

Metamorphism of Precambrian–Palaeozoic schists of the Menderes core series and contact relationships with Proterozoic orthogneisses of the western Çine Massif, Anatolide belt, western Turkey

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Abstract – The tectonic setting of the southern Menderes Massif, part of the western Anatolide belt in western Turkey, is characterized by the exhumation of deeper crustal levels onto the upper crust during the Eocene. The lowermost tectonic units of the Menderes Massif are exposed in the Çine Massif, where Proterozoic basement orthogneisses of the Çine nappe are in tectonic contact with Palaeozoic metasedimentary rocks of the Selimiye nappe. In the southern Çine Massif, orthogneiss and metasedimentary rocks are separated by the southerly dipping Selimiye shear zone, preserving top-to-the-S shearing under greenschist facies conditions. In contrast, in the western Çine Massif, the orthogneiss is deformed and mylonitic near the contact with the metasedimentary rocks. The geometry of the mylonite zone and the observed shear directions change from north to southwest. In the north, the mylonite zone dips shallowly to the north, with top-to-the-N shear sense indicators showing northward thrusting of the orthogneiss over the metasedimentary rocks. In the southwest, the mylonite zone resembles a steep N–S striking strike-slip shear zone associated with top-to-the-SSW sense of shear. Overall, the geometry of the mylonite shear zone is consistent with northward movement of the orthogneiss relative to the metasedimentary rocks. Different shear senses are attributed to strain partitioning.

AFM diagrams and P – T pseudosections with mineral parageneses of metasedimentary rocks of the Selimiye nappe and metasedimentary enclaves within the orthogneiss of the Çine nappe indicate a single Barrovian-type metamorphism. An earlier higher pressure phase is evident from staurolite–chloritoid inclusions in garnets of the Çine nappe, suggesting a clockwise P – T path. A similar path is inferred for the Selimiye nappe. Index minerals and the sequence of mineral parageneses point to a single amphibolite facies metamorphic event affecting metasedimentary rocks of both nappes, which predates Eocene emplacement of the high pressure–low temperature Lycian and Cycladic blueschist nappes. Northward thrusting of the orthogneiss onto the metasedimentary rocks of the Selimiye nappe is coeval with amphibolite facies metamorphism. Recently postulated polymetamorphism cannot be supported by this study. Petrological data provide no evidence for burial of the lower units of the Menderes Massif to depth greater than 30 km during closure of the Neo-Tethys. A major pre-Eocene tectonic event associated with top-to-the-N thrusting and Barrovian-type metamorphism could lend support to the idea of a Neo-Tethys (*sensu stricto*) suture south of the Menderes Massif and below the Lycian nappes.

Keywords: metamorphism, P – T pseudosections, Çine Massif, Menderes Massif, Anatolide belt, western Turkey.

1. Introduction

Metamorphic core zones of orogenic belts display complex interaction of metamorphism and deformation. Asymmetrical fabrics resulting from non-coaxial ductile deformation can be used as shear sense indicators and help reconstruct kinematics of an orogen. However, deformation throughout an orogen is not necessarily homogeneous, as evident in differing senses of shear.

Correlation of varying shear directions with metamorphism and the exhumation history is therefore crucial for understanding the tectonometamorphic evolution of the core zone of an orogen. Heterogeneous shear sense indicators can result from polyphase deformation, but can also reflect strain partitioning during coeval extension and compression. The Montagne Noire in the southern French Massif Central is an example where coeval extension and compression occurred during exhumation (Echtler & Malavieille, 1990; Aerden & Malavieille, 1999; Matte, Lancelot & Mattauer, 1998). Brunel (1986) reported normal faulting and extension

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above thrust faults from the Himalayas. Even in the case of polyphase deformation an *a priori* assumption of polyphased metamorphism or even polymetamorphism cannot be made.

In the western Anatolide belt of western Turkey the lowermost tectonic units of the the Menderes Massif, Proterozoic orthogneiss basement and overlying Palaeozoic metasedimentary rocks, record opposite shear directions developed under amphibolite facies conditions (Ring, Willner & Lackmann, 2001; Régnier *et al.* 2003). The timing of amphibolite facies metamorphism and the nature of contact between basement and metasedimentary cover are still a matter of conjecture (Rimmelé *et al.* 2003). Bozkurt & Park (1994, 1997) argued for a single Barrovian-type metamorphic event, termed Main Menderes Metamorphism, which resulted from Eocene thrusting of high pressure–low temperature (HP–LT) rocks onto the underlying core series and overprinted Proterozoic basement and Palaeozoic sedimentary rocks. Satir & Friedrichsen (1986) and Hetzel & Reischmann (1996) constrain metamorphism in the Proterozoic basement to the Eocene. Régnier *et al.* (2003) used geochronological evidence to argue for coeval Tertiary metamorphism of basement and overlying cover rocks in the southern part of the Menderes Massif, and Neoproterozoic metamorphism of the basement in the northern part. However, the authors cautioned the deduction of two separate metamorphic events based only on geochronological data without convincing petrological evidence for polymetamorphism. The nature of the contact between the Proterozoic basement and the overlying Palaeozoic metasedimentary rocks in the southern part of the Menderes Massif has been interpreted as either tectonic (the Selimiye shear zone of De Graciansky, 1966; Ring *et al.* 1999; Gessner *et al.* 2001a; Régnier *et al.* 2003), and/or intrusive, inferred from granitoid intrusions crosscutting metasedimentary rocks and orthogneiss (Bozkurt, Park & Winchester, 1993; Bozkurt, 2004; Erdoğan & Güngör, 2004).

This study focuses on metamorphic conditions of the southern Menderes Massif across the contact of the orthogneiss with the surrounding metasedimentary rocks. New microstructural data and thermobarometric calculations of the western Çine Massif near Lake Bafa and representative areas of the eastern Çine Massif near Karacasu (Fig. 1), augmented by previous studies by Evirgen & Ataman (1982), Evirgen & Ashworth (1984), Ashworth & Evirgen (1984, 1985), Candan & Dora (1993), Whitney & Bozkurt (2002) and Régnier *et al.* (2003), help develop a refined geodynamic model of the southern Menderes Massif.

2. Geological setting of the Menderes Massif

The Anatolide belt of western Turkey contains three major tectonometamorphic units (Figs 1, 2). The structurally highest unit includes the Lycian nappes and the

İzmir–Ankara suture zone with the Bornova flysch zone comprising ophiolitic mélange and Late Palaeozoic to Mesozoic rift successions deposited during opening of the northern branch of the Neo-Tethys ocean (Collins & Robertson, 1999; Stampfli, 2000). Structurally underneath is the Cycladic blueschist unit with Mesozoic platform carbonates and metaolistostromes. Both units were affected by a single HP–LT metamorphic event resulting from the Late Cretaceous–Eocene closure of the northern branch of the Neo-Tethys (*sensu lato*), and were subsequently thrust southward along the Cyclades–Menderes thrust onto the structurally deepest tectonometamorphic unit, the Menderes core series (Şengör, Satir & Akkök, 1984; Güngör & Erdoğan, 2001; Oberhänsli *et al.* 1998, 2001; Sherlock *et al.* 1999; Collins & Robertson, 2003; Rimmelé *et al.* 2003, 2005).

The Menderes core series, also termed Menderes Massif, is divided by Neogene grabens into northern and central submassifs and the southern Çine Massif. It is interpreted as an Eocene out-of-sequence stacking of the Selimiye, the Çine, the Bozdağ and the Bayındır nappes (Figs 1, 2) (Ring *et al.* 1999; Gessner *et al.* 2001a; Régnier *et al.* 2003). The structurally highest Selimiye nappe comprises Devonian–Carboniferous metapelite, calc-schist, metamarl, marble and quartzite, and was subject to Eocene greenschist to lower amphibolite grade metamorphism (Schuiling, 1962; Çağlayan *et al.* 1980; Satir & Friedrichsen, 1986; Hetzel & Reischmann, 1996; Régnier *et al.* 2003).

The following structurally lower Çine nappe contains deformed orthogneiss and undeformed to weakly deformed metagranites of the Çine Massif, and interlayered amphibolite grade mica schists and partially migmatized sillimanite-bearing paragneisses, with eclogite enclaves recording amphibolite facies overprinting (Candan *et al.* 2001; Dora *et al.* 2001). A late Proterozoic intrusion age (560–540 Ma) is inferred for the orthogneiss protolith (Hetzel & Reischmann, 1996; Loos & Reischmann, 1999).

Structurally below the Çine nappe is the Bozdağ nappe: metapelites with intercalated metapsammite, marble, amphibolite and eclogite of Proterozoic age (Candan *et al.* 2001; Koralay *et al.* 2004). Rocks of the Bozdağ and Çine nappes are locally preserved as klippen on late Cretaceous low-grade phyllite, quartzite, and marble of the structurally lowest nappe, the Bayındır nappe (Figs 1, 2), where consistent top-to-the-S sense of shear is observed (Ring *et al.* 1999; Gessner *et al.* 2001a; Özer & Sözbilir, 2003).

Rocks of the Bozdağ and Çine nappes display amphibolite facies metamorphism associated with top-to-the-N shearing, whereas rocks of the Selimiye nappe are characterized by greenschist to lower amphibolite facies metamorphism coeval with top-to-the-S shearing. All rocks are affected by greenschist facies top-to-the-S shear bands. Amphibolite facies metamorphism is assigned to the Proterozoic, based

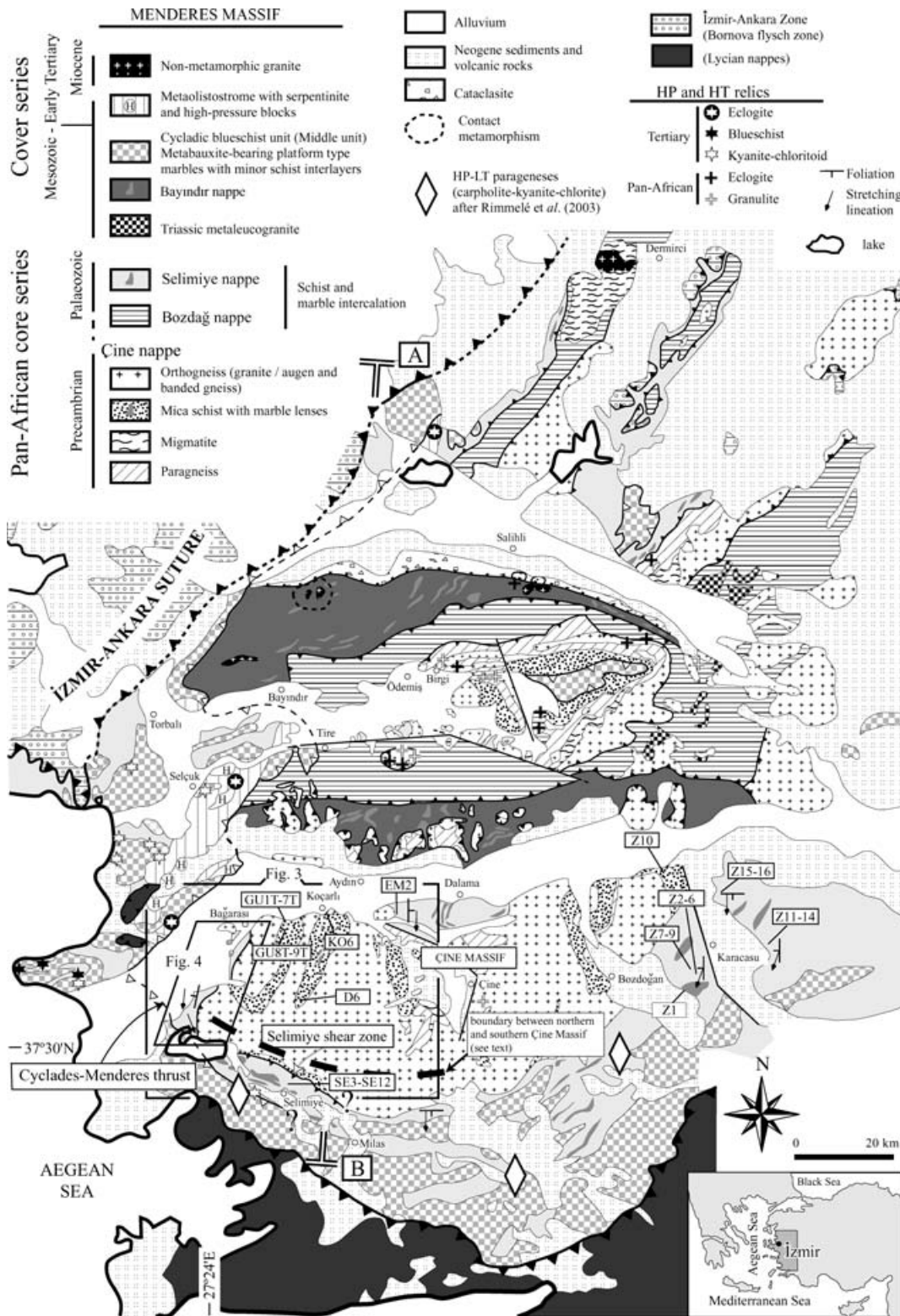


Figure 1. Generalized geologic map of the Menderes Massif in western Turkey modified after Candan & Dora (1998) and Gessner *et al.* (2001a). Locations of samples, detailed maps of the western Çine Massif (Figs 3, 4) and cross-section AB (Fig. 2) are shown. Note that the eastward prolongation of the Cyclades–Menderes thrust and the Selimiye shear zone is unknown (large question marks).

on a SHRIMP zircon age of 566 ± 9 Ma obtained from a metagranite crosscutting the penetrative amphibolite facies foliation in orthogneiss, while stacking of the

nappes is inferred to have occurred during the Tertiary under lower greenschist facies conditions (Gessner *et al.* 2001a, 2004).

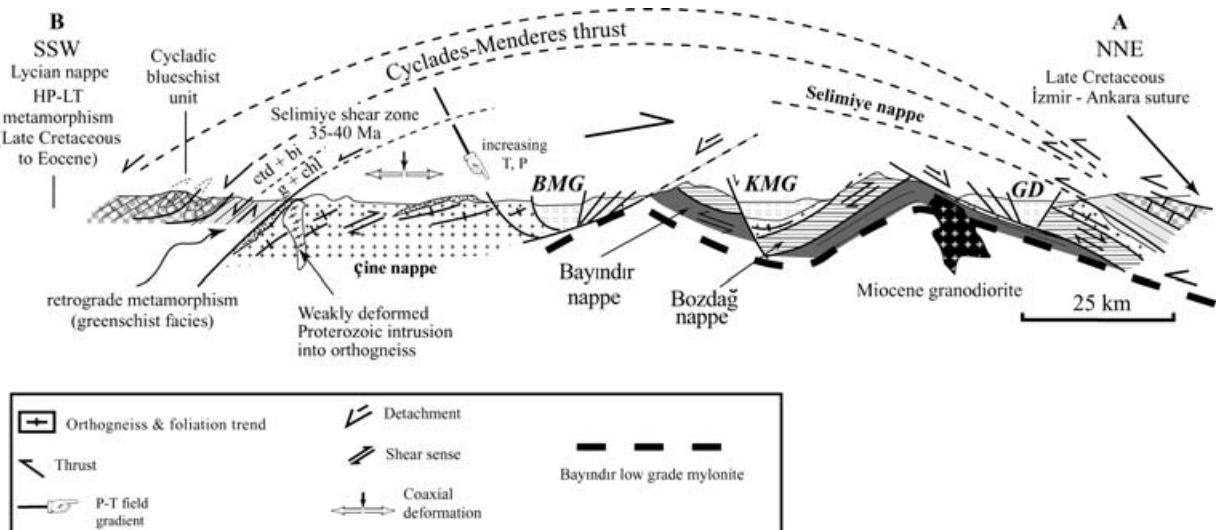


Figure 2. Interpretative cross-section through the central Menderes Massif modified after Gessner *et al.* (2001c), Lips *et al.* (2001) and Régnier *et al.* (2003). ^{40}Ar – ^{39}Ar white mica ages for the Selimiye shear zone are from Hetzel & Reischmann (1996). Thick dashed line outlines Bayındır low grade mylonite zone. BMG: Büyük Menderes graben; GG: Gediz graben; KMG: Küçük Menderes graben. For legend see Figure 1.

3. Structural character of the Çine Massif and nature of contact between orthogneiss and metasedimentary rocks

3.a. Previous studies in the southern Selimiye area

Nearly perpendicular strike directions of the foliation in the Çine nappe orthogneiss (N–S) and the schistosity within the overlying Selimiye nappe (WNW–ESE) are indicative of a major structural break (De Graciansky, 1966; fig. 2a in Régnier *et al.* 2003). Prograde top-to-the-S shearing within the Selimiye nappe is inferred from synkinematic garnet porphyroblasts in equilibrium with chloritoid, biotite and chlorite. Régnier *et al.* (2003) proposed prograde Barrovian-type metamorphism from greenschist to amphibolite facies (4–5 kbar, 350–500 °C), towards the contact with the orthogneiss to the north. An abrupt increase of approximately 2 kbar and 100 °C occurs across the Selimiye shear zone. Régnier *et al.* (2003) assigned metasedimentary rocks east of Lake Bafa, north of the Selimiye shear zone, to the Çine nappe, based on a late Proterozoic metagranite which crosscuts schistosity. Thus, the Selimiye shear zone would represent the contact between the Selimiye and the Çine nappes. Subparallelism of the shear zone, mineral isograds and the main schistosity within the Selimiye nappe suggest the contact originated during post-peak metamorphic motion along the Selimiye shear zone. Previous workers assigned the metasedimentary rocks north of the shear zone to the Selimiye nappe, interpreting intrusive relationships and weakly deformed intrusions close to the orthogneiss near the village of Selimiye (Bozkurt, Park & Winchester, 1993) as preserved intrusive relationships of Proterozoic (Hetzel & Reischmann,

1996; Loos & Reischmann, 1999) and/or younger age (Bozkurt, 2004; Erdoğan & Güngör, 2004).

North of the Selimiye shear zone pressure conditions within the Çine nappe exceeded 7 kbar. Metasedimentary enclaves within the orthogneiss south of the village of Koçarlı preserve staurolite–biotite–kyanite (+ garnet) paragenesis, which corresponds to 9 kbar and 650 °C within the KFMASH system. The prograde character of amphibolite facies metamorphism within the Çine nappe during regional top-to-the-N shearing is supported by the presence of chloritoid in helicitic garnet trails or chloritoid–staurolite inclusions in garnet (Régnier *et al.* 2003).

3.b. Western Çine Massif

The contact between Proterozoic orthogneisses and assumed Palaeozoic schists is best exposed in the western part of the Çine Massif, north of Lake Bafa (Figs 1, 3). Palaeozoic schists are overlain along a tectonic contact by Late Cretaceous metacarbonates of the Cycladic blueschist unit (Özer *et al.* 2001). The foliation of the orthogneiss strikes N–S and was affected by large-scale N-trending folds. In the metasedimentary rocks the foliation orientation is similar, except NE of Çalıköy where it was affected by NW-trending folds (Fig. 4, cross-section GH). Stretching lineations in the orthogneiss trend generally NNE, while in the metasedimentary rocks a NNE trend dominates in the northern part and SSE trends prevail in the southern part around Lake Bafa.

In the westernmost area ductile deformation of metasedimentary rocks is evident from sigmoidal quartz aggregates and C/S fabrics that indicate a

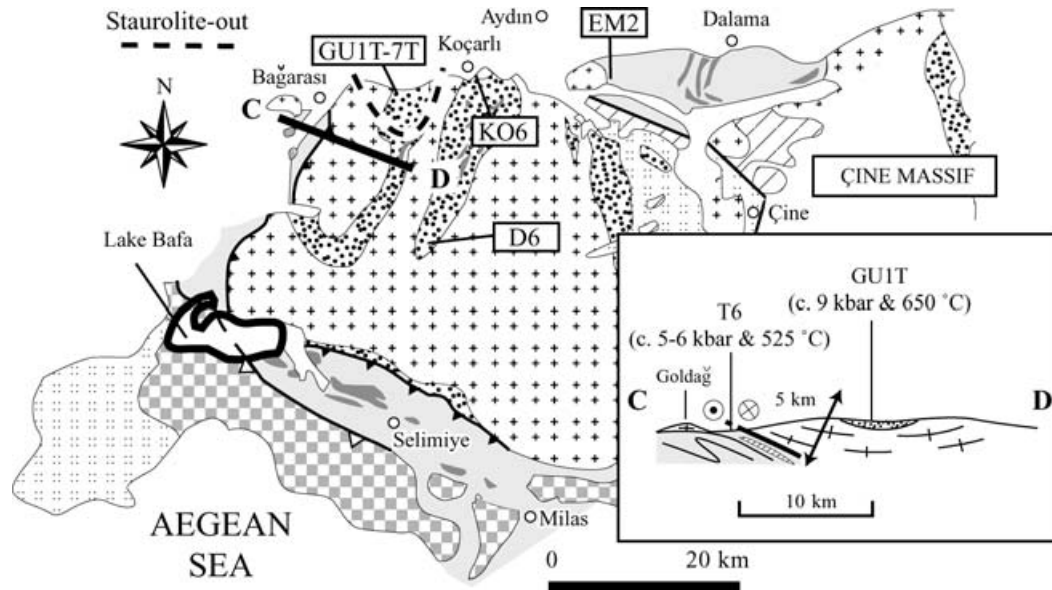


Figure 3. Geological sketch map of the western Çine Massif and cross-section showing contact with orthogneiss and underlying Selimiye nappe. Structural distance, perpendicular to the main foliation, between T6 and GUIT is estimated to be 5 km. Note the staurolite-out isograd north of the cross-section. For legend see Figure 1.

predominant top-to-the-SSW sense of shear (samples R1, R8; Fig. 5a, b), while garnet inclusion trails in albite in pelitic gneiss show consistent top-to-the-S sense of shear (Fig. 6b). Further east, close to the contact with the orthogneiss, sigmoidal quartz pebbles in metasedimentary rocks display mainly top-to-the-N sense of shear (Fig. 5c). Helicitic staurolite in textural equilibrium with chloritoid and chlorite shows top-to-the-N movement (sample T6, Fig. 7a). Greenschist facies top-to-the-S shear bands affected all metasedimentary rocks (Fig. 5c).

In an approximately 100 m thick zone at its base, the orthogneiss experienced strong deformation, partially mylonitic, evident from parallel alignment of finely recrystallized feldspar and quartz grains, as well as decimetre-sized NNE-trending tourmaline aggregates (Figs. 5d, e). Sigma-type feldspars indicate top-to-the-N sense of shear (Fig. 5g). Boudinaged foliation, indicating coaxial deformation, can also be observed (Fig. 5f). Thin sections reveal mica fish with a top-to-the-N sense of shear (sample GU8, Fig. 6c). Less intense deformation with similar characteristics is observed in the upper structural levels of the orthogneiss (Gessner *et al.* 2001a). Similar to the metasedimentary rocks, top-to-the-N shear sense indicators in the mylonitic orthogneiss are overprinted by greenschist facies top-to-the-S shear bands, evident at outcrop scale and in thin sections (Fig. 5g, h; Fig. 6d, sample TE2). In the north, near Goldağ peak, the orthogneiss appears to have been thrust along its mylonitic base northward over metasedimentary rocks which are folded into a recumbent SSW-verging antiform (Fig. 4, cross-section GH; Fig. 6a).

