Contrasting metamorphic evolution of metasedimentary rocks from the Çine and Selimiye nappes in the Anatolide belt, western Turkey

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ABSTRACT

P–T conditions, mineral isograds, the relation of the latter to foliation planes and kinematic indicators are used to elucidate the tectonic nature and evolution of a shear zone in an orogen exhumed from mid-crustal depths in western Turkey. Furthermore, we discuss whether simple monometamorphic fabrics of rock units from different nappes result from one single orogeny or are related to different orogenies. Metasedimentary rocks from the Çine and Selimiye nappes at the southern rim of the Anatolide belt of western Turkey record different metamorphic evolutions. The Eocene Selimiye shear zone separates both nappes. Metasedimentary rocks from the Çine nappe underneath the Selimiye shear zone record maximum P–T conditions of about 7 kbar and > 550 °C. Metasedimentary rocks from the overlying Selimiye nappe have maximum P–T conditions of 4 kbar and c. 525 °C near the base of the nappe. Kinematic indicators in both nappes are related to movement on the Selimiye shear zone and consistently show a top-S shear sense. Metamorphic grade in the Selimiye nappe decreases structurally upwards as indicated by mineral isograds defining the garnet-chlorite zone at the base, the chloritoid-biotite zone and the biotite-chlorite zone at the top of the nappe. The mineral isograds in the Selimiye nappe run parallel to the regional S_R foliation, parallel the Selimiye shear zone and indicate that the Selimiye shear zone formed during this prograde greenschist to lower amphibolite facies metamorphic event but remained active after the peak of metamorphism. ⁴⁰Ar/³⁹Ar mica ages and the tectonometamorphic relationship with the Eocene Cyclades–Menderes thrust, which occurs above the Selimiye nappe in the study area, suggests an Eocene age of metamorphism in the Selimiye nappe.

Metasedimentary rocks of the Çine nappe 20–30 km north of the Selimiye shear zone record maximum P–T conditions of 8–11 kbar and 600–650 °C. An age of about 550 Ma is indicated for amphibolite facies metamorphism and associated top-N shear in the orthogneiss of the Çine nappe. Our study shows that simple monophase tectonometamorphic fabrics do not always indicate a simple orogenic development of a nappe stack. Preservation in some areas and complete overprinting of those fabrics in other areas apparently occur very heterogeneously.

Key words: Anatolide belt; geothermobarometry; metamorphism; P–T grid; shear zone; Turkey.

INTRODUCTION

An interesting problem in polyorogenic settings is that some rocks preserve tectonometamorphic fabrics of an older orogeny extremely well, and that these fabrics are not altered during a severe younger orogenic overprint. A stunning example is recorded in metasedimentary rocks from the Sesia zone in the Italian Alps, which experienced a regional Alpine high-pressure (HP) metamorphism of 500–600 °C and >13 kbar (Pognante, 1989) and associated deformation in the Alpine subduction zone. At Monte Mucrone, metapelite of the Sesia zone shows a Variscan amphibolite facies tectonometamorphic fabric with a pronounced mineral elongation lineation expressed by sillimanite and no signs of Alpine HP metamorphism (Compagnoni, 1977). This metapelite is locally crosscut by cm-wide veins, which contain cm-seized omphacite crystals. In other parts of the Sesia zone, Variscan sillimanite was statically transformed to kyanite (Compagnoni, 1977). Both omphacite and kyanite grew during Alpine HP metamorphism. This Alpine HP metamorphism and associated heterogeneous deformation in the Alpine subduction zone left parts of the Sesia zone mineralogically and/or structurally completely unaffected.

The phenomenon that closely juxtaposed rocks record completely different tectonometamorphic histories, is an important issue for researchers studying orogenic processes. The tectonic significance of faults and shear zones, deduced from the different P–T paths of the rocks above and below the tectonic discontinuity, can be overestimated, when it is automatically assumed that the metamorphic history of the rocks resulted from the same orogenic event.

The Anatolide belt of western Turkey (also referred to as the Menderes Massif), which is part of the Alpine
Hellenide-Anatolide orogen in the eastern Mediterranean, also shows evidence for a polyorogenic history including metamorphism and deformation at the Neoproterozoic/Cambrian boundary (Candan et al., 2001; Gessner et al., 2001a, 2003; Ring et al., 2001). During the Alpine orogeny, the Anatolide belt was assembled; stacking of units with different tectonometamorphic evolutions occurred in the Late Cretaceous and Tertiary (Sengör et al., 1984; Hetzel et al., 1998; Ring et al., 1999a; Gessner et al., 2001a). Nappe tectonism in the western Anatolide belt occurred initially under HP conditions in the upper tectonomorphic units. Subsequently, the HP units were thrust onto the non-high-pressure Menderes nappes (including the Selimiye and Çine nappes) along the Cyclades-Menderes thrust in the Eocene (Gessner et al., 2001b).

The metasedimentary rocks at the southern margin of the Anatolide belt, the Selimiye nappe of Ring et al. (1999a), show a simple monometamorphic greenschist to lower amphibolite facies development during the Tertiary orogeny (Ashworth & Evirgen, 1984). 40Ar/39Ar dating of mica from directly below the Selimiye shear zone (Hetzel & Reischmann, 1996) corroborates an Eocene age for this metamorphism (Gessner et al., 2001b).

In contrast to the Tertiary tectonomorphic development in the Selimiye nappe, the underlying Çine nappe shows widespread evidence for late Neoproterozoic orogenic activity at c. 550 Ma (Loos & Reischmann, 1999; Gessner et al., 2001a; Ring et al., 2001). Gessner et al. (2001a, 2003) showed that undeformed granite, yielding a 207Pb/206Pb single-zircon evaporation age of 547 ± 1 Ma and a SHRIMP zircon age of 566 ± 6 Ma, cuts a penetrative regional-scale amphibolite facies foliation and stretching lineation with associated top-N kinematic indicators in orthogneiss. These data indicate that the main tectonomorphic development, at least in Çine nappe orthogneiss, is of late Neoproterozoic age. This age for penetrative deformation in the Çine nappe is corroborated by 207Pb/206Pb garnet ages from garnet-bearing orthogneiss, which yielded an age of 518 ± 10 Ma for a second garnet generation (Ring et al., 2003a). A fundamental question is how the metasedimentary rocks of the Çine nappe above the orthogneiss and directly below the Selimiye shear zone, which record late Neoproterozoic to Cambrian metamorphism in other parts of the western Anatolide belt (Ring et al., 2001), responded to Eocene nappe stacking in the western Anatolide belt. It seems possible that the anhydrous orthogneiss preserved an old tectonomorphic fabric (the S_PA foliation of Gessner et al., 2001b; the suffix ‘PA’ indicates a pre-Alpine age), whereas the more hydrous metasedimentary rocks may have been at least locally re-equilibrated during the Alpine orogeny.

The Selimiye shear zone separates the Selimiye nappe from the Çine nappe. The tectonic significance of the Selimiye shear zone is debated. Bozkurt & Park (1994) regarded it as a metamorphic-core-complex-type normal fault of Oligocene age. 40Ar/39Ar white mica dating by Hetzel & Reischmann (1996) showed that the Selimiye shear zone is of Eocene age and that orthogneiss of the Çine nappe in its footwall cooled slowly after shearing. Fission-track dating indicates accelerated cooling in the Early Miocene at 23–20 Ma (Gessner et al., 2001c; Ring et al., 2003b). This cooling has been attributed to extensional reactivation of the basal thrust of the Lycian nappes, which occurs tectonically above the Selimiye shear zone (Fig. 1). Ring et al. (1999a) and Gessner et al. (2001b) proposed that the Selimiye shear zone is a thrust, which operated during Eocene nappe stacking, the DA3 event of Gessner et al. (2001b; the suffix ‘A’ indicates an Alpine age). The DA3 event assembled the western Anatolide belt and juxtaposed tectonic units with different tectonomorphic evolutions.

Recently, Whitney & Bozkurt (2002) carried out a tectonomorphic pilot study at the orthogneiss/schist contact at the southern rim of the Anatolide belt. They proposed that the orogenic development resulted from a single Tertiary event, which caused syn-metamorphic top-N thrusting followed by top-S normal faulting during retrograde metamorphism. They based their inference of Alpine top-N shear on top-N kinematic indicators observed in the schist sequence. The assumption of an early Tertiary event causing northward-directed nappe translation has also been invoked by Bozkurt & Park (1994) and Hetzel et al. (1998). In contrast, large-scale tectonic models of the Aegean/western Turkish area argue against Tertiary top-N thrusting (Sengör et al., 1984; Collins & Robertson, 1997, 1998, 2003). Likewise, kinematic studies in the adjacent Aegean also do not support early Tertiary top-N shearing (Ridley, 1984; Ring et al., 1999b).

We have studied metapelite from the Selimiye nappe and also metapelite from the Çine nappe during three field seasons in order to determine P–T conditions and to map mineral isograds. Another major aim is to establish the relationship between isograds, the regional foliation and the Eocene Selimiye shear zone to shed more light on the disputed tectonic nature of the Selimiye shear zone. Special emphasis is put on careful examination of shear-sense indicators in the field and in thin section. We focus on two main areas (Fig. 1): a schist unit south-east of Lake Bafa (southern study area) in which the Selimiye shear zone developed, and a schist unit at the northern rim of the Çine submassif south-west of Aydın (northern study area).

