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The geodynamic evolution of the Alpine orogen in the Cyclades (Aegean Sea, Greece): insights from diverse origins and modes of emplacement of ultramafic rocks

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Abstract: The Alpine orogen in the Cyclades, wherein both high-pressure metamorphic rocks and ultramafic rocks co-occur, is a key area in studying the emplacement of mantle rocks into the crust. Within the Cyclades three distinct ultramafic associations occur: (1) HP–LT ophiolitic mélanges of the Cycladic Blueschist Unit (CBU) on Evia and Syros; (2) meta-peridotites associated with migmatized leucogneisses on Naxos, which represent the deepest exposed levels of the CBU; (3) a greenschist-facies metamorphosed dismembered ophiolite juxtaposed on top of the CBU by an extensional detachment on Tinos. Most of the Cycladic ultramafic rocks were serpentinized prior to Alpine metamorphism, suggesting denudation prior to reburial. The Naxos meta-peridotites preserve, however, relict mantle assemblage and mantle-like oxygen isotope ratios, and thus indicate direct emplacement from the mantle into an underthrust continent during collision and HP metamorphism (M_1). Thus conditions for M_1 in the Naxos leucogneiss core are constrained by ultramafic assemblages to 550–650 °C and $\geq$14 kbar. Mafic blocks of the ophiolitic mélanges in the NW Cyclades span a wide range of chemical compositions indicating derivation from variable oceanic settings and sequential events of alteration and metasomatism. Given the comparable geochemical heterogeneity in the Syros and Evian mélange intervals, the garnet-bearing meta-basites of the Syros mélange record higher M_1 temperatures (450–500 °C) than the garnet-free epidote blueschists of the Evian mélanges (400–430 °C). It follows that going southeastwards from Evia progressively deeper (i.e. hotter) levels of the subducted plate are exposed. Correspondingly, temperatures of the M_2 overprint also increase from pumpellyite-bearing assemblages on southern Evia, through greenschists on Syros to upper-amphibolite, sillimanite-bearing gneisses on Naxos. The diverse P–T paths of the CBU form an array wherein the deeper a rock sequence is buried, the 'hotter' is its exhumation path. Such a pattern is predicted by thermal modelling of tectonically thickened crust unroofed by either erosion or uniform extension.

The occurrence of dense ultramafic rocks, peridotites, the prime constituent of the Earth's mantle, at the surface of the continents requires significant vertical mobility. It is thus not surprising that orogenic belts where continents have collided and vast tectonic movements have taken place host most of the relatively rare peridotites. High-pressure metamorphic rocks best record the vertical movements involved in orogenesis: eclogites and blueschists mostly comprise surface-derived rocks, thus implying a full tectonic cycle of burial and exhumation. Orogenic segments where both high-pressure and ultramafic constituents occur in proximity are thus key areas in answering a fundamental question: how are mantle-derived rocks incorporated into the subduction–exhumation sequence of surficial rocks?

A partial answer to the puzzle of displacement of ultramafic rocks into the crust is given by ophiolite suites representing occasional portions of oceanic plates that escaped destruction at subduction zones and were carried onto the foreland of an adjacent continent. Based on their tectonic setting two major types of ophiolites were distinguished (Moores 1982; Coleman 1984; Wakabayashi & Dilek 2003): (1) ophiolites that occur as thick thrust sheets that rest upon passive margin substrate and are commonly associated with high-temperature metamorphic aureoles at their base (e.g. the Semail ophiolite, Oman; the Pindos ophiolite, Greece); (2) ophiolite bodies that occur as blocks within tectonic blueschist mélange as part of an accretionary prism (e.g. the Franciscan mélange). The differences in the manner of occurrence and tectonic context of the two types reflect their origin and mode of emplacement: Tethyan-type ophiolites formed by thrusting of an oceanic lithosphere slab upon passive continental margin sequences whereas upheaval of oceanic fragments within the accretionary prism of an active margin gave rise to Cordilleran-type ophiolites.

Within the Hellenic segment of the Alpine orogenic belt a major Tethyan-type ophiolite

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emplacement occurred in mid- to late Jurassic times. The 'Eohellenic' ophiolites are interpreted as originating from a Mesozoic Neo-Tethyan oceanic basin, the Pindos Ocean, and subsequently thrust northeastwards onto the Pelagonian passive continental margin (Fig. 1; Robertson *et al.* 1991; Smith 1993). Deep-water sedimentation continued, however, in the Pindos basin until its final closure in the early Tertiary (Jones & Robertson 1991). Within the Cycladic Massif of the Aegean Sea (Fig. 1), a Tertiary high-pressure orogenic segment that lies to the SE of the Hellenides, thin remnants of the Eohellenic ophiolites occur on the island of Paros (Papanikolaou 1980). However, most of the ophiolites in the Cyclades are regionally metamorphosed at variable conditions, they are highly attenuated and dismembered, and are bounded and dissected by low-angle tectonic contacts. The Cycladic ultramafic rocks are associated with a great variety of country rocks including leucogneisses of continental basement origin (the Main Ultramafic Horizon on Naxos; see below), thus raising questions concerning the provenance of peridotites. The diversity in field relations, metamorphic grade and tectonic position of ultramafic rocks makes the Alpine orogen in the Cyclades an attractive terrain to address the questions of their origin and emplacement. Moreover, in a complicated poly-metamorphosed orogenic segment such as the Cyclades, the relative sluggishness of metamorphic reactions in ultramafic rocks turns them into potential preservers of pre- and early metamorphic evolution invariably effaced by later events in other rocks.

In this paper we review the tectonic position and field relations of major ultramafic occurrences in the Cyclades and examine in detail the petrography and chemical compositions of ultramafic and associated rocks. Thus, their origin and mode of emplacement are unveiled and new constraints on the orogenic evolution of the Cyclades are set.

Regional geological setting

The Cycladic Massif

The Cycladic Massif (Fig. 1) records an Alpine orogenic cycle of collisional thickening, collapse and reworking by back-arc extension (Avigad *et al.* 1997). This cycle is well represented by the metamorphic evolution of the dominant tectonic unit of the Massif, the Cycladic Blueschist Unit (CBU, Lower tectonic Unit), and by its position within the orogenic nappe-pile. The CBU underwent regional eclogite- and blueschist-facies metamorphism during Late Cretaceous to Eocene compression (M_1 metamorphism) (Altherr *et al.* 1979; Andriessen *et al.* 1979; Maluski *et al.* 1981, 1987; Wijbrans & McDougall 1988; Wijbrans *et al.* 1990; Bröcker *et al.* 1993; Bröcker & Enders 1999, 2001; Tomaschek *et al.* 2003; Putlitz *et al.* 2005). Today the Cycladic Blueschist Unit forms part of the highly attenuated crust of the Aegean Sea in an extensional back-arc environment. The tectonic sequence of thickening followed by back-arc extension was suggested as an efficient mechanism for rapid exhumation and preservation of high-pressure metamorphic rocks in the Cyclades (Lister *et al.* 1984; Gautier & Brun 1994; Jolivet & Patriat 1999; Trotet *et al.* 2001*a*). Avigad *et al.* (1997) questioned this notion and showed that back-arc extension in the Aegean lagged behind a significant part of the exhumation of the high-pressure metamorphic rocks. Evidence for the initiation of the Aegean back-arc extension in the earliest Miocene is plentiful and includes: (1) activation of normal-sense detachments that juxtaposed sedimentary and low-pressure metamorphic rocks (Upper Tectonic Unit) on top of exhumed high P–T rocks (CBU) (Lister *et al.* 1984; Faure *et al.* 1991; Lee & Lister 1992; Gautier *et al.* 1993); the earliest detachment was observed on Tinos, where both units were intruded by an 18 Ma granite (Avigad & Garfunkel 1989); (2) abundance of north-dipping ductile extensional fabrics in Early Miocene overprinting assemblages (Urai *et al.* 1990; Buick 1991; Gautier & Brun 1994; Jolivet & Patriat 1999); (3) beginning of sedimentation in extension-related basins (Dermitzakis & Papanikolaou 1981; Sanchez-Gömez *et al.* 2002); (4) regional extension-controlled granitic plutonism (Altherr & Siebel 2002).

Placing the Early Miocene initiation of extension within the P–T–t path of the CBU provides convincing evidence for a time gap between exhumation and extension in the Cyclades. Most of the petrological studies of the Cycladic high-pressure rocks concentrated on the islands of Sifnos, Syros and Tinos, where the best-preserved eclogites occur. The three islands form a NE–SW linear array that hereafter will be termed the Central Eclogite Axis (CEX; Fig. 1). The P–T–t paths of the CBU in the CEX include Eocene eclogite-facies metamorphism at 15 ± 3 kbar and 450–500 °C (M_1) that was followed by a 23–21 Ma greenschist-facies overprint (M_2) at similar temperatures and 5–7 kbar (Matthews & Schliestedt 1984; Evans 1986; Schliestedt 1986; Dixon & Ridley 1987; Schliestedt *et al.* 1987; Okrusch & Bröcker 1990; Avigad *et al.* 1992; Bröcker *et al.* 1993, 2004; Bröcker & Franz 1998). Hence, the Early Miocene onset of back-arc extension in the Aegean was coeval with a greenschist-facies metamorphic event (M_2) that overprinted high-pressure rocks in the CEX. It is thus evident that, at the time of overprinting, the Cycladic blueschists and