The mylonitic base of the orthogneiss in the western part of the Çine Massif and the underlying metasedimentary rocks show consistent top-to-the-N sense of shear. The large tourmaline aggregates (Fig. 5e) likely record significant fluid influx during deformation (Dutrow, Foster & Henry, 1999) that may have resulted in localized melting (Erdoğan & Güngör, 2004; Bozkurt, 2004). The geometry of the mylonite zone changes from a steep N–S-striking sinistral strike-slip shear zone near Lake Bafa to a shallowly northerly dipping dip-slip shear zone. Overall, the geometry of the shear zone is consistent with northwards movement of the orthogneiss over the metasedimentary rocks (Fig. 6a). Mylonitic deformation of the metasedimentary rocks generally display top-to-the-N sense of shear, in the southwestern area north of Lake Bafa a top-to-the-SSW sense of shear is associated with the strike-slip dominated part of the shear zone. Conflicting shear directions could result from two separate tectonic events which affected metasedimentary rocks. However, the absence of mylonite within metasedimentary rocks suggests that synkinematic mineral growth was confined to top-to-the-N thrusting and loss of fluid. Mineral growth always results, without major change in pressure, in volume increase and loss of fluid during prograde metamorphism. Thus, different senses of shear could result from rheological contrasts or strain partitioning. Alternatively, observed decimetre-sized a-type folds (Malavieille, 1987) could have inverted initial top-to-the-N kinematic indicators in the southwestern area, so that initial top-to-the-N thrusting is coeval with metamorphism throughout the metasedimentary

Figure 4. Structural map of the western Menderes Massif between Bağarası and Lake Bafa, modified after Schuiling (1962), I. Taskin (unpub. Graduate Thesis, Dokuz Eylül Univ. İzmir, 1981) and Erdoğan & Güngör (2004). Localities of samples and figures are shown. Cross-sections show the geometry of the contact between the orthogneiss of the Çine nappe and metasedimentary rocks in different directions. Temperature gradient and approximate metamorphic zones are shown. The garnet-in isograd is considered to be a 'pseudo-isograd' (see discussion in text).

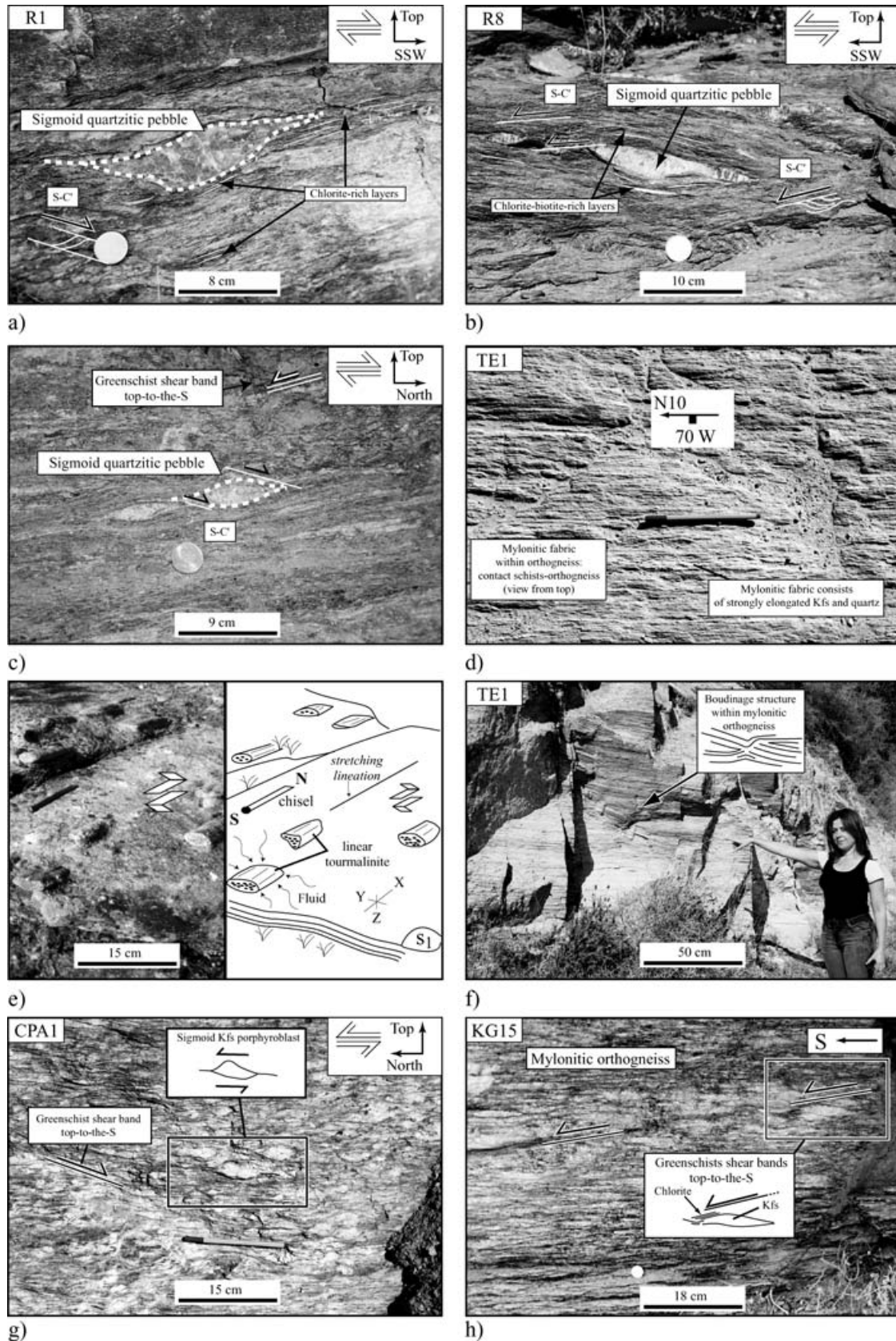


Figure 5. Structural character of metasedimentary rocks and orthogneiss in the western Çine Massif. (a, b) Sigmoidal quartzitic pebbles and S-C' shear bands displaying top-to-the-SSW motion. (c) Top-to-the-N sigmoidal quartz pebble within metasedimentary rocks near the orthogneiss. Greenschist facies shear bands indicate top-to-the-S shearing. (d) Ground view of mylonitic orthogneiss near the contact with metasedimentary rocks. (e) Parallel alignment of tourmalinite megablasts with mineral stretching lineation at the western margin of the orthogneiss. Sense of shear was deduced from thin section. (f) Boudinage foliation within mylonitic orthogneiss. (g, h) Top-to-the-N sigmoidal potassium feldspar (Kfs) porphyroblast and associated top-to-the-S greenschist shear bands. Sample numbers in the upper left corners refer to locations on Figure 4.

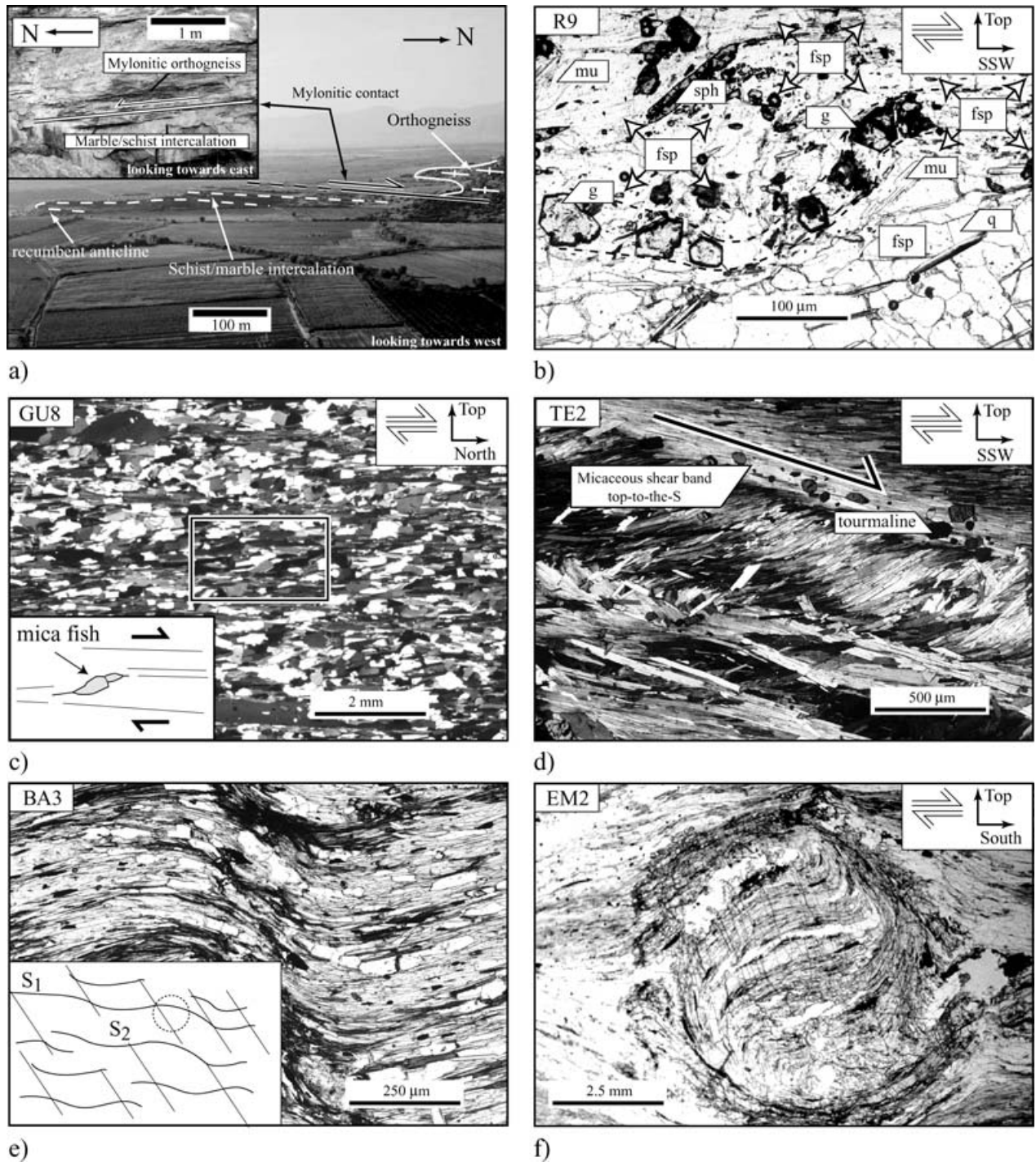


Figure 6. Structural character of metasedimentary rocks and orthogneiss in the Çine Massif. (a) Mylonitic contact between orthogneiss and metasedimentary rocks south of Goldağ peak. See cross-section GH in Figure 4. (b) Photomicrograph of schist from the western Çine Massif with inclusion trails of garnet in albite indicating top-to-the-SSW sense of shear. Plane-polarized light (PPL). For mineral abbreviations see Appendix 1. (c) Photomicrograph of mica schist of the Çine nappe. Mylonitic fabric with mica fish implies top-to-the-N shearing. Cross-polarized light (XPL). (d) Photomicrograph of mica schist from the western Çine Massif near the contact with the orthogneiss. Micaceous shear bands indicate top-to-the-SSW sense of shear. XPL. (e) Photomicrograph of schist from the western Çine Massif north of Lake Bafa. Crenulation cleavage S₂ overprints main S₁ schistosity. XPL. (f) Photomicrograph of metasedimentary rocks of the central Çine massif west of Dalama. Large garnet porphyroblast with helicitic inclusion trails indicates dextral top-to-the-S rotation. PPL. Sample numbers in the upper left corners refer to locations on Figures 3 and 4.

rocks. Absence of disharmonic structures supports this assumption, but a detailed metamorphic study is necessary to constrain deformation history.

A secondary S₂ crenulation cleavage associated with NNE-trending folds overprints the primary S₁ schistosity (Figs 4, 6e). Locally, centimetre-sized recumbent

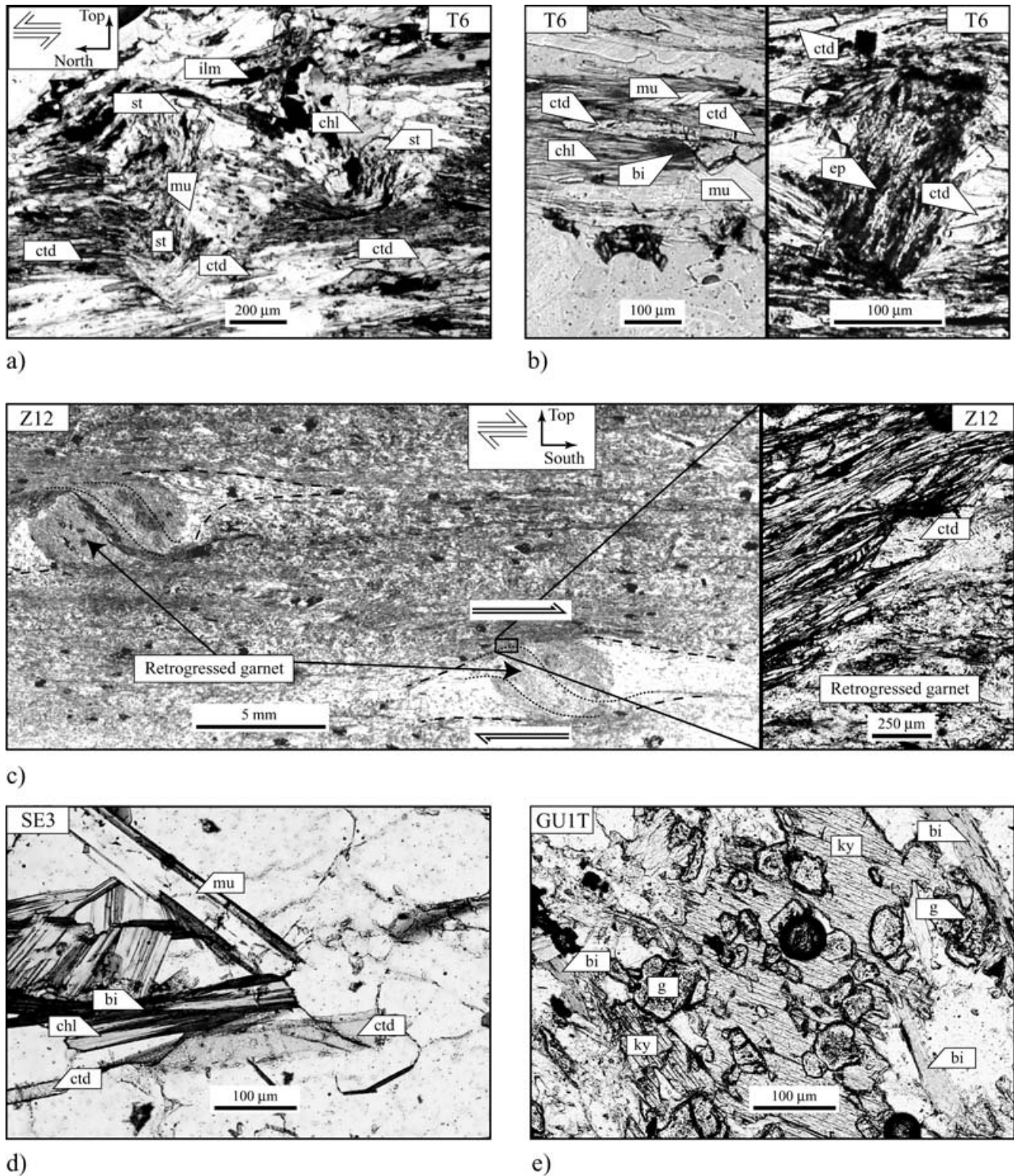


Figure 7. Photomicrographs of observed mineral parageneses. (a) Staurolite–chlorite–chloritoid paragenesis from the western Çine Massif. Helicitic staurolite porphyroblasts display consistent top-to-the-N shear. XPL. (b) Chloritoid–chlorite–biotite (left) and epidote–chloritoid parageneses (right) from the same sample. XPL. (c) Eastern Selimiye nappe near Karacasu: retrogressed garnet porphyroblast with curved inclusion trails and sigma-type quartz pressure shadows indicate top-to-the-S sense of shear. Detail shows garnet–chloritoid paragenesis. Biotite and chlorite occur as traces. PPL. (d) Southern Selimiye nappe: chlorite–biotite–chloritoid paragenesis. PPL. (e) Metasedimentary enclave in the northern Çine nappe: garnet–kyanite–biotite paragenesis. PPL. Sample numbers in the upper left corners refer to locations on Figures 1 and 4. For mineral abbreviations see Appendix 1.

NNE-trending folds, subparallel to stretching lineation, affected metasedimentary rocks. Folding is inferred to postdate juxtaposition of the orthogneiss and metasedimentary rocks.

3.c. Northern and eastern Çine Massif area

Synkinematic porphyroblasts found in enclaves of metasedimentary rocks within the orthogneiss of the northern Çine Massif, south of Koçarlı, record a

Table 1. Single mineral analyses of the Çine Massif

Sample Mineral	SE3 ctd	SE3 mu	SE3 mu	SE3 mu	SE12 g rim	SE12 ctd	SE12 chl	SE12 mu	SE12 mu	SE12 mu	SE12 fsp	GU1T g	GU1T g	GU1T bi	GU1T fsp	GU1T fsp	T6 mu	T6 mu	T6 mu	T6 chl	T6 chl	T6 ctd	T6 ctd	T6 st
SiO ₂	24.83	46.97	48.22	48.85	37.93	24.51	24.77	44.96	47.36	47.33	59.90	38.26	38.90	37.58	61.11	61.36	47.34	45.64	45.14	24.82	23.93	24.87	24.69	28.26
TiO ₂		0.27	0.30	0.31	0.02		0.04	0.11	0.35	0.27	0.01	0.10	0.04	1.39	0.03	0.01	0.27	0.30	0.21	0.04	0.13		0.10	0.23
Al ₂ O ₃	40.24	35.22	33.76	32.39	21.38	40.78	23.14	36.34	35.31	35.97	24.69	21.50	21.72	18.07	23.38	23.16	34.80	36.55	37.82	22.77	22.99	40.81	40.47	53.49
Cr ₂ O ₃		0.12	0.04	0.43	0.09	0.07	0.11	1.74	0.15	0.04	0.06	0.10	0.06	0.12	0.02	0.09	0.08	0.17	0.19	0.45	0.02	0.03	0.08	0.17
Fe ₂ O ₃		0.65	0.35		0.65	0.26		0.65							0.03	0.12	1.28							
FeO	24.57	1.36	1.65	1.95	33.89	22.74	26.31	0.39	1.11	0.97		30.57	30.34	14.29			0.49	0.85	0.72	27.86	28.03	23.33	23.40	8.53
ZnO																								4.93
MnO	0.51	0.02	0.03		0.93	0.10	0.06		0.01	0.03		1.00	0.97	0.08	0.03	0.03				0.05	0.03	0.15	0.14	0.02
MgO	2.22	0.81	1.31	1.51	2.44	3.48	13.54	0.42	0.89	0.76		4.65	4.80	13.66	0.01		0.97	0.48	0.27	11.82	12.02	2.70	2.77	0.95
CaO		0.09		0.11	4.78		0.01	0.06	0.01		3.01	4.25	4.59	0.05	5.20	5.27	0.03	0.03	0.14	0.03	0.03	0.04	0.04	
Na ₂ O		1.19	0.93	0.86	0.01	0.06	0.01	1.43	1.18	1.66	9.89	0.04	0.03	0.20	8.56	8.34	0.96	1.32	1.63	0.02	0.01	0.02	0.01	
K ₂ O	0.02	8.62	9.07	8.85				8.41	8.93	8.58	0.04	0.01	0.01	8.88	0.07	0.07	8.17	9.05	8.38	0.04	0.02		0.02	
Total	92.39	95.32	95.67	95.26	102.12	92.01	88.00	94.52	95.32	95.62	97.79	100.47	101.46	94.33	98.43	98.45	94.41	94.37	94.40	88.01	87.21	91.94	91.71	96.61
Oxygen	6	11	11	11	12	6	14	11	11	11	8	12	12	11	8	8	11	11	11	14	14	6	6	46
Si	1.03	3.10	3.18	3.23	2.99	1.01	2.61	3.00	3.12	3.10	2.72	3.01	3.02	2.80	2.75	2.76	3.13	3.04	3.00	2.64	2.57	1.02	1.02	7.93
Ti			0.02	0.02				0.01	0.02	0.01		0.01		0.08			0.01	0.02	0.01		0.01			0.05
Al	1.96	2.74	2.62	2.52	1.98	1.98	2.87	2.86	2.74	2.78	1.32	1.99	1.99	1.59	1.24	1.23	2.72	2.87	2.96	2.86	2.92	1.98	1.97	17.70
Cr		0.01		0.02	0.01		0.01	0.09	0.01			0.01		0.01				0.01	0.01	0.04				0.04
Fe ³⁺		0.03	0.02		0.04	0.01		0.03									0.06							
Fe ²⁺	0.85	0.08	0.09	0.11	2.23	0.78	2.32	0.02	0.06	0.05		2.01	1.97	0.89			0.03	0.05	0.04	2.48	2.52	0.80	0.81	2.00
Zn																								1.02
Mn	0.02				0.06		0.01					0.07	0.06	0.01						0.01		0.01	0.01	
Mg	0.14	0.08	0.13	0.15	0.29	0.21	2.13	0.04	0.09	0.07		0.54	0.56	1.52			0.10	0.05	0.03	1.87	1.93	0.17	0.17	0.40
Ca		0.01		0.01	0.40						0.14	0.36	0.38		0.25	0.25			0.02					0.01
Na		0.15	0.12	0.11				0.19	0.15	0.21	0.87	0.01		0.03	0.75	0.73	0.12	0.17	0.21	0.01				
K	0.00	0.73	0.76	0.75				0.72	0.75	0.72				0.85			0.69	0.77	0.71	0.01				
Total	3.99	6.94	6.93	6.91	8.00	4.00	9.95	6.96	6.94	6.96	5.06	7.99	7.99	7.76	5.00	4.99	6.87	6.97	6.97	9.92	9.96	3.99	3.99	29.15

Table 1. (Contd.)