**SETTING**

The architecture of the Anatolide belt of western Turkey comprises three major tectonomorphic units (Fig. 1). The Lycian nappes and the İzmir-Ankara suture zone represent the upper tectonomorphic unit. The middle tectonomorphic unit consists of
the Cycladic blueschist unit. These two units were affected only by a single orogeny undergoing a Late Cretaceous to Eocene HP metamorphism during the closure of Neo-Tethys and were subsequently thrust onto the underlying Menderes nappes (Oberhansli et al., 1998, 2001; Sherlock et al., 1999; Ring & Layer, 2003). The lowermost tectonometamorphic unit of the Anatolide belt, the Menderes nappes, comprises from top to bottom: (1) The Selimiye nappe, (2) the Çine nappe, (3) the Bozdağ nappe and (4) the Bayındır nappe (Ring et al., 1999a; Gessner et al., 2001b). The Çine and Bozdağ nappes have a polyorogenic history (Candan et al., 2001; Gessner et al., 2001a). The Menderes nappes are separated from the overlying HP units by the Eocene out-of-sequence Cyclades-Menderes thrust. Above the Menderes nappes, structures related to this huge thrust developed during lower greenschist facies metamorphism (Gessner et al., 2001b).

The Selimiye nappe contains metapelite, calcschists, metamarl, marble and quartzite. Fossil evidence indicates Devonian and Carboniferous protolith ages for some of the metasedimentary rocks (Schuiling, 1962; Çağlayan et al., 1980). Ring et al. (1999a) regarded the entire metasediment sequence above the orthogneiss of the Çine nappe as belonging to the lowermost parts of the Selimiye nappe. However, detailed re-mapping of the Selimiye shear zone in the course of this study indicated that some of the metasedimentary rocks directly south of the orthogneiss occur below the Selimiye shear zone (Fig. 2) and must therefore be part of the Çine nappe. Furthermore, the metasedimentary rocks directly south of the orthogneiss and east of Lake Bafa were intruded by the protolith of a metagranite, which yielded a SHRIMP \(^{206}\)Pb/\(^{238}\)U age of 541 ± 14 Ma (Gessner et al., 2003), indicating a Precambrian
protolith age for the surrounding metapelite (Fig. 3). Because these metasedimentary rocks occur below the Selimiye shear zone and because no Precambrian sedimentary rocks have so far been reported from the Selimiye nappe, we regard the lowermost part of the metasedimentary sequence east of Lake Bafa to be part of the Çine nappe as indicated in Figs 1 and 2 and discussed further below.

Most of the Çine nappe consists of deformed orthogneiss, weakly to undeformed metagranite. Metasedimentary rocks, which in part show migmatitic fabrics, eclogite and amphibolite also occur (Oberhansli et al., 1997; Candan et al., 2001). Protoliths to all dated orthogneisses were intruded at c. 560–540 Ma (Hetzel & Reischmann, 1996; Dannat, 1997; Hetzel et al., 1998; Loos & Reischmann, 1999; Gessner et al., 2001). Some metamorphites have protolith ages of c. 540–530 Ma (Dannat, 1997; Loos & Reischmann, 1999). 207Pb/206Pb single-zircon evaporation dating of migmatites from the Ödemis and Gördes submassifs generally yielded ages of c. 550–540 Ma for magmatization (Dannat & Reischmann, 1999).

Rings et al. (2001) estimated peak-metamorphic conditions from the lowermost Çine nappe of 670–730 °C and c. 6.0–6.5 kbar. Prograde garnet (grt I) growth occurred largely before and during the formation of the regional foliation. Formation of a second garnet generation, which discordantly overgrew grt I, followed at 550–620 °C and c. 6.0–6.5 kbar.

The Bozdag nappe that structurally underlies the Çine nappe is made up of metapelite with intercalated metapsammites, marble, amphibolite and eclogite lenses (Candan et al., 2001). Protolith ages of all rock types are unknown, but a Precambrian age for at least parts of these rocks have been proposed (Candan et al., 2001; Gessner et al., 2001a). Peak-metamorphic conditions vary from 480 to 540 °C and 6.1–7.6 kbar at the base to 610–660 °C and 8.5–10.8 kbar at the top of the nappe and attest to an inverted metamorphic field gradient in the Bozdag nappe (Ring et al., 2001). Differential thermodynamic modelling by the Gibbs method yielded a prograde path for garnet growth during metamorphism (Ring et al., 2001). A 207Pb/206Pb garnet age from metapelite provided an age of 512 ± 56 Ma (Ring et al., 2003a). Shear-sense indicators, especially rotated garnet and associated asymmetric strain shadows around garnet, yielded a consistent top-N sense of shear during prograde amphibolite facies metamorphism. In summary, both the Çine and Bozdag nappes are characterized by systematically top-N ductile shear structures that developed during late Neoproterozoic to Cambrian amphibolite facies metamorphism.

The Bayındır nappe contains phyllite, quartzite, marble and greenschist of inferred Permo-Carboniferous and Mesozoic age (O. Candan, pers. comm. 1999). The rocks were affected by a single greenschist facies metamorphism at 37 Ma (Lips et al., 2001). Syn-metamorphic kinematic indicators yielded a consistent top-N sense of shear (Gessner et al., 2001b).

STRUCTURE

Southern study area

The Eocene Selimiye shear zone consists of strongly to mylonitically deformed metasedimentary rocks, and separates schist of the Selimiye nappe above from schist of the Çine nappe below (Fig. 2). These two schist units strongly resemble each other in the field but record different P–T conditions, which will be discussed in detail below, and are separated by a major shear zone. These are the main reasons why we distinguish both schist units in Fig. 2.

Regional-foliation trajectories vary significantly across the Selimiye shear zone (Fig. 2a) (De Graciansky, 1966). The regional foliation in orthogneiss of the Çine nappe is N–S oriented and folded about large-scale N-trending synforms and antiforms. In metasedimentary rocks of the Çine nappe directly below the Selimiye shear zone, the regional foliation swings into a WNW–ESE orientation parallel to the shear zone (Fig. 2a). In the Selimiye nappe, the regional foliation also strikes WNW–ESE.

Multiple foliations occur in schist of the Selimiye nappe. The penetrative regional foliation is termed Sr (Sr most probably correlates with SA3 of Gessner et al., 2001b) and is related to the Dr deformation. An older foliation is only locally preserved in microlithons between the penetrative Sr foliation (Fig. 4a). The major metamorphic minerals (garnet, chloritoid, amphibole, epidote and mica) grew in Sr. However, where the older foliation is preserved, the same minerals that form Sr also grew parallel to this older foliation. A ‘post-Sr’ foliation crosscuts Sr at a low angle. Thin-section work reveals that these low-angle planes are C shear bands associated with Sr. In addition, Sr is locally deformed by a weak crenulation cleavage, which is associated with late folding (see below).

On Sr a prominent mineral stretching lineation, Lr, formed. Lr is expressed by stretched quartz aggregates, strain shadows around garnet and by the alignment of mica. Lr plunges SSW in the lower parts of the Selimiye nappe and swings in the vicinity of the Cyclades-Menderes thrust into an ESE-WNW trend (Fig. 2b). Shear-sense indicators associated with Sr and Lr are the above mentioned shear bands, asymmetric veins, S-C fabrics as well as rotated garnet, epidote and plagioclase and asymmetric strain shadows around these minerals (Fig. 4a–c). In a number of outcrops shear-sense indicators are imperfectly developed and do not allow an unambiguous determination of the shear sense. Where the kinematic indicators are well developed, they yielded a top-S shear sense (Figs 2c, 4a, 4b & 5a), as has been previously reported by Hetzel & Reischmann (1996) and Gessner et al. (2001a). However, occasionally the same shear-sense indicators show a top-N sense of shear (see below). Kinematic indicators in the upper Selimiye nappe and the overlying Cycladic blueschist unit yielded a top-SE/ESE shear sense. In the Cycladic blueschist unit, the top-SE/ESE kinematic indicators developed during retrograde blueschist facies metamorphism. This is best expressed by pronounced chloritization of various minerals. Chloritization during retrogression also characterizes the kinematic indicators in discrete shear zones in the upper Selimiye nappe (phylolite zones in Fig. 2).

A number of larger scale tight to isoclinal folds with WNW-trending axes occur in the upper Selimiye nappe below the Cyclades-Menderes thrust. These folds fold the Sr foliation and because of their tight to isoclinal character their axial planes are subparallel to the Sr foliation. The relationships between the long and short limbs of the asymmetric folds are consistent with top-S shearing.

