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Tectonic juxtaposition of blueschists and greenschists in Sifnos Island (Aegean Sea)—implications for the structure of the Cycladic blueschist belt

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Abstract—In the Cyclades, kilometre-thick high-pressure rock sequences, displaying blueschist- and eclogite-facies mineral assemblages, overlie a rock sequence which was thoroughly overprinted in the greenschist and amphibolite facies during its exhumation. Both rock sequences, subducted and metamorphosed at high-pressures in Eocene times, have been considered by previous workers to have been exhumed as a coherent rock unit. In contrast, it is suggested here that the preserved high-pressure rock sequences were exhumed more rapidly and prior to the underlying greenschists. Significant metamorphic, structural and geochronological discontinuities exist across the blueschist–greenschist contact and field evidence suggests that the high-pressure metamorphic rocks in Sifnos are tectonically juxtaposed above the greenschists. These two rock sequences were juxtaposed by a low-angle fault subsequent to the Oligocene–Miocene greenschist-facies overprint. Published geochronological data and petrological criteria are used to show that the high-pressure sequence cooled below 350°C when the rocks now immediately underlying it suffered a greenschist-facies overprint at temperatures of *ca* 450°C. The section inferred to absorb this temperature difference is now missing and it is suggested that it has been cut out by the low-angle fault.

INTRODUCTION

THE Cycladic massif is exposed as a series of small islands in the centre of the Aegean Sea (Fig. 1). It consists of several rock units whose stacking is interpreted to be a result of Alpine convergence (Altherr & Seidel 1977, Bonneau 1984, Papanikolaou 1987). High-angle normal faulting and thinning of the Aegean crust is thought to have commenced in the Late Miocene (Le Pichon & Angelier 1979, Angelier *et al.* 1982), whereas investigations of the ductile deformation in Barrovian metamorphic rocks exposed on Ios and Naxos led Lister *et al.* (1984) to conclude that extension, in the mode of core complex tectonics, had already begun in the Aegean region in the Early Miocene. Lister *et al.* (1984) and Avigad & Garfunkel (1991) showed that low-angle normal faults, rather than thrust faults, dominate the structure of the Cycladic massif, juxtaposing low-pressure (LP) metamorphic rocks and sediments above blueschists and eclogites.

The Cycladic massif (Fig. 1) comprises three structural units separated by low-angle faults. The dominant rock unit, which is the subject of the present work, is the Cycladic blueschist unit which occupies an intermediate position within the nappe pile. The blueschist unit is tectonically overlain by an Upper unit which consists of LP metamorphic rocks, ophiolite remnants and sediments of diverse origin (Durr *et al.* 1978). Weakly metamorphosed sequences exposed in several tectonic windows (Katsikatsos 1977, Papanikolaou 1979, Avigad & Garfunkel 1989) indicate that the blueschist unit is allochtonous on a regional scale. This blueschist unit consists of Alpine metasediments and metavolcanic rocks which overlie a reworked Hercynian basement (Henjes-Kunst & Kreuzer 1982, Andriessen *et al.* 1987). High-pressure metamorphism affected the blueschist unit in the Eocene (Altherr *et al.* 1979) as a result of the NE-directed underthrusting of the Apulian microplate beneath the Eurasian plate (Biju-Duval *et al.* 1976). The exhumation of the high-pressure (HP) rocks in Oligocene–Miocene times was accompanied by a regional greenschist-facies overprint and was followed by the emplacement of granites throughout the Miocene (Altherr *et al.* 1982). These granites usually intrude the blueschist unit but on Tinos island (Fig. 1) an Early Miocene monzogranite also penetrates into the Upper unit.

The lower part of the Cycladic blueschist unit is dominated by Oligocene–Miocene greenschist- and amphibolite-facies rocks in which high-pressure rocks occur as relics. The upper parts, exposed in Sifnos and Syros islands (Fig. 1), consists of a more than 1 km thick sequence of eclogite- and blueschist-facies rocks which were weakly overprinted during decompression. Both rock sequences (the overprinted section and the overlying sequence which was preserved almost intact) are thought to comprise a coherent rock unit subducted and metamorphosed at high-pressures in Eocene times, but differentially overprinted at lower pressures during exhumation (Durr *et al.* 1978, Matthews & Schliestedt 1984).

The present study focuses on the structure of the Cycladic blueschist unit exposed in Sifnos island (Fig. 1), where both the high-pressure rocks and the underlying greenschists are exposed, and their metamorphic and geochronological evolutions are well documented