Sample Mineral	T6 st	T6 st	T6 st	T6 st	T6 ilm	T6 ep	T6 fsp	T6 fsp	R9 g core	R9 g rim	R9 fsp core	R9 fsp rim	R9 mu (fsp)	R9 mu	R9 amph (fsp)	R9 amph (fsp)	Z12 ep	Z12 mu	Z12 mu	Z12 mu	Z12 mu	Z12 chl	Z12 chl II
SiO ₂	28.22	28.38	27.50	28.49	0.07	32.41	58.25	59.02	37.82	37.66	67.63	66.45	50.28	48.87	49.19	47.63	38.72	46.37	46.23	46.05	47.03	24.62	24.17
TiO ₂	0.34	0.38	0.47	0.65	52.94	0.03			0.10	0.08		0.02	0.24	0.24	0.23	0.30	0.15	0.12	0.20	0.21	0.23	0.07	0.04
Al ₂ O ₃	54.25	53.71	50.99	53.41	0.03	18.22	24.70	24.82	21.11	21.31	19.54	19.56	26.89	27.59	10.16	12.03	28.35	33.68	34.37	33.46	33.64	22.79	22.82
Cr ₂ O ₃	0.06	0.08	1.75	0.09	0.14	0.39	0.32		0.09	0.06	0.12	0.01	0.67	0.04	0.06	0.05	0.06	0.03	0.02	0.62	0.06	0.11	0.12
Fe ₂ O ₃						11.33	0.02	0.09			0.49	0.07	0.06	0.92	1.38	1.93	6.18	1.65	1.12	1.28	0.66		
FeO	8.48	8.54	8.26	8.59	44.38	0.10			24.99	29.54			2.88	1.72	15.58	15.64	0.06	0.64	0.43	0.49	0.88	28.96	29.43
ZnO	5.05	5.02	4.68	4.82																			
MnO	0.04	0.08		0.08	0.26	0.06	0.04		7.14	1.37	0.07		0.02	0.01	0.09	0.03	0.15	0.02			0.01	0.10	0.15
MgO	0.87	0.93	0.87	0.89		0.24			0.85	0.97		0.02	2.85	2.92	11.38	10.22	0.04	1.07	1.02	1.01	1.03	11.55	11.21
CaO	0.02		0.17	0.01	0.02	12.63	6.52	6.60	8.13	9.23	0.29	0.36	0.12		8.07	7.94	23.13	0.01	0.01	0.10	0.01	0.05	0.02
Na ₂ O					0.03	0.06	7.53	7.64	0.05	0.04	11.63	11.40	0.33	0.45	1.30	1.53	0.02	0.56	0.85	0.63	0.73		—
K ₂ O					0.01		0.06	0.04			0.07	0.08	9.57	9.36	0.20	0.27		8.06	8.35	8.25	8.38	0.01	
Total	97.33	97.11	94.69	97.03	97.86	75.48	97.45	98.22	100.28	100.26	99.83	97.97	93.91	92.12	97.64	97.57	96.87	92.22	92.61	92.11	92.67	88.27	87.96
Oxygen	46	46	46	46	3	12.5	8	8	12	12	8	8	11	11	23	23	12.5	11	11	11	11	14	14
Si	7.86	7.93	7.92	7.96		3.26	2.66	2.68	3.02	3.00	2.97	2.97	3.40	3.36	7.14	6.95	3.03	3.14	3.12	3.13	3.17	2.62	2.60
Ti	0.07	0.08	0.10	0.14	1.02				0.01	0.01			0.01	0.01	0.03	0.03	0.01	0.01	0.01	0.01	0.01	0.01	
Al	17.81	17.68	17.30	17.59		2.16	1.33	1.33	1.99	2.00	1.01	1.03	2.15	2.23	1.74	2.07	2.61	2.69	2.74	2.68	2.68	2.86	2.89
Cr	0.01	0.02	0.40	0.02		0.03	0.01		0.01				0.04		0.01	0.01					0.01	0.01	
Fe ³⁺						0.86					0.02			0.05	0.15	0.21	0.36	0.08	0.06	0.07	0.03		
Fe ²⁺	1.98	2.00	1.99	2.01	0.95	0.01			1.67	1.97			0.16	0.10	1.89	1.91		0.04	0.02	0.03	0.05	2.58	2.64
Zn	1.04	1.04	0.99	0.99																			
Mn	0.01	0.02		0.02	0.01	0.01			0.48	0.09					0.01		0.01					0.01	0.01
Mg	0.36	0.39	0.37	0.37		0.04			0.10	0.12			0.29	0.30	2.46	2.22	0.01	0.11	0.10	0.10	0.10	1.84	1.79
Ca	0.01		0.05			1.36	0.32	0.32	0.70	0.79	0.01	0.02	0.01		1.26	1.24	1.94			0.01	0.01	0.01	
Na						0.01	0.67	0.67	0.01	0.01	0.99	0.99	0.04	0.06	0.37	0.43		0.07	0.11	0.08	0.10		
K													0.83	0.82	0.04	0.05		0.70	0.72	0.72	0.72		
Total	29.15	29.14	29.13	29.10	1.98	7.73	5.00	5.00	7.98	7.99	5.01	5.01	6.93	6.93	15.09	15.12	7.98	6.85	6.89	6.87	6.87	9.94	9.95

Table 1. (Contd.)

Sample Mineral	Z12 g	Z12 ctd (g)	Z12 ctd (g)	Z12 ctd	Z12 ctd	GU10 mu	GU10 mu	GU10 ctd (g)	GU10 ctd (g)	GU10 g rim	GU10 g core	GU10 g rim	GU10 chl	GU10 st	GU10 st	Z7 mu	Z8 mu	KO6 st	KO6 g core	KO6 g rim	KO6 ctd (g)	GU9 ctd (g)	GU9 ctd
SiO ₂	37.53	24.17	24.38	24.23	24.70	47.34	45.33	24.62	24.46	37.58	37.67	37.93	24.73	28.10	27.69	46.25	46.77	26.82	37.59	37.88	24.68	24.86	24.84
TiO ₂	0.07	0.02				0.55	0.46			0.08	0.07	0.11	0.06	0.60	0.41	0.29	0.25	0.37	0.05		0.02		0.01
Al ₂ O ₃	21.26	40.63	40.60	41.00	40.08	32.14	34.65	40.46	40.56	20.85	20.93	21.15	21.59	53.70	49.53	34.85	34.95	54.58	20.88	21.19	40.55	40.79	40.61
Cr ₂ O ₃	0.02	0.04	0.01	0.14	0.10	0.05	0.11	0.03	0.08	0.03	0.08	0.03	0.45	0.07	2.30	0.11	0.05	0.10	0.03	0.10	0.02	0.29	0.11
Fe ₂ O ₃	0.55	0.53		0.30			0.91																
FeO	29.80	23.19	23.62	24.46	24.24	1.31	0.35	23.37	23.69	30.33	30.03	30.23	24.22	8.95	8.54	0.68	0.50	14.28	33.05	36.44	23.85	23.38	23.39
ZnO														5.24	5.13			0.18					
MnO	2.24	0.18	0.28	0.26	0.23		0.03	0.04	0.14	0.43	1.00	0.52	0.04	0.07	0.05	0.02		0.02	1.34	0.04	0.04	0.08	0.05
MgO	1.53	3.05	2.87	2.34	2.55	1.58	1.14	2.61	2.57	1.87	1.10	1.62	14.02	0.92	1.01	0.91	0.86	0.98	0.85	3.59	2.51	2.67	2.74
CaO	7.82	0.01		0.03	0.01	0.05	0.03	0.02	0.02	7.39	8.73	8.27	0.18		0.19	0.03	0.01	0.05	6.21	0.45	0.02	0.05	0.02
Na ₂ O	0.03	0.01				0.86	1.09			0.03	0.02	0.03	0.06			1.05	1.21		0.09	0.05		0.01	
K ₂ O		0.01	0.02	0.01	0.01	8.54	8.36		0.01		0.01	0.01	0.02			8.45	8.69	0.01					0.01
Total	100.85	91.85	91.78	92.77	91.92	92.43	92.47	91.16	91.53	98.59	99.64	99.91	85.38	97.64	94.85	92.64	93.31	97.22	100.09	99.75	91.70	92.14	91.78
Oxygen	12	6	6	6	6	11	11	6	6	12	12	12	14	46	46	11	11	46	12	12	6	6	6
Si	2.98	1.00	1.01	1.00	1.02	3.21	3.07	1.02	1.02	3.03	3.02	3.02	2.67	7.84	8.01	3.12	3.13	7.50	3.03	3.04	1.02	1.02	1.03
Ti			0.00			0.03	0.02			0.01		0.01	0.01	0.13	0.09	0.02	0.01	0.08					
Al	1.99	1.98	1.98	1.99	1.96	2.57	2.77	1.98	1.99	1.98	1.98	1.99	2.75	17.66	16.89	2.77	2.76	18.00	1.98	2.00	1.98	1.98	1.98
Cr				0.01			0.01				0.01		0.04	0.02	0.53	0.01		0.02		0.01		0.01	
Fe ³⁺	0.03	0.02		0.01			0.05																
Fe ²⁺	1.98	0.80	0.82	0.84	0.84	0.07	0.02	0.81	0.82	2.05	2.02	2.02	2.19	2.09	2.07	0.04	0.03	3.34	2.23	2.44	0.83	0.80	0.81
Zn														1.08	1.10			0.04					
Mn	0.15	0.01	0.01	0.01	0.01				0.01	0.03	0.07	0.04		0.02	0.01				0.09				
Mg	0.18	0.19	0.18	0.14	0.16	0.16	0.12	0.16	0.16	0.23	0.13	0.19	2.26	0.38	0.44	0.09	0.09	0.41	0.10	0.43	0.16	0.16	0.17
Ca	0.67									0.64	0.75	0.71	0.02		0.06			0.02	0.54	0.04			
Na						0.11	0.14			0.01		0.01	0.01			0.14	0.16		0.01	0.01			
K						0.74	0.72									0.73	0.74	0.01					
Total	8.00	4.00	4.00	4.00	4.00	6.90	6.93	3.98	3.99	7.97	7.98	7.98	9.94	29.20	29.19	6.91	6.92	29.40	7.98	7.96	3.99	3.98	3.99

Table 1. (Contd.)

Sample Mineral	GU9 mu	GU9 g core	GU9 g rim	Ki6 ctd (g)	Ki6 g core	Ki6 g rim	Ki6 bi	Ki6 mu	Ki6 fsp rim	Ki6 fsp core	EM2 mu	EM2 ctd	EM2 ctd	EM2 ctd (g)	EM2 ctd (g)	EM2 g core	EM2 g rim	EM2 chl	GU6T amph core	GU6T amph rim	GU6T bi	GU6T fsp rim	GU6T fsp core
SiO ₂	47.14	37.68	37.96	24.51	37.13	37.90	36.81	46.40	67.96	68.29	47.41	24.51	24.52	24.45	24.68	37.32	37.49	24.58	45.10	45.83	39.53	59.16	63.28
TiO ₂	0.40	0.06	0.04		0.11	0.03	1.49	0.38			0.35		0.02		0.01	0.05	0.01	0.04	0.40	0.42	1.77		
Al ₂ O ₃	34.41	21.20	20.95	41.23	21.11	21.59	19.28	35.51	18.99	19.33	32.11	40.77	40.67	40.57	40.07	21.05	21.15	23.08	13.35	12.56	16.74	24.84	22.29
Cr ₂ O ₃	0.14	0.04	0.07	0.09	0.09	0.05	0.08	0.10	0.52		0.60	0.07	0.06	0.07	0.08	0.10	0.05	0.12	0.12	0.41	0.50	0.06	
Fe ₂ O ₃						0.02			0.04	0.07	0.22	0.02				0.43			1.93	1.62		0.08	0.08
FeO	1.39	33.54	33.28	23.75	30.08	33.55	17.24	1.13			1.47	23.57	23.12	23.44	23.95	31.08	31.90	26.89	11.33	11.23	13.52		
ZnO																							
MnO		0.55	0.53	0.21	4.81	1.55	0.03	0.02				0.17	0.06	0.07	0.07	0.87	0.88	0.04	0.24	0.19	0.09		0.06
MgO	0.70	1.83	1.83	2.88	1.43	2.23	10.89	0.65	0.02		1.47	3.08	3.05	3.00	2.82	1.78	1.91	12.50	11.23	11.70	13.62		
CaO	0.03	5.29	5.26	0.02	4.94	4.88	0.02		0.20	0.10	0.07	0.02	0.06	0.02	0.01	7.28	6.56	0.04	11.30	11.55	0.49	7.15	4.06
Na ₂ O	0.98	0.02	0.01		0.13		0.21	1.79	11.42	11.77	0.81	0.01		0.01		0.03	0.01		1.35	1.23	0.13	7.31	9.12
K ₂ O	8.61	0.01		0.02	0.01		8.77	8.14	0.06	0.06	8.65		0.01		0.02	0.01		0.02	0.55	0.51	7.54	0.08	0.10
Total	93.80	100.22	99.93	92.71	99.84	101.81	94.83	94.13	99.21	99.62	93.17	92.22	91.56	91.63	91.72	100.01	99.97	87.31	96.91	97.26	93.94	98.68	98.98
Oxygen	11	12	12	6	12	12	11	11	8	8	11	6	6	6	6	12	12	14	23	23	11	8	8
Si	3.15	3.02	3.04	1.00	3.00	2.99	2.76	3.09	3.00	3.00	3.20	1.01	1.01	1.01	1.02	2.99	3.00	2.62	6.61	6.68	2.92	2.67	2.82
Ti	0.02				0.01		0.08	0.02			0.02								0.04	0.05	0.10		
Al	2.71	2.00	1.98	1.99	2.01	2.01	1.71	2.79	0.99	1.00	2.56	1.98	1.98	1.98	1.96	1.99	2.00	2.90	2.31	2.16	1.46	1.32	1.17
Cr	0.01		0.01		0.01		0.01	0.01	0.02		0.03					0.01		0.01	0.01	0.05	0.03		
Fe ³⁺											0.01					0.03			0.21	0.18			
Fe ²⁺	0.08	2.25	2.23	0.81	2.03	2.21	1.08	0.06			0.08	0.81	0.80	0.81	0.83	2.08	2.14	2.40	1.39	1.37	0.83		
Zn																							
Mn		0.04	0.04	0.01	0.33	0.10						0.01				0.06	0.06		0.03	0.02	0.01		
Mg	0.07	0.22	0.22	0.18	0.17	0.26	1.22	0.07			0.15	0.19	0.19	0.19	0.17	0.21	0.23	1.99	2.45	2.54	1.50		
Ca		0.45	0.45		0.43	0.41			0.01	0.01	0.01					0.63	0.56		1.78	1.81	0.04	0.35	0.19
Na	0.13				0.02		0.03	0.23	0.98	1.00	0.11					0.01			0.38	0.35	0.02	0.64	0.79
K	0.74						0.84	0.69			0.75								0.10	0.10	0.71		0.01
Total	6.90	7.98	7.97	4.00	8.00	8.00	7.73	6.96	4.99	5.01	6.91	4.00	3.99	4.00	4.00	8.00	8.00	9.92	15.32	15.30	7.61	4.99	4.99

Mineral abbreviations: amph: amphibole, bi: biotite, chl: chlorite, ctd: chloritoid, ep: epidote, fsp: feldspar, g: garnet, ilm: ilmenite, mu: muscovite, st: staurolite. Mineral formulae are calculated using the program AX by T. J. B. Holland (see text for URL), except for staurolite. SE3: chloritoid–biotite–chlorite zone (Selimiye nappe); SE12: chloritoid–garnet–chlorite zone (Selimiye nappe); GU1T: garnet–biotite–kyanite zone (Çine nappe); T6, R9: western Selimiye nappe; Z12: garnet–biotite–chlorite–chloritoid zone (eastern part); (fsp) indicates that phase occurs as inclusion in feldspar; GU9, GU10, Ki6: Selimiye nappe (western part); Z7, Z8: Selimiye nappe, eastern part; KO6: Çine nappe; (g) indicates that phase occurs as inclusion in garnet; EM2: garnet–biotite–chlorite–chloritoid zone (northern part); GU6T: garnet–biotite–kyanite zone (Çine nappe).

bulk top-to-the-N sense of shear. However, foliation boudinage and coaxial deformation with top-to-the-S sense of shear are also observed (Régner *et al.* 2003). Metasedimentary rocks of the northern central Çine Massif, between the villages of Dalama and Çine, contain rotated garnets with inclusion trails displaying top-to-the-S sense of shear (sample EM2, Fig. 6f). The shear sense, mineral parageneses, and *P-T* conditions are similar to those found in the southern part of the Selimiye nappe. Reconnaissance mapping in the eastern part of the Çine Massif revealed stretching lineations and folds, associated with an *S*₂ schistosity, with similar trends as in the western part of the study area, overprinting a primary schistosity (Fig. 1). Rotated garnets imply top-to-the-S shearing (Fig. 7c).

4. Metamorphic study of the Çine Massif

4.a. Analytical procedures

Pressure and temperature conditions were studied by analyzing 150 samples. Most samples were collected in the western part of the Çine Massif near Lake Bafa. Twelve samples originated from the eastern Çine massif near Karacasu (Figs 1, 4). Due to sparse distribution of index minerals the authors refrained from mapping mineral isograds. Sample localities are shown on Figures 1 and 4. Mineral parageneses are listed in Appendix 1. Individual mineral analyses were obtained with the JEOL Superprobe (JXA 8900RL) of the Johannes-Gutenberg-Universität Mainz, Germany using the following operating conditions: acceleration voltage of 15 kV, beam current of 15 nA, 20 s counting time per element. The following standards were used: wollastonite (Ca, Si), corundum (Al), pyrophanite (Ti), hematite (Fe), MgO (Mg), albite (Na), orthoclase (K), tugtupite (Cl), F-phlogopite (F), Cr₂O₃ (Cr), rhodochrosite (Mn) and ZnS (Zn). Matrix correction was done with a ZAF procedure. The mineral analyses listed in Table 1 are considered to be accurate within a relative error of 3%. Mineral formulae were calculated with the program AX by T. J. B. Holland (<http://www.esc.cam.ac.uk/astaff/holland/ax.html>).

Whole-rock XRF analyses from sample material remaining from thin section preparation, commonly 5×3×3 cm, were made in order to correlate microprobe analyses as precisely as possible with pseudo-section calculations computed with THERMOCALC (version 3.2.1) (Powell, Holland & Worley, 1998). Whole-rock XRF analyses are given in wt % and normalized to mol. % with Fe₂O₃ recalculated to FeO^{total} (Table 2). Mineral abbreviations used in text, figures and tables are adopted from Powell, Holland & Worley (1998) and are listed in Appendix 1.

Table 2. Whole-rock XRF analyses of selected metapelitic rocks of the Çine and Selimiye nappes of the western Menderes Massif

	GU1T	SE3	SE12	T6	Z12	Z8
Weight %						
SiO ₂	67.25	80.63	52.30	64.34	57.09	64.49
TiO ₂	0.90	0.69	1.14	1.03	1.08	1.50
Al ₂ O ₃	13.56	6.43	24.54	18.15	21.86	20.79
Fe ₂ O ₃	5.80	5.33	12.02	7.49	8.53	7.08
MnO	0.08	0.07	0.13	0.04	0.15	—
MgO	3.01	0.33	1.87	0.85	0.94	0.17
CaO	2.17	0.04	0.75	0.88	1.13	0.15
Na ₂ O	2.69	0.22	1.53	0.65	1.02	0.30
K ₂ O	1.67	1.03	1.42	1.43	2.61	1.81
P ₂ O ₅	0.22	0.06	0.05	0.08	0.15	0.10
Fluid	1.79	2.31	4.044	3.65	5.04	3.62
Total	99.14	97.14	99.79	98.59	99.60	100.01
Mol. %						
SiO ₂	69.40	82.15	54.32	66.02	57.39	66.36
TiO ₂	0.70	0.53	0.89	0.79	0.82	1.16
Al ₂ O ₃	8.25	3.86	15.02	10.98	12.95	12.61
FeO _{tot}	4.50	4.09	9.40	5.78	6.45	5.48
MnO	0.07	0.06	0.11	0.03	0.13	—
MgO	4.63	0.50	2.90	1.30	1.41	0.26
CaO	2.40	0.04	0.83	0.97	1.22	0.17
Na ₂ O	2.69	0.22	1.54	0.65	0.99	0.30
K ₂ O	1.10	0.67	0.94	0.94	1.67	1.19
P ₂ O ₅	0.10	0.03	0.02	0.03	0.06	0.04
H ₂ O	6.17	7.86	14.02	12.50	16.91	12.43
Total	100.00	100.00	100.00	100.00	100.00	100.00

4.b. Mineral chemistry of metasedimentary rocks of the Çine Massif

Garnets in metapelites and calc-schists of the Selimiye and Çine nappes are commonly almandine-rich (*X*_{alm} 60–90), with varying abundance of Mg (*X*_{prp} 1–10), Mn (*X*_{sps} 1–10) and Ca (*X*_{grs} 2–30) (Régner *et al.* 2003). Most garnet grains lack zoning (Fig. 8a). Patterns of zoned garnets display core to rim decrease of Ca and Mn and an increase of Mg and Fe (Fig. 8b). Ashworth & Evirgen (1984) and Whitney & Bozkurt (2002) report similar zoning patterns in metasedimentary rocks of the Selimiye nappe. Different zoning patterns are observed in garnets found as inclusions in albite from a pelitic gneiss in the eastern Çine Massif, where Ca and Fe increases from core to rim, while Mn decreases and Mg remains constant (sample R9, Fig. 8c). Garnet alteration to chlorite and pinite is stronger in metasedimentary rocks adjacent to the orthogneiss (Fig. 7c, Appendix 1), than in metasedimentary enclaves within the orthogneiss.

Similarly, biotite from the metasedimentary rocks is commonly chloritized (Ashworth & Evirgen, 1984). Primary chlorite occurs in most of the samples, but secondary chlorite ubiquitously overprints the main foliation or replaces garnet during kelyphytization (e.g. sample Z12). Muscovite forms a solid solution between paragonite, muscovite, celadonite and Fe-celadonite.

Chloritoid is in textural equilibrium with biotite and chlorite (samples SE3, T6, Z9; Fig. 7b, d), garnet–biotite–chlorite (samples SE12, EM2, GU9, Z13;

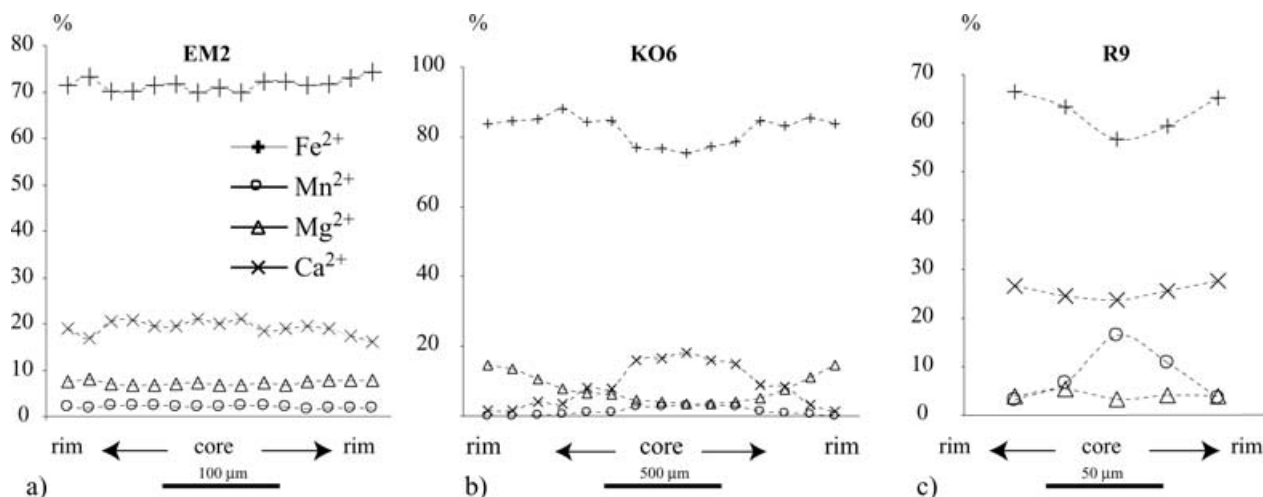


Figure 8. Representative compositional garnet transects from the Çine Massif. (a) Selimiye nappe, central Çine Massif (EM2). (b) Metasedimentary enclave within the orthogneiss, Çine nappe near Koçarlı (KO6). (c) Garnet inclusion in plagioclase from the western Çine Massif (R9). For locations of samples see Figures 1 and 4.