In addition to the open, tight to isoclinal folds, the Sr foliation in the Selimiye shear zone is folded about
Fig. 2. Southern study area: (a) Map of main foliation, S_R, after De Graciansky (1966), Bozkurt & Park (1994) and this study; note that foliation swings from a northerly strike in Çine nappe orthogneiss into a west-north-westerly strike in the Selimiye nappe. (b) Stretching lineation map; lineations swing from a NNE trend into a SE/ESE trend near the Cyclades-Menderes thrust. (c) Shear senses associated with S_R as deduced from rotated garnet, S-C fabrics, shear bands and asymmetric quartz veins; arrows indicate movement direction of tectonic top. (d) Isograd map; isograds parallel S_R foliation in Selimiye nappe; AFM and AKM projections for metapelite of the Selimiye nappe and calc-schist of Çine nappe (amphibolite facies); see text for abbreviations.
tight to isoclinal folds at the centimetre to metre scale. A weak crenulation cleavage is associated with the folds. The axes of the folds are parallel to the $L_R$ stretching lineations (Fig. 5b). These post-$D_R$ folds formed after the peak of the metamorphism and folded the $D_R$ shear-sense indicators. Detailed field work revealed that the top-$S$ shear-sense indicators occur in the upright limbs of the post-$D_R$ tight to isoclinal folds, whereas the top-$N$ fabrics occur in the inverted limbs of the folds (Fig. 5b). Therefore, the top-$N$ shear-sense indicators are folded structures, which originally had a top-$S$ shear sense.

The metasedimentary rocks of the Çine nappe directly below the Selimiye shear zone display similar structures as the schist of the Selimiye nappe; a penetrative foliation and a SSW-plunging stretching lineation associated with top-$S$ kinematic indicators (Fig. 2c). The shear-sense indicators are rotated garnet (Fig. 4c), asymmetric strain shadows around garnet and shear bands. Intrusive late Neoproterozoic orthogneiss at the north-eastern end of Lake Bafa displays, together with the surrounding metapelite, two sets of isoclinal folds. The second generation of these folds have the $S_R$ foliation as an axial-plane cleavage. In structurally deeper levels of the orthogneiss, widespread top-$N$ shear-sense indicators occur (Fig. 5a). The top-$N$ kinematic indicators are asymmetric feldspar porphyroclasts in which K-feldspar was dynamically recrystallized.

Above, we have shown that metamorphism in the Selimiye nappe and the formation of the Selimiye shear zone is of Tertiary age. Therefore, the syn- to post-metamorphic WNW-striking foliation in the Selimiye nappe also has to be of Tertiary age. The Selimiye shear zone formed during top-$S$ shearing associated with the $S_R$ foliation.

Northern study area

In the metasedimentary rocks in the northern part of the Çine nappe south-west of Aydin (Fig. 1), two foliations occur. The main foliation, which we term $S_M$, in the surrounding orthogneiss is parallel to the main foliation in the metasedimentary rocks ($S_M$ most probably correlates with $S_{PA}$ of Gessner et al., 2001b). On $S_M$ a pronounced stretching lineation, $L_M$, occurs (Fig. 6a,b). In orthogneiss, $L_M$ is marked by strongly elongated quartz-feldspar aggregates, whereas in the schist $L_M$ is characterized by an alignment of biotite and stretched quartz aggregates. In the metasedimentary rocks, numerous symmetric foliation-boudinage structures occur, which attest to coaxial deformation (Fig. 6c). Nonetheless, rotated garnet and shear bands are common and indicate top-$N$ shear associated with $S_M$ and $L_M$ in large parts of the metasedimentary rocks of the northern study area (Figs 6c & 7a). In the surrounding orthogneiss, top-$N$ kinematic indicators are common and expressed by asymmetric recrystallized feldspar tails around feldspar porphyroclasts. This top-$N$ shear sense is in marked contrast to the top-$S$ shear sense in the metasedimentary rocks in the southern area (Figs 5–8).

SCOPE OF METAMORPHIC STUDY AND ANALYTICAL PROCEDURES

Metasedimentary rocks in the southern study area were formerly thought to belong entirely to the Selimiye nappe (Ring et al., 1999a). However, as discussed above, the Selimiye shear zone developed within this metasedimentary section and therefore the upper part of this section belongs to the Selimiye nappe, whereas the underlying metasedimentary rocks belong to the Çine nappe. If so, differences in the metamorphic evolution of both metasedimentary units are...
likely. Metasedimentary rocks of the Çine nappe in the northern and southern study area preserve different kinematic indicators (top-S in the south and top-N in the north) and the question arises whether or not the different structures are associated with different metamorphic events.

More than 350 sample were collected in the two study areas in order to map mineral isograds and their relationship to the regional foliation and the Eocene Selimiye shear zone. The extensive sampling also allows us to compare in detail metamorphic development of the different schist units. Location of samples (Fig. S1) and parageneses (Table S1) are available online.

The mineral analyses were obtained with a Jeol Superprobe (JXA 8900RL) at Johannes Gutenberg-Universität Mainz, Germany. Operating conditions were an acceleration voltage of 15 kV, a beam current of 15 nA and 20 s counting time per element. Standards are: wollastonite for Si, corundum for Al, pyrophane for Ti, hematite for Fe, MgO for Mg, wollastonite for Ca, albite for Na, orthoclase for K, Cr₂O₃ for Cr, rhodochrosite for Mn. A ZAF procedure was used for matrix correction. The mineral analyses are considered to be accurate within a range of c. 3% (relative) on any given grain.

Abbreviations used in text, figures and tables are adopted from Powell et al. (1998): als = aluminosilicate, amph = amphibole, bi = biotite, carb = carbonate, cd = cordierite, chl = chlorite, ctd = chloritoid, ep = epidote, fsp = plagioclase, g = garnet, ky = kyanite, mu = muscovite, q = quartz, st = staurolite, tr = tremolite, dol = dolomite, cc = calcite, cz = clinozoisite, an = anorthite, gr =

Fig. 4. Microphotographs from the southern study area: (a) Sample SE3, Selimiye nappe; penetrative regional S_R foliation separating microlithons in which pre-S_R foliation is locally preserved; chloritoid grew in S_R and pre-S_R foliation. Note that most foliation planes in microlithons appear to be related to S_R and form S-C fabrics indicating a top-S shear sense. (b) Sample SE12, garnet-chlorite zone, Selimiye nappe; rotated garnet indicating a top-S sense of shear. (c) Sample OS, Çine nappe; rotated garnet showing a top-S shear sense. (d) Sample N7, Çine nappe; garnet-epidote-amphibole-biotite-muscovite-plagioclase (+ calcite + quartz + fluids) assemblage in calcscist.
grossular, py = pyrope, alm = almandine, andr = andradite, spss = spessartine, phil = phlogopite, clin = clinochlore.

Fig. 5. Cross section DC (for location see Fig. 2d); top-S-displacing Selimiye shear zone (SSZ) separates metasedimentary rocks of the Çine nappe from those of the Selimiye nappe; insert shows that SSZ and SR are folded about N-trending isoclinal folds, which inverted top-S kinematic indicators in overturned limbs (a-type folds; Malavieille, 1987). SPA foliation after Gessner et al. (2001b); the suffix ‘PA’ indicates a pre-Alpine age.

Fig. 6. Northern study area: (a) LMA stretching lineations in amphibolite facies metasedimentary rocks. (b) SMA foliation. (c) Cross section EF; relationship between amphibolite facies DMA shear-sense indicators and overprinting shear bands is shown schematically in the cross section; numerous symmetric boudinage structures suggest areas of local coaxial deformation. AFM projections from analyses (Table 1) and results from thermobarometry are also shown.

$P$–$T$ grids have been calculated with the THERMOCALC software (version 3.1) (Powell et al., 1998). For geothermobarometry, multivariate reactions were
calculated using mode ‘average $P\text{T}$’ of the THERMO-CALC software (Powell & Holland, 1988, 1994; Worley & Powell, 2000). Activities of the end-members were calculated from microprobe analyses (Tables 1 & 2) using the AX software (Holland, 2000). The calculated $P\text{T}$ data are listed in Table 3.

Fig. 7. Northern study area: (a) Sample D13; rotated garnet indicating top-N shear sense. (b) Sample D6; staurolite biotite-kyanite paragenesis. (c) Sample D6; chloritoid and staurolite inclusions in garnet.

Fig. 8. Interpretative cross section AB modified after Lips et al. (2001) (for location of section and patterns see Fig. 1); ages for the Selimiye shear zone are $^{40}\text{Ar}/^{39}\text{Ar}$ white mica ages (Hetzel & Reischmann, 1996), and those for the Güney and Kuzey detachment are fission track ages (Gessner et al., 2001c; Ring et al., 2003b). $P\text{T}$ field gradient for Selimiye and Çine nappes according to this study.

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MINERALOGY AND RELATIONSHIPS BETWEEN DEFORMATION AND MINERAL GROWTH

Southern study area

Selimiye nappe

Garnet has a general grain size of about 5 mm but can be > 1 cm in diameter, and has a composition of almandine (60–70%), pyrope (1–10%), spessartine (1–10%) and grossular (10–30%). Most garnet lacks zoning, while some displays exchange between Ca and Mg from core to rim (Fig. 9a, sample SE12) as previously observed by Whitney & Bozkurt (2002). Mineral formulae do not show significant Fe$^{3+}$ and therefore we assumed that Fe$^{2+} = \text{Fe}^{\text{total}}$.