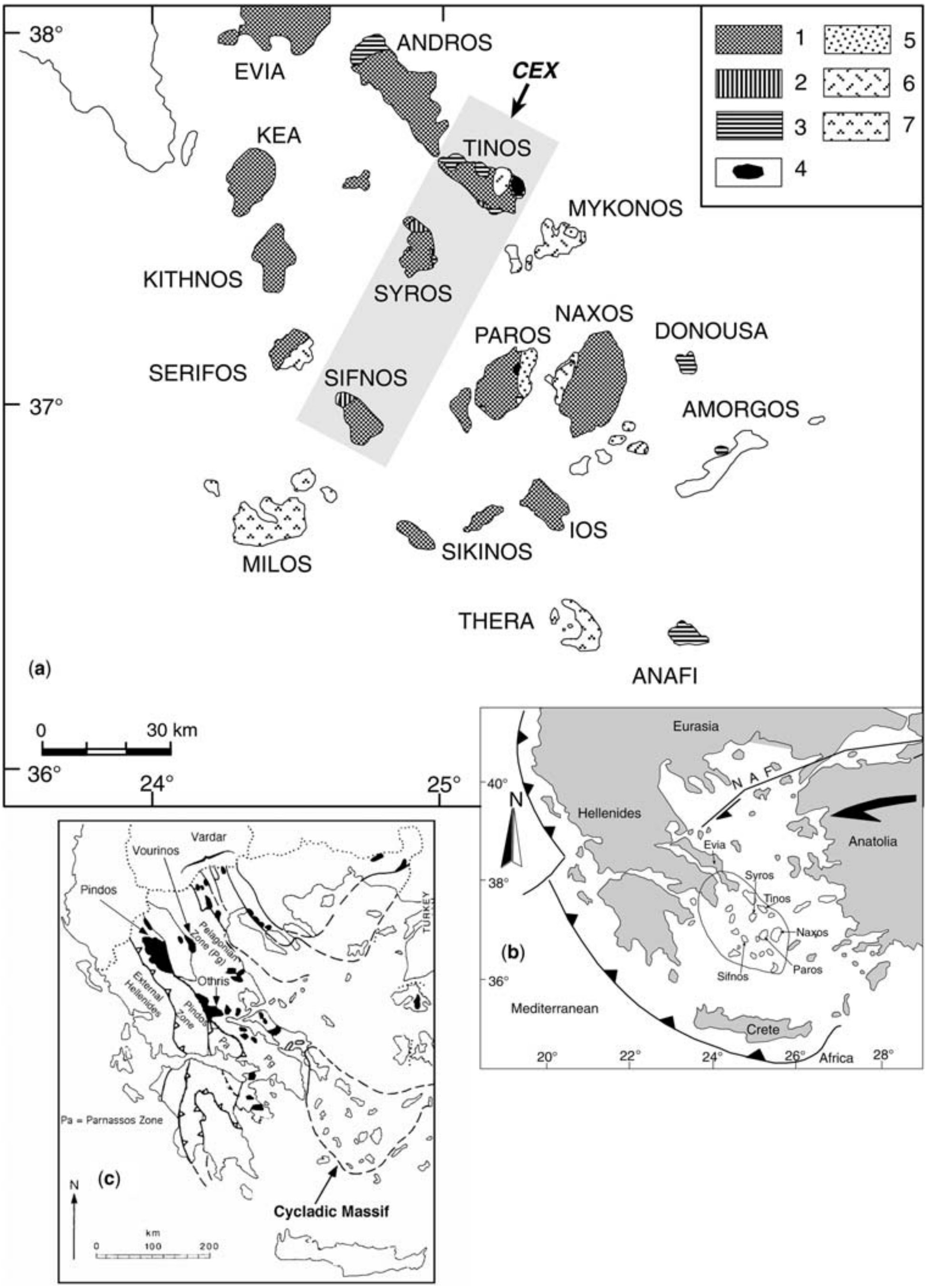

Fig. 1. (**a**) Geological map of the Cycladic Massif (after Avigad & Garfunkel 1991). The Cyclades Blueschist Unit (CBU, the Lower Unit): 1, Eocene high-pressure rocks overprinted at greenschist-facies conditions in the Early Miocene; 2, Eocene eclogite-facies rocks. The Upper Unit: 3, low-pressure metamorphic rocks, mostly of Late Cretaceous age; 4, ophiolites; 5, Early Miocene clastic sediments. Cenozoic igneous rocks: 6, Miocene granitoids; 7, Pliocene to recent volcanic rocks. CEX, Central Eclogite Axis. (**b**) Location map for the Cycladic Massif showing major tectonic features of the Aegean region: back-arc extension in the Aegean is promoted by the southward retreat of the Hellenic Trench and accommodates right-lateral motion on the North Anatolian Fault (NAF). (**c**) Map of Greece showing the geotectonic zones and the locations of main outcrops of Eohellenic ophiolites (from Smith 1993).

eclogites had already isothermally decompressed from their maximum burial depths to much shallower crustal levels. New thermodynamic analysis indicates, however, significant cooling during decompression for Syros and Sifnos eclogites, and suggests variable tectonic scenarios for the early exhumation of the Blueschist Unit (Trotet *et al.* 2001*b*; Schmädicke & Will 2003). Studying the spatial distribution and variation of the P–T paths in the Cyclades can lead to better understanding of the exhumation-related tectonic processes in the time interval between high-pressure metamorphism in Cretaceous–Eocene times (M_1) to overprinting in the Early Miocene (M_2). Thus a major objective of this research is to study the petrological evolution of ultramafic and associated rocks of the Lower tectonic Unit both to the NW (Evia) and to the SE (Naxos) of the CEX. Additionally, a dismembered ophiolite of the Upper Unit that was juxtaposed on top of the blueschists on Tinos is studied to provide insight into processes that affected the overburden that covered the CBU during exhumation.

Ultramafic rock associations

The Alpine orogen in the Cyclades comprises three main tectono-metamorphic units separated by low-angle faults (Fig. 1; Dürr *et al.* 1978*a*; Avigad & Garfunkel 1991). The two upper units contain ultramafic rocks. The dominant Cycladic Blueschist Unit (CBU, the Lower Unit), the evolution of which is discussed above, is delimited from above by flat-lying normal faults that omit a significant part of the overburden that covered it since Early Miocene times (Lister *et al.* 1984; Ridley 1984*a*; Avigad & Garfunkel 1989; Gautier & Brun 1994; John & Howard 1995; Patriat & Jolivet 1998; Bröcker & Franz 1998; Ring *et al.* 2003). These detachments juxtaposed the Upper tectonic Unit, which comprises lithologically diverse thin remnants of the overburden, on top of the CBU. Tectonic windows on Evia and Tinos expose the lowermost weakly metamorphosed Basal Unit (Almyropotamos Unit), which consists of a thick platform carbonate sequence topped by Eocene flysch, beneath the blueschists (Dubois & Bignot 1979; Avigad & Garfunkel 1989). The apparent inverted age and metamorphic sequences indicate that the tectonic contact between the Lower and Basal Units is a major thrust fault (Katsikatsos *et al.* 1986; Matthews *et al.* 1999; Shaked *et al.* 2000).

The CBU is characterized by a change in protoliths from a sequence of continental origin in the central Cyclades (Naxos and Paros) to an oceanic or basinal sequence in the NW Cyclades (Syros, Tinos, Andros) and southern Evia. The Lower Unit on Naxos consists of pre-Alpine granitic and quartzo-feldspathic precursors overlain by a Mesozoic sedimentary sequence dominated by bauxite-bearing shelf carbonates (Jansen & Schuiling 1976; Feenstra 1985; Keay *et al.* 2001). Its counterpart in the NW Cyclades consists mainly of Mesozoic quartz-poor clastic and volcanic protoliths with moderate to minor amounts of carbonates that probably accumulated in a deep basin. Strikingly, ultramafic rocks are associated both with the leucogneisses and the overlying platformal sediments on Naxos and with the basinal sequences of the NW Cyclades, where they are associated as mélange constituents with other ophiolitic lithologies.

The Upper Unit comprises varied sequences of low- to medium-pressure metamorphic rocks and unmetamorphosed rocks. Metamorphic sequences include amphibolite- to greenschist-facies basites, acidites, pelites and marbles on the small islands of Donousa, Nikouria and Anafi on the eastern edge of the Cyclades and on Syros and Tinos (Dürr *et al.* 1978*a*; Reinecke *et al.* 1982; Maluski *et al.* 1987; Patzak *et al.* 1994; Ring *et al.* 2003; Beeri, Y., pers. comm.). High-pressure mineralogy was never found in the metamorphic rocks of the Upper Unit and metamorphism was dated as Late Cretaceous (Dürr *et al.* 1978*b*; Reinecke *et al.* 1982; Maluski *et al.* 1987; Altherr *et al.* 1994; Patzak *et al.* 1994). Both grade and age of metamorphism indicate that the rocks of the Upper Unit were at shallow crustal levels since the Late Cretaceous and escaped the Eocene high-pressure metamorphism that affected the CBU. Rare meta-serpentinites occur on Anafi; however, the largest ultramafic exposure occurs in the Upper Unit of Tinos: a metamorphosed dismembered ophiolite composed of mafic and ultramafic slices (Katzir *et al.* 1996). The tectono-stratigraphic position of the Tinos dismembered ophiolite and its sliced structure greatly differs from those of its counterpart ultramafic associations in the CBU, suggesting a distinctive tectonic setting for its deformation and metamorphism and possibly for its origin.

Three major ultramafic associations are thus focused on: (1) HP–LT ophiolitic mélanges in the NW Cyclades (Syros and Evia); (2) meta-peridotites associated with migmatized leucogneisses on Naxos; and (3) metamorphosed dismembered ophiolite in the Upper Unit of Tinos.

In addition to comparing their Alpine metamorphic evolution, we also use the field relations, pre-metamorphic textural and mineralogical relicts, and whole-rock chemical composition to shed light on the petrogenesis of the igneous constituents of the diverse ultramafic associations and on the tectonic setting in which they were assembled.

HP–LT ophiolitic mélanges (NW Cyclades)

In the basinal association of the NW Cyclades ultramafic rocks occur as sheets and lenses of meta-serpentinite or as meta-serpentinitic envelopes to metabasites within HP–LT ophiolitic mélanges. The two major sequences of ophiolitic mélange in the CBU occur on Syros and on southern Evia (Blake *et al.* 1981). On Syros the mélange interval occurs in the uppermost part of the Blueschist Unit and includes ‘knockers’ of metagabbro enclosed in a meta-greywacke sequence (Bonneau *et al.* 1980; Dixon & Ridley 1987; Bröcker & Enders 1999, 2001; Tomaschek *et al.* 2003). Thin and highly metasomatized serpentinite envelopes occur at the contacts between the metabasite blocks and the enclosing metasediments. On southern Evia, however, two intervals of ophiolitic mélange occur (Fig. 2). The Ochi ophiolitic mélange occurs at the upper levels of the Blueschist Unit, and according to its position, thickness and composition might be correlated to the Syros mélange (Table 1). The Tsaki ophiolitic mélange occurs at the base of the Lower Unit immediately above the basal thrust. Unlike the Syros mélange, both mélange intervals on Evia include large bodies of meta-serpentinite (up to several hundred metres in diameter).

Origin

In Table 1 the major pre-metamorphic characteristics of the ophiolitic mélange on Syros and Evia are described.

On both Syros and Evia, gabbros preserve a partial record of igneous and hydrothermal mineral assemblages and textures that formed in the oceanic crust, prior to high-pressure metamorphism. However, such preservation in the other mélange components is unique to Evia. The ultramafic protoliths of the meta-serpentinite lenses in the Evian mélanges probably represent two different levels in the oceanic lithosphere: whereas the cumulate wehrlitic lens and the gabbro-hosted wehrlite dyke of Ochi represent the mantle–crust transition zone, the bastite-bearing serpentinite of Tsaki was probably derived from mantle harzburgite. The thin, intensively sheared Syros serpentinites bear no igneous mineralogical or textural relicts, thus their origin is not disclosed. Likewise, the occurrence of metavolcanic blocks in the Evian mélanges is evident by the clearly observed pillowed structure and vesicles. On Syros, however, a volcanic origin for some of the metabasite blocks cannot be demonstrated unambiguously (Seck *et al.* 1996).

