe originated in an internal position, whereas the occurrence of the Katavasis complex even below the Kerketas nappe indicates an external origin. This palinspastic inference is not corroborated by the regional geology
(Jacobshagen, 1986). (2) The cross-section in Fig. 17(b) is an alternative explanation in which both cataclastic shear zones link up. The knife-sharp fault on top of the Katavasis complex would be a splay of the underlying D3 Kallithea detachment. Consequently this detachment system, and therefore the emplacement of the Kallithea nappe, would be < 10 Ma. This interpretation allows us to treat the Kallithea nappe and the Katavasis complex as one paleogeographic entity before emplacement onto the Kerketas nappe and the Cycladic blueschist unit. It would also give this entity an originally internal position, which is compatible with regional geologic constraints. Therefore, we envision scenario (2) to be more realistic. However, the late-D3 emplacement of the Kallithea nappe and the Katavasis complex between 9 and 10 Ma (the lower age limit for the Kallithea detachment is given by the maximum age for D4, see below) was top-to-the-NW/NNW and therefore in a different direction as the general D3 transport deep in the nappe pile.

The envisioned < 10 Ma emplacement of the Kallithea nappe took place when the rocks of the Cycladic blueschist unit had already reached the surface (clasts from the Agios Nikolaos, Ampelos and Selçuk nappes in the Serravallian Basal Conglomerate Formation). Therefore, the extensional emplacement of the Kallithea nappe did not aid exhumation of the Cycladic blueschist unit on Samos Island. Because no pebbles from the Kerketas nappe were observed as components in the Basal Conglomerate Formation, we infer that this tectonically deepest unit was not exposed in pre-Tortonian times and the Kallithea detachment might have played a role in its exhumation.

Deformation/metamorphism relationships indicate that some D3 structures developed before the Early Miocene climax of M3 and should therefore be Oligocene in age (see also below). Several lines of evidence suggest that D3 continued into the Middle and Late Miocene: (1) the Serravallian onset of sedimentation in the graben, (2) the inferred emplacement of the Kallithea nappe between 9 and 10 Ma, (3) microprobe work which indicates that D3 structures developed at and after the peak of Early Miocene greenschist-facies M3 metamorphism, (4) the Aegean was subjected to pronounced regional crustal extension during the Miocene.

6.3. D4 contraction and D5 extension

D4 and D5 are characterized by faulting along moderate to steep faults, which may have caused limited block rotations across the island. Furthermore, the dip pattern of bedding in the vicinity of and away from faults in the graben sediments supplies a measure for the degree of reorientation during faulting. We are confident that the majority of D4 and D5 faults have not been rotated to any great degree after their formation. Consequently, D4 results from horizontal crustal contraction and D5 from horizontal crustal extension.

Based on the youngest age for the Hora Formation and the oldest age for the unconformably overlying Mytilini Formation, Weidmann et al. (1984) placed the D4 event into the period between 9 and 8.6 Ma. However, since this unconformity did apparently not develop at the same time across the entire basin, the D4 event might have lasted somewhat longer than 8.6 Ma. Contractual deformation caused a major change in the sedimentary facies pattern; deep lacustrine sediments of the Hora Formation were replaced by fluvial deposits of the Mytilini Formation.
suggesting that contraction was associated with pronounced uplift. Despite a large component of coaxial deformation, the general tectonic transport during D₄ appears to be top-to-the-W. We believe that the presently E-dipping structural grain (Figs. 2, 4 and 5b) of Samos Island was caused by D₄ reverse faulting and west-vergent folding.

The structural relationships on the western side of the Mytilini graben at the north coast indicate that D₄ reverse faulting postdated graben formation. This and the relationship between ductile D₃ structures and graben formation in the Pyrgos area are thought to supply strong evidence against a transpressional origin of the Miocene graben on Samos Island.

In the latest Miocene, extensional deformation (D₅) revived and continues until the present. The pattern of S-dipping listric and planar normal faults in the southern half of the island and N-dipping faults in the northern half suggest that D₅ largely controlled the present-day topography. Therefore, we regard Samos Island as a tilted horst block within the Aegean Sea.

7. Discussion

7.1. Interpretation of P–T data and subduction-zone retreat

The shape of the P–T path in Fig. 6 illustrates a major change in the thermal structure at ~30–35 Ma. The onset of high-pressure metamorphism in Crete at about 25 Ma (Seidel et al., 1982) occurred afterwards. Thomson et al. (1998) estimated an age of 32–36 Ma for the initiation of subduction in Crete. The present plate-convergence rate of Africa with respect to Eurasia is ~15 mm/a (Livermore et al., 1985), whereas the Aegean region is moving southward to Africa at a rate of 30–40 mm/a (Le Pichon et al., 1995). This results in subduction rates of ~50 mm/a at the modern Hellenic subduction zone. Assuming an average geometry of an accretionary wedge (basal décollement ~7°, Davis et al., 1983), and that the present subduction rates are largely representative for the last ~35 Ma, we would expect that subduction to a depth of ~35 km (as deduced from P–T data on Crete, Seidel et al., 1982) would have occurred in about 6–7 Ma. The latter data suggest that the high-pressure rocks of Crete started to subduct at about 31–32 Ma. We note that the estimates for the onset of subduction beneath Crete of Thomson et al. (1998), our crude estimate, and the change in the thermal profile in the Cyclades/Samos region at ~30–35 Ma are similar. We believe that this time marks the obstruction of subduction in the Cyclades and the retreat of the subduction zone towards the south. The distribution of Miocene high-pressure rocks (Fig. 1) suggests that the subducting slab retreated only in the region south of the Aegean Sea. Asthenospheric return flow associated with the southward retreat of the subduction zone may have facilitated the profound Miocene magmatism in the Cyclades which in turn may have caused the increased thermal profile and prograde M3 metamorphism. Because extensional deformation is commonly linked in some fashion to the retreat of the subduction zone (Lister et al., 1984; Buick, 1991), horizontal crustal extension should have commenced in the Early Oligocene, i.e. at about 30–35 Ma. Structures due to crustal extension were indeed dated at 32–34 Ma by Raouzais et al. (1996) from Sifnos Island.