In all assemblages white mica is a solid solution between paragonite, muscovite and celadonite depending on whole rock composition, degree of metamorphism and probably retrograde reactions. Margarite occurs in calc-schists and occasionally in meta-aptites as inclusions in garnet. Phengitic substitution (Massonne & Schreyer, 1987) increases from 2.9 to 3.17 per formula unit (pfu) towards the orthogneiss, but changes unsystematically from one sample to another. Secondary white mica occasionally occurs and has a lower Si content than primary white mica.

Primary chlorite occurs in all assemblages (Table S1) but secondary chlorite is ubiquitous. Garnet is often totally retrograded to chloride, which implies an important stage of fluid circulation. Secondary chloride porphyroblasts cut the main $S_R$ foliation and secondary chloride replaced biotite that grew in the $S_R$ foliation.

Biotite is present in all samples, even in the calc-schists, and in general is severely chloritized. Biotite is also found as inclusions in garnet.

Chloritoid is in equilibrium with biotite and chloride in the upper parts of the nappe (samples SE2, SE3, C7, G12, G14, G16, E3, E4, E5, O28, Fig. 4a). At the base of the Selimiye nappe, chloritoid is stable with garnet, chlorite and biotite (Fig. 4b) (Table S1). Inclusions of chloritoid in garnet suggest that chloritoid grew before garnet during prograde metamorphism. On the basis of six-oxygen structural formula, the mole fraction of Mg in chloritoid increases from the SSW to the NNE (Mg content varies from 0.132 to 0.213 pfu), and is related to an increase in temperature (Table 1, samples SE3 and SE12).

Plagioclase appears as albite in the pelitic schists but as oligoclase (An$_{30}$) in the calc-schists (Table 2, sample C9P and SE14). In the calc-schists, we focused our study on impure layers in which carbonate minerals mainly consist of solutions between dolomite, calcite and subordinate siderite or magnesite. Epidote minerals are zoisite and clinozoisite. Carbonate minerals and epidote are commonly present as inclusions in garnet. Accessory minerals in the Selimiye nappe are opaques (rutile, ilmenite, hematite, graphite), titanite, apatite, zircon and tourmaline.

Deformation/metamorphism relationships indicate that top-$S$ shearing during $D_R$ occurred during prograde to peak metamorphism. Rotated garnet consistently indicates top-$S$ shear. Asymmetric strain shadows around garnet contain primary chlorite, primary white mica and biotite. These sheet silicates also grew in $D_R$ shear bands.

Secondary white mica and biotite cut across the $S_R$ foliation in some places. Secondary chloride porphyroblasts also cut across $S_R$, but also occasionally replace biotite that grew in the $S_R$ foliation suggesting that fluids circulated along $S_R$ planes during this retrograde phase of metamorphism. Retrograde chloride-forming reactions in $S_R$ are in part associated with top-$S$ shearing. Phyllonite zones with top-$S$ kinematic indicators developed in metapelite of the chlorite-biotite zone and are characterized by very fine-grained layers of quartz and plagioclase interlayered with biotite and chlorite. Small porphyroblasts of chlorite occasionally appear in the matrix.

In a quartzite conglomerate with intercalated pelitic layers of the Cycladic blueschists unit directly above the Selimiye nappe south of Selimiye (Fig. 2d), chloritoid, kyanite, chlorite, white mica, quartz and opaques occur (T. Will, written communication 2002). Similar rocks with the same mineral assemblage yielded $P$–$T$ conditions of 15 kbar and 500 °C on nearby Samos Island in the Aegean Sea (Will et al., 1998). On Samos Island and on Dilek Peninsula, the minerals of this HP assemblage form a penetrative foliation. In the mylonite zone of the Cyclades–Menderes thrust, this HP assemblage is completely retrograded and zones enriched in the secondary chlorite are common.

Metasedimentary rocks of the Çine nappe

Pelitic gneiss directly beneath the Selimiye shear zone shows assemblages of quartz, muscovite, plagioclase, biotite, garnet and accessory minerals (tourmaline, rutile, ilmenite, hematite, apatite, zircon). Tourmaline can be strongly concentrated forming black layers parallel to the foliation. Chloritoid has never been observed in these assemblages.

Garnet has a similar composition as in the Selimiye nappe but appears more homogeneous in both pelitic gneisses and in the calc-schists. Garnet composition is almandine (63%), pyrope (8%), spessartine (2%) and grossular-andradite (25%) (Fig. 9b,c). Primary chlorite only occurs in calc-schists (Table S1) and secondary chlorite is absent. Phengitic substitution in white mica (Massonne & Schreyer, 1987) is up to 3.3 pfu but changes unsystematically from one sample to another. Biotite is not altered and can be used for thermodynamic calculations.

In the pelitic gneisses, plagioclase zoning does not show obvious disequilibrium between core and rim but does show a slight compositional core-to-rim variation from An$_{14}$ to An$_{16}$ (Sample SE25, Table 2) in relationship to the degree of metamorphism. Sample SE18...
shows coexisting albite and oligoclase in equilibrium (peristerite gap, Table 2) (Ashworth & Evirgen, 1985).

Amphibole appears in calcshists as tschermakite or magnesioclnhornblende (Leake et al., 1997) and coexists with andesine, white mica, biotite, garnet, epidote, quartz and carbonate (Fig. 4d). Amphibole occurs along a horizon extending for 7 km along strike (Fig. 2d). On the basis of 23-oxygen structural formula, total Al content varies from 3 (core) to 2.5 pfu (rim; for instance sample A11, Table 1). This variation is mainly correlated with an increase of Si in the T1 tetrahedral site. Mg increases from 1.8 (core) to 2.3 pfu.

Table 2. Plagioclase analyses for different metamorphic zones. C9P: schist, Selimiye nappe. SE14: calcshist, Selimiye nappe. SE25: pelitic gneiss, Çine nappe. SE18: schist, Çine nappe (fsp1 = oligoclase; fsp2 = albite). Mineral formulae have been calculated with ax software (Holland, 2000).

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<th>C9P fsp/core</th>
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<th>SE14 fsp/core</th>
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Garnet zoning SE12 (garnet-chlorite zone, Selimiye nappe)
Garnet zoning A11 (calcshist, Çine nappe)
Garnet zoning C22 (pelitic gneiss, Çine nappe)
Garnet zoning D23 (schist, staurolite-biotite-kyanite or chlorite-staurolite zone, Çine nappe)

Fig. 9. Garnet zoning patterns from different metamorphic zones in Selimiye (a) and Çine nappe (b–d); zoning is depicted by profiles across garnet (diagrams at the top), pyrope-grossular/andradite-almandine/spessartine distribution (diagrams in the centre), and X_Mg vs. X_Ca (diagrams at the bottom); cations are given on the basis of 12-oxygen structural formula. All garnet is close to almandine end member; some show a slight exchange between Ca and Mg indicating prograde metamorphism.
(rim), whereas Ca and Fe$^{2+}$ decrease in the octahedral sites. Epidote minerals are zoisite and clinozoisite. In the calcschists plagioclase shows zoning from oligoclase (core) to andesine (rim, An$_{17}$, Sample A11, Table 2). Accessory minerals are opaques (rutile, ilmenite, hematite, graphite), titanite, apatite, zircon and tourmaline.

Rotated garnet crystals show a top-S sense of shear associated with $S_R$ (Fig. 2c). Asymmetric top-S strain shadows around garnet and plagioclase with oligoclase rims contain white mica with high Si values and biotite. In shear bands in $S_R$ white mica with the highest Si contents and biotite grew. In calcschists, primary chlorite also occurs in these shear bands. These relationships indicate that top-S shearing occurred during prograde to peak metamorphism.

Northern study area

The most notable difference between metasedimentary rocks of the Çine nappe in the north and those in the southern study area is the occurrence of kyanite and staurolite in the north. In addition, calcschists do not contain amphibole. The retrograde reactions are again less important than in the Selimiye nappe and only sporadic secondary chlorite is observed. Chloritoid appears only as inclusions in garnet except for one sample (Table S1, sample D16bis), where chloritoid occurs with biotite in the matrix (the Mg content of this chloritoid is around 0.225 pfu). Garnet has the same chemistry as in the southern study area, with a slight exchange between Ca and Mg from core to rim (Fig. 9d, sample D23). Plagioclase is oligoclase or andesine and does not show variations from core to rim (e.g. sample D42, Table 1).

Staurolite is in equilibrium with biotite, garnet and kyanite (sample D6, Fig. 7b) and also with chloritoid as inclusions in garnet (sample D6, Fig. 7c). Therefore, staurolite grew during prograde metamorphism and is also stable with garnet at the peak of metamorphism (equilibrium at $T_{\text{max}}$ with garnet, kyanite and biotite). Epidote and opaques are common in calcschist and appear in metapelitic as inclusions in garnet.

In contrast to the southern study area, rotated garnet, locally with staurolite inclusions, depict a top-N sense of shear (see below) associated with $S_M$. Staurolite in the matrix grew after the main $S_M$ foliation formed. The relationships between garnet rotation during the formation of $S_M$ and later growth of staurolite indicates that $S_M$ formed during prograde amphibolite facies metamorphism and that peak metamorphic conditions were attained after the $S_M$-forming deformation.