Further characterization of the tectonic setting in which the igneous protoliths of the blocks in the mélange were formed is possible by geochemical analysis (Seck *et al.* 1996; Katzir *et al.* 2000). Mineralogical, geochemical and isotope evidence indicates that the mélange constituents had been hydrothermally altered in an oceanic environment and possibly also metasomatized during subduction (Dixon & Ridley 1987; Katzir *et al.* 2000; Putlitz *et al.* 2000; Bröcker & Enders 2001; Tomaschek *et al.* 2003). The rare earth elements (REE) are considered immobile during fluid–rock interaction and are thus useful tools in studying the origin of Evia and Syros metabasites. Chondrite-normalized REE patterns given in Figure 3 show significantly higher REE abundances for both Ochi gabbros and basalts relative to their Tsaki equivalents, indicating different sources. In either of the mélange intervals on Evia the REE contents of metabasalts are significantly higher than those of the associated metagabbros. Basalts are particularly light REE (LREE)-enriched relative to gabbros, whereas heavy REE (HREE) contents of some of the metagabbros are almost equivalent to those of metabasalts. Assuming that the blocks in each mélange interval were assembled from a single source terrane, and given the lack of relict cumulate texture and positive Eu anomalies in Evian gabbros, the most plausible explanation for the depletion in incompatible elements in gabbros relative to the adjacent basalts is the loss of residual melt by compaction. REE abundances in metabasalts are thus a more reliable indicator of the tectonic setting of the source areas of the Evian mélanges. Ochi metabasalts are highly fractionated and LREE enriched: the La content is up to 120 times chondrite values and $(La/Yb)_{CN}$ values range from five to 10. These features indicate an enriched mantle source for Ochi basalts and are comparable with those of enriched mid-ocean ridge basalt (E-MORB) (Sun & McDonough 1989). Consistently lower REE abundances and less fractionated patterns are observed for the Tsaki metabasalts: La abundances are 20–30 times chondrite and $(La/Yb)_{CN}$ values are 2–3. These values are characteristic of normal MORB (N-MORB) to transitional MORB (T-MORB), or of tholeiitic basalts from back-arc setting. Syros metagabbros are the least REE-enriched rocks; however, REE abundances of other HP–LT metabasites of the Syros mélange are highly variable. The slightly upward-convex REE patterns of the Syros metagabbros were interpreted as representing a mixture of REE-poor cumulate phases and N-MORB intercumulus melt (Seck *et al.* 1996). Because of the lack of pre-metamorphic relicts, it is difficult to define the protoliths of Syros glaucophanites and eclogites. However, glaucophanites are fine to medium grained and highly foliated, whereas eclogites (and particularly garnet glaucophanites) show variable grain size and texture

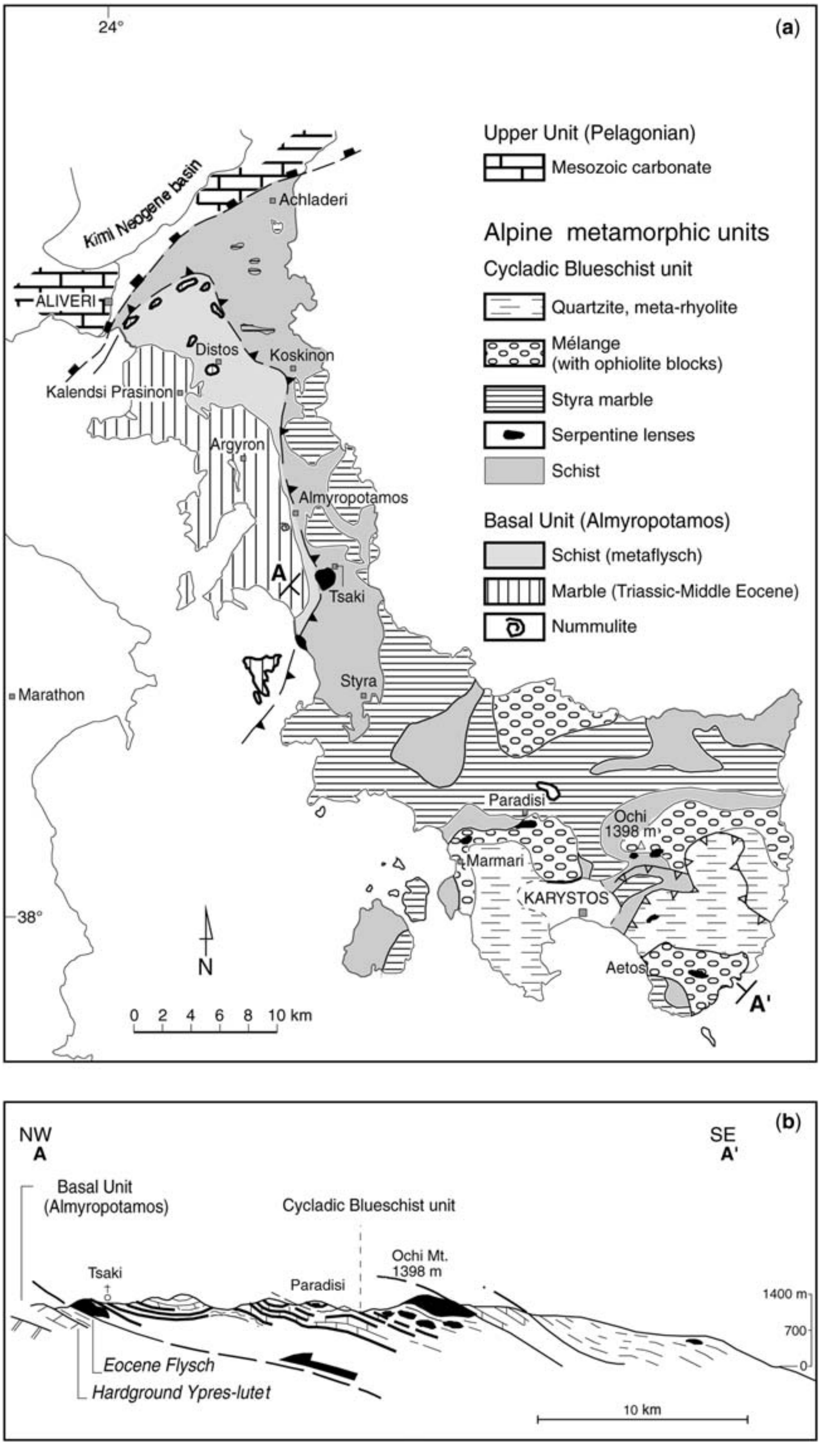

Fig. 2. Geological map (**a**) and cross-section (**b**) of southern Evia after Jacobshagen (1986), showing the tectonic contacts at the base and the top of the Cycladic Blueschist Unit and several sub-units within it.

Table 1. *Lithological and textural characteristics of HP–LT ophiolitic mélanges in the Cyclades with emphasis on pre-metamorphic features*

		Syros	Ochi	Tsaki
Lithological components	Blocks	Metagabbro (≤700 m), meta-acidite (granophyre?), clast-supported meta-igneous breccia of sedimentary origin	Metagabbro (≤200 m), meta-serpentinized wehrlite, metabasalt	Meta-serpentinite (≤500 m), metagabbro, metabasalt
	Matrix	Pelitic schist, meta-tuffites, metasomatized serpentinite, rare meta-chert, meta-ironstone	Semi-pelitic schist, meta-chert, meta-ironstone	Pelitic schist
Underlying sequence		Schist–marble (meta-flysch)	Semi-pelitic schist, meta-tuffite	Basal Unit (Almyropotamos)
Overlying sequence		Pelitic–psammitic schist, marble	Marble, quartzite, volcanogenic schist, meta-rhyolite	Pelitic schist
Pre-metamorphic relicts		Gabbro: pseudomorphs of actinolite after cpx (sub-sea-floor hydrothermal alteration) including rare augite cores; cumulate texture preserved by topotactic growth of metamorphic minerals (gln, ep) over igneous precursors (cpx, plg); intrusive relations between basic and acid rocks	Gabbro: aug, aug replaced by hbd Wehrlite: cumulate texture of cpx poikiloblasts enclosing mesh-textured serpentine after olivine Basalt: pillows zoned from omp and vesicle-rich core to gln-rich and non-vesicular rims Wehrlite dyke in gabbro	Serpentinite: bastite pseudomorphs after opx Gabbro: coarse-grained hbd Basalt: vesicular; sub-ophitic texture; Ti-aug cores in omp
Metamorphic textures related to pre-metamorphic features		Flaser gabbro: porphyroclasts of act or omp Metasomatic rinds at ultramafic contacts: omphacitite, jadeitite	Flaser gabbro: porphyroclasts of aug	Basalt: vesicles filled with various metamorphic assemblages

act, actinolite; aug, augite; cpx, clinopyroxene; ep, epidote; gln, glaucophane; hbd, hornblende; omp, omphacite; opx, orthopyroxene; plg, plagioclase.
Data from Syros and Evia are based on the observations of Dixon & Ridley (1987) and Katzir *et al.* (2000), respectively.

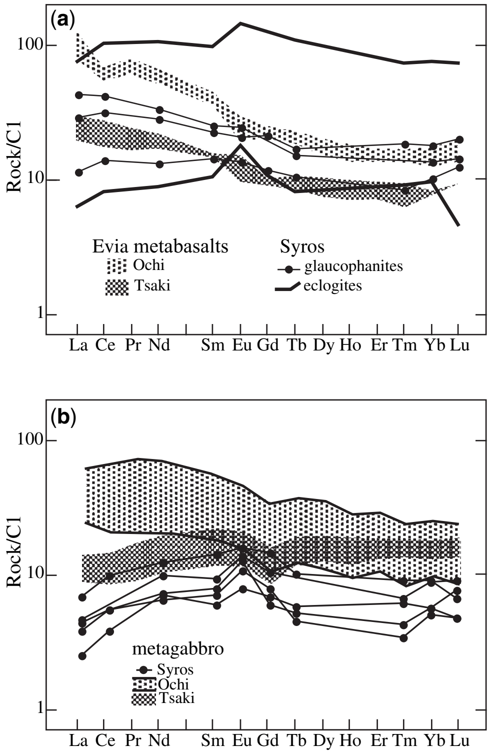

Fig. 3. Chondrite-normalized REE patterns of (**a**) metabasalts and (**b**) metagabbros from the HP–LT ophiolitic mélanges on Syros and southern Evia. Data for Syros and Evia are from Seck *et al.* (1996) and Katzir *et al.* (2000), respectively.

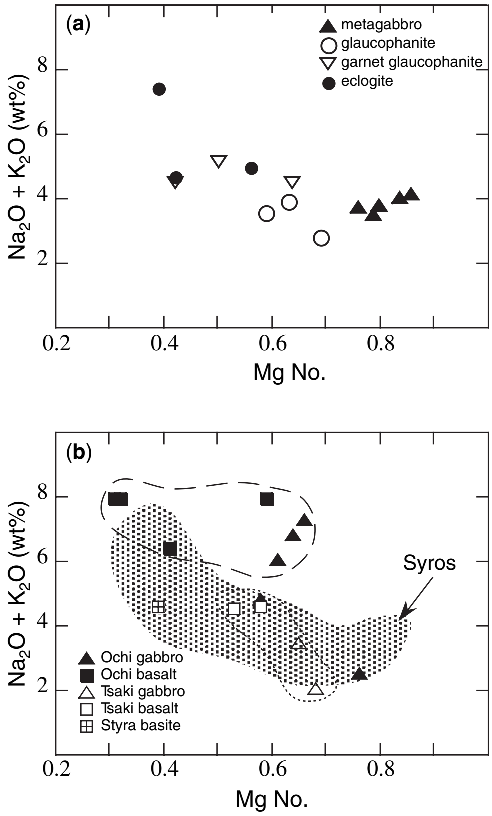

Fig. 4. Mg-number (atomic ratio (Mg/(Mg + Fe)) v. total alkali content ($Na_2O + K_2O$ wt%) in metabasites from (**a**) Syros and (**b**) Evia HP–LT ophiolitic mélanges. Data for Syros and Evia are from Seck *et al.* (1996) and Katzir *et al.* (2000), respectively.

including massive to foliated rocks (Seck *et al.* 1996). Based on textural criteria and the resemblance of their REE patterns to those of the Tsaki metabasalts, the Syros glaucophanites may have been derived from similar, moderately enriched basaltic protoliths. The textural variability of Syros eclogites and garnet glaucophanites is also reflected in their REE patterns. The coarser-grained, massive members have low to moderate total abundances of REE and positive Eu anomalies that suggest cumulate gabbro precursors. However, the most REE-rich eclogites (100 times chondrite) are foliated, possibly representing highly differentiated basalts. To sum up, the basaltic components of the NW Cyclades ophiolitic mélanges span a wide range of REE compositions. The REE patterns are characteristic of variable oceanic environments including N-, T- and E-MORB. These may represent either different segments of a single spreading centre or adjacent small oceanic basins.