Fig. 9a, b), staurolite–chlorite (sample T6; Fig. 7a), and as inclusions in garnet (samples D6, EM2, GU9, GU10, KG3, KG6, Ki6, KO6, Z12; Fig. 9b, c; Appendix 1). In metasedimentary rocks of the Selimiye nappe X_{Mg} , based on a 6-oxygen structural formula, increases towards the contact with the orthogneiss from 0.13 to 0.21 (Régner *et al.* 2003). In the western part (samples T6, GU10, GU9) and in the eastern part (sample Z12) X_{Mg} ranges from 0.16 to 0.19. There is no significant difference in X_{Mg} between chloritoid as inclusion in garnet and in the matrix. In metasedimentary enclaves within the orthogneiss chloritoid inclusions in garnet have X_{Mg} as high as 0.22 (sample D6, Régner *et al.* 2003).

Plagioclase composition is near albitic in lower grade rocks, and oligoclase–andesine at higher metamorphic grades (Table 1; see also Régner *et al.* 2003). Significant zoning of plagioclase (oligoclase core and andesine rim) has been observed by Régner *et al.* (2003) solely in calc-schists, but Whitney & Bozkurt (2002) reported this from metapelites. Furthermore, oligoclase is observed in apparent textural equilibrium with albite (peristerite gap: Ashworth & Evirgen, 1985; Régner *et al.* 2003).

Staurolite is in textural equilibrium with chloritoid and chlorite in the western part of the Çine Massif (sample T6; Fig. 7a). Régner *et al.* (2003) reported the staurolite–chloritoid paragenesis (textural equilibrium) as inclusions in garnet south of Koçarlı, but never in matrix. A small porphyroblast of staurolite is observed in equilibrium with chlorite and garnet (sample GU10; Fig. 9c). In both cases staurolite is enriched in Zn (≤ 5 wt %). In comparison, a staurolite from the schist enclaves south of Koçarlı has a significantly lower Zn content (~ 0.2 wt %) (sample KO6; Fig. 1, Table 1).

Kyanite is a common phase in the schist enclaves south of Koçarlı (sample D6; Régner *et al.* 2003). It is in equilibrium with staurolite–biotite and garnet. Sample GU1T provides the key paragenesis kyanite–biotite–garnet, without staurolite (Fig. 7e). This study reports the first observations of kyanite in equilibrium with muscovite–ilmenite in the surrounding metasedimentary rocks (Selimiye nappe, samples Z7, Z8; Fig. 9d).

Amphibole in calc-schists near Selimiye coexists with garnet–biotite–zoisite–muscovite–plagioclase–chlorite–carbonate (Whitney & Bozkurt, 2002; sample A11 in Régner *et al.* 2003). Actinolite has been described within low-grade calc-schists near Milas (Fig. 1; Ashworth & Evirgen, 1985). In addition, near Çalıköy small calcic amphibole porphyroblasts were observed as inclusions in plagioclase in pelitic gneiss (samples R9, TE9; Appendix 1). In the metasedimentary enclaves south of Koçarlı amphibolite consists mainly of calcic amphibole and biotite ($\pm$ epidote $\pm$ sphene) (samples GU4T, GU6T, GU8T, GU9T; Appendix 1). The compositional range of amphibole is shown on a Si–Mg/(Mg + Fe²⁺) diagram (Fig. 10) after Leake *et al.* (1997). Sample R9 plots near the transition ferrohornblende–magnesianhornblende, whereas sample GU6T is located near the transition magnesianhornblende–tschermakite, indicating a higher Al content on the T1 tetrahedral site (Table 1). A similar composition to GU6T has been observed near the orthogneiss north of Selimiye (sample A11 in Régner *et al.* 2003).

Zoisite and clinozoisite are common epidote phases in the calc-schists. Epidote also occurs in metapelite enriched in Fe³⁺ and Ca (samples T6, Z12, Appendix 1). Common accessory minerals are sphene, ilmenite, hematite, rutile, tourmaline, apatite and zircon. Some

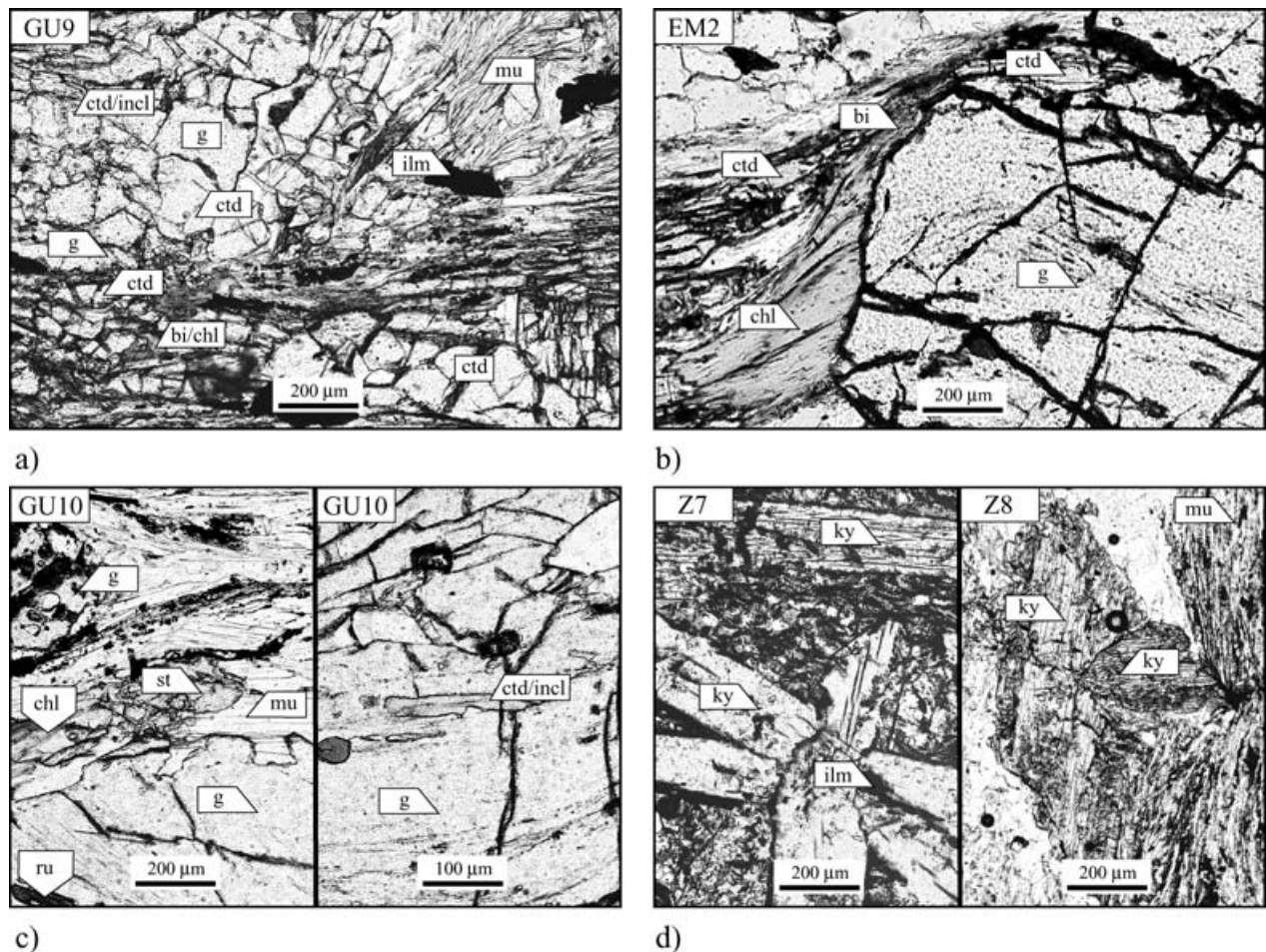


Figure 9. Photomicrographs of mica schist samples from the Çine Massif. (a) Western zone: garnet–chlorite–biotite–chloritoid paragenesis. Chloritoid forms inclusions in garnet. Biotite and chlorite occur as traces. PPL. (b) Central zone: chlorite–biotite–garnet–chloritoid paragenesis. Biotite occurs as trace. PPL. (c) Western zone: staurolite–chlorite–garnet paragenesis (left) and chloritoid inclusion in garnet (right). PPL. (d) Eastern zone: kyanite–ilmenite–muscovite paragenesis. Kyanite displays polysynthetic macles. PPL. Sample numbers in the upper left corners refer to locations on Figures 1 and 4. For mineral abbreviations see Appendix 1.

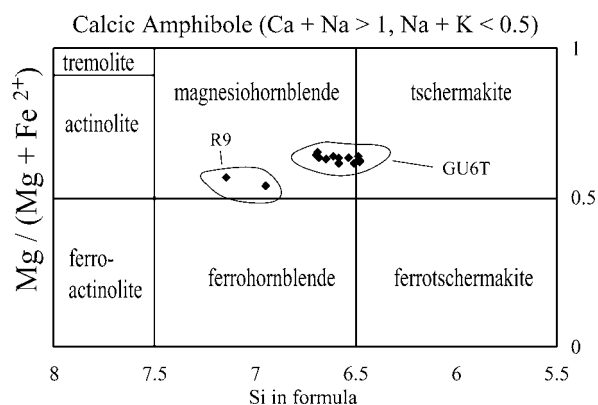


Figure 10. Si^{total} – $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ diagram (Leake *et al.* 1997) of amphiboles from the western Çine Massif (R9) and the meta-sedimentary enclave of the Çine nappe (GU6T). Lower amphibolite facies R9 plots near the transition ferrohornblende–magnesiohornblende and amphiboles from the upper amphibolite facies sample GU6T plot near the transition magnesiohornblende–tschermakite. The regional pressure and temperature increase correlates with Al content (8–Si) in the T1 tetrahedral site.

samples, e.g. Z12, contain stilpnomelane overgrowing the main foliation, as a result of retrograde metamorphism or alteration.

4.c. Implications of mineral composition and parageneses

Similar zoning patterns in garnets of the metasedimentary rocks of the Selimiye nappe and enclaves in the Çine orthogneiss have also been observed in the central Menderes Massif near Ödemiş (Ashworth & Evirgen, 1985). In some samples bulk rock chemistry controls zoning. Absence of observed inherited garnet cores argues against polymetamorphism (Ashworth & Evirgen, 1984, 1985).

In a pure KFMASH system the paragenesis staurolite–chlorite–garnet results from the breakdown of chloritoid (Spear & Cheney, 1989; Powell, Holland & Worley, 1998). Abundant Zn stabilizes staurolite at lower temperatures than pure Fe-staurolite (Soto & Azañón, 1994). This is evident in sample GU10 where

staurolite is in equilibrium with garnet and chlorite. Nearby sample GU9 still contains abundant chloritoid in equilibrium with garnet. Finally, the internally consistent thermodynamic dataset used by THERMOCALC (see Section 5.a) does not take the Zn end-member into account, and therefore we will assume in calculation that $\text{Fe}^{2+} = \text{Fe}^{2+} + \text{Zn}$ for staurolites of GU10 and T6.

The occurrence of the garnet–biotite–kyanite paragenesis in metasedimentary enclaves within the orthogneiss south of Koçarlı (sample GU1T) implies conditions above 630°C and 8 kbar in the KFMASH system (Spear & Cheney, 1989; Powell, Holland & Worley, 1998). Other parageneses in metasedimentary enclaves include biotite–kyanite–staurolite, conspicuously lacking chloritoid in the matrix. In contrast, the metasedimentary rocks east of the Çine Massif are characterized by stable chloritoid in the kyanite field (samples Z7, Z8). The presence of chloritized biotite in low-grade metapelites complicates conventional thermodynamic considerations and does not allow projection of biotite in an AFM diagram (+ muscovite + quartz + H₂O).

Hornblende from schist of the upper amphibolite facies (GU5T) has lower Si-number than that from lower amphibolite facies metapelite (R9), suggesting a correlation of increasing Al content on the T1 tetrahedral site (decreasing Si-number) of calcic amphibole with increasing metamorphic grade (Fig. 10, Table 1), as proposed by Johnson & Rutherford (1989) and Blundy & Holland (1990). Although the erosional level in the study area is non-isobaric, the Al content in the T1 site can be considered a good approximation of a general increase of the *P–T* conditions, which is also observed near the orthogneiss north of Selimiye (Régner *et al.* 2003).

The sequence of mineral parageneses in the metapelitic rocks around the Çine Massif displays always the same pattern: chlorite–biotite (within the kyanite stability field, cf. samples Z7, Z8) at low grade, followed by chlorite–biotite–chloritoid or chloritoid–staurolite–chlorite, and then garnet–biotite–chloritoid–chlorite at higher grade (Ashworth & Evirgen, 1984). These mineral isograds are mapped in detail by Régner *et al.* (2003) in the Selimiye nappe. In the western part of the Çine Massif isograd mapping is nearly impossible because of the lack of suitable critical parageneses. However, at first approximation three main metamorphic zones can be distinguished from observed parageneses (see Appendix 1): a chlorite–biotite zone, a lower amphibolite facies garnet zone and an amphibolite facies zone near the orthogneiss, more or less correlative with the occurrence of staurolite in sample GU10 (Fig. 4).

Conspicuously absent in the metasedimentary rocks of the Çine Massif are typical contact metamorphic parageneses, e.g. cordierite–andalusite and garnet–cordierite, supporting the notion of a Barrovian-

type metamorphism. A similar type of metamorphism and associated sequences of parageneses have been reported from the central Menderes Massif around Ödemiş (chloritoid, staurolite–kyanite, sillimanite, kyanite) by Evirgen & Ataman (1982), and south of Dermici (staurolite–andalusite–garnet–biotite, staurolite–garnet–kyanite–sillimanite, garnet–kyanite–sillimanite–biotite) by Candan & Dora (1993).

Index minerals, the sequence of mineral parageneses, and the staurolite–chloritoid paragenesis associated with top-to-the-N sense of shear around the Çine Massif and as inclusions in garnet of the metasedimentary enclaves, all point to a single metamorphic event affecting metasedimentary rocks of the Selimiye and Çine nappes. In contrast to Catlos & Çemen (2005), our observations do not lend support to polymetamorphism, based either on garnet zonation or on the paragenetic study of our samples. In accordance with Whitney & Bozkurt (2002) and Régner *et al.* (2003) we find no evidence for polymetamorphism in the Menderes Massif and suggest coeval development of lower to upper amphibolite facies metamorphism.

5. Pseudosections, *P–T* estimates, and *P–T* path

5.a. Applied systems

P–T conditions of individual samples are estimated by means of pseudosections computed with THERMOCALC (version 3.2.1), using the May 2001 update of the internally consistent thermodynamic data set of Holland & Powell (1998). Models used for THERMOCALC calculations are described in Appendix 2.

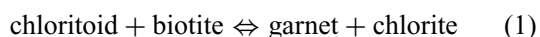
MnO is negligible on first approximation in all analysed samples without garnet. TiO₂ is relatively abundant, and mainly incorporated in ilmenite (e.g. samples SE12, Z12, Z8, T6; Table 1). White *et al.* (2000) show that large amounts of TiO₂ in metapelite have a restricted effect on KFMASH phase equilibria. Therefore, it is assumed at first approximation that all opaques are 'in excess'.

Five representative samples (SE3, SE12, T6, Z12, GU1T) were selected for calculation of *P–T* pseudosections in the KFMASH (SE3, T6), NCKFMASH (T6) and MnNCKFMASH (SE12, Z12, GU1T) systems. The NCKFMASH system is suitable for samples lacking garnet, but with epidote, albitic plagioclase and paragonite end-members participating in muscovite solid solution (sample T6). The MnNCKFMASH system is applied when an additional garnet phase is present (SE12, Z12, GU1T). Since minor additional MnO can have a strong influence on garnet growth it is considered as well (Droop & Harte, 1995; Tinkham, Zuluaga & Stowell, 2001). Low abundance of K₂O, Na₂O and CaO prohibits projection from muscovite or plagioclase throughout *P–T* space (Table 2). Isopleths of phengitic substitution help in some cases to constrain

pressure or temperature conditions (Wei & Powell, 2003, 2004). In addition, garnet isopleths (core) are used to constraint initial P – T conditions of garnet growth. For plagioclase the model 4T (C1 structure) of Holland & Powell (1992) was chosen, since the anorthite component of plagioclase in this study does exceed An_{30} . Other activity models used are listed in Appendix 2.

5.b. Sample SE3, southern Selimiye nappe

This sample from the chloritoid–biotite–chlorite zone north of Selimiye is characterized by lack of garnet and low content of CaO , Na_2O and MnO (Table 2). Abundant SiO_2 (> 80 wt %) is due to a quartz-rich layer in the sample material used for whole-rock analyses. Since pseudosections are projected from quartz this has no influence on thermodynamic calculations, although a loss of accuracy has to be taken into account. The paragenesis chloritoid–chlorite–biotite (+ muscovite + quartz) is observed in the pelitic part of the thin section (Fig. 7d). P – T conditions in pseudosection for the observed chloritoid–chlorite–biotite (+ muscovite + quartz + H_2O) paragenesis modelled in the KFMASH system are in the order of 475–550 °C and 1–5 kbar (Fig. 11). However, the stability field of this paragenesis may be wider, due to the large uncertainties at low pressure and the inability of THERMOCALC to compute a large Fe-rich field in the KFMASH system (Roger Powell, written communication, 2003). Phengitic substitution (Si^{total} in muscovite) ranges from 3.10 to 3.23 with an average of 3.15 (Table 1). Since values above 3.10 are considered unstable, the lowest value was used for calculation of an AFM diagram, where the whole-rock composition of the sample with corresponding calculated phases plots at 4.5 kbar and 540 °C (Fig. 11, inset). Possible interpretation is a decompressional P – T path and/or heterogeneity of the bulk composition at small scale. Chloritized biotite prevents comparison between calculated and observed phases. However, in the KFMASH system the chlorite–chloritoid–garnet paragenesis occurs at 5 kbar and 540 °C by the univariant reaction (Fig. 11):



5.c. Sample SE12, southern Selimiye nappe

This sample originates 1.5 km north of SE3 and contains the paragenesis chlorite–biotite–garnet–chloritoid–muscovite–quartz (Fig. 1; Appendix 1, Régnier *et al.* 2003). Plagioclase (An_{14}) is rare and not in equilibrium with garnet or chloritoid (plagioclase–muscovite–quartz paragenesis). Epidote occurs as an accessory mineral (Table 1). P – T conditions were estimated with pseudosections in the NCKFMASH and MnNCKFMASH systems (Fig. 12). Both systems contain a part that cannot be calculated at low T and

medium P , due to the fact that zoisite is not predicted as a stable phase. The chloritoid–chlorite–garnet (+ muscovite + plagioclase + quartz + H_2O) paragenesis is stable at 8 kbar and 560 °C in the NCKFMASH system and at 6–8 kbar and 525–550 °C in the MnNCKFMASH system (Fig. 12). Stability of biotite with garnet, chlorite and chloritoid is not predicted in either system. Phengitic substitution ranges from 3.00 to a maximum of 3.12 (Table 1), yielding maximum P – T conditions of 7.5 kbar and 550 °C. The pressure estimates are considered too high, since phengitic substitution averages of 3.03 for muscovite indicate low or medium pressure. Furthermore, the large grain size of garnets, approximately 0.5 cm, implies significant fractionation of the bulk rock composition. Absence of plagioclase in textural equilibrium with garnet may also result from fractionation of the bulk rock composition, since bulk rock contents of CaO (0.83 mol. %) and Na_2O (1.54 mol. %; Table 2) are low. Ca is preferentially concentrated in garnet during nucleation, while Na is incorporated mainly in muscovite, preventing coexisting plagioclase and garnet. Presence of sporadic paragonite is also possible.

The garnet core composition (X_{alm} 68, X_{grs} 22, X_{prp} 5 and X_{sps} 5) is considered to represent chemical conditions of initial garnet growth. Calculated $X_{Fe-Mg-Ca}$ isopleths for such a garnet composition intersect at approximately 8 kbar and 530 °C (Fig. 12b, inset), which correlates well with maximum phengitic values. We consider this as the P – T condition of initial garnet growth, since non-negligible fractionation due to garnet nucleation will change effective bulk rock composition in the KFMASH, NCKFMASH or MnNCKFMASH systems. Within the classic scheme of Barrovian-type metamorphism, these conditions correspond to P_{max} followed by decompression and equilibrium at T_{max} in the matrix. P – T conditions at T_{max} are likely those predicted by the KFMASH system (+ quartz + muscovite + H_2O) for the univariant reaction (1) at 5 kbar and 540 °C (Fig. 12b), coinciding with the chloritoid–chlorite–biotite paragenesis of sample SE3 discussed above. A possible P – T path for SE12 is shown in Figure 12b.