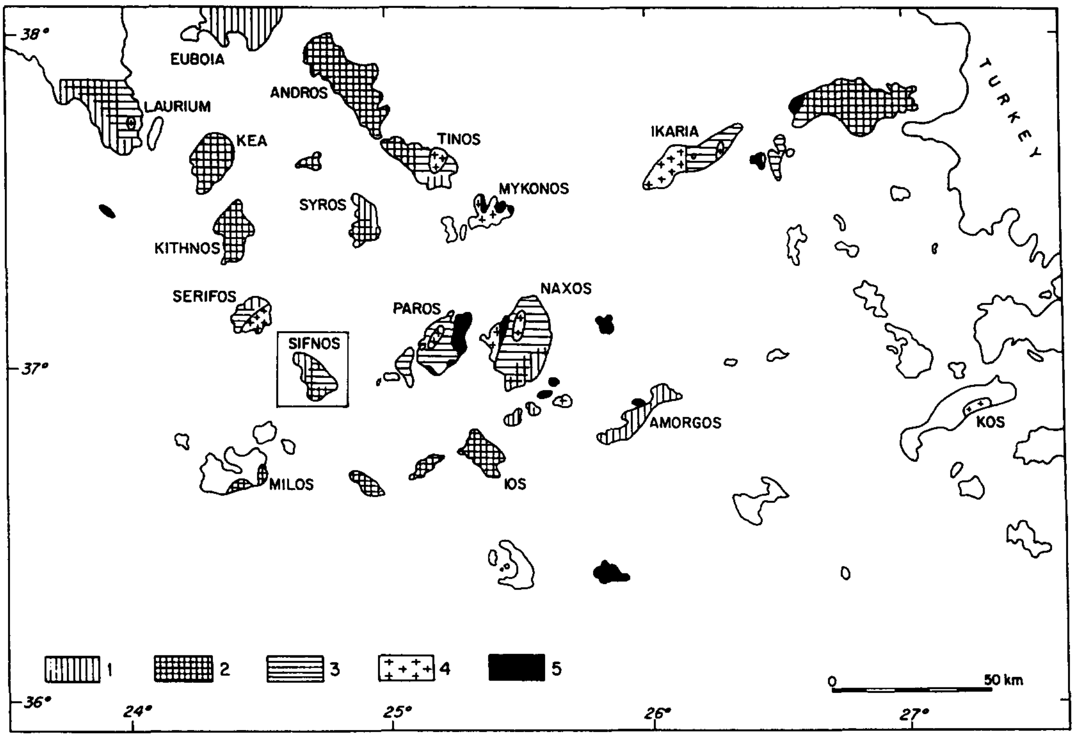


Fig. 1. Geological map of the Cycladic massif (after Altherr *et al.* 1979, 1982). (1) Eocene high-pressure metamorphic rocks. (2) Eocene high-pressure metamorphic rocks overprinted by the Oligocene–Miocene low- to medium-pressure metamorphism. (3) Oligocene–Miocene low- to medium-metamorphic rocks. (4) Miocene granitoids. (5) Undifferentiated rocks of the upper tectonic unit.

(Okrusch *et al.* 1978, Altherr *et al.* 1982, Matthews & Schliestedt 1984, Schliestedt *et al.* 1987, Wijbrans *et al.* 1990, Avigad *et al.* 1991). Field evidence presented here shows that the eclogite–blueschist-facies rocks and the underlying greenschists were tectonically juxtaposed by a low-angle fault. Integration of these field observations with published data show that the low-angle fault marks an important petrological and geochronological break within the Cycladic blueschist unit. The movement sense along the low-angle fault and the relationship between this fault and ductile structures which were formed during decompression from peak *P–T* conditions are described in the following sections.

TECTONOSTRATIGRAPHY

Sifnos Island (Figs. 1 and 2) exposes an approximately 2.5 km thick metamorphic rock sequence of the Cycladic blueschist unit as a series of N-dipping tilted blocks (Fig. 3) produced by the Late Cenozoic faulting in the Aegean Sea. The metamorphic sequence comprises a 'Greenschist unit' (GSU) at the base, and an overlying 'Eclogite–Blueschist unit' (EBU) at the top (Figs. 2 and 3).

The GSU, approximately 1000 m thick, whose base is not exposed, crops out in central Sifnos. The rocks of this unit include metapelites, metabasites and carbonates dominated by Oligocene–Miocene greenschist-facies metamorphic assemblages (Altherr *et al.* 1982, Matthews & Schliestedt 1984). Evidence of the earlier high-pressure metamorphism in this unit is given by the presence of relic eclogite facies rocks and glaucophane–crossite epidote–albite-bearing metabasites showing partial breakdown to lower pressure assemblages (Avigad *et al.* 1991). The EBU consists of a 800 m thick marble sequence (the Main Marble) with prominent interbeds of metavolcanics in its base. Blueschists, eclogites and jadeite gneisses formed by high-pressure metamorphism of a series of alternating acid and basic metavolcanics, metapelites and quartzites overlie the Main Marble. The section ends with a 300 m thick sequence of calcite marbles (Upper Marble) containing layers of dolomite, quartzites and metapelites. High-pressure assemblages are usually well preserved in this unit (Davis 1966, Okrusch *et al.* 1978) but schist intercalations in the Main Marble were partially overprinted (Schliestedt & Matthews 1987).

The contact between the EBU and the GSU lies at the base of the Main Marble which is dipping shallowly (*ca* 15°) to the north. Examination of the contact (Fig. 4) reveals fault gouge, up to 3 m thick (for example, along the road east of Kamares, Fig. 2), dismembered schist layers and quartz veins tectonically mixed with faulted blocks of the underlying GSU, intensive brecciation of the marble, and locally (road east of Kamares, Fig. 2) sub-horizontal slickensides trending 190°/10°. The contact is usually concordant with the lithological layering, but in the southern part of the island in the area of Vathi (Fig. 2) it dips to the south-west, whereas lithological layering and the greenschist-facies foliation in the GSU dip to the north-east. A 5 m thick layer of serpentinite, a few metres beneath the marble (Fig. 2), was probably tectonically incorporated into the section during juxtaposition of the two units. These features of brittle deformation indicate that the contact between the two units has been active as a fault plane with approximately N–S-directed movement. In the following section the existing petrological and geochronological data are used to emphasize the metamorphic and age discordancies between the EBU and GSU, with an attempt to better define the scale of this fault and the sense of movement along it.