Whereas REE abundances are considered as attained during igneous crystallization and retained through post-magmatic events, major element composition is susceptible to hydrothermal and metasomatic modifications. In Figure 4 the alkali element contents of Syros and Evia metabasites are plotted against Mg-number (Mg/(Mg + Fe) atomic ratio). In all three mélange sequences alkali element and iron enrichments are positively correlated. These correlations were interpreted in various ways. Based on mineralogical and geochemical variations between core and rim of pillow basalts in the Evian mélanges, Katzir *et al.* (2000) attributed the Na–Fe trends to various degrees of hydrothermal alteration in the oceanic crust. On Syros, however, where monomineralic reaction rinds (glaucophane, omphacite, actinolite and chlorite) between blocks and serpentinite envelopes are well developed, high-pressure metasomatism was seen as responsible for Na and Fe variations, and in some cases for extreme desilicification (Dixon & Ridley 1987). According to Seck *et al.* (1996), metasomatism in the Syros mélange is rather limited, and Fe-rich eclogites represent strongly differentiated basalts that crystallized in small magma chambers. Regardless of whether the major element variations in Syros and Evia metabasites are the result of igneous differentiation, oceanic hydrothermal alteration, synmetamorphic fluid–rock interaction or some combination of these processes, a major conclusion may be drawn: in all three mélange intervals wide ranges of major element compositions exist (SiO_2 46–55 wt%; Mg-number 0.31–0.76; $Na_2O + K_2O$ 4.5–8 wt%) allowing the crystallization of diverse, yet comparable metamorphic assemblages.

Formation of mélanges

The occurrence of glaucophane and omphacite metasomatic reaction zones at the serpentinite–mafic blocks and serpentinite–metasedimentary matrix contacts on Syros indicates that juxtaposition of the mélange components preceded peak M_1 metamorphism (Dixon & Ridley 1987). On Evia chlorite and tremolite blackwalls envelop the ultramafic bodies and follow their current geometry, thus also indicating pre- to syn-M_1 incorporation of the igneous bodies into the enclosing sedimentary sequence. The boudin shape of most blocks, including the largest ones, suggests that the last stage of the evolution of the mélanges involved significant flattening by extensional strain during metamorphism. However, the process by which the mélange components were initially assembled cannot be unequivocally deduced. The Evian and Syros mélanges may have formed either as olistostrome horizons within thick sedimentary flysch sequences (Dixon & Ridley 1987; see also Mukhin 1996) or as shear zones that separate distinct thrust sheets (Bröcker & Enders 2001). A combined scenario of a structurally weak serpentinite-bearing stratigraphic layer that later developed into a shear zone and was reshaped as a tectonic mélange is also possible (Bröcker & Enders 2001). The Evian and Syros mélanges differ totally, however, from Franciscan-type trench mélanges: unlike their Franciscan counterparts, which form sequences several kilometres thick exposed regionally, ophiolitic mélanges in the Cyclades are rare and occur in thin horizons. Also, the metamorphic conditions recorded in the igneous blocks and sedimentary matrix of the Cycladic mélanges are homogeneous, whereas the Franciscan-type complexes are characterized by high variability of metamorphic temperatures and pressures in the blocks, often included in non- or weakly metamorphosed clastic sediments. Thus, models suggested for the formation of trench mélanges such as return laminar flow in the subduction trench are definitely inapplicable to the HP–LT ophiolitic mélanges in the Cyclades.

High-pressure metamorphism (M_1)

Early field and petrographic studies in the Cycladic Blueschist Unit suggested an increase in M_1 metamorphic grade from southern Evia southeastwards through Andros to the Central Eclogite Axis (Blake *et al.* 1981; Bonneau & Kienast 1982). Southern Evia was considered a lower-grade blueschist terrane because of the occurrence of lawsonite and the absence of garnet in glaucophane-rich metabasites. In comparison, garnet is abundant in Syros eclogites and glaucophanites, and the occurrence of epidote–white mica pseudomorphs after lawsonite confines its stability to pre-peak M_1 prograde conditions (Ridley 1984*b*; Putlitz *et al.* 2005). A temperature range of 450–500 °C was determined for M_1 on Syros by Fe–Mg exchange thermometry of garnet and omphacite (Okrusch & Bröcker 1990). Similar temperatures were obtained by cation exchange thermometry for Sifnos and Tinos eclogites (Schliestedt 1986; Okrusch & Bröcker 1990) and confirmed by oxygen isotope thermometry on Sifnos and Syros (Matthews & Schliestedt 1984; Matthews 1994; Putlitz *et al.* 2000). More recent studies have argued, however, for higher peak M_1 temperatures: ≤ 580 °C (Trotet *et al.* 2001*b*) and 550–600 °C (Schmädicke & Will 2003) (Table 2). The first estimate is based on multi-equilibrium approach using TWEEQ whereas the latter uses the same cation thermometry as in earlier studies. We find no reason to prefer the recent, higher-temperature estimates to the previous ones. On the contrary, peak M_1 temperatures of 450–500 °C have been advocated by numerous petrological and isotope studies on a variety of rock compositions (Evans 1986; Schliestedt 1986; Schliestedt & Okrusch 1988; Matthews 1994; Putlitz *et al.* 2000, 2005; see Table 2).

Our observations on southern Evia showed that the dominant high-pressure assemblage in Ochi and Tsaki metabasites is epidote–glaucophane–omphacite, which defines the M_1 grade as epidote blueschist (Katzir *et al.* 2000). Relict lawsonite enclosed in various high-pressure and retrograde minerals records an earlier, prograde part of the *P–T* path. The reaction lawsonite + jadeite = zoisite + paragonite + quartz + H_2O is thus a lower temperature limit for the peak M_1 assemblages on both Syros and southern Evia. Notwithstanding, the absence of garnet in Evian blueschists is either temperature or composition dependent. It has been shown above that Evian and Syros metabasites span a rather wide and comparable range of compositions. This includes SiO_2, MgO, FeO and CaO, the activities of which strongly affect the garnet-in reaction in blueschists of the haplobasaltic system: chlorite + quartz + epidote = garnet + actinolite + H_2O (Evans 1990). Thus, peak M_1 temperatures on

Table 2. *Temperature and pressure estimates for the M_1 (Eocene) and M_2 (Miocene) metamorphic events in the Central Eclogite Axis of the Cyclades (Sifnos and Syros) and on southern Evia*

	Sifnos			Syros			Southern Evia		
	P–T	References	Method	*P–T*	References	Method	*P–T*	Reference	Method
M_1									
T (°C)	470 ± 30	1, 2, 3	ce, pd	450–500	3, 7	ce, pd	400–430	10	pd
	480 ± 25	4	oxi	500 ± 30	8	oxi	400	11	pd
	≤580	5	pd	450	9	pd			
	550–600	6	ce	≤580	5				
P (kbar)	15 ± 3	1, 3	pd	14–20	3, 7	pd	>12	10	pd
	14–18	2	pd	≤20	5	pd	>10–12	12	pd
	20	5, 6	pd						
M_2									
T (°C)	≤450	13, 14	pd	370–430	9		300–350	10	pd
	400–500	15	oxi	≤520	5				
	380–440	6	pd						
P (kbar)	5–7	13, 14	pd	6–9	9		4–8	10	pd
	9	6	pd	≤11	5				

References: 1, Schliestedt (1986); 2, Evans (1986); 3, Okrusch & Bröcker (1990); 4, Matthews (1994); 5, Trotet *et al.* (2001*b*); 6, Schmädicke & Will (2003); 7, Ridley (1984*b*); 8, Putlitz *et al.* (2000); 9, Putlitz *et al.* (2005); 10, Katzir *et al.* (2000); 11, Reinecke (1986); 12, Shaked *et al.* (2000); 13, Schliestedt & Matthews (1987); 14, Avigad *et al.* (1992); 15, Matthews & Schliestedt (1984). ce, cation exchange thermometry; oxi, oxygen isotope thermometry; pd, phase diagram calculations.

Evia were not high enough to allow garnet to crystallize in metabasites. The preservation of relict igneous minerals and lawsonite in Evia metabasites also indicates that M_1 temperatures were lower than for the CEX; they were estimated at 400–430 °C (Katzir *et al.* 2000).

The lower and upper pressure limits on Syros are defined, respectively, by the occurrence of Jadeite$_{92}$ + quartz in meta-acidites and of paragonite instead of omphacite–kyanite in eclogites (Ridley 1984*b*; Schliestedt *et al.* 1987; Okrusch & Bröcker 1990). At 470 °C the corresponding reactions of jadeite + quartz = albite and paragonite = jadeite + kyanite+ H_2O define a pressure range of 12–20 kbar (Schliestedt 1986; Schliestedt & Okrusch 1988; Okrusch & Bröcker 1990). The maximum jadeite content of sodic clinopyroxene in Evian meta-acidites is equal to that of their counterparts on Syros, indicating similar minimum M_1 pressures of ≥12 kbar (Schliestedt *et al.* 1987; Katzir *et al.* 2000).

There are two possible ways to interpret the distribution of the M_1 *P–T* estimates in the NW Cyclades and on southern Evia. In the first interpretation, peak pressures, like temperatures, were lower on southern Evia compared with the CEX. Consequently, Evian blueschists were subducted to shallower levels than the Cycladic eclogites, and both terranes experienced similar *P–T* trajectories. Alternatively, Evian blueschists and Cycladic eclogites represent the same depth interval of the subducted plate; however, the latter remained longer at deep crustal levels, which allowed prolonged heating.

Metamorphic overprint (M_2)

The M_2 metamorphic overprint occurred throughout the CEX at greenschist-facies conditions. A typical M_2 assemblage in metabasites includes albite–chlorite–epidote–phengite–calcic amphibole. The presence of barroisitic amphibole in some greenschists and the scarcity of biotite in metapelites constrain M_2 conditions to 6–7 kbar and ≤450 °C, respectively (Schliestedt *et al.* 1987; Bröcker *et al.* 1993). Oxygen isotope thermometry in Sifnos (quartz–epidote; quartz–phengite) and Tinos (quartz–magnetite) greenschist-facies rocks gave temperatures of 400–500 °C and 440–470 °C, respectively (Matthews & Schliestedt 1984; Bröcker *et al.* 1993). The temperature ranges of the M_1 high-pressure metamorphism and the M_2 greenschist overprint overlap, thus indicating that within the M_1–M_2 time interval the high-pressure rocks of the CEX experienced isothermal decompression. This two-point isothermal path was further established by a detailed petrological study of a Sifnos rock sequence that described several stages of metamorphic equilibration at successively decreasing pressures: relict eclogite-facies rocks transformed into albite-bearing, garnet-free epidote blueschist to be finally equilibrated at greenschist-facies conditions (Avigad *et al.* 1992). Recent studies claimed to show, however, that high-pressure rocks of the CEX experienced significant cooling during decompression (Table 2). Based on local equilibria in chlorite–phengite-bearing assemblages, Trotet *et al.* (2001*b*) argued that eclogites at the top of the CBU on Sifnos and Syros cooled during decompression whereas the lower, strongly overprinted part of the section decompressed isothermally. THERMOCALC-based petrogenetic grids calculated for the well-preserved eclogites and for the underlying greenschists on Sifnos show, however, decompressional cooling for both sequences (Schmädicke & Will 2003). The temperatures estimated for the retrograde M_2 overprint in both studies do not appreciably differ from the 'traditional' petrological and isotope thermometry estimates: they all converge at 400–500 °C (Table 2). The apparent cooling postulated for the Cycladic eclogites mostly stems from new peak temperatures assigned to the M_1 event: 550–600 °C. These estimates disagree with numerous well-established petrological and isotope studies (see above) and thus isothermal decompression still seems the most probable path for exhumation in the CEX.