5.d. Sample T6, western Selimiye nappe

This sample from the western Çine Massif displays the following textural equilibria: chlorite–chloritoid–staurolite–muscovite–quartz, chloritoid–biotite–chlorite–muscovite–quartz and epidote–chloritoid–muscovite–quartz (Table 1, Appendix 1). It mainly consists of chloritoid (80 vol. %), while plagioclase (An_{32}) is rare and not in equilibrium with staurolite or epidote (plagioclase–muscovite–quartz paragenesis). In the KFMASH petrogenetic grid of Powell, Holland & Worley (1998), stable parageneses of chlorite–chloritoid–staurolite and chlorite–chloritoid–biotite are restricted to relatively low pressure by reactions (1)

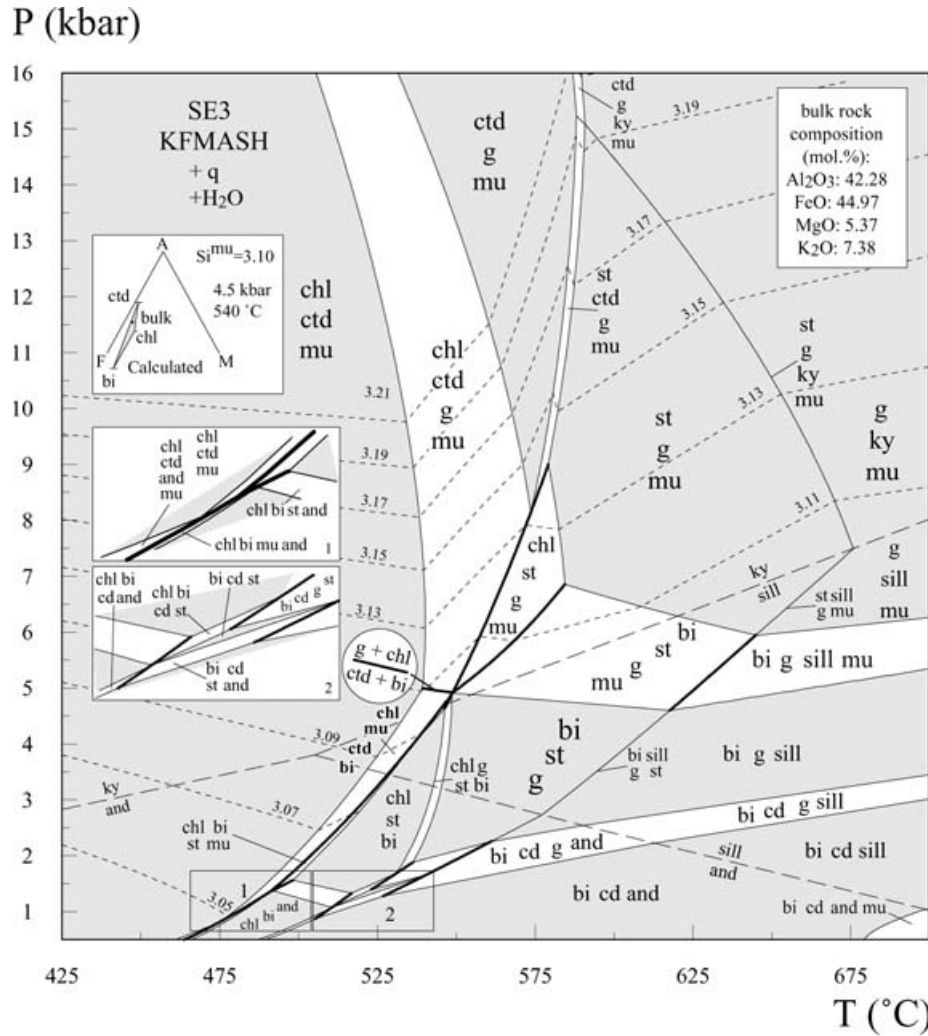
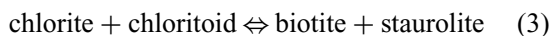
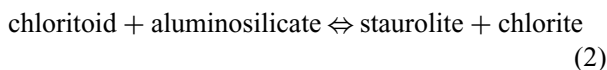


Figure 11. P – T pseudosection of SE3 (southern Selimiye nappe) in the KFMASH system (+ quartz + H_2O) with detailed sections enlarged (insets). The position of the reaction $ctd + bi = g + chl$ is obtained from the 2001 update of the dataset of Holland & Powell (1998). Phengitic substitution (Si^{total}) ranges from 3.10 to 3.23 (Table 1), P – T estimates for the observed chlorite–biotite–chloritoid paragenesis are approximately 4.5 kbar and 540 °C for Si^{total} in muscovite = 3.10. Insets show AFM diagram of calculated phases and bulk whole-rock composition. Chloritized biotite prevents representation of observed phases in the AFM diagram. Muscovite with high Si value is considered metastable. Heterogeneity of bulk composition and/or P – T path in decompression is envisioned.

and (Régner *et al.* 2003):



In the KFMASH grid of Spear & Cheney (1989) these parageneses are stable over a wider pressure range. Observations of natural assemblages corroborate the stability of the chlorite–chloritoid–biotite paragenesis in the andalusite and kyanite stability fields (Spear & Cheney, 1989; Likhanov *et al.* 2001; Likhanov, Reverdatto & Selyatitski, 2004, 2005). The discrepancy between observed and calculated stability fields originates from the uncertainty of the location of the invariant point [cd, als] (Powell, Holland & Worley, 1998) and the slope of reaction

(1). To constrain pressure estimates the phengitic substitution of muscovite was calculated with THERMOCALC and correlated with microprobe analyses. Si^{total} in muscovites from this sample ranges from 3.00 to 3.13 and averages 3.08 (Table 1). A calculated AFM diagram using the bulk whole-rock composition plots the chlorite–chloritoid–staurolite paragenesis at 530 °C and 5 kbar (Fig. 13, inset). However, the P – T pseudosection in KFMASH does not show a stable field for chloritoid–biotite–chlorite. This could be explained by small-scale spatial variation in the whole-rock composition. The presence of coexisting plagioclase and epidote allows calculation of a NCKFMASH pseudosection. It differs little from the KFMASH pseudosection at moderate pressures (Fig. 14). The chlorite–chloritoid–staurolite paragenesis is stable at 530 °C and 5 kbar, which also correlates

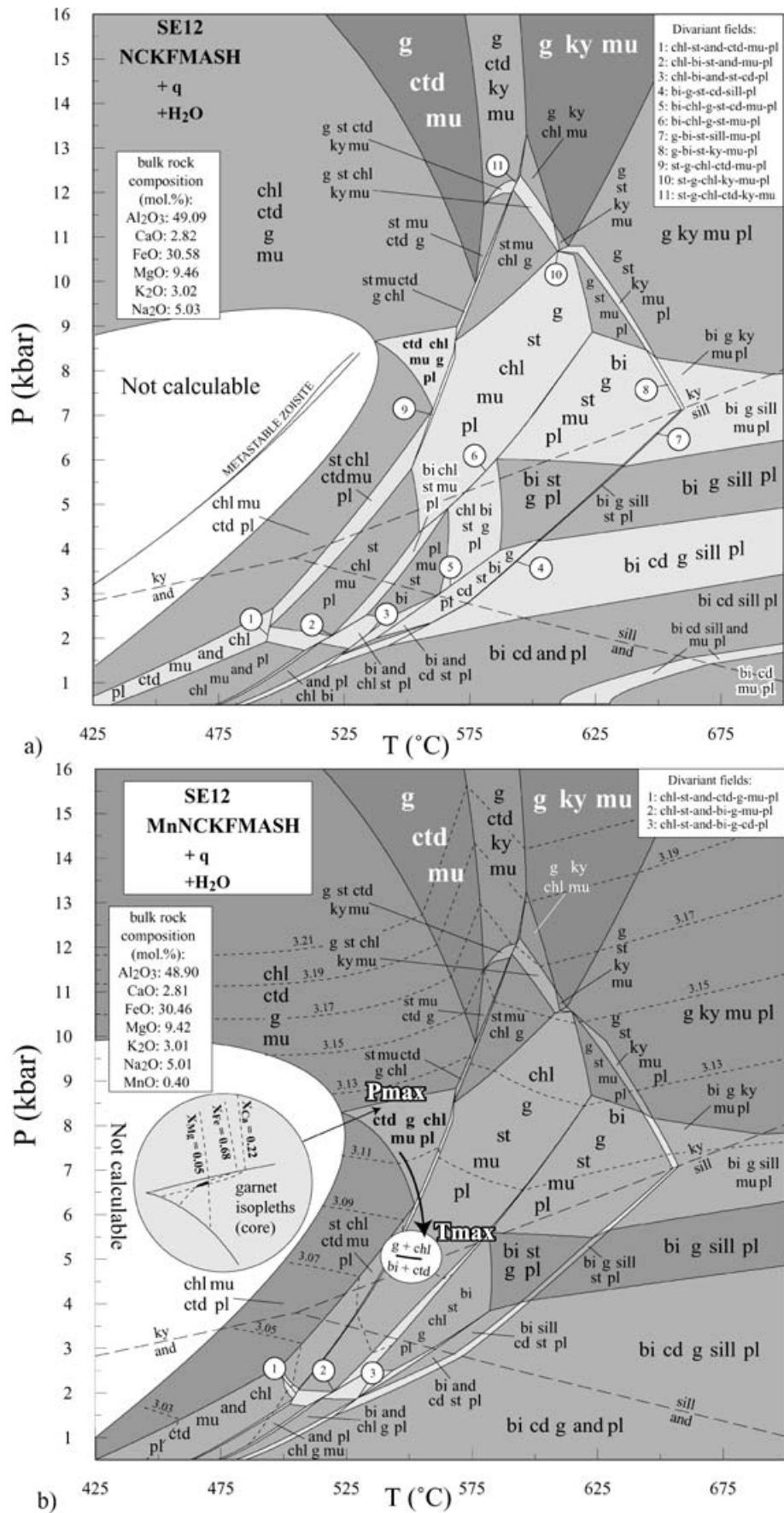


Figure 12. For legend see facing page.

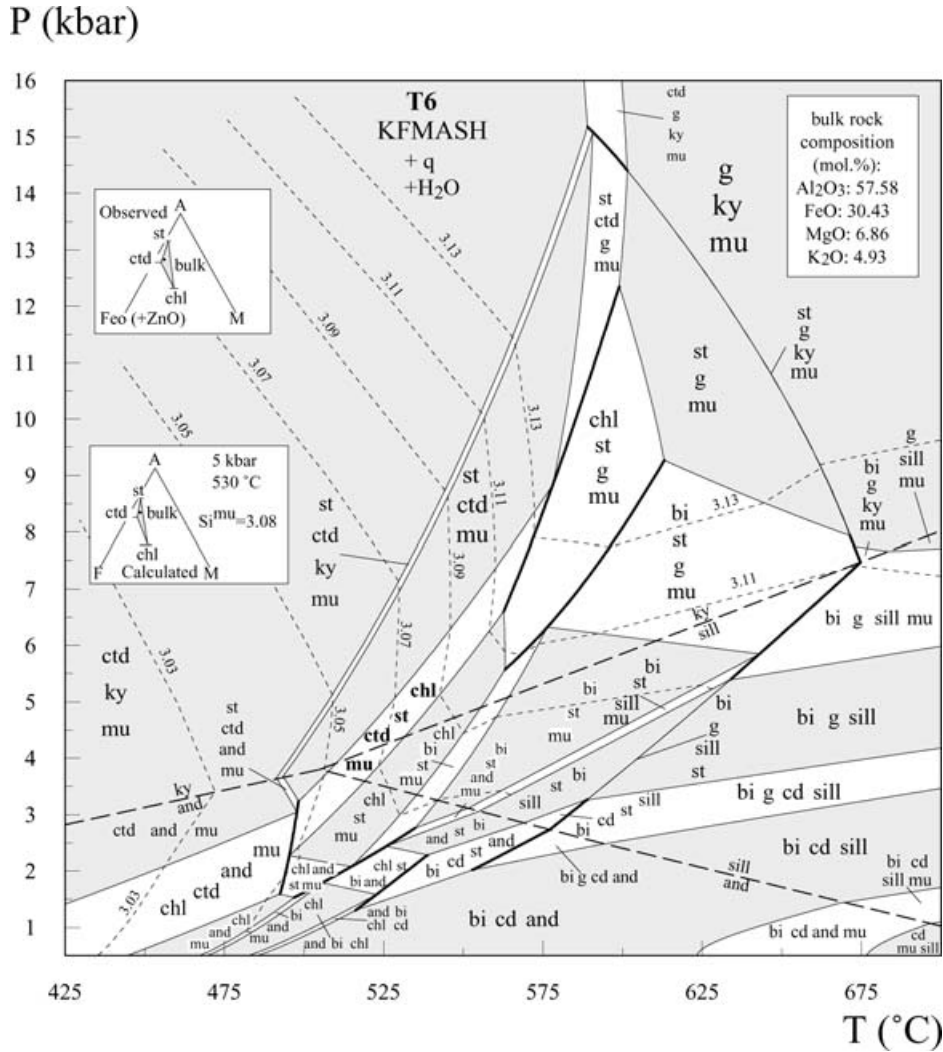


Figure 13. P - T pseudosection of sample T6 (western Çine Massif) in the KFMASH system (+ quartz + H_2O). Insets show AFM diagrams of observed and calculated phases and bulk whole-rock composition. FeO has been corrected in staurolite for the amount of ZnO. Phengitic substitution (Si^{total}) ranges from 3.00 to 3.13 (Table 1). P - T estimates for chlorite–staurolite–chloritoid are approximately 5 kbar and 530 °C for Si^{total} average of 3.08 for muscovite. The P - T pseudosection does not account for stability of observed chlorite–biotite–chloritoid paragenesis. Zoisite stability cannot be calculated in this system (see Fig. 14). Muscovite with high Si value is considered metastable. Heterogeneity of bulk composition and/or P - T path in decompression is envisioned.

well with the phengitic substitution average of 3.08 (Fig. 14a). Abundant Zn in staurolite may extent the stability field to lower temperatures (Table 1). Zoisite–chloritoid–chlorite–muscovite–plagioclase is stable at relatively high pressures (7.5 kbar) for Ab_{70} in plagioclase (Fig. 14b). However, growth of epidote and

plagioclase (e.g. Fe^{3+} , Ca^{2+}) may have resulted from small-scale variation of the bulk rock composition. Since Fe^{3+} is not considered in the NCKFMASH system, epidote growth conditions cannot be computed. The crucial paragenesis chlorite–chloritoid–staurolite–muscovite–quartz (+ plagioclase, not in equilibrium

Figure 12. Pseudosections of SE12 from the southern Selimiye nappe. (a) NCKFMASH system (+ quartz + H_2O): the chloritoid–chlorite–garnet–plagioclase–muscovite paragenesis is stable around 8 kbar and 560 °C. (b) MnNCKFMASH system (+ quartz + H_2O): the chloritoid–chlorite–garnet–plagioclase–muscovite paragenesis is stable around 6–8 kbar and 525–550 °C. Phengitic substitution ranges from 3.00 to a maximum of 3.12 (Table 1), leading to maximum P - T conditions around 7.5 kbar and 550 °C. Similar P - T conditions are deduced from calculated/observed isopleth intersection for garnet core. Garnet growth in this sample likely involves fractionation of bulk rock composition leading to a paragenesis at T_{max} better described in the KFMASH system. The pseudosection in the MnNCKFMASH system is therefore suitable for initial P - T conditions of garnet growth whereas the KFMASH system seems more suitable to explain the presence of biotite at T_{max} . A decompressional P - T path accounts for biotite in textural equilibrium with garnet–chlorite–chloritoid in matrix.

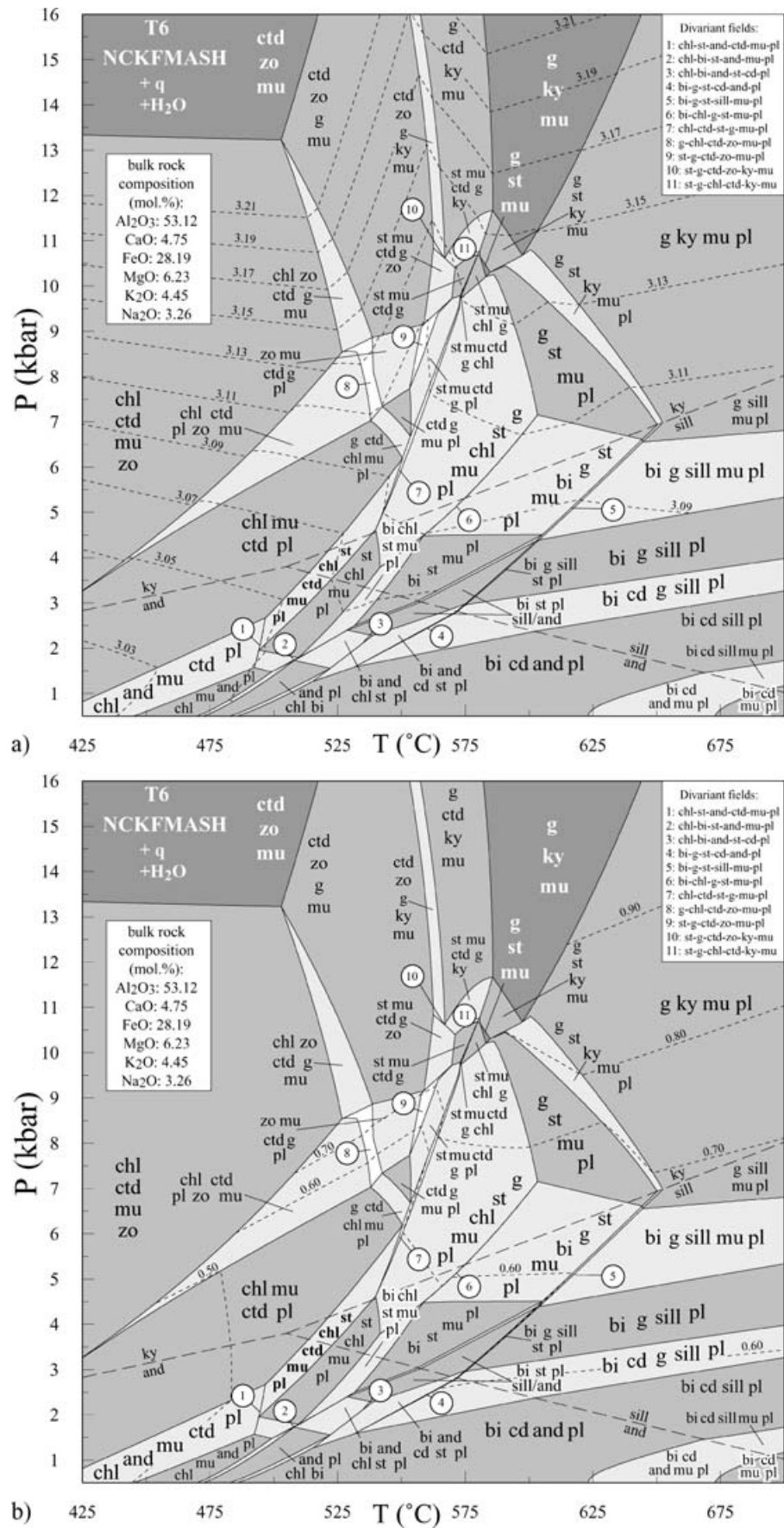


Figure 14. For legend see facing page.

here) corresponds to P – T conditions in the kyanite field near the aluminosilicate triple point, which is near reaction (1) within the KFMASH system as well.

5.e. Sample Z12, eastern Selimiye nappe

This sample from the Selimiye nappe in the eastern part of the Çine Massif is characterized by alteration of garnet to chlorite and pinite. Garnet relics have compositions (X_{alm} 66, X_{grs} 22, X_{prp} 6, X_{spss} 5; Table 1) similar to garnet cores of samples SE12 and EM2 (Fig. 8a). It is possible that Mn in the garnet core can stabilize the grain at low temperatures even after considerable retrogression. The observed parageneses are chlorite–chloritoid–garnet–biotite–muscovite–quartz and chloritoid–epidote–muscovite–quartz (Table 1, Appendix 1). Chlorite and biotite occur as traces. Almost complete kelyphitization and pseudomorphism of garnet by secondary chlorite can be observed. Plagioclase was not observed in thin section. With the given bulk rock composition, calculations show that the chloritoid–garnet paragenesis without staurolite or kyanite does not exist in the KFMASH system. A more suitable system for garnet-bearing samples is MnNCKFMASH (Fig. 15a). This does not predict stable biotite. Observed parageneses correspond to the quadrivariant field of chlorite–chloritoid–garnet–zoisite (+ muscovite + quartz + H_2O) without plagioclase, which covers a pressure range of 8–16 kbar (Fig. 15a). Epidote may be stabilized by Fe^{3+} at medium pressure. The absence of plagioclase may result from fractionation of the bulk rock composition during nucleation of large garnets (c. 0.5 cm; Fig. 7c). Phengitic substitution values in muscovite (3.12–3.17) correspond to pressures of 7–10 kbar, with an average of 3.15 corresponding to 8.5 kbar for T_{max} at 525 °C. Garnet isopleths form a field in P – T space at 6–9 kbar and 525 °C (Fig. 15a, dark grey field). Almost total garnet kelyphitization indicates retrograde metamorphism beyond garnet field stability (Fig. 15a, white arrow).

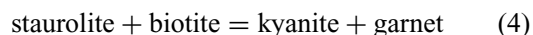
5.f. Sample Z8, eastern Selimiye nappe

This sample from the overlying metasedimentary rocks SW of Karacasu contains synkinematic kyanite porphyroblasts in textural equilibrium with muscovite, ilmenite/hematite and quartz (Fig. 9d, Appendix 1). The presence of previously unreported kyanite provides important constraints on the pressure estimation. A P – T pseudosection cannot be calculated since the sample lacks a suitable paragenesis in KFMASH.

Whole-rock analysis shows a non-negligible content of TiO_2 (Table 2), which suggests that most of FeO – Fe_2O_3 is incorporated in iron oxides (ilmenite, hematite), suppressing growth of other Al–Fe silicates (White *et al.* 2000). First order estimation using the phengitic substitution in muscovite ($\text{Si}^{\text{total}} = 3.13$; Table 1) suggests similar pressure conditions as for sample Z12.

5.g. Sample GU1T, Çine nappe

This sample from the metasedimentary enclaves within the orthogneiss of the Çine Massif SW of Koçarlı contains the paragenesis garnet–kyanite–biotite (Fig. 7e). Staurolite is absent, muscovite occurs as traces, oligoclase is not zoned. Due to low abundance of K_2O biotite and muscovite do not coexist. This paragenesis indicates P – T conditions of 630 °C and more than 8 kbar in the KFMASH petrogenetic grid (Spear & Cheney, 1989; Powell, Holland & Worley, 1998). A pseudosection in the MnNCKFMASH system, possible due to significant contents of Na_2O and CaO , indicates conditions of 8–9 kbar and above 630 °C for the biotite–garnet–kyanite–plagioclase ($\pm$ muscovite) field (Table 2, Fig. 15b). This is in good agreement with 600–630 °C and 8–11 kbar estimated for the concomitant staurolite–biotite–kyanite zone south of Koçarlı (samples D1, D30, D42; Appendix 1). However, the boundary between these zones is marked by the staurolite-out isograd. The corresponding univariant reaction in the KFMASH system (Régner *et al.* 2003):



occupies a narrow biotite–garnet–staurolite–kyanite–plagioclase quadrivariant field in MnNCKFMASH in a similar P – T space.

6. P – T paths, metamorphic gradients and tectonic model of the Menderes Massif

Metasedimentary enclaves in orthogneiss of the western Çine massif record temperatures of 600–640 °C and pressures around 8–10 kbar (e.g. sample GU1T), similar to samples obtained south of Koçarlı (D1, D6, D30, D42; Régner *et al.* 2003). Metasedimentary rocks (T6) west of, and 5 km structurally underneath, the orthogneiss experienced lower temperatures (525–540 °C) and pressures (5–6 kbar). This results in an inverted temperature field gradient of approximately 25 °C/km (Fig. 3, inset). The difference of 2–3 kbar across the contact between the orthogneiss of the Çine

Figure 14. P – T pseudosections of sample T6. (a) NCKFMASH system (+ quartz + H_2O) with calculated isopleths for Si in muscovite (b) and albite content in plagioclase. Stability of observed chlorite–chloritoid–biotite–muscovite–quartz paragenesis is not predicted. P – T conditions for the stable chlorite–chloritoid–staurolite–muscovite (+ plagioclase, not in equilibrium in thin section) paragenesis are around 530 °C and 5 kbar and correlate well with average of phengitic substitution ($\text{Si}^{\text{total}} = 0.8$). P – T conditions are near the reaction $\text{bi} + \text{ctd} = \text{g} + \text{chl}$ in the KFMASH system (see Fig. 11).