METAMORPHISM

P–T–t paths illustrating the metamorphic evolution of the EBU and the GSU are presented in Fig. 6. Peak *P–T* conditions in the EBU were determined by Okrusch *et al.* (1978) and Schliestedt (1986) whereas the details of the *P–T* path of the Greenschist unit have been presented by Avigad *et al.* (1991). Although the shape of the overall *P–T* paths of both units are similar, several important differences in the degrees and, perhaps, the age of the overprinting and the cooling histories of these units need to be emphasized. In the EBU typical mineral assemblages include: (a) in eclogites: garnet + ompha-

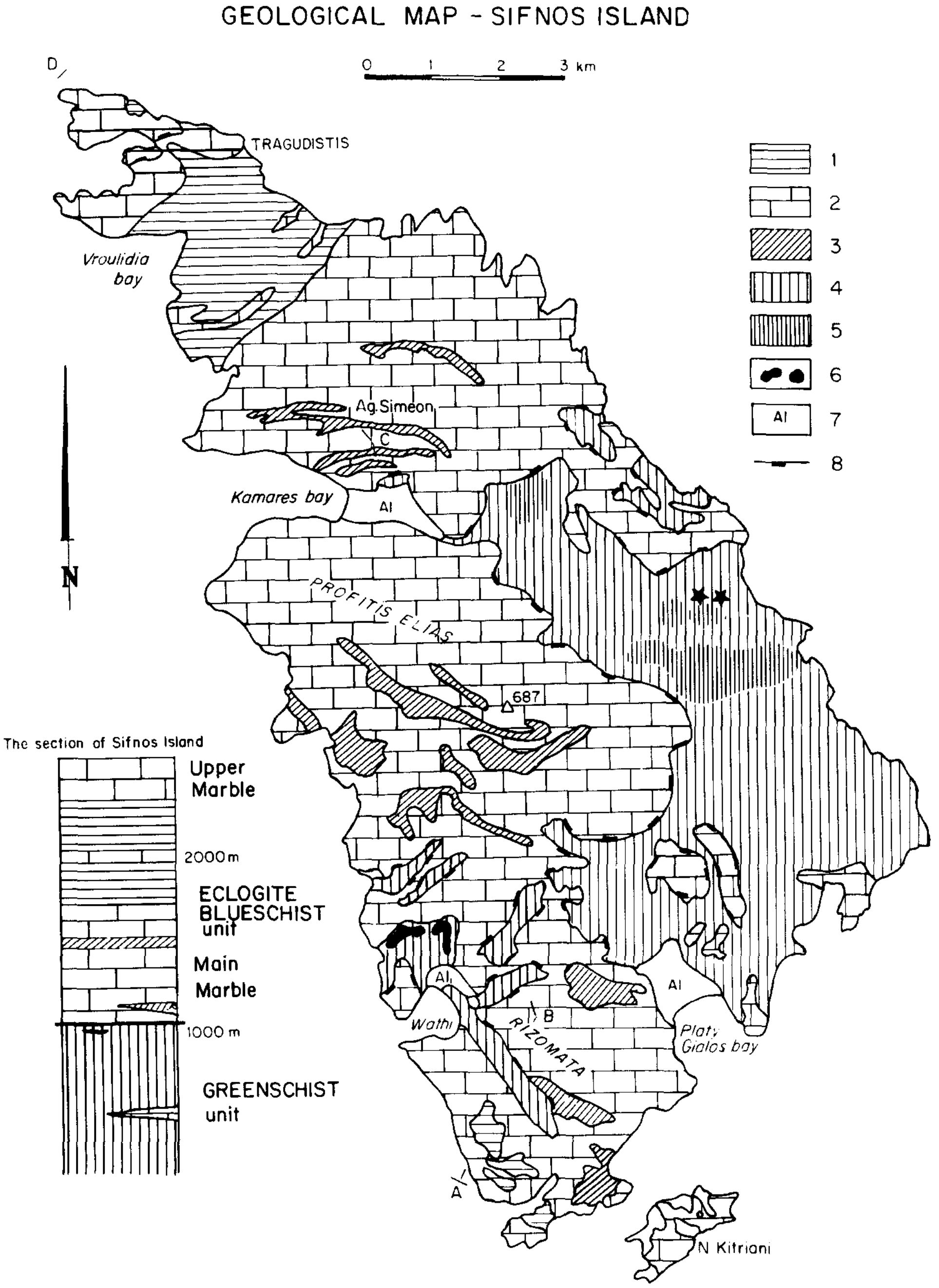


Fig. 2. Geological map of Sifnos Island (modified from Davis 1966) and a schematic lithological section. Legend: (1) eclogite facies schists and gneisses; (2) marbles; (3) partly overprinted schist layers within marbles; (4) greenschist-facies rocks; (5) basic albite-bearing epidote–blueschist-facies rocks in the Greenschist unit; (6) serpentinite; (7) alluvium; (8) low-angle tectonic contact. Stars: relict mica–eclogites in the Greenschist unit.

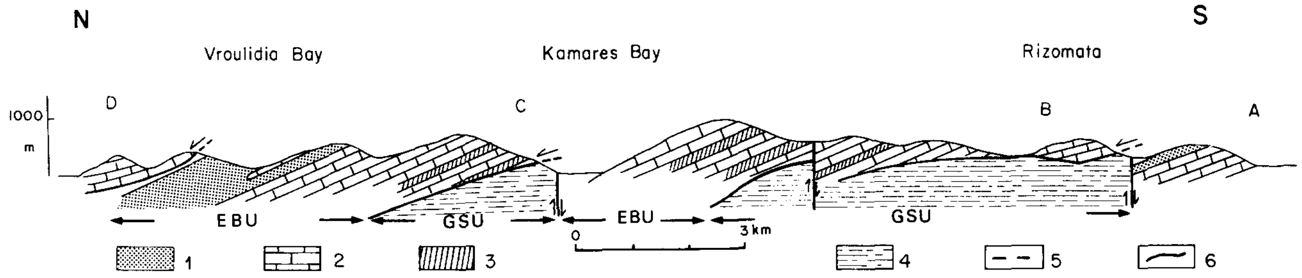


Fig. 3. Approximately N–S structural cross-section of the island of Sifnos (A–B–C–D in Fig. 2). Legend: (1) eclogite facies schists and gneisses; (2) marbles; (3) metamorphically overprinted schists layers within marbles; (4) greenschist-facies rocks; (5) serpentinite; (6) low-angle fault.