The M_2 assemblage of metabasites in the Evian mélanges, actinolite–pumpellyite–epidote–chlorite–albite–phengite, is characteristic of the pumpellyite–actinolite facies (Banno 1998). Phase diagram calculations using mineral compositions in Evian metabasites limit this paragenesis to pressures of 4–8 kbar and temperatures of 300–350 °C (Table 2; Katzir *et al.* 2000). The M_2 assemblage in Tsaki metagabbros differs, however, from its Ochi counterpart: it includes sodic augite, which indicates lower temperatures relative to the clinopyroxene-free Ochi assemblage (e.g. Maruyama & Liou 1985).

By studying mineral assemblages in ophiolitic mélanges, a clear distinction between the metamorphic histories of the Lower tectonic Unit of Evia and the CEX can be made: whereas Syros eclogites ($T = 450$–500 °C) decompressed isothermally through greenschist-facies conditions, decompression of the lower-grade Evian blueschists ($T = 400$–430 °C) involved cooling to pumpellyite–actinolite-facies conditions ($T = 300$–350 °C). Moreover, Tsaki metabasites at the base of the section on Evia were overprinted at lower temperatures compared with their Ochi equivalents that occur 2 km upsection. This decrease in M_2 temperatures downwards towards the base of the Blueschist Unit can be explained by the underthrusting of the Late Eocene sedimentary rocks of the Basal Unit (Almyropotamos platform), which either caused cooling by conduction or resulted in a more prolonged retrograde metamorphism impelled by the infiltration of fluids from below.

Leucogneiss-associated peridotites (Naxos)

Both the protoliths and the metamorphic evolution of the sequence that hosts ultramafic horizons in the Lower Unit of Naxos are very different from the NW Cycladic mélanges. On Naxos an Early Miocene (18 Ma; Andriessen *et al.* 1979; Wijbrans & McDougall 1988; Keay *et al.* 2001). Barrovian-type overprint (M_2) occurred during extension and exhumation of former high-pressure rocks (M_1) and almost totally effaced their assemblages and fabrics. Naxos (Fig. 5) is a mantled gneiss dome whose core consists of migmatites formed during M2 metamorphism of quartzofeldspathic ortho- and para-gneisses (Buick 1988; Pe-Piper *et al.* 1997; Keay *et al.* 2001). Overlying the leucogneiss core (Buick 1988) is a 7 km thick metasedimentary envelope dominated by siliciclastic schists and gneisses in its lower part ('Lower Series' of Jansen & Schuiling 1976) and by meta-bauxite-bearing marbles in its upper part ('Upper Series'). The grade and intensity of M_2 metamorphism increase with increasing structural depth from greenschist-facies rocks containing relict M_1 assemblages at the top of the sequence to upper amphibolite-facies rocks in the core. The approximate range of M_2 temperatures spanned by the Barrovian facies series is 400–700 °C at pressures of 5–7 kbar (Jansen 1977; Feenstra 1985). Further petrological studies refined the peak M_2 conditions that have occurred at the leucogneiss core to 6 ± 2 kbar and 670–700 °C (Buick & Holland 1989, 1991; Katzir *et al.* 1999, 2002; Matthews *et al.* 2003). However, remnants of the former high-pressure mineralogy (M_1) have not been found in the leucogneiss core of Naxos, thus its exhumation P–T path remains speculative. A first view of the pre-M_2 evolution of the Naxos core was made possible by petrological and oxygen isotope study of peridotite lenses hosted by the silicic gneisses (Katzir *et al.* 1999, 2002).

Field relations and M_2 metamorphism

Ultramafic horizons occur at four structural levels within the metamorphic sequence of Naxos (Fig. 5): (1) the Main Ultramafic Horizon (MUH) occurs within upper amphibolite-facies rocks at the transition from the leucogneiss core to the Lower Series; it is composed of 1–10 m sized lenses of massive to moderately foliated medium- to coarse-grained meta-peridotites; (2) the Agia ultramafic horizon occurs in NW Naxos within sillimanite-grade rocks of the Lower Series; (3) sporadic ultramafic bodies occur within staurolite–kyanite-grade rocks at the transition from the Lower to the Upper Series; and (4) within greenschist-facies rocks of the Upper Series.

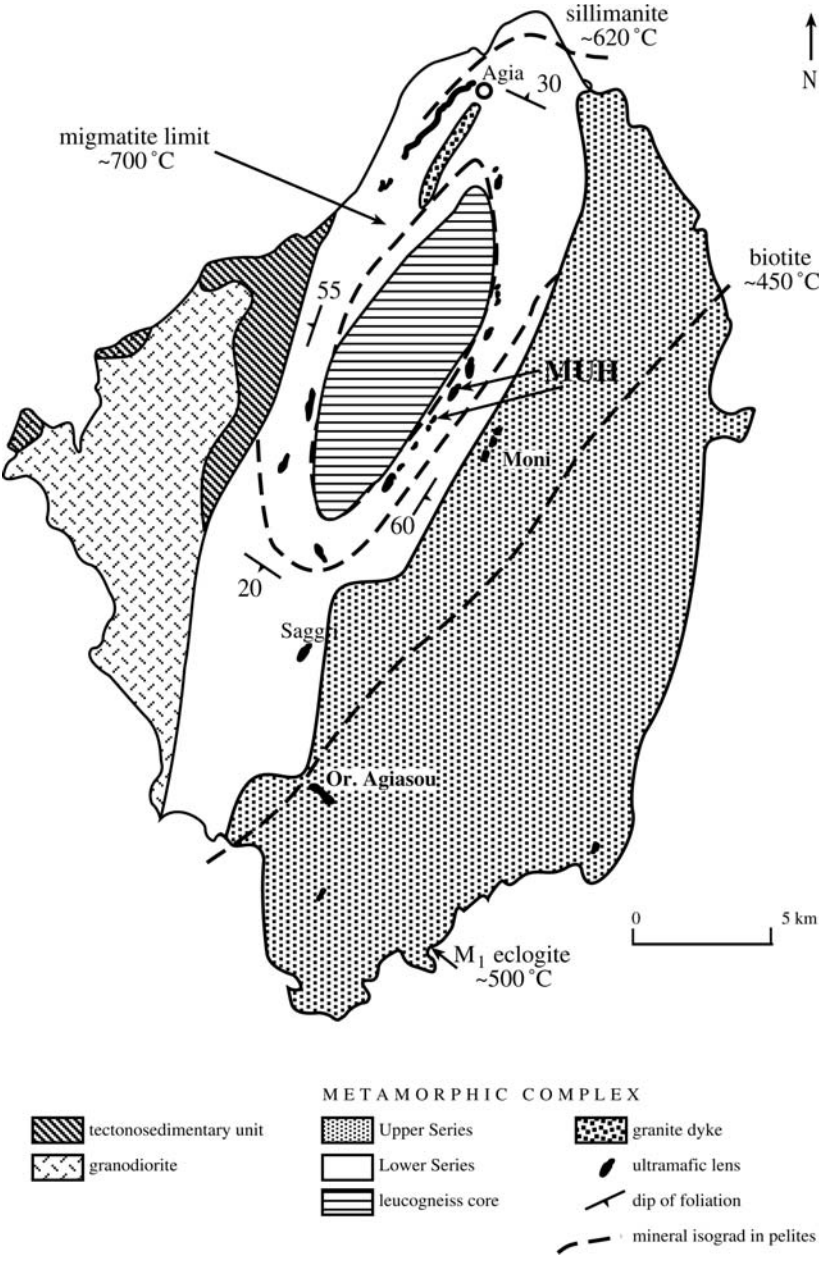

Fig. 5. A simplified geological map of Naxos after Jansen & Schuiling (1976) showing the metamorphic complex of the CBU (subdivided), the tectonosedimentary Upper Unit and a 12 Ma granodiorite. Ultramafic rocks (in black) occur within the metamorphic dome at four structural levels: (1) the Main Ultramafic Horizon (MUH) occurs at the transition from the leucogneiss core to the overlying Lower Series; (2) the Agia ultramafic horizon within the Lower Series; (3) at the transition from the Lower to Upper series (Moni and Saggri exposures); (4) within the Upper Series (Ormos Agiasou exposure). The M_2 mineral isograds mapped by Jansen & Schuiling (1976) and modified by Feenstra (1985) and Buick (1988) are shown. The occurrence of particularly well-preserved M_1 high-pressure rocks at the top of the section on SE Naxos is shown (Avigad 1998).

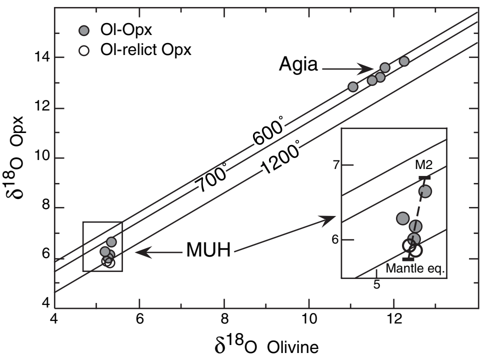

Fig. 6. Plot of δ^{18}O (orthopyroxene) v. δ^{18}O (olivine) in meta-peridotites of the MUH and the Agia ultramafic horizon on Naxos. Isotherms are calculated according to Rosenbaum *et al.* (1994). The analytical uncertainty is smaller than the symbol size.

A non-deformed talc–enstatite assemblage and extremely high δ^{18}O values of olivine and orthopyroxene (11–14‰; Fig. 6) indicate that recrystallization of the Agia meta-peridotites occurred during post peak M_2 infiltration of silica-rich fluids from pegmatites (Katzir *et al.* 2002; Matthews *et al.* 2003). This fluid-flow event erased the former mineralogy of the Agia peridotites, thus precluding further investigation of their origin and emplacement. Notwithstanding, both the MUH meta-peridotites and the ultramafic bodies that occur at the base and within the Upper Series are structurally concordant and isofacially metamorphosed with their host rocks. A synkinematic olivine–orthopyroxene–hornblende–chlorite–spinel assemblage is dominant in the MUH peridotites. Anthophyllite–talc and antigorite–talc schists characterize the two ultramafic occurrences of the Upper Series, respectively. Temperature estimates of the ultramafic assemblages agree well with those of the host M_2 Barrovian series: *c.* 700 °C in the sillimanite schist-associated MUH, *c.* 580 °C in anthophyllite–talc schists enclosed by staurolite–kyanite-grade pelites and $<$450 °C in antigorite–talc schists of the biotite zone. Highly foliated metasomatic reaction zones are well developed at the ultramafic–felsic rocks interface: phlogopite–actinolite–anthophyllite in the MUH and chlorite–tremolite in the structurally higher located bodies. The variable mineralogy of the metasomatic zones is mostly dependent on the chemical potentials of various ions, but is also consistent with increase in temperature going downsection from the Upper Series to the leucogneiss core.

Field and petrological evidence thus indicates that the Naxos meta-peridotites have experienced the Barrovian M_2 metamorphism with their host rocks.