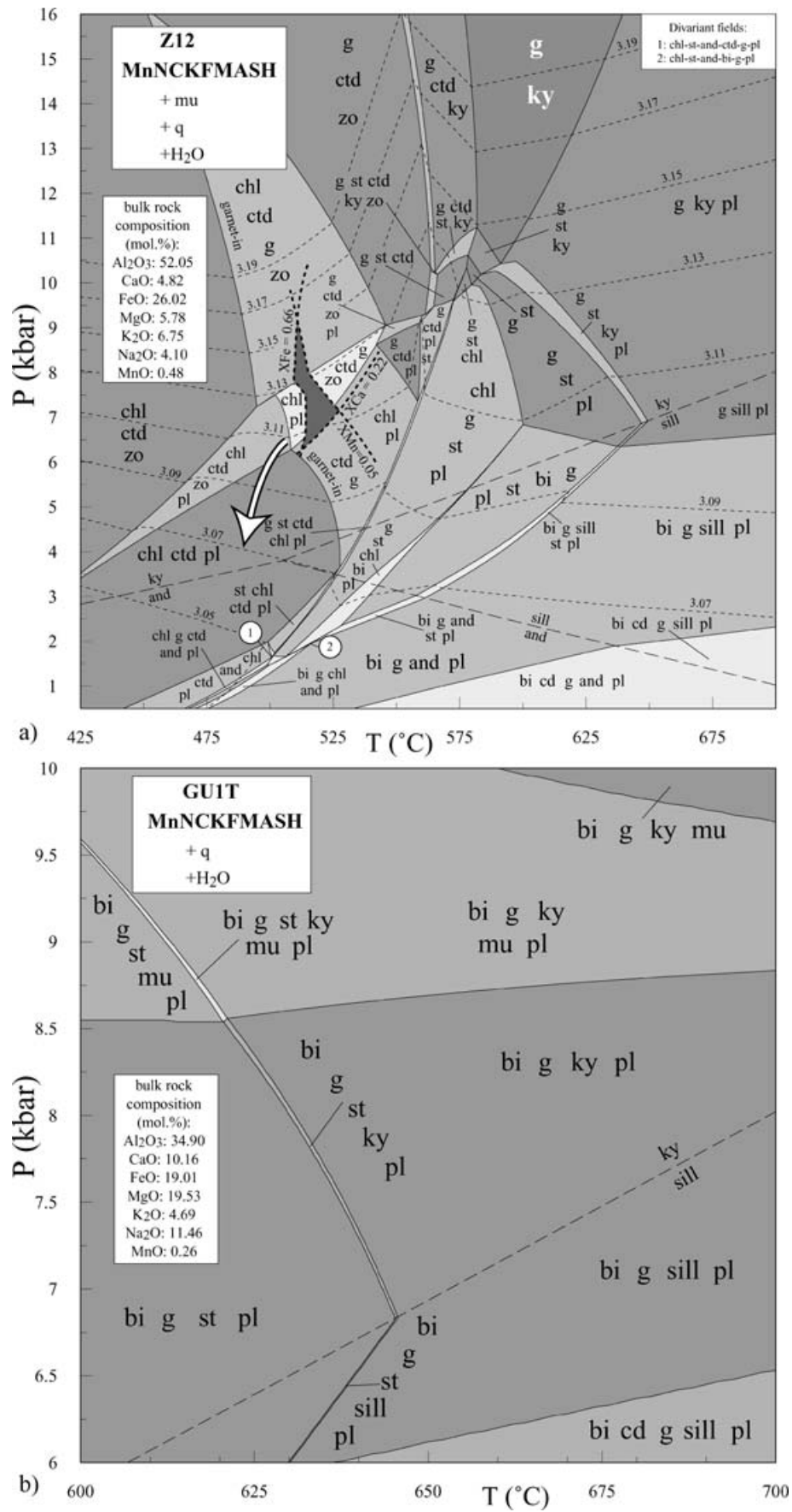


Figure 15. For legend see facing page.

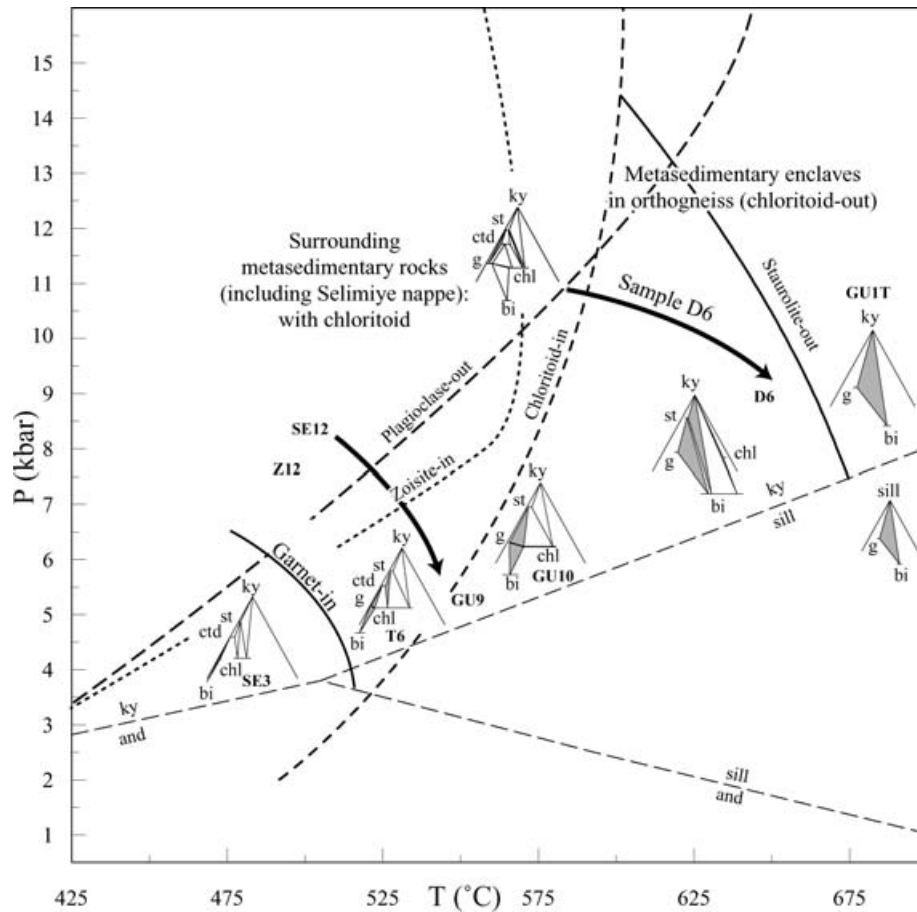


Figure 16. Summary of parageneses observed in the Çine Massif (see also Régner *et al.* 2003). The KFMASH system has been chosen in order to simplify the representation of parageneses with calculated AFM diagrams. Observed parageneses are shown in grey. The sequence of parageneses describes a typical Barrovian field gradient. Chloritoid-in and staurolite-out isograds are well constrained independently of chemical system. The location of other isograds depends on whole-rock analyses of individual samples. Clockwise P - T path is deduced from pseudosections and paragenetic analysis in KFMASH system. Surrounding metasedimentary rocks are characterized by the stability of chloritoid, which had disappeared completely in the metasedimentary enclaves. Sillimanite occurs in the central Menderes Massif near Tire (Evirgen & Ataman, 1982) and in the North Menderes Massif, south of Dermici (Candan & Dora, 1993, Bozdağ nappe). Black arrows indicate P - T path.

nappe and metasedimentary rocks of the Selimiye nappe implies that approximately 7 km of crust is missing throughout the Selimiye shear zone.

The observed parageneses are summarized in Figure 16 with AFM diagrams of the KFMASH system, which works well for 'normal' metapelitic rocks. Only small differences in pseudosections were observed by using different chemical systems, with the exception of the Mn-sensitive garnet stability field. Independent of the chemical system, excellent thermometers are provided by the chloritoid-out and staurolite-out iso-

grads (Wei & Powell, 2003). Parageneses within the orthogneiss and surrounding metasedimentary rocks, including the Selimiye nappe, fall on the 'typical' geotherm for 'normal' continental crust, characteristic of Barrovian-type metamorphism (Spear, 1993, pp. 37–9). Evidence of an earlier high-pressure metamorphism is provided by chloritoid and chloritoid–staurolite inclusions in garnets from the Çine nappe (sample D6), implying a clockwise P - T path. A similar path is suggested for metasedimentary rocks of the Selimiye nappe (sample SE12).

Figure 15. (a) Sample Z12 from the eastern Selimiye nappe near Karacasu. P - T pseudosection in MnNCKFMASH system (+ μ + quartz + H_2O). According to the phengitic substitution average (S_i^{total} around 3.15; Table 1) and garnet–chloritoid–chlorite equilibrium, P - T estimates are in the order of 8.5 kbar and 525 °C. Calculated–observed (core?) garnet isopleths yield P - T conditions around 6–8 kbar and 525 °C (see intersection field). Retrogression produces almost complete kelyphitization of garnet in chlorite–chloritoid–muscovite ($\pm$ plagioclase $\pm$ epidote) field (see arrow). (b) Sample GU1T from the northern Çine nappe metasedimentary enclave. P - T pseudosection in MnNCKFMASH system (+ quartz + H_2O). Traces of muscovite approximate pressures of 8–9 kbar at a temperature above 630 °C. Note that the garnet–biotite–kyanite ($\pm \mu \pm pl + q + H_2O$) paragenesis implies, as in the KFMASH system, a complete breakdown of staurolite.

If metamorphism of the Çine nappe were older than in the overlying Selimiye nappe, where chloritoid is stable, as proposed by Gessner *et al.* (2001b), retrograde metamorphism within the chloritoid stability field should be observed within the metasedimentary enclaves of the Çine nappe. However, despite a sufficient amount of water available, such retrograde overprinting of the staurolite–kyanite–biotite parageneses is not observed. Thus the metamorphic parageneses are consistent with a single metamorphic event and can be regarded as coeval for the reconstruction of the deformation history (Fig. 17). General top-to-the-N thrusting of the Çine orthogneiss accounts for different senses of shear. Within the orthogneiss a top-to-the-N sense of shear prevails, whereas top-to-the-S is observed in the overlying metasedimentary rocks. In the lateral areas to the east and west deformation is accommodated by strike-slip shear zones. The occurrences of sillimanite within the Bozdağ nappe in the central Menderes Massif near Tire (Evirgen & Ataman, 1982) and in the North Menderes Massif south of Dermici (Candan & Dora, 1993) are associated with top-to-the-N shear criteria (Hetzel *et al.* 1998) and concur with northward crustal thickening.

The postulated prograde inverse metamorphic gradient north of the study area near Birgi–Bozdağ (Hetzel *et al.* 1998) with occurrence of sillimanite – and presumably cordierite in paragneiss or pyroxene–garnet–plagioclase in calc-silicate rocks – at the base of the Çine nappe, requires top-to-the-N thrusting (Figs 1, 18). Rocks of the underlying Bozdağ nappe (see Fig. 2), characterized by staurolite–kyanite–biotite and garnet–biotite parageneses, record a decrease in temperature across the nappe boundary (Hetzel *et al.* 1998; Dora *et al.* 2001). This succession describes a typical prograde inverse field gradient (Fig. 18) (e.g. Scheuvens, 2002). However, Dora *et al.* (2001) and Ring, Willner & Lackmann (2001) interpreted this inverted field gradient as a normal prograde field gradient folded during the Eocene.

South of Dermici, Candan & Dora (1993) describe index mineral sequences consistent with Barrovian-type metamorphism, which are exposed along a normal prograde field gradient. Additionally, Régnier *et al.* (2003) reported a normal prograde erosional level near Selimiye. It seems plausible that thrusting of the lower unit coeval with Barrovian-type metamorphism does not record a real prograde inverse field gradient; however, detailed isograd mapping is necessary to solve this problem in the Birgi–Bozdağ area.

Widespread greenschist shear bands are assumed to be related to greenschist facies top-to-the-S emplacement of the Çine, Bozdağ and Selimiye nappes over the Bayındır nappe (Fig. 17). Finally, common retrograde metamorphism within the Selimiye nappe, reported by Ashworth & Evirgen (1984) and Régnier *et al.* (2003), is correlated to a major tectonic event.

7. Discussion

Based on the stratigraphy of the Menderes Massif, Okay (2001) interpreted the southern Menderes submassif as a large N-vergent recumbent fold, without considering isograd mapping in metapelites. His model was contested by the discovery of Cretaceous rudist species in the Göktepe Formation, previously assigned to the Permo-Carboniferous (Özer & Sözbilir, 2003). Bozkurt & Park (1994) associated Barrovian metamorphism with Eocene burial of the Menderes Massif under the HP–LT Cycladic blueschist unit and Lycian nappes, an interpretation which does not seem plausible since these nappes are typically cold during their exhumation. Rimmelé *et al.* (2003) argued for burial of the lower units of the Menderes Massif to a minimum depth of 30 km during the closure of the northern branch of the Neo-Tethys. However, our study has not reported any evidence for a HP metamorphic overprinting. Régnier *et al.* (2003) correlate major retrograde metamorphism in the Selimiye nappe and greenschist facies shear zones in the Menderes Massif with thrusting of the HP–LT units onto the core series (Figs 17b, 18b, c). Exhumation of lower structural levels has likely been initiated with top-to-the-S movement along the Bayındır nappe under lower greenschist facies conditions. However, in contrast to our study from the southern Çine massif (Régnier *et al.* 2003), new evidence from parageneses from the western part, especially the staurolite–chlorite–chloritoid and staurolite–garnet–chlorite parageneses, strongly suggests that amphibolite facies metamorphism throughout the Menderes Massif is coeval. This implies that Barrovian metamorphism predated nappe emplacement and HP–LT metamorphism (Fig. 18b; Güngör & Erdoğan, 2001).

Gessner *et al.* (2001a, 2004) and Ring, Willner & Lackmann (2001) suggested that regional top-to-the-N deformation exclusively occurred during Proterozoic amphibolite facies metamorphism. In addition, Gessner *et al.* (2001a) correlated regional top-to-the-S shearing with Eocene greenschist metamorphism. Data of the western Çine Massif discussed in this study refute such simplification. Top-to-the-N shearing under lower amphibolite facies and top-to-the-S shearing under amphibolite facies conditions have been observed in the southern Çine nappe. In addition, coaxial deformation is observed at the top of the orthogneiss unit (Régnier *et al.* 2003). Index minerals and deformation within the Menderes Massif point to a major tectonic event, northward thrusting of Proterozoic orthogneisses onto the lower metasedimentary units, prior to the Eocene (Fig. 18b) (Hetzel *et al.* 1998; Lips *et al.* 2001). Top-to-the-N shearing was accommodated in the underlying metasedimentary rocks in the central Menderes Massif (Hetzel *et al.* 1998; Ring, Willner & Lackmann, 2001), while coaxial deformation prevails at the top

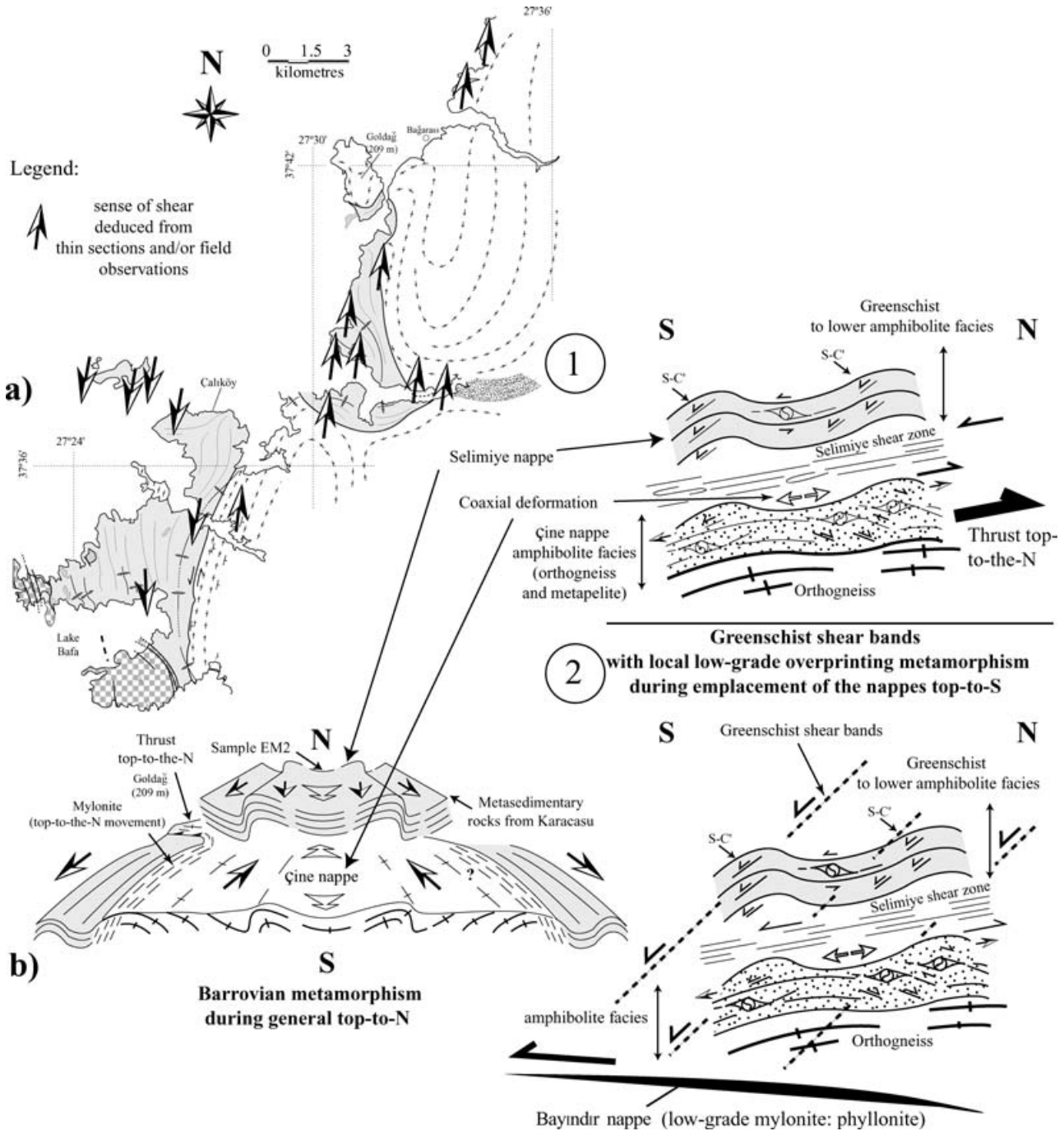


Figure 17. Interpretative sketch of the deformation in the Çine massif. Sequence of parageneses is consistent with a single tectonic event. (a). Senses of shear in the western Çine massif. (b) 1. Top-to-the-N thrusting of orthogneiss involved different senses of shear in metasedimentary enclaves or within surrounding metasedimentary rocks. Within the Selimiye nappe a S-directed shear sense prevails, whereas deformation in the orthogneiss and metasedimentary enclaves displays mainly top-to-the-N fabrics with local coaxial deformation on the top of the structure. Lateral deformation is more complex with possible partition of the strain due to strike-slip shear zone. 2. Top-to-the-S greenschist shear bands and possibly local greenschist metamorphic overprinting most likely occurred during emplacement of nappes by thrusting onto the Bayındır nappe.

of the orthogneiss. Top-to-the-S fabrics in the Selimiye nappe are interpreted to be related to southward back-thrusting onto the Proterozoic orthogneiss (Figs 17b, 18b).

Metamorphism in the upper metasedimentary unit of the thrust can involve a clockwise P - T path just as in the lower thrust unit if dissipated heat is taken in account (England & Molnar, 1993). Unloading implies

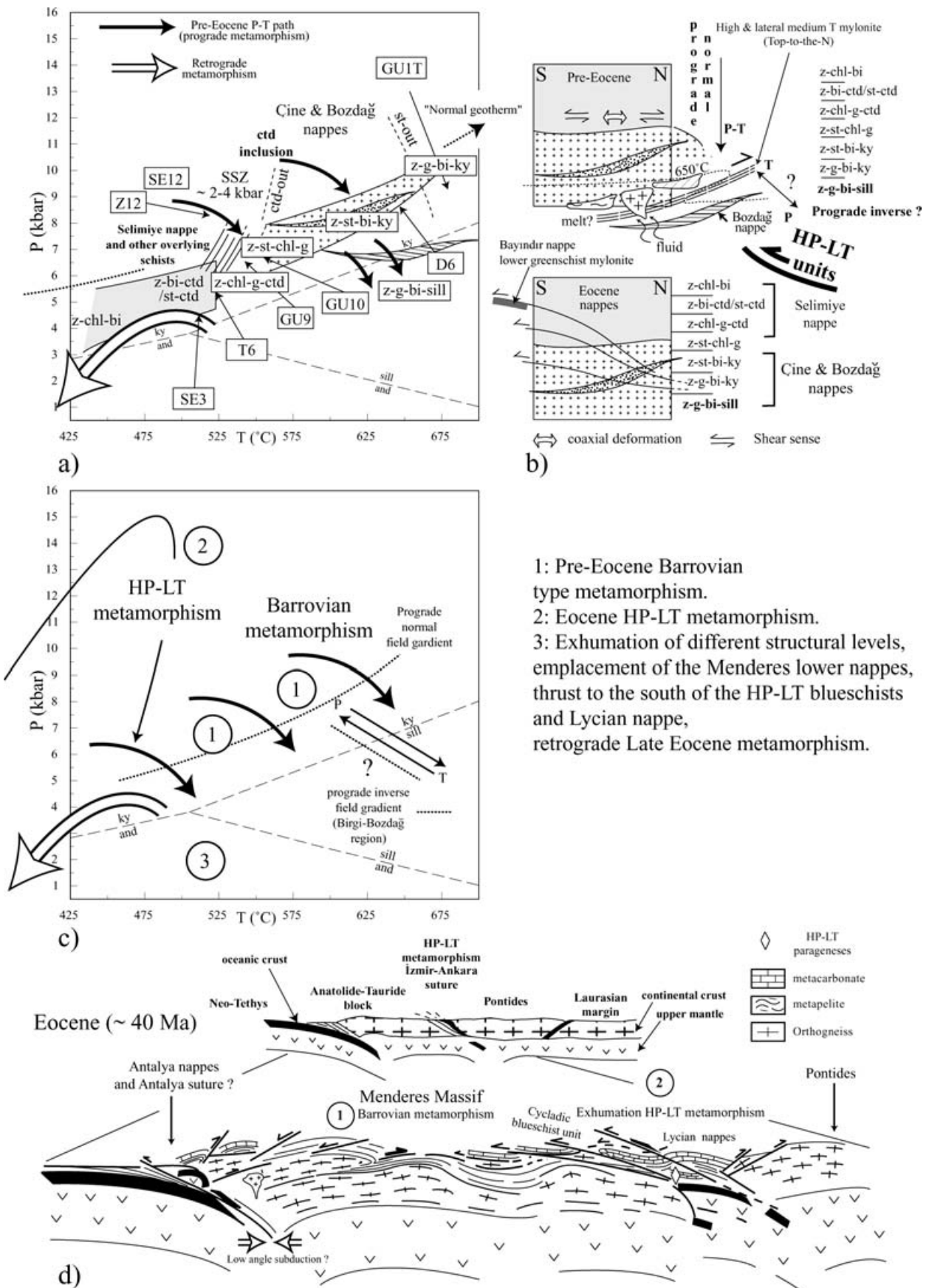


Figure 18. For legend see facing page.

decompression followed by heating of the upper unit, while compression during loading is followed by heating of the lower unit (Spear, 1993). The only difference will be the presence of an inverted field gradient in the lower unit, as described controversially in the Birgi–Bozdağ area.