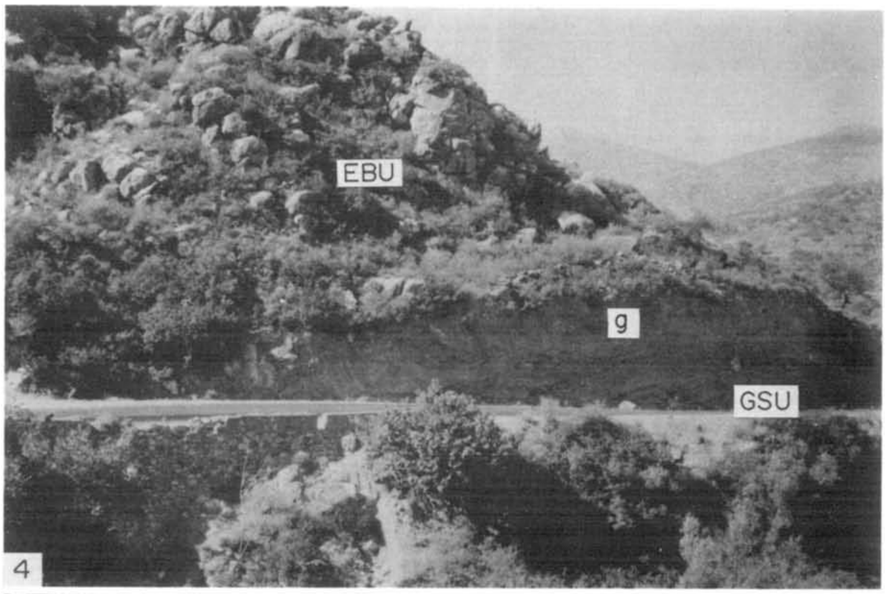
cite + rutile + epidote + phengite; (b) in blueschists: glaucophane + garnet + epidote + phengite + sphene; and (c) in acid gneisses: jadeite + quartz + garnet + phengite. Temperatures of 440–500°C are reported for the high-pressure metamorphism of this unit using garnet–pyroxene Mg/Fe exchange thermometry on eclogites (Okrusch *et al.* 1978, Schliestedt 1986) and oxygen isotope thermometry (Matthews & Schliestedt 1984, revised Matthews 1988). Corresponding pressures averaging at 15 kbar were deduced from the occurrence of jadeite + quartz in acid gneisses and the presence of deerite (Okrusch *et al.* 1978, Evans 1986, Schliestedt 1986). The exhumation path is characterized by crystallization of albite as a result of the breakdown of jadeite and omphacite. In some metabasic rocks, the new albite coexists with glaucophane indicating a transition into the albite-bearing epidote–blueschist facies (cf. Evans 1990, fig. 4). Further transformation at decreasing pressures involved the breakdown of glaucophane and the formation of actinolite and chlorite in the greenschist-facies. The resulting approximate pressure–temperature (*P–T*) path for the EBU (Fig. 6) is consistent with the general preservation of the high-pressure assemblages, and indicates that temperatures did not increase during the exhumation of these rocks. The metamorphic evolution of the GSU is characterized by several stages of metamorphic equilibration along the exhumation path, at successively decreasing pressures and temperatures (Avigad *et al.* 1991). Peak *P–T* metamorphic conditions corresponding to the early high-pressure metamorphic event were deduced from rare eclogite-facies relics. These relics, usually micaschist eclogites, consist of chloromelanite (jd-30%), garnet, phengite and quartz, which are constrained by the petrogenetic grid and by GEOCALC-PTX (Berman 1988, Brown *et al.* 1988 and THERMOCALC (Holland & Powell 1985, Powell & Holland 1988) softwares calculations, to coexist at pressures in excess of 12 kbars at temperatures of 480–520°C (Avigad *et al.* 1991). Metabasic rocks consisting of glaucophane–crossite, albite and epidote equilibrated along the exhumation path in the field of the albite-bearing epidote–blueschist facies, at pressures of 8–10 kbar (Avigad *et al.* 1991). These rocks were further replaced by chlorite, actinolite, albite and epidote, giving rise to a typical greenschist-facies assemblage. The *P–T* conditions of the final equilibration in the greenschist-facies have been estimated by oxygen isotope studies and mineral equilibria indicating temperatures of 400–500°C and pressures of 5–7 kbar (Matthews & Schliestedt 1984, revised Matthews 1988, Avigad *et al.* 1991).