Summary

The mineralogical development in both study areas is characterized by a single prograde metamorphism. This is shown in the southern study area by an increase in Mg from core to rim in garnet and amphibole, and an increase in Ca from the core to rim in plagioclase (samples A11 & SE25, Table 2). Inclusion of chloritoid in garnet in the northern study area also indicates a prograde metamorphism. No evidence for polymetamorphism has been found in either study area.

Deformation/metamorphism relationships in the southern Çine nappe are the same as those in the Selimiye nappe and indicate top-S shearing during prograde greenschist to amphibolite facies metamorphism. Along the Cyclades–Menderes thrust top-S shearing took place during lower to middle greenschist facies conditions and phyllonite zones developed. Shearing in the metasedimentary rocks of the northern Çine nappe also developed during prograde metamorphism; however, here the sense of shear is top-N.

In the Selimiye nappe, secondary chlorite, biotite and white mica cut across the regional $S_R$ foliation, and garnet was retrograded to chlorite after the formation of $S_R$. This is interpreted to be the result of retrograde metamorphism associated with pronounced fluid circulation after $D_R$. Evidence has been given that top-S shearing in the Selimiye nappe locally continued during retrogression (see also Whitney & Bozkurt, 2002). This retrogression complicates thermodynamic considerations and necessitates the study of single subsystem models and Schreinemakers’ rules. The coexistence of oligoclase with albite (peristerite gap) described from metasedimentary rocks of the Çine nappe also make $P$–$T$ estimations in the southern study area problematic (Evirgen & Ataman, 1982; Evirgen & Ashworth, 1984; Ashworth & Evirgen, 1984; Ashworth & Evirgen, 1985; Whitney & Bozkurt, 2002). In the northern study area, parageneses and mineral analyses are more suitable for estimating $P$–$T$ conditions of amphibolite facies metamorphism.

PARAGENESES AND THERMOBAROMETRY

Southern study area

Selimiye nappe

Analysis of the metamorphic parageneses (Table S1) allows mapping of three isometamorphic zones in the Selimiye nappe (Fig. 2d): the chlorite–biotite zone in the upper Selimiye nappe, the chloritoid–biotite zone and the garnet–chlorite zone at the base of the nappe. However, two samples containing garnet (G2 & G8) have been found in the chlorite–biotite zone close to the Cyclades–Menderes thrust and will be discussed further below. The boundaries of the different metamorphic zones are parallel to $S_R$ over a distance of about 30 km between Lake Bafa and Milas (Fig. 2d).

The calcschists always display the same quartz–muscovite–plagioclase–biotite–chlorite–epidote–carbonate ($\pm$ garnet) assemblage, which is not suitable to distinguish different metamorphic zones. Therefore, the KFMASH system has been used because it
provides a framework to study both calcschists and metapelites in a comparable fashion. Minerals used are chlorite, biotite, garnet, staurolite, cordierite, chloritoid and aluminosilicates. They are projected into an AFM diagram from quartz, muscovite and H2O. Fluid (H2O) pressure is assumed to be equal to total pressure. Accessory minerals are not taken into account in the reactions observed. Structural formulae and activity models are shown in the Appendix (available online). The resulting P–T grid for the KFMASH system is given in Fig. 10(a).

The occurrence of chloritoid in the KFASH system is related to the continuous reaction

\[ \text{Fe-chlorite} \leftrightarrow \text{annite} + \text{Fe-chloritoid}. \quad (1) \]

This reaction appears around 500 °C allowing for the occurrence of the chlorite–biotite–chloritoid paragenesis in the KFMASH system (Sample SE3, Figs 2d, 4a & 10a). In samples SE2, SE3, C7, G12, G14, G16, E3, E4, E5 and O28, chloritoid is in equilibrium with chlorite and biotite. The occurrence of garnet (e.g. sample SE12, Figs 2d, 4b & 10a) is assumed to be related to the discontinuous reaction in the KFMASH system:

\[ \text{biotite} + \text{chloritoid} \leftrightarrow \text{garnet} + \text{chlorite}. \quad (2) \]

Numerous chloritoid inclusions in garnet corroborate the inference (Table S1) that chloritoid grew before garnet during increasing temperature. The position of reaction (2) in P–T space essentially depends on the invariant [cd, als] point (Fig. 10a), the position of which is subject to considerable uncertainty. (Powell et al., 1998) or in the kyanite field (Spear & Cheney, 1989). The coexistence of minerals involved in reaction (2) with the different aluminosilicates should be deduced in thin sections, but unfortunately no aluminosilicates have been found in the Selimiye nappe. Therefore, metamorphic pressure is poorly constrained. Recently, Likhanov et al. (2001) described reaction (2) in the andalusite field at around 3 kbar. This would place the invariant [cd, als] point close to the Al2SiO5 triple point. If so, the pressure of about 4 kbar and a temperature around 500 °C would be implied for the Selimiye nappe.

According to the phengitic substitution observed in the Selimiye nappe (between 2.90 and 3.17 pfu), pressure can be estimated around 4 kbar (Massonne & Schreyer, 1987). Therefore, it seems probable that the P–T conditions in the lower Selimiye nappe are around 500 °C and 4 kbar as previously suggested by Ashworth & Evirgen (1984) and Whitney & Bozkurt (2002).

The occurrence of the paragenesis chloritoid-biotite before garnet–chloritoid with increasing temperature could be related to Fe-rich and Mn-poor bulk compositions of the rocks (e.g. low Mn contents in garnet in sample SE12, Table 1). It has been shown by a number of studies that the occurrence of garnet in greenschist facies pelites may be a factor of the Mn content of the rock (Spear & Cheney, 1989; Droop & Harte, 1995). However, two samples (G2 & G8) in the chlorite-biotite zone contain garnet (without chloritoid inclusions). These may be explained either by a local Mn-rich bulk composition allowing the stability of garnet, or by thrust imbrication of slices from the overlying Cycladic blueschist unit.

According to the P–T grid (Fig. 10a) the continuous reaction (1) appears at the high-T side of the discontinuous reaction:

\[ \text{chloritoid} + \text{andalusite} \leftrightarrow \text{chlorite} + \text{staurolite}, \quad (3) \]

despite this, staurolite is not observed in these rocks. Whitney & Bozkurt (2002) argued that the absence of staurolite is a temperature indicator rather than a bulk-compositional effect. However, staurolite is stable in KFMASH at considerably lower temperatures than the 560 °C suggested by these authors (Fig. 10a). We therefore suggest that the absence of staurolite could be also related to a compositional effect rather than a consequence of lower temperature.

Finally, chloritoid inclusions in the garnet-chlorite zone (for instance samples SE6, SE18, SE11, A20, A18, B22) could be indicative of a clockwise P–T path: chlorite–biotite followed first by chloritoid–biotite, then garnet–chloritoid at higher pressure, and finally by garnet–chloritoid–chlorite–biotite at T_max (Fig. 10a).

Metasedimentary rocks of Çine nappe

Thermodynamic calculations using independent sets of reactions have been carried out for samples SE18 and SE25. Sample SE25 is pelitic gneiss above the orthogneiss and sample SE18 is a metapelite directly below the Selimiye shear zone. Analyses and results are reported in Tables 1 and 3. Temperatures of 590 ± 54 °C (sample SE18) are 675 ± 62 °C (sample SE25) have been obtained for a pressure of c. 10 kbar. However, the pressure is probably overestimated, because mineral analyses from sample SE18 show the coexistence of oligoclase with albite (peristerite gap, Table 2) and X_ca in plagioclase from sample SE25 is low (Table 1).

In the vicinity of the orthogneiss, tschermakitic-magnesiohornblende amphibole occurs in calcschist (samples A11, A13, N6, N7) and coexists with andesine, white mica, biotite, garnet, epidote, quartz and carbonate. The plane marking the initial growth of amphibole in the calcschists parallels Sr over a distance of about 7 km (Fig. 2d). We used this amphibole horizon to constrain P–T conditions (sample A11,
Temperature estimates using the Ca-amphibole-plagioclase thermometer (Blundy & Holland, 1990; Holland & Blundy, 1994) are between 500 and 640 °C for a pressure range of 6–9 kbar. Pressure estimates using the calibration of Johnson & Rutherford (1989) related to the Al\text{total} content of amphibole (3–2.5 pfu), range from c. 9 kbar for amphibole cores to c. 6.3 kbar for the rims.