Pre-M_2 evolution

Relict phases in ultramafic rocks of different structural levels on Naxos indicate two distinct pre-M_2 histories. In the upper two ultramafic bodies mesh textured serpentine associated with fine-grained magnetite is overgrown by the M_2 talc-bearing assemblages. The occurrence of early serpentine fabrics overprinted by M_2 mineralogy indicates that the Upper Series ultramafic rocks were first denuded and hydrated prior to Alpine metamorphism either on the sea floor or during ophiolite emplacement. In contrast, the MUH peridotites show no signs of early serpentinization and instead preserve their mantle oxygen isotope signature (δ^{18}O (Ol) = 5.2‰) and some of their original mantle assemblage. Large, moderately deformed porphyroclasts of olivine and orthopyroxene are randomly oriented within the M_2 recrystallized matrix. The pre-kinematic orthopyroxenes have high-Al_2O_3 and high-CaO cores (up to 5.5% and 0.9%, respectively) and contain exsolution lamellae of Cr-spinel. Other relicts of the pre-M_2 peridotite include rare grains of high-Al green spinel (60 wt% Al_2O_3). Aluminium and Ca-in-Opx thermometry of relict orthopyroxene, olivine and spinel yields temperatures of *c.* 1050 °C (Katzir *et al.* 1999). Oxygen isotope Opx–Ol thermometry in the MUH meta-peridotites gives a bimodal distribution of temperatures, grouped at 700 °C and 1200 °C, and indicates partial oxygen exchange during M_2 superimposed on previous mantle fractionation (Fig. 6; Katzir *et al.* 2002).

Because emplacement of peridotites into shallow crustal levels invariably involves serpentinization it is concluded that the MUH peridotites were directly transported from the mantle and tectonically interleaved with the continental crustal section of Naxos at depth. Probably, the incorporation of the ultramafic rocks into the upper crustal section occurred while the latter was underthrust and buried to great depth during Alpine collision and high-pressure metamorphism (M_1). Their emplacement at the base of the orogenic wedge is inferred to have involved isobaric cooling from temperatures of *c.* 1050 °C within the spinel lherzolite field to eclogite-facies temperatures higher than 500 °C, the upper stability limit of serpentine. Further support for the deep origin of the Naxos core is given by down-section extrapolation: M_1 pressures of $\geq$12 kbar and temperatures of $\leq$500 °C were estimated for relict eclogite at the top of the section on SE Naxos, corresponding to a regional average gradient of *c.* 15 °C km^{-1} (Fig. 5; Avigad 1998). Bearing in mind that the present-day 7 km thick section separating SE Naxos from the leucogneiss core underwent post-M_1 ductile thinning (Buick 1991), minimum pressures of *c*. 14 kbar and temperatures of 600 °C may be estimated for M_1 at the core.

A dismembered ophiolite (Tinos Upper Unit)

On the island of Tinos several slices of the Upper Unit were tectonically juxtaposed over the CBU before the emplacement of an 18 Ma old granite, which intrudes the contact between them (Fig. 7). A detailed study of the Upper Unit occurrences revealed that they consist mostly of a dismembered ophiolite sequence (Katzir *et al.* 1996). This includes tens-of-metres thick slices of strongly sheared phyllite (metabasalt), mostly at the base, sheared and massive serpentinites and metagabbros (Fig. 8). These slices reach a total thickness of up to 200–300 m. The strong slicing greatly reduced the thickness of the ophiolite and also disrupted the typical order of the various lithologies.

In spite of their common oceanic affinity, the Tinos dismembered ophiolite and the NW Cyclades HP–LT ophiolitic mélanges differ in major aspects of their evolution.

(1) Lower than igneous oxygen isotope ratios in relict oceanic actinolite of metagabbros from the Syros mélange (3–5‰; Putlitz *et al.* 2000) are similar to ratios measured in modern oceanic gabbros and in the intact, fully developed ophiolite suites of Oman and Troodos (Heaton & Sheppard 1977; Gregory & Taylor 1981). These isotope ratios indicate high-temperature interaction with seawater during the generation of new oceanic crust. In contrast, hornblende of sub-sea-floor hydrothermal origin in metagabbros of the Tinos

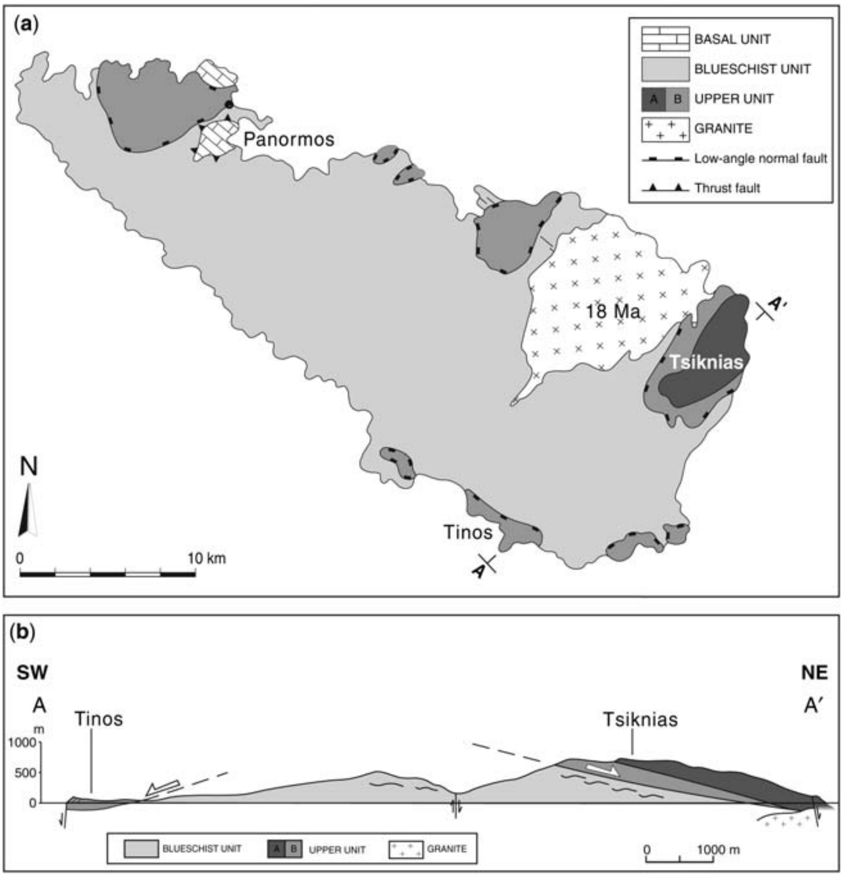

Fig. 7. (**a**) Geological map of Tinos (after Avigad & Garfunkel 1989). The Upper Unit includes slices of a dismembered ophiolite: serpentinites and metagabbros (A, dark grey) overlying mafic phyllites (B, bright grey). (**b**) A structural cross-section (A–A′ in (a)) of Tinos showing the main detachment surface that separates the Upper Tectonic Unit in the hanging wall from the underlying CBU. An 18 Ma granite that intrudes both tectonic units near Tsiknias gives an upper time constraint for their juxtaposition.

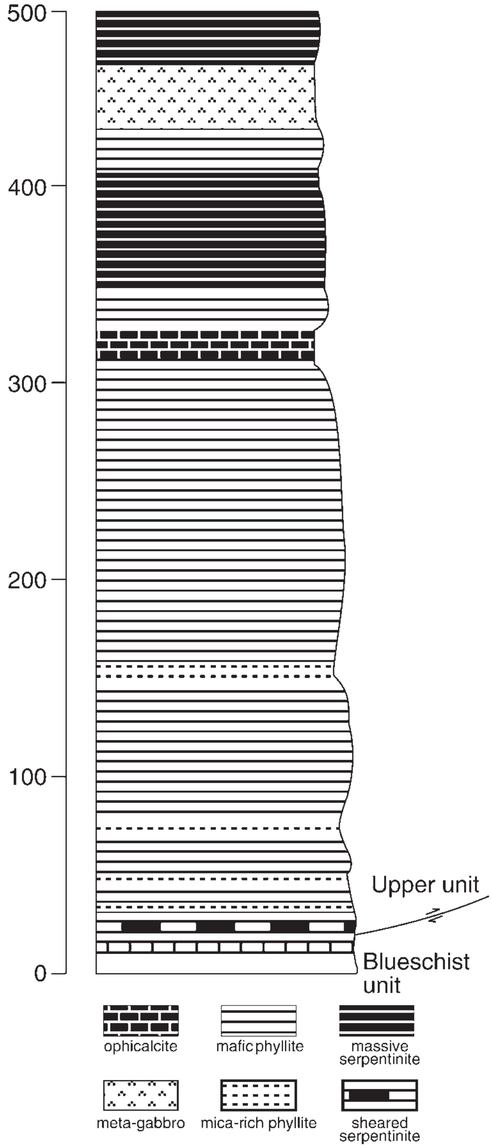

Fig. 8. General lithological columnar section of the dismembered ophiolite of the Upper Unit on Tinos (after Katzir *et al.* 1996; Zeffren *et al.* 2005). On the basis of textural and geochemical observations, mafic phyllites were interpreted as metabasalts.

Upper Unit has higher than igneous $\delta^{18}O$ values (5.5–7.5‰, Putlitz *et al.* 2001). The high $\delta^{18}O$ values of the hornblendes are compatible with interaction of oceanic gabbros with seawater that had previously been enriched in $^{18}O/^{16}O$ by isotope exchange at high temperatures. Isotope exchange is thought to have occurred by deep penetration of seawater during early, hot intra-oceanic thrusting (Putlitz *et al.* 2001). Thus, the Syros and Tinos metagabbros represent oceanic floor segments that have undergone different sea-floor alteration: the Syros metagabbros represent 'typical' sea-floor alteration whereas the Tinos metagabbros represent a different environment, affected by tectonic disturbance.

(2) In contrast to the random occurrence of ophiolitic lithologies in the mélanges, the Tinos ophiolite consists of coherent slices, each slice dominated by a single oceanic lithology and bounded by a low-angle tectonic contact (Fig. 8). The sequence of slices does not follow the primary order of the ophiolite suite, but sometimes repeats or reverses the original order. Dissection of the primary oceanic crust by low-angle reverse faults can account for the disturbed order of the oceanic lithologies. However, reverse faulting cannot account for the absence of primary components in the Tinos ophilolite (deep-sea sediments), or for the significant reduction in its thickness. Omission of major portions of the ophiolite suite requires, instead, a post-thrusting phase of normal faulting with flat-lying faults that cut out substantial portions of the original slice pile. Such extension agrees well with the strong slicing within the ophiolite and with the normal faulting at the base of the Upper Unit (Avigad & Garfunkel 1989, 1991; Patriat & Jolivet 1998).

(3) Unlike the HP–LT mélanges of the NW Cyclades, the Tinos ophiolite was never buried to great depths. However, it was metamorphosed at greenschist-facies conditions in the course of the Alpine orogenesis. Greenschist-facies mineralogy overprints the early oceanic fabrics in all major rock types of the ophiolite and is associated with penetrative deformation in the phyllites and sheared serpentinites and gabbros. Some of the metamorphic mineral growth in the Upper Unit exposure of Mt. Tsiknias, including crystallization of olivine neoblasts in serpentinite, was considered to occur by contact metamorphism induced by the intrusion of the granite at 18 Ma (Fig. 8; Stolz *et al.* 1997). However, this cannot account for a major penetrative regional metamorphism observed in all rock types and exposures of the Tinos Upper Unit. Given the field evidence for early oceanic thrusting, Katzir *et al.* (1996) suggested that metamorphism was induced by continued thrusting and piling of nappes that have created the necessary overburden to cause greenschist-facies conditions. Based on K–Ar ages on amphiboles extracted from an amphibolite-facies slice that constitutes the topmost part of the Upper Unit on Tinos (Patzak *et al.* 1994), a Late Cretaceous age was assigned to the metamorphic event that affected the entire sequence of the Upper Unit (Katzir *et al.* 1996). Late Cretaceous ages were also determined

for Upper Unit gneisses and amphibolites on several other islands in the Cyclades (Reinecke *et al.* 1982; Maluski *et al.* 1987; Altherr *et al.* 1994). The first geochronological results for the phyllites of the Upper Unit were obtained by Rb–Sr dating, which yielded ages between 21 and 92 Ma (Bröcker & Franz 1998). The youngest age was obtained for a sample collected near the contact with the CBU. Recently obtained $^{40}Ar/^{39}Ar$ ages on synkinematic white micas from phyllites at the base of the Upper Unit were concentrated in the Oligocene–Miocene (31–21 Ma; Zeffren *et al.* 2005). These ages suggest that the rocks represent an extensional shear zone that operated in the Late Oligocene and juxtaposed the Upper Unit on top of the partially exhumed high-pressure rocks of the CBU. It is thus concluded that the lower part of the Upper Unit experienced recrystallization and age resetting in a time interval that corresponds to the M_2 overprint in the CBU. The upper part of the dismembered ophiolite records, however, an older, pre-Tertiary metamorphic history.