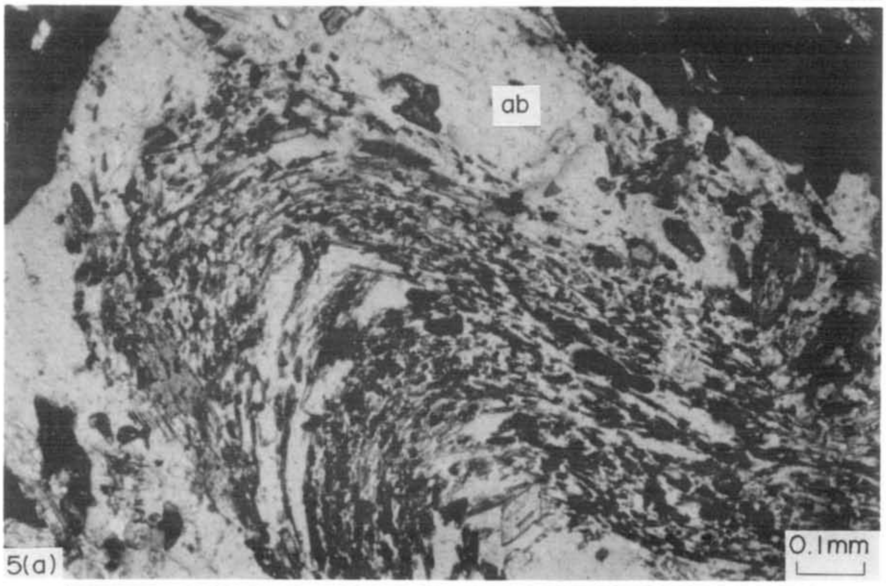
The *P–T* paths of both units are characterized by near isothermal decompression. They differ from *P–T* paths predicated by thermal models which assume whole-crust uplift and erosion (England & Richardson 1977) and are consistent with previous studies which suggest active tectonics as a main cause for the exhumation of the high-pressure metamorphic rocks in the Cyclades (Lister *et al.* 1984, Avigad & Garfunkel 1991, Faure *et al.* 1991). A different *P–T* path has been proposed on the basis of a geochronological study by Wijbrans *et al.* (1990) who considered the rocks exposed on Sifnos as belonging to a coherent unit and that the greenschist-facies overprint has been caused by a thermal pulse at *ca* 19 Ma.

GEOCHRONOLOGICAL CONSTRAINTS

Eocene (*ca* 42 Ma) K–Ar, Ar–Ar and Rb–Sr ages from phengites and paragonites in the gneiss–schist sequence of the EBU, are reported for the high-pressure event (Altherr *et al.* 1979, Wijbrans *et al.* 1990), and probably record cooling below Ar-closure temperature. Wijbrans *et al* (1990) obtained younger ^{40}Ar–^{39}Ar single crystal laser probe ages of *ca* 31–35 Ma from phengites occurring in schists of the Main Marble, near the base of the EBU, and interpreted these ages as reflecting gradual cooling of the sequence during uplift. According to the above data, the base of the EBU cooled below 350°C (which is the argon closure temperature estimated in Wijbrans *et al.* 1990) *ca* 30 Ma ago. Age determinations of the partial transformation in the albite-bearing epidote–blueschist and greenschist facies during the exhumation of the EBU are not available, but should be older than 30 Ma. The age of the high-pressure metamorphism in the GSU has not yet been determined. K–Ar and Rb–Sr Early Miocene ages (24–21 Ma) obtained on phengites predominate in this unit (Altherr *et al.* 1979), and are interpreted as dating the greenschist-facies overprint.

The Eocene ages retained by the EBU, and the excellent preservation of the high-pressure assemblages in it, indicate that the EBU has a different metamorphic history than the underlying GSU. Whereas in the Early





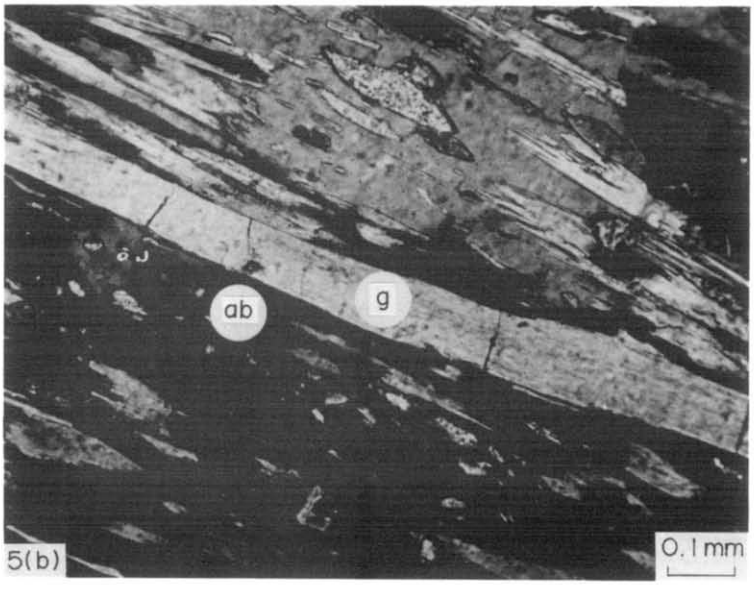


Fig. 4. The tectonic contact between the EBU and the GSU is marked by flat-lying fault gouge (g) at the top of the GSU below the base of the Main marble (road east of Kamares).

Fig. 5. (a) Undeformed albite porphyroblast encloses an earlier microfold defined by relic high-pressure minerals (crossed polarizers). (b) The fabric of a NNE-trending glaucophane lineation in the GSU (crossed polarizers). Note sharp grain boundaries between glaucophane (g) and albite (ab) which suggests that the two minerals coexisted in equilibrium and hence that this lineation may reflect the kinematics of the exhumation of the GSU.

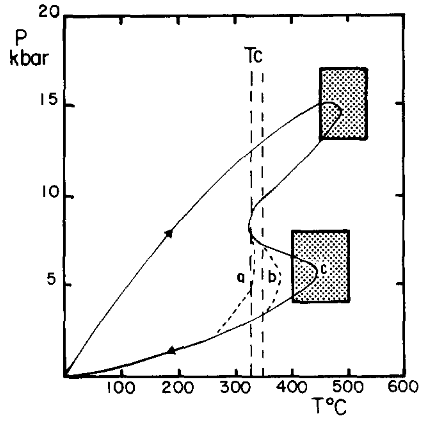


Fig. 7. *P–T* paths for the EBU and GSU proposed for Sifnos by Wijbrans *et al.* (1990). **a** and **b** show the effect of the 19 Ma greenschist pulse (Wijbrans *et al.* 1990) on the schist–gneiss sequence of the EBU and on schist intercalations in the Main Marble, respectively, **c** is the effect of the greenschist pulse on the GSU (added). At present, points **a** and **c** are separated by a *ca* 800 m thick rock section.