**Table 1.**

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**Fig. 10** $P$–$T$ grids calculated with THERMO-CALC v3.1 (Powell & Holland, 2001), KFMASH system. $\text{H}_2\text{O} = \text{fluid (mu, q, } \text{H}_2\text{O in excess)}$. [bi, cd]: $P = 13.93 \pm 1.2$ kbar, $T = 607.8 \pm 10$ °C. [cd, ctd]: $P = 11.2 \pm 1.4$ kbar, $T = 639.6 \pm 14$ °C. [cd, g]: $P = 1.25 \pm 1.6$ kbar, $T = 490 \pm 26$ °C. [cd, als]: $P = 4.22 \pm 4.2$ kbar, $T = 542.3 \pm 46$ °C. Aluminosilicate invariant point: $P = 3.8 \pm 0.2$ kbar, $T = 505 \pm 16$ °C. (a) $P$–$T$ grid for rocks of Selimiye nappe. (b) $P$–$T$ grid for rocks of Çine nappe (northern and southern study areas). Reactions discussed in the text have been labelled.
Because CO₂ fluids play an important role in the calcschists, we attempted to determine P–T conditions related to the occurrence of amphibole, using THERMOCALC software in the K₂O–CaO–MgO–Al₂O₃–SiO₂–CO₂–H₂O subsystem (KCMAS-CH). The end-member minerals used are grossular, phlogopite, dolomite, tremolite, clinohlore, anorthite, clinozoisite and muscovite. These minerals are projected into the Al₂O₃-K₂O-MgO diagram from quartz, calcite, and fluids (H₂O-CO₂). Fluid pressure (H₂O, CO₂) is assumed to be equal to total pressure.

Before considering P–T conditions related to amphibole crystallization, the mole fraction of CO₂ in the rocks needs to be estimated. Xₐₙ was determined using a P–Xₐₙ pseudosection. Activities of the end members were calculated from the analyses using the AX software (Holland, 2000) (Table 1, sample A11). A minimum temperature of about 500 °C can be estimated via the Ca-amphibole-plagioclase thermometer used herein. Figure 11(a) displays the P–Xₐₙ pseudosection grid for a pressure range of 2–10 kbar. The inset in Fig. 11(a) shows the occurrence of all phases except clinohlore around the invariant points [clin, mu, phl, tr, dol] and [clin, cz, tr] at Xₐₙ = 0.1 and around 7 kbar. This is consistent with the complex parageneses observed in thin section (Table S1). At 600 °C only the invariant point [clin, cz, tr] is stable and moves to higher Xₐₙ, while the reaction tremolite → dolomite is shifted to higher pressure where clinozoisite becomes unstable. We believe that the P–Xₐₙ pseudosection below 500 °C provides a good estimate for the mole fraction of CO₂ (Xₐₙ = 0.1), which appears reasonable for the stabilization of garnet (grossular end member). Given this CO₂ mole fraction and the activities of the end members from the analyses, a P–T grid has been drawn (Fig. 11b). The stability of clinozoisite–garnet ( grossular end member) at 540 °C resulting from the reaction in the KCMAS-CH subsystem:

\[
dolomite \leftrightarrow tremolite \] or
\[
clinohlore \leftrightarrow grossular + tremolite.\]

In addition, thermodynamic calculations using independent sets of reactions have been carried out using the same activities for the end members, yielded 502 ± 32 °C and 7 ± 1.2 kbar (Table 3). An increase in Xₐₙ up to 0.5 yielded a temperature and pressure of c. 550 °C and c. 8 kbar but increased the uncertainties of the estimates. It appears that amphibole in the calcschists near the orthogneiss formed at higher pressure than reaction (2) in the KFMASH subsystem which we estimated to be > 7 kbar. Therefore, thermobarometric calculations from sample SE18 probably yielded too high pressures (peristerite gap) but temperatures in the metasedimentary rocks of the Çine nappe close to the orthogneiss are likely to be higher than 550 °C (e.g. sample SE25). The lack of staurolite in the metasedimentary rocks of the Çine nappe in the
Table 3. P–T calculations. Abbreviations: an = anorthite, ab = albite, phl = phlogopite, ann = annite, east = eastonite, py = pyrope, gr = grossular, alm = almandine, mu = muscovite, pa = paragonite, cel = celadonite, mst = Mg-staurolite, fst = Fe-staurolite, ky = kyanite, cc = calcite, dol = dolomite, cz = clinocicates, tr = tremolite, clin = clinohlorite, q = quartz, sd = standard deviation, corr = correlation, sigl = t2 test. Activities have been calculated with the AX software using following models (Holland, 2000; Holland & Powell, 1998); amphibole: nonideal mixing model for Ca-amphiboles (Holland & Blundy, 1994), plagioclase: model 1 from Holland & Powell (1992), biotite: Al-M1 ordered; site-mixing model; symmetric formalism (Powell & Holland, 1993, 1999), garnet: 2-site mixing; regular solution gammas, muscovite: non ideal interactions (Holland & Powell, 1998), epidote: symmetric formalism, carbonate: disordered calcite structure, site-mixing, chlorite: Al-M4 ordered model (Holland et al., 1998), staurolite: ideal Fe-Mg mixing.

Sample SE18 (Selimiyne nappe, garnet-chlorite zone), T_{max}:

Activities and their uncertainties from analyses (Table 1):

<table>
<thead>
<tr>
<th></th>
<th>an</th>
<th>ab</th>
<th>phl</th>
<th>ann</th>
<th>east</th>
<th>py</th>
<th>gr</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>0.400</td>
<td>0.780</td>
<td>0.0644</td>
<td>0.0530</td>
<td>0.9350</td>
<td>0.00217</td>
<td>0.0160</td>
</tr>
<tr>
<td>sd(a)/a</td>
<td>0.10800</td>
<td>0.05000</td>
<td>0.36935</td>
<td>0.35086</td>
<td>0.39650</td>
<td>0.68205</td>
<td>0.50357</td>
</tr>
</tbody>
</table>

Results:

T = 590°C, sd(T) = 27, P = 9.7 kbar, sd(P) = 1.0, cor = 0.820, sigf = 0.51

Sample SE25 (southern study area, pelitic gneiss), T_{max}:

Activities and their uncertainties from analyses (Table 1):

<table>
<thead>
<tr>
<th></th>
<th>mu</th>
<th>pa</th>
<th>cel</th>
<th>py</th>
<th>gr</th>
<th>alm</th>
<th>an</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>0.630</td>
<td>0.542</td>
<td>0.0400</td>
<td>0.00279</td>
<td>0.00085</td>
<td>0.300</td>
<td>0.280</td>
</tr>
<tr>
<td>sd(a)/a</td>
<td>0.10000</td>
<td>0.10000</td>
<td>0.38400</td>
<td>0.66444</td>
<td>0.57014</td>
<td>0.15000</td>
<td>0.15552</td>
</tr>
</tbody>
</table>

Results:

T = 675°C, sd(T) = 31, P = 11.7 kbar, sd(P) = 1.2, cor = 0.790, sigf = 0.93

Sample A11 (southern study area, calcic schist), T_{max}:

Activities and their uncertainties from analyses (Table 1):

<table>
<thead>
<tr>
<th></th>
<th>clin</th>
<th>mu</th>
<th>aln</th>
<th>gr</th>
<th>cz</th>
<th>phl</th>
<th>tr</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>0.0500</td>
<td>0.700</td>
<td>0.500</td>
<td>0.2000</td>
<td>0.700</td>
<td>0.0700</td>
<td>0.100</td>
</tr>
<tr>
<td>sd(a)/a</td>
<td>0.36167</td>
<td>0.10000</td>
<td>0.07500</td>
<td>0.47772</td>
<td>0.05000</td>
<td>0.32451</td>
<td>0.28053</td>
</tr>
</tbody>
</table>

Results:

T = 502°C, sd(T) = 16, P = 7.0 kbars, sd(P) = 0.6, cor = 0.884, sigf = 0.75

Sample D1 (northern study area, staurolite-biotite-kyanite zone), T_{max}:

Activities and their uncertainties from analyses (Table 1):

<table>
<thead>
<tr>
<th></th>
<th>phl</th>
<th>ann</th>
<th>east</th>
<th>py</th>
<th>gr</th>
<th>alm</th>
<th>mu</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>0.00800</td>
<td>0.00250</td>
<td>0.00700</td>
<td>0.00400</td>
<td>0.00240</td>
<td>0.360</td>
<td>0.670</td>
</tr>
<tr>
<td>sd(a)/a</td>
<td>0.30859</td>
<td>0.45063</td>
<td>0.34770</td>
<td>0.63695</td>
<td>0.67514</td>
<td>0.15000</td>
<td>0.15000</td>
</tr>
</tbody>
</table>

Results:

T = 643°C, sd(T) = 21, P = 9.7 kbar, sd(P) = 1.1, cor = 0.773, sigf = 0.75

Sample D30 (northern study area, staurolite-biotite-kyanite zone), T_{max}:

Activities and their uncertainties from analyses (Table 1):

<table>
<thead>
<tr>
<th></th>
<th>mu</th>
<th>pa</th>
<th>cel</th>
<th>py</th>
<th>gr</th>
<th>alm</th>
<th>an</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>0.710</td>
<td>0.930</td>
<td>0.0170</td>
<td>0.590</td>
<td>0.670</td>
<td>0.06600</td>
<td>0.00440</td>
</tr>
<tr>
<td>sd(a)/a</td>
<td>0.10000</td>
<td>0.10011</td>
<td>0.45374</td>
<td>0.50504</td>
<td>0.05000</td>
<td>0.50403</td>
<td>0.66291</td>
</tr>
</tbody>
</table>