However, a number of studies (Urai et al., 1990; Buick, 1991; Faure et al., 1991; Lee and Lister, 1992; Forster and Lister, 1999) showed pronounced Middle Miocene to Recent crustal extension (e.g. formation of Cretan Sea) which can be observed across the entire present-day arc. The onset of graben development across the Aegean largely coincides with the inception of the North Anatolian fault (Barka and Hancock, 1984) and the westward extrusion of Anatolia (Le Pichon et al., 1995). We concur with Le Pichon et al. (1995) that initial retreat of the Hellenic slab in the Oligocene and Early Miocene exerted a pull on Anatolia which subsequently controlled its lateral extrusion. We propose that the extruding Anatolian microplate in turn impinged on the Hellenic slab and caused a second phase of retreat of the Hellenic slab from a position underneath Crete to its present position north of the Libyan coast.

The present-day relative plate motion (Le Pichon et al., 1995) in the upper-plate behind the magmatic arc, i.e. in the Cyclades, is divergent indicating that the Cyclades are a continental rift. According to the flow-line modelling of retreating subduction zones by Garfunkel et al. (1986), the asthenospheric flow in the arc and back-arc region behind retreating slabs is forced to accelerate in the horizontal to fill the free space causing plate divergence (rifting) (Fig. 18). We speculate that this process probably was and still is the
dominant cause for extending the upper plate in the Aegean.

7.2. Differential crustal extension during D3

D3 was characterized by a large degree of vertical shortening which might have caused the variable senses of shear within the Ampelos nappe. Nonetheless, the two pre-10 Ma D3 shear zones, which overprinted the Pythagoras and Selçuk thrusts, had a relatively uniform top-to-the-ENE sense of shear. We propose that the simplest interpretation of the D3 movement pattern is that these two D3 shear zones are the main structures along which crustal-scale extension was localized. We interpret the D3 kinematic indicators to reflect, in general, top-to-the-ENE tectonic transport. This transport direction for extensional deformation is in marked contrast to the uniform NNE–SSW-trending transport in the Cyclades and in the Menderes Massif of western Turkey. As shown above, both regions underwent considerably different degrees of crustal stretching. Consequently, a NNE-trending sinistral wrench corridor that accommodated differential crustal extension should have separated the Cyclades and the Menderes Massif (Fig. 19).

We propose that this wrench corridor largely controlled D3 extension in Samos. The local direction of extension in such a sinistral wrench corridor would be ENE; however, during ongoing deformation early formed stretching lineations would rotate anticlockwise and would finally occupy a similar orientation as those in the Cyclades and the Menderes Massif. The Miocene paleomagnetic rotations of about 30° reported by Kissel and Laj (1988) might have been caused in this wrench corridor. Bartley and Glazner (1991) and Janecke et al. (1991) have reported similar patterns of rotation in displacement transfer zones within extensional belts. We envision that differential extension was caused by the localized subduction-zone retreat in the Aegean which eventually created the Aegean Sea and the Hellenic arc.

If our interpretation of the existence of a sinistral wrench corridor was correct, it would lend strong support into a transtensional origin of the Miocene graben on Samos Island. The rapid subsidence recorded by the relatively thick section of the Hora Formation, the pronounced lateral facies changes from shallow to deeper lacustrine conditions recorded by the Pythagorion and Hora Formations and the difference in the occurrence of the unconformity are typical features of pull-apart basins (Christie-Blick and Biddle, 1985).

8. Conclusions

We reached the following conclusions:

1. Five generations of structures (D1–D3) have been recognized on Samos Island. D3 caused the dominant regional deformation. D2, D4 and D5 caused attenuation of the tectonostratigraphic pile.
2. D1/2 structures caused nappe stacking and were subsequently highly overprinted. Therefore the kinematics of early nappe stacking remains controversial.
3. D3 crustal extension commenced in the Early Oligocene and is largely coeval with a marked change in the thermal structure. We propose that this change in the thermal structure coincided with the retreat of the subduction zone towards the external Hellenides. The emplacement of the Kalilthea nappe as part of the Cycladic ophiolite nappe occurred during the final stages of D3 crustal extension between 9 and 10 Ma. In general, tectonic transport in the deep parts of the nappe pile was top-to-the-ENE, whereas the late, shallow-level emplacement of the Kalilthea nappe was top-to-the-NW/NNW directed.

Fig. 19. Proposed sinistral wrench corridor, which is thought to have accommodated differential extension between the Cyclades and western Turkey. Paleomagnetically determined rotations from Kissel and Laj (1988) are indicated. Vertically ruled line separates areas of top-to-the-N and top-to-the-S sense of shear during crustal extension in the Aegean and the Menderes Massif, respectively; note that this boundary has been sinistrally displaced by ~150 km and that the Vardar–Izmir–Ankara suture is bent to the south.
4. The geometry of pre-10 Ma D3 extension contrasts with ~N–S extension in the neighbouring regions and might have been controlled by a sinistral wrench corridor that accommodated differential crustal extension between the Cyclades and the Menderes Massif.

5. D4 contraction was a short-lived event, which caused local inversion of the Miocene graben. The cause of this event remains unknown.

6. After the Upper Tortonian, extensional deformation (D5) revived and continues until the present. D5 extension produced W/NW-striking faults that largely control the present-day geomorphology of the island.

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