A metagranite crosscutting regional-scale amphibolite facies foliation in the orthogneiss, and which was only deformed by greenschist shear bands, yields a SHRIMP zircon age of 566 ± 9 Ma. This led Régner *et al.* (2003) and Gessner *et al.* (2004) to propose a Proterozoic age for amphibolite facies metamorphism. Protolith ages for part of the metasedimentary rocks around the Çine Massif could be older than previously assumed (Çağlayan *et al.* 1980), Proterozoic instead of Palaeozoic in age (Koralay *et al.* 2004). Alternatively, zircons from the metagranites could be inherited. If melting occurred with a strong fluid participation, temperatures attained will not be sufficiently high to cause rim overgrowth of the zircon grain. Thus, some intrusions in the Çine Massif could be younger (Bozkurt, 2004). Eocene Rb–Sr and ^{40}Ar – ^{39}Ar ages obtained from mica of the Çine and Selimiye nappes, locally associated with top-to-the-N displacement deformation (Satır & Friedrichsen, 1986; Hetzel & Reischmann, 1996) argue against Gessner *et al.* (2004). An Eocene ^{40}Ar – ^{39}Ar age of mica associated with top-to-the-N shearing in mylonitic granitic gneiss at the base of the Çine nappe further supports a Tertiary age for amphibolite facies metamorphism (Lips *et al.* 2001). These ^{40}Ar – ^{39}Ar ages could be interpreted as true cooling ages. However, it appears unlikely for amphibolite facies metamorphism to take 500 Ma to cool below 400 °C. In addition, Th–Pb ion microprobe dating of *in situ* monazite also yielded an Eocene age for a part of the metasedimentary rocks of the Menderes Massif (Catlos & Çemen, 2005). However, monazite ages are difficult to interpret without knowledge of the *P*–*T* conditions during monazite crystallization.

These results outline a serious geochronological problem which cannot be solved solely by a meta-

morphic study (Bozkurt, 2004). The possibility of a major tectonic event prior to the Eocene and during the Cretaceous–Tertiary, associated with top-to-the-N thrusting and Barrovian-type metamorphism, has major implications on the palaeocontinental reconstruction of the eastern Mediterranean region and could lend support to the idea of a Neo-Tethys (*sensu stricto*) suture south of the Menderes Massif and below the Lycian nappes (Stampfli, 2000).

8. Conclusions

Parageneses observed in Proterozoic(?)–Palaeozoic metasedimentary rocks of the southern Menderes Massif are indicative of a single Barrovian-type amphibolite facies metamorphism related to crustal thickening and northwards thrusting of the lower unit which is exposed in the Çine Massif. This could have occurred during the pre-Eocene Tertiary or the Mesozoic. Emplacement of the lower nappes (Çine, Bozdağ, Selimiye) postdated the main metamorphic event by thrusting onto the so-called Bayındır nappe during lower greenschist facies conditions. The Bayındır nappe can be interpreted as a major greenschist facies mylonitic shear zone developed at the base of the lower thrust nappe. Subsequent late Eocene overthrusting of the HP–LT units following their exhumation resulted in retrograde metamorphism observed in the immediately underlying Selimiye nappe. Petrological data provide no evidence for burial of the lower units of the Menderes Massif deeper than 30 km during closure of the Neo-Tethys.

A key problem to be solved is the cause and the age of Barrovian metamorphism. Although an early Tertiary age for widespread amphibolite facies metamorphism in the Menderes Massif is possible, further geochronological studies of schists and metapelite enclaves in the orthogneisses in the western Çine Massif, especially Ar–Ar dating of calc-schist amphiboles from the enclaves, are necessary to test this hypothesis.

Figure 18. Summary of parageneses encountered in the Çine Massif and timing of different metamorphic events in the Menderes Massif. For legend see Figure 1. (a) Different parageneses outline a Barrovian field gradient (Spear, 1993). Additional *P*–*T* estimates are from Régner *et al.* (2003). Chloritoid inclusions in Çine Massif garnets constrain clockwise *P*–*T* path during thickening. The contact of the Çine and Selimiye nappes is tectonic, evident from an abrupt increase in pressure by 2–4 kbar, depending on the area. The Selimiye nappe and surrounding schists in the western Çine Massif are characterized by *P*–*T* conditions within the chloritoid–kyanite field near the aluminosilicate triple point. The Çine nappe is characterized by *P*–*T* conditions from the chloritoid-out isograd to the staurolite-out isograd (biotite–garnet–kyanite field, sample GU1T). Retrograde metamorphism occurs mainly within the Selimiye nappe. The prefix ‘z-’ indicates metamorphic zone. ‘SSZ’ denotes the Selimiye shear zone. (b) Top sketch: pre-Eocene Barrovian metamorphism induced by northward thrusting of the Proterozoic orthogneiss (including the Çine nappe) onto metasedimentary rocks (Hetzel *et al.* 1998; Lips *et al.* 2001). Lower sketch: syn- to post-Eocene exhumation during closure of Neo-Tethys and emplacement of HP–LT units (Cycladic blueschist unit and Lycian nappe) onto the Menderes Massif. Parageneses from this study are also shown; the garnet–sillimanite–biotite paragenesis is from Candan & Dora (1993). (c) *P*–*T* path history proposed in this study correlating emplacement of the HP–LT units over the Menderes Massif with major retrogression observed in rocks of the Selimiye nappe. The hypothetical prograde inverse field gradient from the Birgi–Bozdağ region is also shown (Hetzel *et al.* 1998; Dora *et al.* 2001). (d) Interpretative cross-section summarizing metamorphic events. Note that the top-to-N thrust and associated Barrovian metamorphism may be due to low angle subduction of the Neo-Tethys.

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Sample	N (37 °)	E (27 °)	q	fsp	mu	carb	chl	ctd	bi	g	st	ky	ep	amph	op	to	ap	zi
BA1	33°32"	23°58"	X	O	X	—	?	—	O	—	—	—	—	—	O	—	T	T
BA2	33°36"	24°09"	X	X	X	—	II	—	X	—	—	—	O	—	X	T	—	T
BA3	33°39"	24°14"	X	X	X	—	T(p1)	—	X(p1)	—	—	—	—	—	O	T	T	T
BA4	33°45"	24°03"	X	X	X, T(i)/fsp	—	—	—	X	—	—	—	—	—	X	T	T	T
BA5	33°46"	24°06"	X	X	X	—	X(p1)	—	X(p1)	—	—	—	—	—	O	T	T	T
BA6	33°52"	24°05"	X	X	X, T(i)/fsp	—	O(p1)	—	X(p1)	—	—	—	—	—	X	O	—	T
									i/fsp									
BA7	33°55"	24°02"	X	X	X	—	X(p1)	—	T(p1)	—	—	—	T	—	O	T	T	T
BA8	33°55"	24°00"	X	X	X	—	—	—	—	—	—	—	—	—	X	—	—	—
BA9	34°15"	24°14"	X	O	X	—	—	—	X	—	—	—	—	—	O	O	T	—
BA10P	32°56"	23°33"	T	—	—	X	—	—	—	—	—	—	—	—	T	—	—	—
BA11	33°08"	23°27"	X	O	X	X	X(p1)	—	X(p1)	—	—	—	X(p1)	—	X	T	T	T
BA12	33°13"	23°35"	X	X	X	T(i)/fsp	X(p1)	—	X(p1)	—	—	—	—	—	O	—	—	T
BA13	33°17"	23°44"	X	X	X	T(i)/fsp	O(p1), II	—	X(p1)	—	—	—	T	—	O	O	T	T
BA14	33°12"	24°12"	X	X	X	—	O(p1)	—	X(p1)	—	—	—	O	—	X	O	T	T
BA15	33°08"	24°20"	X	X	X	—	T(p1)	—	X(p1)	—	—	—	—	—	O	T	—	T
BA16	33°05"	24°29"	X	X	X	—	X(p1), II	—	X(p1)	A	—	—	—	—	O	O	T	T
BA17	33°13"	24°39"	X	X	X	—	T(p1), II	—	X(p1)	—	—	—	—	—	O	O	T	T
BA18	33°13"	24°48"	X	X	X	—	—	—	T	—	—	—	—	—	X	O	T	T
BA19	33°16"	25°06"	X	X	X	—	2	—	O	—	—	—	—	—	X	O	T	T
BA20	33°14"	25°13"	X	X	X	—	T(p1), II	—	O(p1)	—	—	—	—	—	X	O	—	T
BA21	33°13"	25°16"	X	X	X	—	T(p1), II	—	X(p1)	—	—	—	—	—	X	T	T	T
BA22	33°17"	25°29"	X	X	X	—	T(p1), II	—	T(p1)	—	—	—	O	—	X	O	O	T
BA23	33°20"	25°39"	X	X	X	—	X(p1)	—	T(p1)	—	—	—	—	—	X	O	T	T
BA24	33°24"	25°50"	X	X	X	—	T(p1), II	—	T(p1)	A	—	—	—	—	O	O	O	T
BA25	33°25"	25°55"	X	X	X	—	T(p1), II	—	O(p1)	—	—	—	—	—	O	O	—	T
BA26	33°28"	26°02"	X	X	X	—	—	—	T	—	—	—	—	—	X	T	—	T
L2P	32°05"	26°15"	X	X	X	—	X(p1)	—	O(p1)	—	—	—	—	—	O	T	T	T
L3	32°07"	26°16"	X	X	X	—	X(p1)	—	O(p1)	—	—	—	—	—	O	X	T	T
L4	32°11"	26°23"	X	O	X	—	X(p1)	—	T(p1)	T?	—	—	O(p1)	—	O	—	—	T
L5	32°13"	26°30"	X	O	X	—	2?	—	O(p1)	—	—	—	—	—	T	T	T	T
L6P	32°13"	26°30"	X	X	X	—	X(p1)	—	T(p1)	O(p1)	—	—	—	—	T	X	T	T
YE1	34°43"	27°22"	X	X	X	—	II	—	X	—								

Appendix 1. (Contd.)

Sample	N (37 °)	E (27 °)	q	fsp	mu	carb	chl	ctd	bi	g	st	ky	ep	amph	op	to	ap	zi
R8	37°51''	26°11''	X	X	X, T(i)/fsp	T(i)/fsp	X(p1), II	—	X(p1)	O(p1) (i)/fsp	—	—	T	—	O	T	O	T
R8P	37°51''	26°11''	X	X	X, T(i)/fsp	T(i)/fsp	X(p1), II	—	X(p1)	O(p1) (i)/fsp	—	—	—	—	O	X	O	T
R9	37°04''	26°37''	X	X	X, T(i)/fsp	T(i)/fsp	X(p1), II	—	X(p1)	X(p1) (i)/fsp	—	—	T	T(i)/fsp	O	O	T	T
R10	37°00''	26°28''	X	X	X, T(i)/fsp	—	O(p1), II	—	O(p1)	—	—	—	—	—	O	O	O	T
GU1	37°25''	34°27''	X	X	X	X	T(p1), II	—	O(p1)	X(p1)	—	—	O(p1)	—	O	T	—	T
GU2	37°29''	34°29''	X	O	X	—	—	—	O(p1)	—	—	—	O(p1)	—	X	—	O	T
GU3	37°34''	34°11''	X	O	O	X	—	—	T	—	—	—	X	—	O	—	—	T
GU3b	37°34''	34°11''	O	O	O	X	—	—	T	—	—	—	X	—	X	—	—	T
GU4	37°36''	33°56''	O	X	O	X	—	—	T	—	—	—	—	—	O	T	T	T
GU5	37°31''	34°00''	X	X	X	—	—	—	X	—	—	—	—	—	T	—	O	T
GU6	37°31''	33°54''	X	O	X	—	—	—	X	—	—	—	—	—	T	X	O	T
GU7	37°32''	33°49''	X	X	X	—	T(p1)	—	X(p1)	O(p1)	—	—	—	—	T	X	O	T
GU8	37°27''	33°36''	X	O	X	—	—	—	X	—	—	—	—	—	T	O	T	T
GU9	37°15''	32°54''	X	O	O	—	T(p1)	X(p1), O(i)/g	T(p1)	O(p1)	—	—	—	—	X	—	—	T
GU10	37°14''	32°20''	X	X	X, T(i)/g	—	T(p1), II	T(i)/g	O(p1,2)	X(p1,2) T(i)/g	T(p2)	—	—	—	X	O	T	T
GU11	37°19''	31°16''	X	X	X	—	II	—	X(p1)	X(p1)	—	—	—	—	X	T	T	T
GU12	37°35''	30°51''	X	X	X	—	O(p1), II	—	X(p1)	X(p1)	—	—	—	—	T	T	T	T
Ki1	37°26''	32°23''	X	X	X	—	—	—	O	—	—	—	—	—	O	O	T	T
Ki2	37°32''	32°17''	X	X	X	—	—	—	X(p1)	X(p1)	—	—	—	—	O	—	T	T
Ki3	37°32''	32°04''	X	X	X	—	—	—	X	—	—	—	—	—	O	T	O	T
Ki4	37°30''	31°52''	X	X	X	—	II	—	X(p1)	X(p1)	—	—	—	—	O	T	O	T
Ki5	37°43''	31°30''	X	X	X	—	O(p1), II	—	X(p1,2)	O(p2)	—	—	—	—	O	T	T	T
Ki6	37°56''	31°21''	X	X	X	—	II	T(i)/g	X(p1)	X(p1)	—	—	—	—	O	T	T	T
Ki7	38°20''	30°31''	X	X	X, T(i)/fsp	—	O(p1) T(i)/fsp, II	—	X(p1,2) T(i)/fsp	X(p2) (i)/fsp	—	—	—	—	O	—	T	T
Ki8	38°23''	30°28''	X	X	X, T(i)/fsp	T(i)/fsp	O(p1) T(i)/fsp, II	—	X(p1,2) T(i)/fsp	O(p2) (i)/fsp	—	—	—	—	O	O	O	T
Ki9	38°36''	30°27''	X	X	O, T(i)/fsp	—	X(p1) T(i)/fsp, II	—	X(p1) (i)/fsp	O(p1)	—	—	—	—	O	—	T	T
Ki10	38°41''	30°55''	X	X	X	—	O(p1), II	—	X(p1)	—	—	—	—	—	X	O	T	T
KG1	41°54''	30°39''	X	X	X	—	—	—	O	—	—	—	—	—	T	X	O	T
KG2	40°57''	31°08''	X	X	X	—	—	—	O	—	—	—	—	—	O	T	T	T
KG3	40°41''	30°50''	X	X	X	—	X(p1), II	T(i)/g	X(p1)	X(p1)	—	—	—	—	X	—	T	T
KG4	42°24''	31°00''	X	X	X	—	—	—	O	—	—	—	—	—	T	—	T	T
KG5	41°42''	31°39''	X	O	X	—	—	—	X	—	—	—	—	—	T	T	O	T
KG6	41°27''	31°45''	X	X	X	—	X(p1), II	T(i)/g	X(p1)	X(p1)	—	—	—	—	O	X	T	T
KG6bis	41°27''	31°45''	X	X	X	—	X(p1), II	T(i)/g	X(p1)	X(p1)	—	—	—	—	O	X	T	T
KG7	40°22''	31°41''	O	O	X	—	II	—	X	—	—	—	—	—	O	—	O	T
KG8	38°47''	31°00''	X	X	X	—	X(p1), II	—	X(p1)	O(p1)	—	—	—	—	O	—	T	T
KG9	39°15''	30°55''	X	X	X	—	X(p1), II	—	X(p1)	X(p1) (i)/fsp	—	—	—	—	O	—	T	T
KG10P	39°36''	31°00''	X	X	X	—	X(p1), II	—	X(p1)	—	—	—	—	—	O	—	T	T
KG11	40°38''	31°42''	X	X	X	—	X(p1), II	—	O(p1)	—	—	—	—	—	O	—	O	T
KG12	41°14''	31°51''	X	X	X	—	—	—	X(p1)	X(p1)	—	—	—	—	X	—	T	T
KG13	41°54''	32°10''	X	X	X	—	—	—	O	—	—	—	—	—	O	—	T	T
KG14	42°24''	32°54''	X	X	X	—	—	—	O	—	—	—	?	—	O	—	T	T
KG15	43°45''	33°38''	X	X	X	—	—	—	X	—	—	—	—	—	T	—	O	T
CPA1	44°25''	34°11''	X	T	O	—	—	—	X	—	—	—	—	—	T	—	T	T
CPA2	46°21''	36°59''	X	X	X	—	—	—	O	—	—	—	—	—	T	—	T	T
GU1T	45°20''	37°27''	X	X	T	—	II	—	X(p1)	X(p1)	—	X(p1)	—	—	O	—	T	T
GU2T	45°10''	37°16''	X	X	X	—	—	—	X(p1)	O(p1)	—	—	—	—	T	—	T	T
GU3T	44°59''	37°17''	X	X	X	—	II	—	X(p1)	X(p1)	—	O(p1)	—	—	O	—	O	T
GU4T	44°49''	37°14''	X	X	—	—	—	—	X(p1)	—	—	—	—	X(p1)	O	—	T	T
GU5T	44°06''	37°06''	X	X	X	—	II	—	X(p1)	O(p1)	—	—	—	—	O	—	O	T
GU6T	43°42''	36°58''	O	O	T	—	—	—	X(p1)	—	—	—	—	X(p1)	O	—	T	T
GU7T	42°52''	37°06''	X	X	O	—	II	—	X(p1)	X(p1)	—	—	—	—	O	X	O	T
GU8T	42°03''	37°31''	T	T	T	—	—	—	X(p1)	—	—	—	O(p1)	X(p1)	O	—	T	T
GU9T	41°53''	37°32''	T	—	—	—	—	—	T(p1)	—	—	—	O(p1)	X(p1)	O	—	T	T
GU10T	44°13''	37°21''	X	X	O	—	II	—	X(p1)	X(p1)	—	—	—	—	O	—	T	T
KO1	45°53''	42°40''	X	T	O	—	—	—	T	—	—	—	O	—	T	—	—	T
KO2	45°37''	39°43''	X	X	X	—	—	—	X(p1)	O(p1)	—	—	—	—	T	—	T	T
KO3	45°48''	39°42''	X	X	X	—	II	—	X(p1)	X(p1)	—	—	—	—	O	O	T	T
KO4	46°03''	39°41''	X	X	X	—	—	—	X	—	—	—	—	—	T	—	T	T
KO5	46°07''	39°50''	X	X	X	—	—	—	X(p1)	O(p1)	—	—	O(p1)	—	O	—	T	T

Appendix 1. (Contd.)

Sample	N (37 °)	E (27 °)	q	fsp	mu	carb	chl	ctd	bi	g	st	ky	ep	amph	op	to	ap	zi
KO6	46°02''	43°22''	X	X	X	–	–	T(i)/g	T(p1)	X(p1)	X(p1)	–	–	–	O	–	–	T
KO6b	46°02''	43°22''	X	X	X	–	–	–	O(p1)	X(p1)	X(p1)	–	–	–	X	–	T	T
KO7	46°01''	44°09''	X	X	X	–	–	–	X(p1)	X(p1)	–	–	–	–	O	O	O	T
D6	36°15''	42°13''	X	X	X, T(i)/g	–	–	T(i)/g	X(p2,3)	X(p1,3)	X(p1,2,3)X(p1,2)	T(i)/g	–	–	O	O	T	T
EM1	45°48''	50°10''	X	X	X	X, (i)/g	–	–	O(p1)	X(p1)	–	–	T(i)/g	–	X	T	–	T
EM1b	45°49''	50°10''	X	X	X	O(p1)	–	–	T	X(p1)	–	–	O(i)/g	–	X	T	–	–
EM2	45°12''	51°45''	X	O	X, T(i)/g	–	T(p1)	X(p1)	O(p1)	X(p1)	–	–	–	–	X	T	–	T
EM1P	45°12''	51°45''	X	X	X, T(i)/g	O	–	–	T	X	–	–	O(i)/g	–	X	O	–	T
SE3	25°25''	39°09''	X	O	X	–	T(p1), II	X(p1)	X(p1)	–	–	–	–	–	X	T	–	T
SE12	26°27''	39°37''	X	O	X	–	T(p1), II	X(p1)	X(p1)	X(p1)	–	–	T	–	X	T	T	T
T1	40°35''	31°58''	X	X	X	–	II	–	X(p1)	X(p1)	–	–	–	–	X	–	T	T
T2	40°33''	31°29''	X	X	X	–	–	–	X	–	–	–	–	–	T	–	T	T
T3	40°33''	31°38''	X	X	X	–	–	–	O	–	–	–	–	–	T	T	T	T
T5	40°25''	31°35''	X	X	X	–	–	–	X	–	–	–	–	–	T	–	T	T
T6	40°22''	31°37''	X	O	X	–	O(p1,2)	X(p1,2,3)	T(p1)	–	O(p2)	–	O(p3)	–	X	T	T	T
Z1	35°03''	31°19''	X	X	O	O	–	–	T	–	–	–	O	–	O	–	–	–
Z2	38°26''	34°41''	X	X	X	–	X(p1)	–	O(p1)	–	–	–	T	–	O	T	–	–
Z3	38°47''	34°48''	X	X	X	–	O(p1), II	–	O(p1)	X(p1)	–	–	–	–	X	–	–	–
Z4	38°56''	34°55''	X	X	X	–	O(p1,2), II	–	T(p2)	O(p1)	–	–	T	–	X	T	T	T
Z5	38°43''	34°15''	X	X	X	O	–	–	O	–	–	–	T	–	X	T	–	–
Z6	38°29''	33°59''	X	X	X	–	T(p1)	–	X(p1)	–	–	–	T	–	X	T	–	T
Z7	40°22''	32°57''	X	–	T	–	–	–	–	–	–	X	–	–	X	–	–	–
Z8	40°21''	33°03''	X	–	X	–	–	–	–	–	–	X	–	–	X	–	–	T
Z9	40°26''	33°49''	X	O	X	–	T(p1), II	X(p1)	T(p1), II	–	–	–	–	–	X	T	–	T
Z10	51°11''	32°09''	X	X	T	–	II	–	X(p1)	X(p1)	–	–	–	–	O	–	O	T
Z11	43°50''	44°34''	X	X	O	–	T(p1), II	–	O(p1)	–	–	–	–	–	O	T	–	T
Z12	44°33''	45°12''	X	–	X, II	–	T(p1), II	X(p1,2)	T(p1), II	R(p1)	–	–	O(p2)	–	X	–	–	T
Z13	44°33''	44°56''	X	O	X	–	II	T(i)/g	T(p1)	O(p1)	–	–	O	–	X	T	–	T
Z14	43°49''	43°34''	X	O	O	O	–	–	T	–	–	–	T	–	X	T	–	T
Z15	47°04''	40°43''	X	O	X	–	II	–	O(p1)	X(p1)	–	–	–	–	–	T	O	T
Z16	46°53''	40°38''	X	X	X	–	II	–	T(p1)	X(p1)	–	–	–	–	O	T	T	T

X: >10 vol. %; O: 1–10 vol. %, T: <1 vol. %; –: absent; p(+numeral): main paragenesis; R: retrograded garnet; II: secondary phase.