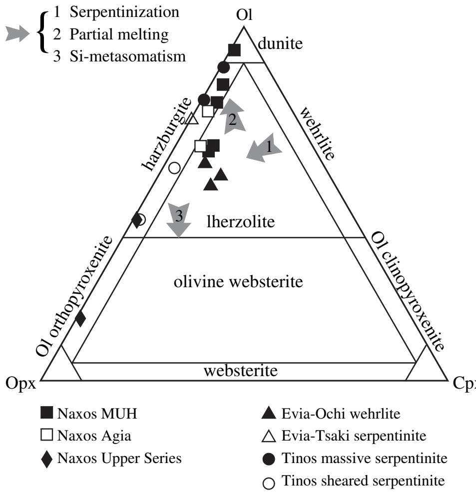

Fig. 9. Spinel peridotite norm compositions of the ultramafic rocks of the Cyclades plotted on the ultramafic classification diagram. Arrows indicate compositional trends resulting from serpentinization, partial melting and Si-metasomatism.

Discussion

Assembly and evolution of the ultramafic associations

The origin and manner of assembly of the major ultramafic rock-bearing associations in the Cyclades are diverse. A clear distinction exists between the Naxos association, where thin ultramafic horizons occur within continental quartzofeldspathic gneisses (MUH) and continental platform metasediments (the Agia and Upper Series horizons), and the other two associations, where ultramafic rocks are involved with oceanic crust. The oceanic associations are further distinguished by their manner of assembly: ophiolitic olistotrome intervals within basinal sedimentary sequences in the Lower Unit compared with sliced dismembered ophiolite in the Upper Unit. The ultramafic rocks of the Lower Unit, both on Naxos and in the NW Cyclades, were incorporated into their respective platformal and basinal host sequences prior to or during the high-pressure M_1 metamorphism. Since then they have experienced with their host rocks the whole cycle of Alpine Tertiary orogenesis including compression, exhumation and metamorphic overprint during extension. The ultramafic slices of the Tinos Upper Unit, however, escaped most of the Tertiary tectonometamorphic cycle, but could not avoid the extensional overprint that resulted in their final juxtaposition against other ophiolitic slices in the Early Miocene (Zeffren *et al.* 2005).

The combined effect of the processes that shaped the Cycladic ultramafic rocks since their derivation from the mantle, through polymetamorphism to denudation and serpentinization (or vice versa) is reflected in their whole-rock geochemical compositions. Partial melting in the mantle, metasomatism by Si-rich fluids derived from crustal host rocks, and serpentinization have distinct geochemical signatures that are clearly seen on major element variation diagrams. Spinel peridotite norms of the Cycladic ultramafic rocks were calculated and plotted on the peridotite classification diagram (Fig. 9). The peridotites of the MUH on Naxos form a roughly vertical trend going from the lherzolite field into the harzburgite and dunite fields. This trend and the negative linear correlations of SiO_2, CaO and Al_2O_3 with MgO (Fig. 10) are compatible with mantle depletion caused by partial melt extraction. The MUH samples with the lowest MgO contents are very close in composition to primitive upper mantle estimates (Hart & Zindler 1986; Hofmann 1988), whereas the others could form by different degrees of partial melting. The compositions of Agia meta-peridotites plot close to the depletion trend defined by the MUH, suggesting a common origin. Excluding the Naxos MUH and Agia rocks, most other ultramafic rocks plot along the Ol–Opx join of the triangular diagram (Fig. 9). This is a clear indication of serpentinization, which always involves major loss of Ca. Even the Ochi wehrlites, which still have significant Ca

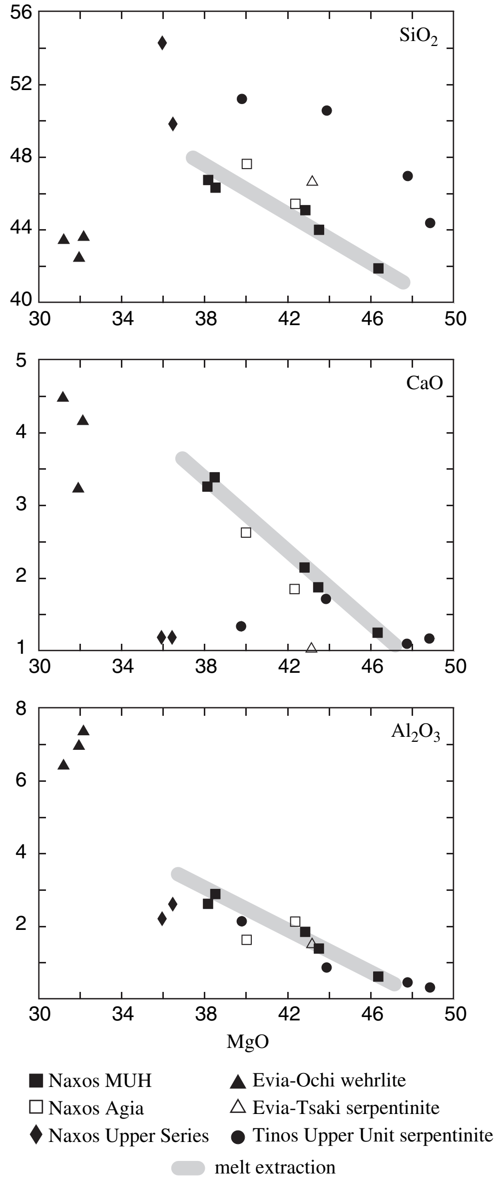

Fig. 10. Variation diagrams of SiO_2, CaO and Al_2O_3 v. MgO for the ultramafic rocks of the Cyclades.

contents, plot within the lherzolite field far to the left of their original compositions, as a result of a high degree of serpentinization. The geochemical modifications during serpentinization, including significant to complete Ca loss and increase in the Si/Mg ratio, are also illustrated in the variation diagrams (Fig. 10). The Tsaki, Naxos Upper Series and Tinos Upper Unit meta-serpentinites have almost no CaO regardless of the MgO content. For a given MgO content they have higher SiO_2 than the Naxos meta-peridotites, which have not experienced serpentinization. Except for the Ochi wehrlites, which are not true mantle peridotites, all the other rocks form a general negative correlation trend on the Al_2O_3–MgO variation diagram. This is not surprising, as Al behaves as a conservative element during serpentinization and its content may be used as a qualitative criterion of mantle fertility. The Naxos Upper Series serpentinites plot on the extreme left side of the peridotite classification diagram; however, their norms, which have less than 50% olivine, cannot be accounted for by serpentinization alone (Fig. 9). Their relatively high SiO_2 contents indicate that after Ca loss in serpentinization, they were silica enriched by externally derived fluids. The occurrence of synkinematic talc in their M_2 metamorphic assemblages and the blackwalls at their contacts with the host quartzofeldspathic rocks suggest that they were metasomatized by Si-rich fluids derived from metamorphic reactions in the adjacent country-rocks. A similar scenario may apply to one sample of Tinos serpentinite taken from a thin, highly sheared talc-rich horizon at the base of the Upper Unit.

Geochemical trends thus indicate a fundamental difference between the Naxos MUH meta-peridotites and the rest of the ultramafic rocks in the Cyclades. Whereas the latter were first exhumed and serpentinized at near-surface conditions prior to Alpine metamorphism, the MUH meta-peridotites were incorporated at depth into the orogenic wedge, thus avoiding low-temperature hydration.

P–T *paths in the Cyclades*

The equilibrium P–T conditions for blueschists and eclogites usually lie at higher pressures and lower temperatures than most normal geothermal gradients. Thermal modelling suggests that only exceptionally rapid uplift can prevent reinstatement of an equilibrium geothermal gradient with consequent heating and destruction of the high-pressure mineral assemblages (England & Thompson 1984; Thompson & Ridley 1987). Abundant field and thermometric data from the Cyclades and many other high-pressure orogenic belts show, however, that high P–T rocks decompressed adiabatically with almost no heating, or even with cooling (Ernst 1988; Dunn & Medaris 1989; Platt 1993). In particular, the petrological and geochronological studies on Sifnos, Syros and Tinos have shown that the Cycladic eclogites decompressed isothermally at *c.* 450–500 °C from 15 ± 3 kbar at *c.* 50–45 Ma to 5–7 kbar at *c.* 23–21 Ma. Extensional tectonism has thus been suggested as responsible for the rapid unroofing and exhumation of eclogites in the Cyclades (Lister *et al.* 1984; Jolivet & Patriat 1999; Trotet *et al.* 2001*a*).

The new petrological analysis of metamorphic assemblages in ultramafic and associated rocks on Evia and Naxos allows redrawing of the *P–T* paths of the Lower tectonic Unit away from the CEX and offers a broader perspective on the exhumation processes in the Cyclades (Fig. 11). Because of the occurrence of pre-Alpine basement in a lowermost structural position, the leucogneiss core of Naxos is considered as the deepest exposed levels of the Alpine orogen in the Cyclades. However, its *P–T* path remains speculative, as anatexis and deformation during exhumation (M_2) totally destroyed any evidence for its evolution during collision and M_1 metamorphism. The occurrence of non-serpentinized fertile spinel lherzolites, probably representing the subcontinental mantle, within the leucogneiss core supports its deep origin. The peridotites were interleaved with the subducted upper crustal section during M_1 and can thus serve as indicators of peak M_1 temperatures. The lack of serpentinization and the preservation of relict >1000 °C mantle assemblage indicate that the peridotites cooled to temperatures of *c*. 500–650 °C during M_1 (Katzir *et al.* 1999). Comparable temperatures and pressures >14 kbar are given by down-section extrapolation from relict eclogites on SE Naxos (Avigad 1998). Given the well-constrained temperatures of 670–700 °C calculated for the M_2 high amphibolite overprint (Buick & Holland 1989, 1991), a *P–T* path of decompressional heating emerges for the leucogneiss core on Naxos (Fig. 11).