Miocene (24–21 Ma ago) the GSU underwent a thorough metamorphic recrystallization at temperatures between 400 and 500°C (Matthews & Schliestedt 1984, Avigad *et al.* 1991), the high-pressure assemblages in the EBU remained relatively intact. Moreover, the geochronological data indicate that at 30 Ma the entire EBU had already cooled below 350°C. An important conclusion is that in the Early Miocene, the EBU was at least 50–150°C cooler than the GSU. This can also be inferred from the *P–T* diagram presented by Wijbrans *et al.* (1990) (Fig. 7) which shows a difference of more than 50°C between the GSU and the EBU at 19 Ma. This fact is interpreted here as indicating that, at that time, the EBU was located at shallower crustal levels, escaping the greenschist-facies overprint which thoroughly affected the GSU still at depth. The presence of a temperature difference between the EBU and the GSU implies that a crustal interval, perhaps several kilometres thick, separated the EBU from the GSU in the Early Miocene. This inferred rock section whose thickness depends on the steepness of the Early miocene thermal gradient is now missing, and it is proposed that it has been removed by the low-angle fault that juxtaposed the two units.

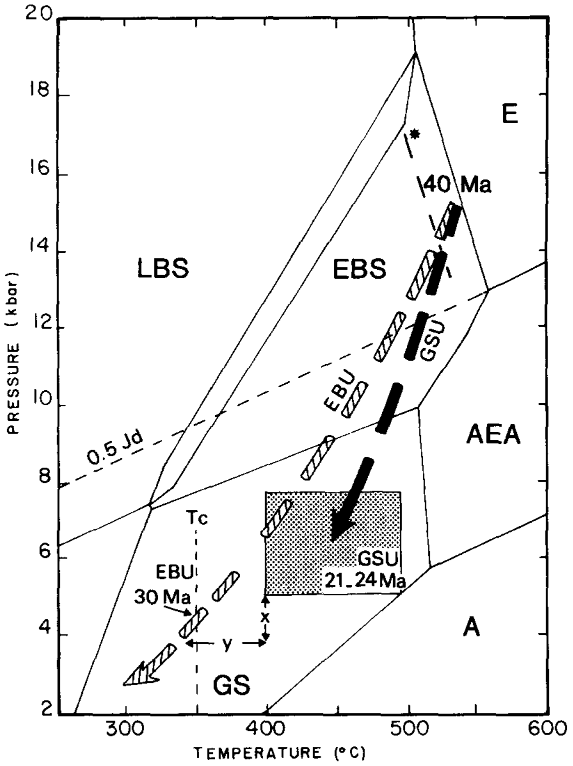


Fig. 6. *P–T*–time paths of the EBU and the GSU in Sifnos. The facies grid is from Evans (1990). Age data is from Altherr *et al.* (1979) and Wijbrans *et al.* (1990). The base of the EBU cooled below *ca* 350°C at 30 Ma (Wijbrans *et al.* 1990). The greenschist-facies overprint affected the GSU at 21–24 Ma (Altherr *et al.* 1979) at temperature conditions of 400–500°C (Matthews & Schliestedt 1984, Avigad *et al.* 1991). The peak *P–T* estimate for the EBU is taken from Schliestedt (1986). *P–T* conditions of the GSU are based on estimates by Matthews & Schliestedt (1984) and Avigad *et al.* (1991). **x** and **y**, respectively, represent estimates of the minimum pressure and temperature differences between the EBU and the EGU in the Oligocene–Miocene, based on geochronological data discussed in the text. Facies abbreviations: EBS, epidote–blueschist; LBS, lawsonite blueschist; E, eclogite; GS, greenschist; AEA, albite–epidote amphibolite; A, amphibolite. Star-eclogite, epidote–blueschist transition calculated for glaucophane composition #2 in Evans (1990).

TECTONIC IMPLICATIONS

The different metamorphic and thermal histories of the EBU and the GSU as well as the brittle nature of the low-angle fault suggest that these units were tectonically juxtaposed after the greenschist-facies overprint (i.e. later than 24–21 Ma ago). The excellent preservation of the high-pressure assemblages in the EBU suggests a relatively rapid exhumation of this unit with respect to the GSU where HP minerals are rarely preserved. This is consistent with the evidence of a temperature difference of at least 50–150°C between the two units in the Early Miocene. This temperature difference is interpreted here to indicate that the EBU was located at shallower crustal levels than the GSU, and escaped the greenschist-facies overprint which thoroughly affected the GSU while still at depth. This interpretation, which assumes a linear correlation between temperature and depth, needs to be justified, because folded isotherms and inverse thermal gradients may characterize the thermal structure of orogenic belts (Oxburgh & Turcotte 1974, Peacock 1987). Temperatures may decrease with depth and a warmer rock unit may overlie a cooler one. If this is the initial situation, juxtaposition of the deeper unit over the shallower unit is supposed to cause pressure increase in the footwall as the thickness of the overburden above it increases. It will also cause decreasing temperatures as the footwall is juxtaposed with rocks of lower temperatures and the result is a counterclockwise *P–T* path in the footwall. By contrast, a broad clockwise *P–T* path is expected in the hangingwall which moves upwards to overlie higher temperature rocks. None of these features are observed in Sifnos Island where cooling and continuous decompression accompanied the exhumation of the EBU and the GSU. The cooling of the EBU during its decompression implies that this unit has been juxtaposed with rocks of progressively lower temperatures, i.e. emerged through a rock pile having a normal, rather than inverse, thermal structure. The *P–T* path of the GSU does not show evidence for an intermediate period of pressure increase and is consistent with continuous uncovering of this rock unit. These arguments support the suggested Oligocene–Miocene configuration whereby the EBU is of more shallow level than the GSU. The low-angle fault brought the GSU from a depth into the base of the previously exhumed EBU. It cut out the inferred crustal interval between these units, causing omission of a rock sequence, and hence it is assigned a normal sense of motion.