Results:

T = 610°C, sd(T) = 19, P = 8.3 kbar, sd(P) = 0.9, cor = 0.750, sigf = 0.38

Sample D42 (northern study area, staurolite-biotite-kyanite zone), T_{max}:

Activities and their uncertainties from analyses (Table 1):

<table>
<thead>
<tr>
<th></th>
<th>an</th>
<th>ab</th>
<th>py</th>
<th>gr</th>
<th>alm</th>
<th>pa</th>
<th>mst</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>0.640</td>
<td>0.640</td>
<td>0.00050</td>
<td>0.00050</td>
<td>0.320</td>
<td>0.880</td>
<td>0.00040</td>
</tr>
<tr>
<td>sd(a)/a</td>
<td>0.05000</td>
<td>0.05016</td>
<td>0.60264</td>
<td>0.60565</td>
<td>0.15000</td>
<td>0.10000</td>
<td>71.42857</td>
</tr>
</tbody>
</table>

Results:

T = 600°C, sd(T) = 13, P = 7.9 kbar, sd(P) = 0.7, cor = 0.567, sigf = 0.49
southern study area, as well as in the schists of the Selimiye nappe, has been used by Whitney & Bozkurt (2002) to indicate temperatures < 560 °C. We think, especially in the metasedimentary rocks of the Çine nappe, that the lack of staurolite (as well as the lack of aluminosilicates) is related to bulk composition.

In conclusion, $P-T$ estimates from the metasedimentary rocks of the Çine nappe below the Selimiye shear zone are of the order of 7 kbar and > 550 °C, which are higher than those from the metasedimentary rocks of the Selimiye nappe above the shear zone (c. 4 kbar and < 525 °C). This finding supports our structural observations in the field that both units are separated by a tectonic contact (Figs 2 & 5a).

**Northern study area**

Due to the occurrence of kyanite and staurolite, metamorphic conditions in this area are better constrained than in the southern study area. The KFMASH system is again used to describe the petro-genetic relationships, with the $P-T$ grid shown in Fig. 10(b). Analysis of the parageneses (Table S1) shows that kyanite–biotite–staurolite is a critical paragenesis in this area. Considering the $P-T$ grid and this critical paragenesis, $P-T$ estimates are between 8 and 11 kbar and > 600 °C.

As already observed, the retrograde reactions are less important than in the southern part and only mica shows retrograde alteration. Indeed analyses sometimes point to lower K contents in biotite (Table 1, samples D6, D16bis) than those expected (c. 10% of $K_2O$). Chloritoid always appears as inclusions in garnet, occasionally in textural equilibrium with staurolite (samples D6, D8, D19). This indicates prograde metamorphism (Figs 6c, 7c & 10b). In sample D16bis, one small porphyroblast of chloritoid was found in contact with biotite in the matrix. Because of the paragenesis garnet–biotite–staurolite in the same thin section, this mineral should be considered as relic and therefore metastable. The occurrence of kyanite in sample D23 could be due to the discontinuous reaction in the KFMASH subsystem:

\[
\text{chlorite + staurolite $\leftrightarrow$ biotite + kyanite.} \tag{4}
\]

Reaction (4) allows the stability of the paragenesis staurolite–biotite–kyanite in the KFMASH system and the disappearance of chlorite towards the Mg-chlorite end member (Fig. 10b). Because of the low Mg contents of chlorite in samples D23 and D30 (Table 1), these minerals are considered to be secondary.

Thermodynamic calculations using independent sets of reactions have been undertaken for samples D1, D30 and D42. Analyses and results are reported in Tables 1 and 3. Temperatures of 600–650 °C and pressures of 8–11 kbar were obtained. The results are consistent with the $P-T$ grid.

**DISCUSSION**

**Results of the metamorphic study**

East of Lake Bafa, the Selimiye shear zone separates two metasedimentary units, which record different $P-T$ conditions but largely similar structures. Metasedimentary rocks of the Çine nappe experienced amphibolite facies metamorphism shown by the occurrence of the amphibole–garnet paragenesis and $P-T$ conditions of about 7 kbar and > 550 °C. This schist sequence is overlain by lower grade metasedimentary rocks with maximum $P-T$ conditions of about 4 kbar and 525 °C. These greenschist to lower amphibolite facies rocks belong tectonically to the Selimiye nappe. Prograde metamorphism in the Selimiye nappe decreases structural upwards as indicated by mineral isograds defining the garnet–chlorite zone at the base, the chloritoid–biotite zone and the biotite–chlorite zone at the top of the nappe. Due to uncertainties in the thermodynamic calculations, the metamorphic break between the metasedimentary rocks of the Çine and Selimiye nappe is not very well defined but appears to be of the order of 2 kbar. The $P-T$ break indicates that about 7 km of metamorphic section is missing and therefore a tectonic contact, the Selimiye shear zone. Metamorphic mineralogy and fabric relationships demonstrate that the Selimiye shear zone developed during prograde metamorphism. However, the $P-T$ break demonstrates that movement along the Selimiye shear zone continued during peak and/or retrograde metamorphism (see also Whitney & Bozkurt, 2002).

Amphibolite facies schists in the northern study area south-west of Aydın record maximum $P-T$ conditions of 8–11 kbar and 600–650 °C. The inclusions in garnet indicate prograde metamorphism. Because the schists south-west of Aydın and those east of Lake Bafa both occur below the Selimiye shear zone and no major tectonic contact occurs between the two studied schist units, we propose that these two schist sequences belong to the same tectonic unit, the Çine nappe. However, the metasedimentary rocks in the northern study area record different structures than those in the south. We have found no evidence for regional-scale polymetamorphism in the metasedimentary rocks of either the Çine or Selimiye nappes.

**Implications for the nature of the Selimiye shear zone**

In the southern study area, mineral isograds in the Selimiye are parallel to the regional $S_R$ foliation and formed during prograde metamorphism. The $S_R$ foliation and associated top-S kinematic indicators are related to the formation of the Selimiye shear zone and show that the Selimiye shear zone formed during prograde to peak metamorphism. This strongly suggests that the Selimiye shear zone originated during N–S crustal shortening, not extension.
Gessner et al. (2001b) showed that the Eocene Cyclades–Menderes thrust resulted from N–S crustal shortening. Because the Selimiye shear zone also resulted from N–S tectonic movement, is also of Eocene age (Hetzel & Reischmann, 1996) and runs subparallel to the overlying Cyclades–Menderes thrust, it appears most probable that both shear zones formed during the same N–S shortening event and are thrusts.

The break in metamorphism across the Selimiye shear zone with lower grade rocks occurring tectonically above higher grade rocks is often considered to be evidence that the tectonic contact is extensional. However, it is important to note that a number of detailed studies (Wheeler & Butler, 1994; Ring & Brandon, 1994; Ring, 1995; Ring et al., 1999c) showed that this type of metamorphic break is not indicative for normal faulting. Indeed, Gessner et al. (2001b) showed that Eocene stacking of nappes in the western Anatolide belt was out-of-sequence thrusting, which can explain the occurrence of lower grade rocks tectonically above higher-grade rocks.

Strong evidence against an interpretation that the Selimiye shear zone is a metamorphic-core-complex-type normal fault, as proposed by Bozkurt & Park (1994), comes from the slow cooling recorded in the footwall of the Selimiye shear zone (Hetzel & Reischmann, 1996). The footwalls of core-complex-bounding normal faults show extremely fast cooling rates (Foster & John, 1999; Ring et al., 2003c). Rapid cooling at 23–20 Ma at the southern rim of the Anatolide belt as recorded by fission-track ages is remarkably similar to the pattern and timing of cooling at the northern end of the Anatolide belt (Ring et al., 2003b). In the north, fast cooling was caused by Early Miocene extensional faulting along the Simav detachment (İşik & Tekeli, 2001; Ring et al., 2003b). The Simav detachment reactivated the basal thrust of the Cycladic blueschist unit. The similarities of the fission-track cooling pattern suggests that there should also be an Early Miocene extensional detachment at the southern rim of the Anatolide belt and it has been speculated that the basal thrust of the Lycian nappes was reactivated as an extensional fault (Ring et al., 2003b). It might be feasible that some of the retrograde top-S shear-sense indicators in the upper Selimiye nappe formed during the proposed Early Miocene extensional reactivation of the Lycian nappes. Alternatively, the retrograde top-S kinematic indicators might be due to out-of-sequence thrusting along the Cyclades–Menderes thrust (see below).

**Cause of Eocene metamorphism across the Selimiye shear zone**

In the Cycladic blueschist unit of Dilek Peninsula, 40Ar/39Ar dating on phengite yielded an Eocene age of 40.1 ± 0.4 Ma (Oberhansli et al., 1998), which is most probably related to movement on deep-seated parts of the Cyclades–Menderes thrust (Gessner et al., 2001b; Ring & Layer, 2003). The age for greenschist to lower amphibolite facies metamorphism in the Selimiye nappe is also Eocene (> 43–37 Ma; Hetzel & Reischmann, 1996). The source of heating could be related to a complicated redistribution of the isotherms during the stacking of the Menderes nappes (lower units) involving lower and middle crustal levels. Subsequently, emplacement of the middle and upper tectonometamorphic units onto the Menderes nappes would have caused the retrograde metamorphism observed. The fact that the Menderes nappes have no Tertiary HP–LT overprint indicates that the middle and upper units were significantly exhumed before emplacement on the Menderes nappes.