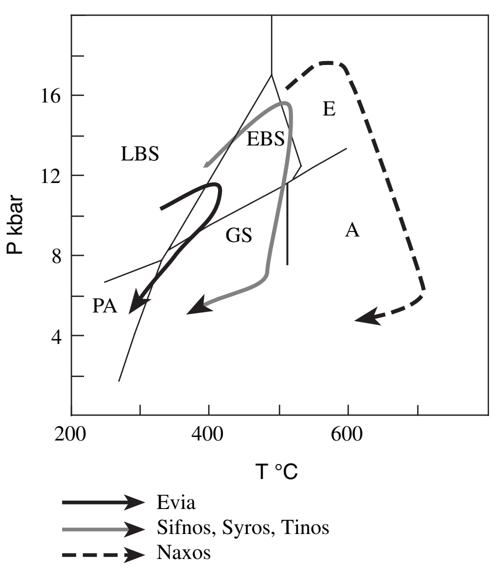

Fig. 11. Pressure–temperature paths for the Cycladic Blueschist Unit (CBU) in the Central Eclogite Axis (Tinos, Syros and Sifnos), on Naxos and on southern Evia plotted on the metamorphic facies diagram of Evans (1990). Abbreviations for metamorphic facies: A, amphibolite; E, eclogite; EBS, epidote blueschist; GS, greenschist; LBS, lawsonite blueschist; PA, pumpellyite–actinolite.

The high-pressure metamorphic rocks of the Cyclades show diverse *P–T* paths (Fig. 11). Whereas eclogites on Sifnos, Syros and Tinos decompressed isothermally, blueschists on Evia cooled and the deepest-buried, peridotite-associated gneisses on Naxos were heated during decompression. The pattern formed by the *P–T* paths in the Cyclades is not random: the higher the maximum pressure is, the 'hotter' is the exhumation path. This pattern was predicted by thermal modelling of thickened continental crust unroofed by erosion (England & Richardson 1977). The *P–T* path of any single metamorphic rock is governed by two concurrently competing processes: temperature increase by conductive relaxation and decompression by erosion of the orogenic pile. Assuming relatively rapid and uniform thickening of the crust, the magnitude of temperature increase is dependent on the time that elapses before the rock is exhumed sufficiently to be affected by the proximity of the cold upper boundary (England & Thompson 1984). Thus the deeper the rock is buried within the orogenic pile, the longer will be the period during which its temperature increases, and the hotter will be the overprint during exhumation. Although developed for erosion, the assumptions and heat transfer equations of the model hold regardless of the actual mechanism responsible for bringing the rocks nearer the surface. Thus the Eocene to Miocene removal of *c*. 30 km of rock overburden implied by the diverse *P–T–t* paths of the Cycladic Blueschist Unit could have been accomplished by erosion, uniform 'pure shear' extension, or a combination of both. However, according to the England & Thompson model, cooling during decompression is possible only if exhumation of the underthrust rocks began very early, simultaneously with the onset of conductive heating. Mid-Eocene flysch that tops the Basal Unit in the Almyroptamos tectonic window of southern Evia indicates such early synorogenic erosion. Generally, however, thick sequences of Eocene or younger clastic sediments are not exposed in the Cyclades. An efficient mechanism that can account for the decompressional cooling of the Evian HP–LT rocks is accretion of relatively cold rocks underneath them (Rubie 1984). Underthrusting of the Basal Unit beneath the Blueschist Unit can also account for the inverted M_2 temperature gradient within the Blueschist Unit towards the basal thrust on Evia (Katzir *et al.* 2000). Cooling by conductive heat transfer is also indicated by thermometry of marbles across the basal thrust on Tinos (Matthews *et al.* 1999). Notwithstanding, the cooling during decompression inferred by Trotet *et al.* (2001*b*) for the uppermost part of the Blueschist Unit in the CEX was attributed to early exhumation by non-uniform, 'simple-shear' extension manifested by deep ductile shear zones (Ruppel *et al.* 1988; Ruppel 1995).

In summary, no single $P-T$ path should be assigned to the Cycladic Blueschist Unit. Instead, a common $P-T$ trajectory for the M_1 high-pressure metamorphism can be drawn. Going southeastwards from southern Evia, progressively deeper levels of the subducted plate are exposed. Correspondingly, temperatures of the M_2 overprint also increase from pumpellyite-bearing assemblages on southern Evia, through greenschists on the CEX, to upper-amphibolite, sillimanite-bearing gneisses on Naxos. The array of enveloping $P-T$ loops thus given is compatible with exhumation by uniform attenuation of the crust accomplished by either erosion or extension. Post-M_1 thrusting of the Blueschist Unit on top of the Basal Unit exerts excess cooling at its base on southern Evia. Restricted cooling at the top of the Blueschist Unit on Sifnos and Syros was explained by local 'simple shear'-type extension that resulted in early exhumation.

Provenance of the Cycladic ultramafic rocks

The relative positions of the Cycladic Massif and adjacent terraines on both sides of the Aegean Sea at the beginning of the Neogene are shown in Figure 12 (from Garfunkel 2004). This palaeogeographical reconstruction, which corrects for the Neogene extension and block rotation in the Aegean Sea, is used as a guide for names and locations of major rock units and terranes in the following discussion.

Several models integrating the tectono-metamorphic evolution of the Cycladic Massif into the Alpine history of the Hellenides in continental Greece have been proposed (Biju-Duval *et al.* 1977; Bonneau 1984; Papanikolaou 1987). All of the models emphasize two orogenic events: (1) the Eohellenic event of Late Jurassic age that involved emplacement of ophiolites from the Pindos Ocean over the Pelagonian continental margins (Smith 1993); (2) the Meso-Hellenic final closure of the Mesozoic Pindos basin by subduction and eventual collision of the Apulian and Pelagonian microplates in Early to Middle Eocene times (Robertson & Dixon 1984). This collisional event is thought to have caused the high-pressure metamorphism of rock sequences of the underthrust Apulian plate, including the Pindos deep-water sediments (possibly the NW Cyclades) and the Apulian continental basement and platform (possibly Naxos; Blake *et al.* 1981; Papanikolaou 1987).

Diverse origins and modes of emplacement have been deduced in this study for the ultramafic rocks of the Cyclades. However, the field relations and history of the majority of ultramafic occurrences in the Cyclades hardly fit the above two-stage evolution of the Hellenides. The simplest rocks to correlate are the serpentinites of the Paros Upper Unit, which are covered by transgressive Barremian limestone and thus probably represent the Eohellenic event. Apparently, the high $P-T$ ophiolitic

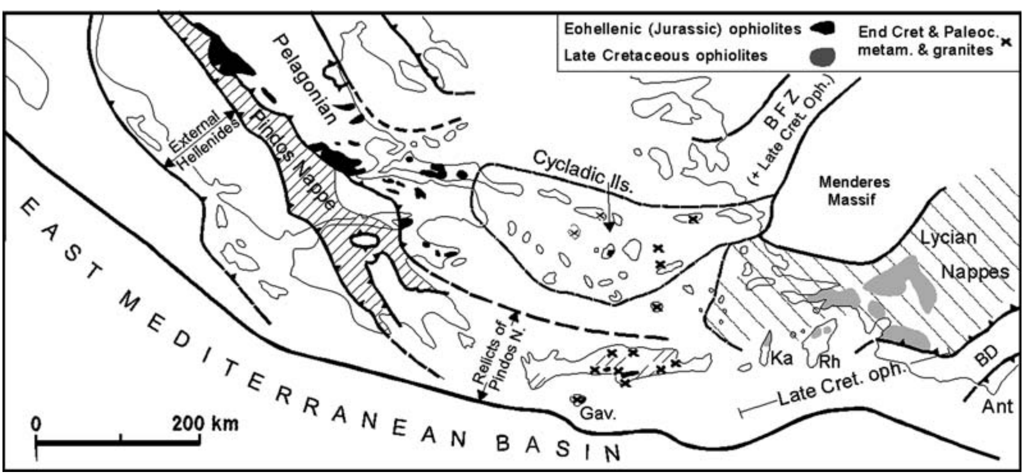

Fig. 12. Major continental terranes on the two sides of the southern Aegean Sea at the beginning of the Neogene (from Garfunkel 2004). Ant, Antalya complex; BD, Bey Daglari platform; BFZ, Bornova Flysch Zone; Gav, Gavdos; Ka, Karpathos; Rh, Rhodes. The Cyclades Islands are interpreted as underlain by a distinct continental fragment, of unclear provenance, which has Variscan age crust.

mélanges on southern Evia and Syros embedded in a pelagic clastic sequence might be correlated to the subducted Pindos Ocean. Ion-probe study of zircon crystals separated from an eclogitized metagabbro block from the Syros mélange yielded Late Cretaceous ages of *c.* 80 Ma (Tomaschek *et al.* 2003). These were interpreted as the crystallization age of the protolith; however, Late Cretaceous oceanic magmatism has not been recognized anywhere in continental Greece. The residual Cretaceous Pindos basin is a possible source for the Cycladic eclogites, but other oceanic basins that show unequivocal evidence for Late Cretaceous sea-floor spreading, such as those represented by the ophiolites of the Lycian nappes to the east (Fig. 12), should also be considered as potential source terranes for the Syros ophiolitic mélange.

The rocks of the structurally lowest ultramafic horizon on Naxos (MUH) are considered to be mantle flakes that intermingled with an upper continental crustal section while the latter was buried to great depth during the Eocene continental collision. Nevertheless, correlating the underthrust Naxos continental section including ortho- and paragneisses, pelite schists and thick karst-bauxite-bearing marbles with equivalent unmetamorphosed sequences in the Hellenides or elsewhere is not straightforward. Variscan and younger pre-Alpine ages are dominant in zircons separated from various gneisses in the Naxos core (Keay *et al.* 2001). Given its Pan-African basement, correlation with the Menderes Massif of SW Turkey is ruled out (Fig. 12). The provenance of the Naxos section should be sought in areas with Variscan continental basement, such as the Apulian and Pelagonian zones of the external and internal Hellenides, respectively. Pelagonian sequences tectonically overlie the CBU on Evia and are the main source of clasts for the Upper Unit conglomerates on Mykonos (Shaked *et al.* 2000; Sanchez-Gömez *et al.* 2002). It is hard to envision how the Pelagonian section can form both the uppermost and lowermost parts of the present orogenic pile in the Cyclades. Likewise, the occurrence of serpentinized ultramafic rocks within the Upper Series of Naxos makes its correlation with the ophiolite-devoid shallow-water sedimentary cover of Apulia questionable.

The ophiolitic slices of the Tinos Upper Unit were metamorphosed at greenschist- to amphibolite-facies conditions in Late Cretaceous times (Avigad & Garfunkel 1991; Patzak *et al.* 1994; Bröcker & Franz 1998). This period of time was characterized by renewed pelagic sedimentation in the Pindos basin and is regarded as a 'calm' interval between the Eohellenic and Meso-Hellenic events (Robertson *et al.* 1991). The Tinos ophiolite thus either originated in an area outside the Hellenic realm (Avigad & Garfunkel 1989) or indicates remobilization of previously emplaced Eohellenic ophiolites. The Tinos ophiolite slices may be related to a rock assemblage of high-temperature metamorphic rocks, greenschists and granites of Late Cretaceous to Paleocene age that occurs at the eastern edge of the Cyclades (Fig. 12; Dürr *et al.* 1978*b*; Reinecke *et al.* 1982; Altherr *et al.* 1994). Like the Tinos ophiolite they overlie the CBU and possibly record the former existence of a large metamorphic terrane in the central Aegean.

Summing up, the ultramafic rocks and their host sequences in the Cyclades show bimodal provenance. Several ultramafic associations share characteristics with rock sequences in the Hellenides, whereas others have clear non-Hellenic origin. These include ophiolites formed and deformed at *c.* 80–65 Ma, a time of sea-floor spreading, but overall plate convergence, in the Anatolian domain to the east of the Aegean Sea. The mixing of elements from domains of entirely different history located west and east of the Aegean Sea supports the idea that the Cycladic Massif represents a major discontinuity along which these domains were juxtaposed in the Tertiary (Ring *et al.* 1999; Garfunkel 2004).

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