Mineral abbreviations: alm: almandine; als: aluminosilicate; amph: amphibole; and: andalusite; ap: apatite; bi: biotite; carb: carbonate; cd: cordierite; chl: chlorite; ctd: chloritoid; ep: epidote; fsp: feldspar; g: garnet; grs: grossular; ilm: ilmenite; i, incl: inclusion(s); Kfs: potassium feldspar; ky: kyanite; mu: muscovite; op: opaques; p: plagioclase; prp: pyrope; q: quartz; ru: rutile; sill: sillimanite; sph: sphene; sps: spessartine; st: staurolite; to: tourmaline; zi: zircon; zo: zoisite.

Appendix 2. Activity models

A.2.a. Activity models used with THERMOCALC
v. 3.2.1 for the KFMASH system

In addition to the solid solution minerals described below, the pure phases quartz, andalusite, sillimanite, kyanite, and an H₂O fluid phase were used in calculations.

Biotite: $K_1^A[Al^{3+}, Fe^{2+}, Mg^{2+}]_1^{M1}[Mg^{2+}, Fe^{2+}]_2^{M2}[Si^{4+}, Al^{3+}]_2^{T1}Si_2O_{10}(OH)_2$

Biotite is modelled in the system KFMASH involving order–disorder of Fe and Mg between one M1 site and two M2 sites (Powell & Holland, 1999). There are four independent end-members: phlogopite (phl), annite (ann), eastonite (east), and the ordered end-member (obi). Biotite mixing is described by the following three variables:

$$x = \left(\frac{Fe^{2+}}{Fe^{2+} + Mg^{2+}} \right)^{tot}; \quad y = X_{Al}^{M1}; \quad N = 3(x - X_{Fe}^{M2})$$

Site fractions in terms of compositional variables are:

$$X_{Al}^{M1} = y; \quad X_{Fe}^{M1} = x(1 - y) + \frac{2N}{3};$$

$$X_{Mg}^{M1} = (1 - y)(1 - x) - \frac{2N}{3}; \quad X_{Mg}^{M2} = (1 - x) + \frac{N}{3};$$

$$X_{Fe}^{M2} = x - \frac{N}{3}; \quad X_{Si}^{T1} = \frac{1 - y}{2}; \quad X_{Al}^{T1} = \frac{1 + y}{2}$$

The ideal activities of end-members are expressed as:

$$a_{phl}^{ideal} = 4X_{Mg}^{M1} (X_{Mg}^{M2})^2 X_{Al}^{T1} X_{Si}^{T1}$$

$$a_{ann}^{ideal} = 4X_{Fe}^{M1} (X_{Fe}^{M2})^2 X_{Al}^{T1} X_{Si}^{T1}$$

$$a_{east}^{ideal} = X_{Al}^{M1} (X_{Mg}^{M2})^2 (X_{Al}^{T1})^2$$

$$a_{obi}^{ideal} = 4X_{Fe}^{M1} (X_{Mg}^{M2})^2 X_{Al}^{T1} X_{Si}^{T1}$$

The proportions of each end-member in the biotite phase are defined as:

$$p_{phl} = (1 - x)(1 - y) - \frac{2N}{3}$$

$$p_{ann} = x - \frac{N}{3}$$

$$p_{east} = y$$

$$p_{obi} = -xy + N$$

Non-ideality is expressed using symmetric formalism (Powell & Holland, 1993) with interaction parameters from Powell & Holland (1999) for KFMASH biotites. The interaction parameters are (kJ/mol end-member): $W_{\text{phl-ann}} = 9$; $W_{\text{phl-east}} = 10$; $W_{\text{phl-obi}} = 3$; $W_{\text{ann-east}} = -1$; $W_{\text{ann-obi}} = 6$; $W_{\text{east-obi}} = 10$. The Darkens Quadratic Formalism (DQF) parameter for the ordered end-member obi is (kJ/mol end-member): $I_{\text{obi}} = -10.73$ (Powell & Holland, 1999; <http://www.earthsci.unimelb.edu.au/tpg/thermocalc/>).

Chlorite: $[\text{Fe}^{2+}, \text{Mg}^{2+}]_4^{\text{M23}} [\text{Al}^{3+}, \text{Mg}^{2+}, \text{Fe}^{2+}]_1^{\text{M1}} [\text{Al}^{3+}, \text{Mg}^{2+}, \text{Fe}^{2+}]_1^{\text{M4}} [\text{Si}^{4+}, \text{Al}^{3+}]_2^{\text{T2}} \text{Si}_2\text{O}_{10}(\text{OH})_8$

Chlorite is modelled in the system FMASH involving four end-members: Al-free chlorite (afchl), clinocllore (clin), daphnite (daph) and amesite (ames) with:

$$x = \left(\frac{\text{Fe}^{2+}}{\text{Fe}^{2+} + \text{Mg}^{2+}} \right)^{\text{tot}}; \quad y = X_{\text{Al}}^{\text{T2}}; \quad N = \frac{X_{\text{Al}}^{\text{M4}} - X_{\text{Al}}^{\text{M1}}}{2}$$

Site fractions in terms of compositional variables are:

$$\begin{aligned} X_{\text{Fe}}^{\text{M23}} &= x; & X_{\text{Mg}}^{\text{M23}} &= 1 - x; & X_{\text{Al}}^{\text{M1}} &= y - N; \\ X_{\text{Fe}}^{\text{M1}} &= x(1 - y + N); & X_{\text{Mg}}^{\text{M1}} &= (1 - x)(1 - y + N) \\ X_{\text{Al}}^{\text{M4}} &= y + N; & X_{\text{Fe}}^{\text{M4}} &= x(1 - y - N); \\ X_{\text{Mg}}^{\text{M4}} &= (1 - x)(1 - y - N); & X_{\text{Al}}^{\text{T2}} &= y; & X_{\text{Si}}^{\text{T2}} &= 1 - y \end{aligned}$$

The ideal activities of end-members are expressed as:

$$\begin{aligned} a_{\text{afchl}}^{\text{ideal}} &= (X_{\text{Mg}}^{\text{M23}})^4 X_{\text{Mg}}^{\text{M1}} X_{\text{Mg}}^{\text{M4}} (X_{\text{Si}}^{\text{T2}})^2 \\ a_{\text{clin}}^{\text{ideal}} &= 4 (X_{\text{Mg}}^{\text{M23}})^4 X_{\text{Mg}}^{\text{M1}} X_{\text{Al}}^{\text{M4}} X_{\text{Al}}^{\text{T2}} X_{\text{Si}}^{\text{T2}} \\ a_{\text{daph}}^{\text{ideal}} &= 4 (X_{\text{Fe}}^{\text{M23}})^4 X_{\text{Fe}}^{\text{M1}} X_{\text{Al}}^{\text{M4}} X_{\text{Al}}^{\text{T2}} X_{\text{Si}}^{\text{T2}} \\ a_{\text{ames}}^{\text{ideal}} &= (X_{\text{Mg}}^{\text{M23}})^4 X_{\text{Al}}^{\text{M1}} X_{\text{Al}}^{\text{M4}} (X_{\text{Al}}^{\text{T2}})^2 \end{aligned}$$

The proportions of each end-member in the chlorite phase are defined as:

$$\begin{aligned} p_{\text{afchl}} &= 1 - y - N \\ p_{\text{clin}} &= 2N - \left(\frac{2x}{5} \right) (3 - y) \\ p_{\text{daph}} &= \left(\frac{2x}{5} \right) (3 - y) \\ p_{\text{ames}} &= y - N \end{aligned}$$

Non-ideality is expressed using symmetric formalism (Powell & Holland, 1993) with interaction parameters from Holland, Baker & Powell (1998) for FMASH chlorite. The interaction parameters are (kJ/mol end-member): $W_{\text{afchl-clin}} = 18$; $W_{\text{afchl-daph}} = 14.5$; $W_{\text{afchl-ames}} = 20$; $W_{\text{clin-daph}} = 2.5$; $W_{\text{clin-ames}} = 18$; $W_{\text{daph-ames}} = 13.5$.

White mica: $\text{K}_1^{\text{A}} [\text{Al}^{3+}, \text{Fe}^{2+}, \text{Mg}^{2+}]_1^{\text{M2}} [\text{Si}^{4+}, \text{Al}^{3+}]_2^{\text{T1}} \text{AlSi}_2\text{O}_{10}(\text{OH})_2$

White mica is modelled in the system KFMASH involving three end-members: muscovite (mu), celadonite (cel) and Fe-celadonite (fccl). Ideal mixing is assumed (Massonne &

Schreyer, 1987; Holland & Powell, 1998). The compositional variables are:

$$x = X_{\text{Si}}^{\text{T1}}; \quad y = \left(\frac{\text{Fe}^{2+}}{\text{Fe}^{2+} + \text{Mg}^{2+}} \right)^{\text{tot}}$$

Site fractions in terms of compositional variables are:

$$\begin{aligned} X_{\text{Al}}^{\text{M2}} &= 2(1 - x); & X_{\text{Fe}}^{\text{M2}} &= y(2x - 1); \\ X_{\text{Mg}}^{\text{M2}} &= 2 \left(x - \frac{1}{2} \right) (1 - y); & X_{\text{Si}}^{\text{T1}} &= x; & X_{\text{Al}}^{\text{T1}} &= 1 - x \end{aligned}$$

The ideal activities of end-members are expressed as:

$$\begin{aligned} a_{\text{mu}}^{\text{ideal}} &= 4 X_{\text{Al}}^{\text{M2}} X_{\text{Al}}^{\text{T1}} X_{\text{Si}}^{\text{T1}} \\ a_{\text{cel}}^{\text{ideal}} &= X_{\text{Mg}}^{\text{M2}} (X_{\text{Si}}^{\text{T1}})^2 \\ a_{\text{fccl}}^{\text{ideal}} &= X_{\text{Fe}}^{\text{M2}} (X_{\text{Si}}^{\text{T1}})^2 \end{aligned}$$

The proportions of each end-member in the white mica phase are defined as:

$$\begin{aligned} p_{\text{mu}} &= 2(1 - x) \\ p_{\text{cel}} &= 2 \left(x - \frac{1}{2} \right) (1 - y) \\ p_{\text{fccl}} &= y(2x - 1) \end{aligned}$$

Cordierite: $[\text{Fe}^{2+}, \text{Mg}^{2+}]_2^{\text{M}} \text{Al}_4\text{Si}_5\text{O}_{18}[\Omega, \text{H}_2\text{O}]_1^{\text{W}}$

Cordierite is modelled in the system FMASH involving three end-members: cordierite (crd), Fe-cordierite (fcrd) and hydrous cordierite (hcrd). Ideal mixing is assumed. The compositional variables and site fractions in terms of compositional variables are:

$$X_{\text{Fe}}^{\text{M}} = x; \quad X_{\text{H}_2\text{O}}^{\text{W}} = h; \quad X_{\text{Mg}}^{\text{M}} = 1 - x; \quad X_{\Omega}^{\text{W}} = 1 - h$$

The ideal activities and proportion of each end-member in the cordierite are expressed as:

$$\begin{aligned} a_{\text{crd}}^{\text{ideal}} &= (X_{\text{Mg}}^{\text{M}})^2 (X_{\Omega}^{\text{W}}); & p_{\text{crd}} &= 1 - (x + h) \\ a_{\text{fcrd}}^{\text{ideal}} &= (X_{\text{Fe}}^{\text{M}})^2 (X_{\Omega}^{\text{W}}); & p_{\text{fcrd}} &= x \\ a_{\text{hcrd}}^{\text{ideal}} &= (X_{\text{Mg}}^{\text{M}})^2 (X_{\text{H}_2\text{O}}^{\text{W}}); & p_{\text{hcrd}} &= h \end{aligned}$$

The mixing model is after Holland & Powell (1998).

Staurolite: $[\text{Fe}^{2+}, \text{Mg}^{2+}]_4^{\text{M}} \text{Al}_{18}\text{Si}_{7.5}\text{O}_{48}\text{H}_4$

Staurolite is modelled in the system FMASH involving two end-members: Fe-staurolite (fst) and Mg-staurolite (mst). Ideal activities of end-members:

$$a_{\text{fst}}^{\text{ideal}} = (X_{\text{Fe}}^{\text{M}})^4; \quad a_{\text{mst}}^{\text{ideal}} = (X_{\text{Mg}}^{\text{M}})^4$$

Non-ideality is expressed using symmetric formalism (Powell & Holland, 1993) with the interaction parameter (kJ/mol end-member): $W_{\text{fst-mst}} = -8$ (<http://www.earthsci.unimelb.edu.au/tpg/thermocalc/>).

Chloritoid: $[\text{Fe}^{2+}, \text{Mg}^{2+}]_1^{\text{M}} \text{Al}_2\text{SiO}_5(\text{OH})_2$

Chloritoid is modelled in the system FMASH involving two end-members: Fe-chloritoid (fctd) and Mg-chloritoid (mctd). Ideal activities of end-members:

$$a_{\text{fctd}}^{\text{ideal}} = X_{\text{Fe}}^{\text{M}}, \quad a_{\text{mctd}}^{\text{ideal}} = X_{\text{Mg}}^{\text{M}}$$

Non-ideality is expressed using symmetric formalism (Powell & Holland, 1993) with the interaction parameter (kJ/mol end-member): $W_{\text{fctd-mctd}} = 1$ (<http://www.earthsci.unimelb.edu.au/tpg/thermocalc/>).

Garnet: $[\text{Fe}^{2+}, \text{Mg}^{2+}]_3^{\text{A}} \text{Al}_2\text{Si}_3\text{O}_{12}$

Garnet is modelled in the system FMAS involving two end-members: almandine (alm) and pyrope (py). Ideal activities of end-members:

$$a_{\text{alm}}^{\text{ideal}} = (X_{\text{Fe}}^{\text{A}})^3; \quad a_{\text{py}}^{\text{ideal}} = (X_{\text{Mg}}^{\text{A}})^3$$

Non-ideality is expressed using symmetric formalism (Holland & Powell, 1998) with the interaction parameter (kJ/mol end-member): $W_{\text{alm-py}} = 2.5$.

A.2.b. Activity models used with THERMOCALC v. 3.2.1 for the NCKFMASH system

We added plagioclase and zoisite as new phases in the NCKFMASH system. All other phases use similar activity models except for white mica and garnet. The pure phases quartz, andalusite, sillimanite, kyanite, and an H_2O fluid phase were used as well in calculations.

White mica: $[\text{K}^+, \text{Na}^+]_1^{\text{A}} [\text{Al}^{3+}, \text{Fe}^{2+}, \text{Mg}^{2+}]_1^{\text{M}2} [\text{Si}^{4+}, \text{Al}^{3+}]_2^{\text{T}1} \text{AlSi}_2\text{O}_{10}(\text{OH})_2$

White mica is modelled in the system NCKFMASH involving four end-members: paragonite (pa), muscovite (mu), celadonite (cel) and Fe-celadonite (fcel). The compositional variables are:

$$x = X_{\text{Si}}^{\text{T}1}; \quad y = \left(\frac{\text{Fe}^{2+}}{\text{Fe}^{2+} + \text{Mg}^{2+}} \right)^{\text{M}2}; \quad z = \left(\frac{\text{Na}^+}{\text{Na}^+ + \text{K}^+} \right)^{\text{A}}$$

Site fractions in terms of compositional variables are:

$$X_{\text{Al}}^{\text{M}2} = 2(1 - x); \quad X_{\text{Fe}}^{\text{M}2} = y(2x - 1);$$

$$X_{\text{Mg}}^{\text{M}2} = 2 \left(x - \frac{1}{2} \right) (1 - y); \quad X_{\text{Si}}^{\text{T}1} = x;$$

$$X_{\text{Al}}^{\text{T}1} = 1 - x; \quad X_{\text{Na}}^{\text{A}} = z; \quad X_{\text{K}}^{\text{A}} = 1 - z$$

The ideal activities of end-members are expressed as:

$$a_{\text{phl}}^{\text{ideal}} = 4X_{\text{Na}}^{\text{A}} X_{\text{Al}}^{\text{M}2} X_{\text{Al}}^{\text{T}1} X_{\text{Si}}^{\text{T}1}$$

$$a_{\text{mu}}^{\text{ideal}} = 4X_{\text{K}}^{\text{A}} X_{\text{Al}}^{\text{M}2} X_{\text{Al}}^{\text{T}1} X_{\text{Si}}^{\text{T}1}$$

$$a_{\text{cel}}^{\text{ideal}} = X_{\text{K}}^{\text{A}} X_{\text{Mg}}^{\text{M}2} (X_{\text{Si}}^{\text{T}1})^2$$

$$a_{\text{fcel}}^{\text{ideal}} = X_{\text{K}}^{\text{A}} X_{\text{Fe}}^{\text{M}2} (X_{\text{Si}}^{\text{T}1})^2$$

The proportions of each end-member in the white mica phase are defined as:

$$p_{\text{pa}} = z$$

$$p_{\text{mu}} = 1 - z - (2x - 1)$$

$$p_{\text{cel}} = 2 \left(x - \frac{1}{2} \right) (1 - y)$$

$$p_{\text{fcel}} = y(2x - 1)$$

Non-ideality is expressed using symmetric formalism with the interaction parameters of Holland & Powell (1998; <http://www.esc.cam.ac.uk/astaff/holland/thermocalc.html>). The interaction parameters are (kJ/mol end-member): $W_{\text{pa-mu}} = 12 + 0.4P$; $W_{\text{pa-cel}} = 14 + 0.2P$; $W_{\text{pa-fcel}} = 14 + 0.2P$. The DQF parameter for the end-member paragonite is (kJ/mol end-member, P in kbar): $1.42 + 0.4P$.

Garnet: $[\text{Fe}^{2+}, \text{Ca}^{2+}, \text{Mg}^{2+}]_3^{\text{A}} \text{Al}_2\text{Si}_3\text{O}_{12}$

Garnet is modelled in the system CFMAS involving three end-members: almandine (alm), grossular (gr) and pyrope (py). Ideal activities of end-members:

$$a_{\text{alm}}^{\text{ideal}} = (X_{\text{Fe}}^{\text{A}})^3; \quad a_{\text{gr}}^{\text{ideal}} = (X_{\text{Ca}}^{\text{A}})^3; \quad a_{\text{py}}^{\text{ideal}} = (X_{\text{Mg}}^{\text{A}})^3$$

Non-ideality is expressed using symmetric formalism (Worley & Powell, 1998) with the interaction parameter (kJ/mol end-member): $W_{\text{gr-py}} = 33$.

Plagioclase: $[\text{Na}^+, \text{Ca}^{2+}]_1^{\text{A}} [\text{Si}^{4+}, \text{Al}^{3+}]_4^{\text{T}} \text{O}_8$

Plagioclase is modelled with the binary albite (ab)–anorthite (an) solution model 4T ($\text{C}\bar{1}$ structure) of Holland & Powell (1992). The compositional variable is:

$$x = \left(\frac{\text{Na}^+}{\text{Na}^+ + \text{Ca}^{2+}} \right)^{\text{A}}$$

Site fractions in terms of compositional variables are:

$$X_{\text{Na}}^{\text{A}} = x; \quad X_{\text{Ca}}^{\text{A}} = 1 - x; \quad X_{\text{Si}}^{\text{T}} = \frac{1}{2} + \frac{1}{4}x; \quad X_{\text{Al}}^{\text{T}} = \frac{1}{2} - \frac{1}{4}x$$

The ideal activities of end-members are expressed as:

$$a_{\text{ab}}^{\text{ideal}} = \frac{256}{27} (X_{\text{Na}}^{\text{A}}) (X_{\text{Al}}^{\text{T}}) (X_{\text{Si}}^{\text{T}})^3$$

$$a_{\text{an}}^{\text{ideal}} = 16 (X_{\text{Ca}}^{\text{A}}) (X_{\text{Al}}^{\text{T}})^2 (X_{\text{Si}}^{\text{T}})^2$$

Regular solution model interaction parameter, from Worley & Powell (1998) is (kJ/mol): $W_{\text{ab-an}} = 5.5$. The DQF parameter for the end-member anorthite is (kJ/mol end-member, T in Kelvin): $4.31 - 0.00217T$.

A.2.c. Activity models used with THERMOCALC v. 3.2.1 for the MnNCKFMASH system

MnO is assumed mainly concentrated in staurolite, chloritoid and garnet. All other phases use similar activity models (see

NCKFMASH system). The pure phases quartz, andalusite, sillimanite, kyanite, and an H₂O fluid phase were used as well in calculations.

Staurolite: $[\text{Fe}^{2+}, \text{Mg}^{2+}, \text{Mn}^{2+}]_4^{\text{M}} \text{Al}_{18}\text{Si}_{7.5}\text{O}_{48}\text{H}_4$

Staurolite is modelled in the system MnFMASH involving three end-members: Fe-staurolite (fst), Mg-staurolite (mst) and Mn-staurolite (mnst). Ideal activities of end-members:

$$a_{\text{fst}}^{\text{ideal}} = (X_{\text{Fe}}^{\text{M}})^4; \quad a_{\text{mst}}^{\text{ideal}} = (X_{\text{Mg}}^{\text{M}})^4; \quad a_{\text{mnst}}^{\text{ideal}} = (X_{\text{Mn}}^{\text{M}})^4$$

Non-ideality is expressed using symmetric formalism (Powell & Holland, 1993) with the interaction parameter (kJ/mol end-member): $W_{\text{fst-mst}} = -8$ (<http://www.earthsci.unimelb.edu.au/tpg/thermocalc/>).

Garnet: $[\text{Fe}^{2+}, \text{Ca}^{2+}, \text{Mg}^{2+}, \text{Mn}^{2+}]_3^{\text{A}} \text{Al}_2\text{Si}_3\text{O}_{12}$

Garnet is modelled in the system MnCFMAS involving four end-members: almandine (alm), grossular (gr), pyrope (py) and spessartine (spss). Ideal activities of end-

members:

$$a_{\text{alm}}^{\text{ideal}} = (X_{\text{Fe}}^{\text{A}})^3; \quad a_{\text{gr}}^{\text{ideal}} = (X_{\text{Ca}}^{\text{A}})^3; \\ a_{\text{py}}^{\text{ideal}} = (X_{\text{Mg}}^{\text{A}})^3; \quad a_{\text{spss}}^{\text{ideal}} = (X_{\text{Mn}}^{\text{A}})^3$$

Non-ideality is expressed using symmetric formalism (Worley & Powell, 1998; Wood, Hackler & Dobson, 1994) with the interaction parameter (kJ/mol end-member): $W_{\text{gr-py}} = 33$; $W_{\text{py-spss}} = 4.5$.

Chloritoid: $[\text{Fe}^{2+}, \text{Mg}^{2+}, \text{Mn}^{2+}]_1^{\text{M}} \text{Al}_2\text{SiO}_5(\text{OH})_2$

Chloritoid is modelled in the system MnFMASH involving three end-members: Fe-chloritoid (fctd), Mg-chloritoid (mctd) and Mn-chloritoid (mnctd). Ideal activities of end-members:

$$a_{\text{fctd}}^{\text{ideal}} = X_{\text{Fe}}^{\text{M}}, \quad a_{\text{mctd}}^{\text{ideal}} = X_{\text{Mg}}^{\text{M}}, \quad a_{\text{mnctd}}^{\text{ideal}} = X_{\text{Mn}}^{\text{M}}$$

Non-ideality is expressed using symmetric formalism (Powell & Holland, 1993) with the interaction parameter (kJ/mol end-member): $W_{\text{fctd-mctd}} = 1$.