**Orogenic implications**

The most fundamental question concerns the age of the described tectonometamorphic fabrics in the Çine nappe. Above it was shown that prograde metamorphism in the Selimiye nappe has to be Tertiary in age. However, the age of metamorphism in the underlying Çine nappe is much less clear. The lack of evidence for polymetamorphism might suggest that metamorphism in the entire Çine nappe has the same age as that in the Selimiye nappe. However, the example from the Sesia zone in the Alps given above shows that such a simple explanation might be misleading. Furthermore, Gessner et al. (2001a, 2003) showed that the amphibolite facies S_PA foliation and associated top-N kinematic indicators in orthogneiss are of late Neoproterozoic to Cambrian age and the 207Pb/206Pb garnet ages (Ring et al., 2003a) from the orthogneiss and from underlying metapelite of the Bozdag nappe support this conclusion.

Our preferred interpretation is that amphibolite facies metamorphism in the metasedimentary rocks of the southern Çine nappe, directly below the Selimiye nappe is also of Tertiary age and thus unrelated to the tectonometamorphic fabrics in the orthogneiss below these rocks. We believe that the strongest support for this inference are the DR structures with their consistent top-S kinematic indicators and the deformation/metamorphism relationships, which show that the kinematic indicators developed during prograde metamorphism and can be related to the Eocene Selimiye shear zone. The variation in foliation strike between the schist units in the hanging wall and footwall of the Selimiye shear zone and the orthogneiss (Fig. 2a) would then imply that the N–S-striking foliation in the orthogneiss is of late Neoproterozoic to Cambrian age and the WNE-striking foliation in the schist is of Eocene age.

Another possibility is that amphibolite facies metamorphism in the metasedimentary rocks of the southern Çine nappe is of the same late Neoproterozoic to Cambrian age as amphibolite facies metamorphism in the orthogneiss. Support for this interpretation would be that the schist and the orthogneiss record a single
amphibolite facies metamorphism. However, because the kinematic indicators in the orthogneiss indicate top-N shear with local coaxial deformation and those in the metasedimentary rocks consistently indicate top-S shear, this interpretation does not appear likely.

The age of amphibolite facies metamorphism in the metasedimentary rocks of the northern Çine nappe is even more problematic. Because the metasedimentary units in the south and in the north record amphibolite facies metamorphism and occur above the orthogneiss, one could assume that their metamorphism is of the same age, and, according to our preferred interpretation given above, of Tertiary age. However, in this case it is necessary to explain the different kinematic indicators associated with garnet growth, i.e. top-N rotated garnet in the north and top-S rotated garnet in the south. Furthermore, the main foliation in the northern metasedimentary rocks does not show a discordant pattern with the underlying orthogneiss. Because of this, and because the orthogneiss and the northern metasedimentary rocks depict the same prograde top-N kinematic indicators, we prefer the interpretation that amphibolite facies metamorphism in the metasedimentary rocks of the northern Çine nappe is of late Neoproterozoic to Cambrian age. This inference is supported by the metamorphic study of Ring et al. (2001) from the lowermost Çine nappe and the Bozdag nappe. These authors showed that deformation/metamorphism relationships between staurolite and garnet growth in these two nappes are very similar to those in the metasedimentary rocks of the Çine nappe in our northern study area. Ring et al. (2001) concluded that the staurolite-garnet relationships resulted from late Neoproterozoic to Cambrian metamorphism, which is corroborated by the \(^{207}\text{Pb}/^{206}\text{Pb}\) garnet ages (Ring et al., 2003a).

Our preferred interpretation implies that both studied metasediment outcrops in the Çine nappe have a monometamorphic amphibolite facies overprint of different age. Earlier studies (Candan et al., 2001; Gessner et al., 2001a; Ring et al., 2001) showed that the Çine and Bozdag nappes have widespread evidence of amphibolite, and local eclogite and granulite facies metamorphism of late Neoproterozoic to Cambrian age. Nappe stacking during the Tertiary orogeny caused pronounced retrogression along the nappe boundaries during greenschist facies metamorphism. Locally, deformation zones associated with top-S kinematic indicators formed during this greenschist facies retrogression within the Çine and Bozdag nappes. Apart from these localised zones at the nappe boundaries and within the nappes, the Çine and Bozdag nappes were not affected by Tertiary deformation and metamorphism. According to this interpretation the metasedimentary rocks of the Çine nappe directly below the Selimiye shear zone were severely affected by Alpine shearing. Note that only these southern metasedimentary rocks are close to a Tertiary thrust (Fig. 8). We speculate that deformation-related fluid flow obliterated evidence for older metamorphism in the metasedimentary rocks of the Çine nappe immediately below the Selimiye nappe.

Although this interpretation is speculative, it is based on a detailed metamorphic and microstructural study of more than 350 samples and a number of age constraints. This interpretation highlights that simple tectonometamorphic fabrics do not a priori indicate a simple orogenic development of a heterogeneous nappe stack. Preservation in some areas and complete overprinting of those fabrics in other areas is apparently controlled by deformation-related fluid flow associated with Tertiary nappe stacking. Detailed dating of tectonometamorphic fabrics is essential to constrain the orogenic history more closely.

Alpine top-N shearing in the Anatolide belt

A number of workers (Bozkurt & Park, 1994; Hetzel et al., 1998; Whitney & Bozkurt, 2002) interpreted top-N amphibolite facies shearing in the Bozdag and Çine nappes as an early phase of Tertiary deformation related to northward nappe movement. Gessner et al. (2001a, 2003) showed that the top-N kinematic indicators in orthogneiss were cut by granite yielding an age of 550 Ma (see above). \(^{207}\text{Pb}/^{206}\text{Pb}\) garnet dating in both nappes (Ring et al., 2003a) also yielded ages of > 500 Ma. These data indicate that amphibolite facies shearing in the Bozdag nappe and in orthogneiss of the Çine nappe is of latest Neoproterozoic to Cambrian age and is not related to the Alpine orogeny.

For the amphibolite facies top-N shearing event in the metasedimentary rocks of the Çine nappe in our northern study area we also suggested a pre-Tertiary age. However, there are no geochronological data for this rock sequence to substantiate this inference, and therefore a Tertiary age for top-N shearing, although unlikely, cannot be ruled out.

Bozkurt & Park (1994) and Whitney & Bozkurt (2002) based their inference of Tertiary top-N shearing in part on work in the Selimiye nappe, a tectonic unit, which has no proven pre-Alpine orogenic history. It has been shown in this study that occasional top-N sense-of-shear indicators in the Selimiye shear zone are inverted top-S kinematic indicators and are therefore not indicative for an early Tertiary phase of top-N shear. Furthermore, there is no evidence in the Selimiye shear zone, that the top-N shear-sense indicators formed before the top-S kinematic indicators as assumed by Bozkurt & Park (1994) and Whitney & Bozkurt (2002).

Bozkurt & Park (1994) also mentioned early north-verging folds in the Selimiye nappe. We have also mapped tight to isoclinal folds with WNW-trending axes and S-dipping axial planes. The structural facing direction of these asymmetric folds is unknown. The relationships between long and short limbs favour a top-S shear sense.
In summary, it is concluded that so far no evidence for early Alpine top-N nappe movement in the western Anatolide belt exists. Large-scale tectonic models also do not supply any hint for such an event (Şengör et al., 1984; Collins & Robertson, 1997, 1998, 2003).

CONCLUSIONS

The major conclusions are:
(1) The Selimiye shear zone developed during prograde greenschist- to lower amphibolite facies metamorphism, which resulted from burial below the upper and middle tectonometamorphic units of the Anatolide belt in the Eocene.
(2) The Selimiye shear zone results from N-S crustal shortening and is not an extensional shear zone.
(3) Metamorphism in the Selimiye nappe decreases structurally upwards as indicated by mineral isograds defining the garnet-chlorite zone at the base, the chloritoid-biotite zone and the biotite-chlorite zone at the top of the nappe.
(4) Despite the very fact that metamorphism in the underlying Čine nappe is relatively simple and monophase it appears unlikely that all studied metasedimentary rocks in the Čine nappe record the same Eocene metamorphism. Our study emphasises that rocks, which have relatively simple metamorphic fabrics, may reflect complex tectonic evolutions resulting from more than one single orogeny. A detailed geo-chronological study is needed to constrain the age of metamorphism in the Čine nappe.
(5) There is no evidence for an early Alpine phase of top-N thrusting.

ACKNOWLEDGEMENTS

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SUPPLEMENTARY MATERIAL

The following material is available online from www.blackwellpublishing.com/products/journals/summat/JMG/JMG473/JMG473sm.htm.

Fig S1. Location of samples

Table S1. Parageneses of studied samples.

Appendix. Activity models used with THERMOCALC v3.1 for the KFMASH system.

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