EXHUMATION-RELATED STRUCTURES

Ductile structures in the EBU unit are usually associated with the recrystallization of high-pressure assemblages and reflect the style and kinematics of the subduction process. The scarcity of exhumation-related ductile structures suggests that this rock unit has been exhumed as a rigid body. The GSU is characterized by the presence of a penetrative regional cleavage which dips shallowly to the north. The fabric of this regional cleavage comprises greenschist-facies mineral assemblages which are usually not deformed. This fabric is a result of a mimetic growth of the greenschist-facies assemblages over a former high-pressure fabric which is locally preserved as relics in undeformed albite porphyroblasts (Fig. 5a). Thus, in the bulk of the GSU rock sequence, the effect of a ductile deformation during and subsequent to the greenschist-facies overprint is minor. Exceptions are found at the vicinity of large, NW-trending folds below the Main Marble. These folds, absent from the schist–gneiss sequence of the EBU, are the most prominent structural feature developed during the exhumation of the GSU (see also Gournellos 1980). They are well developed in the centre of the island (Fig. 8) and comprise the latest fold generation superimposed over at least two earlier folding phases. The NW-trending folds have open to recumbent hinges, are tens of metres in size, overturned to the north-east and their axial planes dip to the south-west. The axial planar fabric developed in the hinges of these folds is a crenulation cleavage defined by white mica domains which wrap around rotated albite poikiloblasts. In metabasic rocks which were affected by this folding phase glaucophane–crossite recrystallized in the hinge zone either in a direction parallel to the fold axes or randomly, and it coexists with albite. These relations indicate that the NW-trending folds developed while the rocks were within the albite-bearing epidote–blueschist-facies field, at pressures of 8–10 kbar. The direction of overturning of the NW-trending folds indicates 'top-to-the-NNE' movement.

A down-dip mineral lineation defined by 020°-trending glaucophane–crossite lying within shallowly dipping cleavage planes is sometimes observed in the GSU. This lineation has been considered by previous workers as 'subduction related' (Blake *et al.* 1981) but the microstructures associated with it cast some doubt on this interpretation. The lineation is defined by preferentially oriented glaucophane crystals which have sharp grain boundaries with albite (Fig. 5b). These boundaries and the absence of a reaction texture between glaucophane and albite suggests that these two minerals are texturally and chemically equilibrated. This suggests that the glaucophane lineation has formed in the stab-

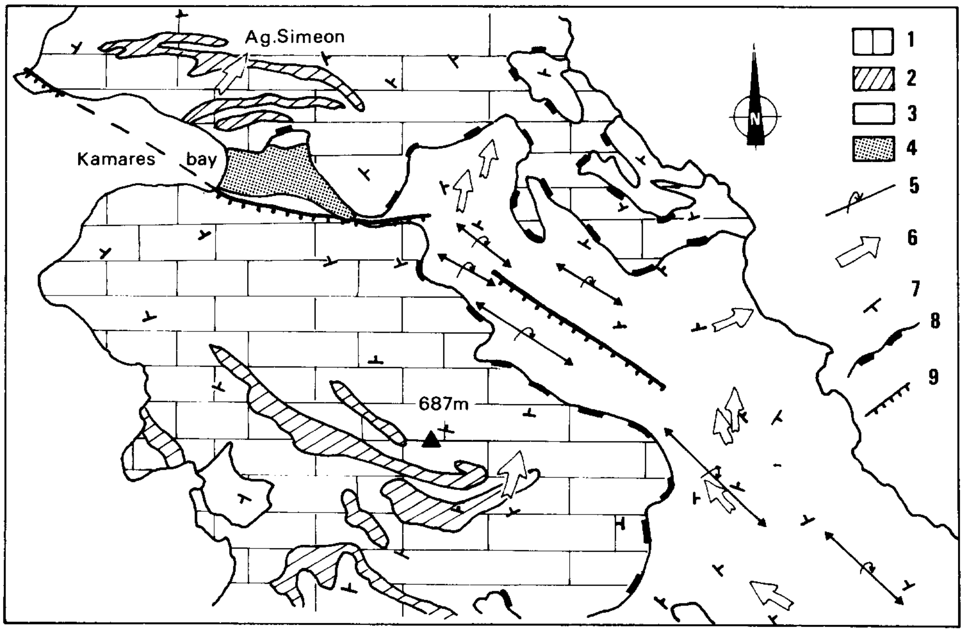


Fig. 8. A structural map of the central part of Sifnos island showing the orientation and location of late, large-scale fold hinges formed during the exhumation of the GSU. Legend: 1, Main Marble; 2, schist layers in the Main Marble; 3, GSU; 4, alluvium; 5, fold axis; 6, glaucophane–crossite lineation; 7, orientation of cleavage; 8, low-angle normal fault; 9, high-angle normal fault.

ility field of albite, perhaps under P–T conditions of the albite-bearing epidote–blueschist facies (in the sense of Evans 1990), which are certainly lower than peak P–T conditions estimated for this unit. Thus, the 020° trending glaucophane lineation in the GSU may have formed during exhumation and could mark the kinematics of the emergence process. Lister *et al.* (1984) and Buick (1991) have ascribed mylonitic foliations and associated 010–020°-trending stretching lineations on Naxos as being related to exhumation, perhaps in a crustal-scale shear zone. On Naxos, Eocene high-pressure metamorphic assemblages were almost completely obliterated by the Oligocene–Miocene overprint and the stretching fabric is developed in Barrovian assemblages. Jansen & Schuiling (1976) and Buick (1988) deduced pressures of 6–7 kbars and temperatures of around 600°C for the Barrovian stage which produced kyanite-bearing rocks. Thus, the 020° stretching lineations in Sifnos and Naxos may both indicate approximately N–S ductile extension during exhumation (see also Faure *et al.* 1991). The more penetrative development of mylonites on Naxos relative to Sifnos can be explained by a higher ductility caused by the high-temperature conditions on Naxos in the Oligocene–Miocene. In Sifnos, the N–S movement along the low-angle fault deduced from the trend of slickensides at the base of the Main Marble is consistent with the orientation of the earlier glaucophane–crossite–albite lineation. Both structural elements attest to a N–S extension coupled with subvertical thinning of the rock pile, and may indicate a prolonged extensional deformation history.

SUMMARY AND CONCLUSIONS

The upper part of the Alpine Cycladic blueschist belt comprises kilometre-thick blueschist- and eclogite-facies rocks in the islands of Sifnos and Syros, whereas the lower part is dominated by greenschist-facies rocks. Although both units may have shared an early process of subduction and high-pressure metamorphism in Eocene times, they greatly differ in their exhumation-related evolution. The geochronological data suggest that the EBU cooled prior to the GSU and that in the Oligocene–Miocene a temperature difference of at least 50°C existed between the units. The greenschist-facies overprint which effaced most of the high-pressure assemblages in the GSU led only to a local and incomplete recrystallization in the EBU. The study on Sifnos Island shows that the contact between the EBU and the underlying GSU is a low-angle fault. The brittle nature of the contact and the differences in the metamorphic and geochronological evolution between the two units indicate that their juxtaposition occurred subsequent to the greenschist-facies overprint which took place in the Oligocene–Miocene. Matthews & Schliestedt (1984) assumed that the EBU and GSU form a coherent rock section and that a pervasive fluid infiltration into the GSU caused the greenschist-facies overprint whereas the impermeable Main Marble at the base of the EBU prevented fluid infiltration into the EBU which remained intact. However, the field evidence show that the EBU and GSU were juxtaposed subsequent to the greenschist-facies overprint, and therefore, that these units were derived from different structural levels. A similar tectonometamorphic structure is observed in Syros Island (Fig. 1), where a rock sequence comparable to that of Sifnos is exposed, and Eocene high-pressure rocks (Maluski *et al.* 1987) overlie greenschists with a low-angle fault (Hecht 1977). Thus, rather than a coherent rock unit, the Cycladic blueschist belt should be considered as comprising two major tectonometamorphic units: the Eclogite–Blueschist unit at the top and the Greenschist unit at the base. The post-metamorphic nature of the low-angle fault indicates that the juxtaposition of the two units occurred during the exhumation history, and post-dated the convergence and crustal thickening which led to the Eocene high-pressure metamorphism. The tectonic emplacement of the high-pressure blueschists and eclogites over greenschists could suggest that an important phase of thrusting affected the Cycladic blueschist belt in post Oligocene–Miocene times. However, rather than causing crustal thickening and continuous burial, the Oligocene and Miocene tectonics in the Cyclades has led to the emergence and unroofing of the high-pressure metamorphic rocks. The exhumation of these rocks implies that the Eocene orogenic wedge has been attenuated and that the overburden which has overlain the high-pressure rocks at peak P–T conditions has been removed. In an attempt to define the sense of motion along the low-angle fault it was suggested that the excellent preservation of the high-pressure assemblages in the EBU indicate that this unit has been rapidly exhumed. Restoring to the Oligocene–Miocene, the EBU was placed above the GSU as might also be indicated by the fact that this unit was then at lower temperatures. The exact thickness of this section depends on the steepness of the Oligocene–Miocene isotherms, and it cannot be transferred to absolute pressures because the isotherms in convergence zones tend to be inclined. This intermediate crustal interval has been cut out by the low-angle fault which brought the GSU from a depth into the base of the previously exhumed EBU. As this low-angle fault omits a section, it was assigned a normal sense of motion. Normal fault tectonics can better explain the attenuation of the Alpine orogenic belt which is implied by the exhumation of the high-pressure rocks. Former studies (Lister *et al.* 1984, Avigad & Garfunkel 1989, 1991, Faure *et al.* 1991) recognized that the contact of the Cycladic blueschist unit with the overlying rocks of the Upper unit ('Asterusia nappe' of Bonneau 1984, 'Superficial nappe' in Faure & Bonneau 1988) is a low-angle normal fault. In the present study we show that normal faults occur also within the Cycladic blueschist unit dividing it into the Eclogite–Blueschist and the Greenschist units. With one exception, post-metamorphic low-angle faults in the entire Cycladic massif cut out section and hence operated as normal faults. The exception is the lower-most fault at the base of the Cycladic blueschist unit, which is a thrust fault placing the blueschist unit over LP–LT rock sequences exposed in Evia, Samos and Tinos. As the LP–LT rocks were not affected by the high-pressure metamorphism, we suggested (Avigad & Garfunkel 1989, 1991) that the Cycladic blueschist unit has been uncovered prior to overthrusting. As the uncovering of the high-pressure rocks was a result of normal faulting, I conclude that the overthrusting post-dated low-angle normal faulting. Hence, convergence and thrust tectonics in the area of the Cyclades did not cease when the emergence and the uncovering of the high-pressure rocks began.

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