

Metamorphic evolution of blueschists of the Altınekin Complex, Konya area, south central Turkey

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The Altınekin Complex in south central Turkey forms part of the south-easterly extension of the Tavşanlı Zone, a Cretaceous subduction complex formed during the closure of the Neo-Tethys ocean. The protoliths of metamorphic rocks within the Altınekin Complex include peridotites, chromitites, basalts, ferruginous cherts and flysch-facies impure carbonate sediments. Structurally, the complex consists of a stack of thrust slices, with massive ophiolite tectonically overlying a Cretaceous sediment-hosted ophiolitic mélange, in turn overlying a sequence of Mesozoic sediments.

Rocks within the two lower structural units have undergone blueschist-facies metamorphism. Petrographic, mineral-chemical and thermobarometric studies were undertaken on selected samples of metasedimentary and metabasic rock in order to establish the time relations of deformation and metamorphism and to constrain metamorphic conditions. Microstructures record two phases of plastic deformation, one predating the metamorphic peak, and one postdating it. Estimated peak metamorphic pressures mostly fall in the range 9–11 kbar, corresponding to burial depths of 31–38 km, equivalent to the base of a continental crust of normal thickness. Best-fit peak metamorphic temperatures range from 375 to 450°C. Metamorphic fluids had high H₂O:CO₂ ratios. Peak metamorphic temperature/depth ratios (*T/d* values) were low (*c.* 10–14°C/km), consistent with metamorphism in a subduction zone. Lawsonite-bearing rocks in the southern part of the ophiolitic mélange record lower peak temperatures and *T/d* values than epidote blueschists elsewhere in the unit, hinting that the latter may consist of two or more thrust slices with different metamorphic histories. Differences in peak metamorphic conditions also exist between the ophiolitic mélange and the underlying metasediments.

Rocks of the Altınekin Complex were subducted to much shallower depths, and experienced higher geothermal gradients, than those of the NW Tavşanlı Zone, possibly indicating dramatic lateral variation in subduction style.

Retrograde *P–T* paths in the Altınekin Complex were strongly decompressive, resulting in localized overprinting of epidote blueschists by greenschist-facies assemblages, and of lawsonite blueschists by pumpellyite-facies assemblages. The observation that the second deformation was associated with decompression is consistent with, but not proof of, exhumation by a process that involved deformation of the hanging-wall wedge, such as gravitational spreading, corner flow or buoyancy-driven shallowing of the subduction zone. Copyright © 2005 John Wiley & Sons, Ltd.

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1. INTRODUCTION

Ketin (1966), in his classic paper, subdivided Turkey geologically into four major belts. These are, from north to south, the Pontides, Anatolides, Taurides and Border folds. Şengör and Yılmaz (1981) and Şengör (1984) classified the major tectonic elements of Turkey in terms of plate tectonics, and showed that the geology of Turkey was shaped mainly by the collision of Laurasia and Gondwanaland and their fragments during the Late Mesozoic and Tertiary. The İzmir–Ankara Suture Zone, the main Neo-Tethyan suture in Turkey, divides Turkey into two main tectonic units, the Pontides and Anatolide–Tauride Platform (Figure 1(a); Şengör and Yılmaz 1981; Okay

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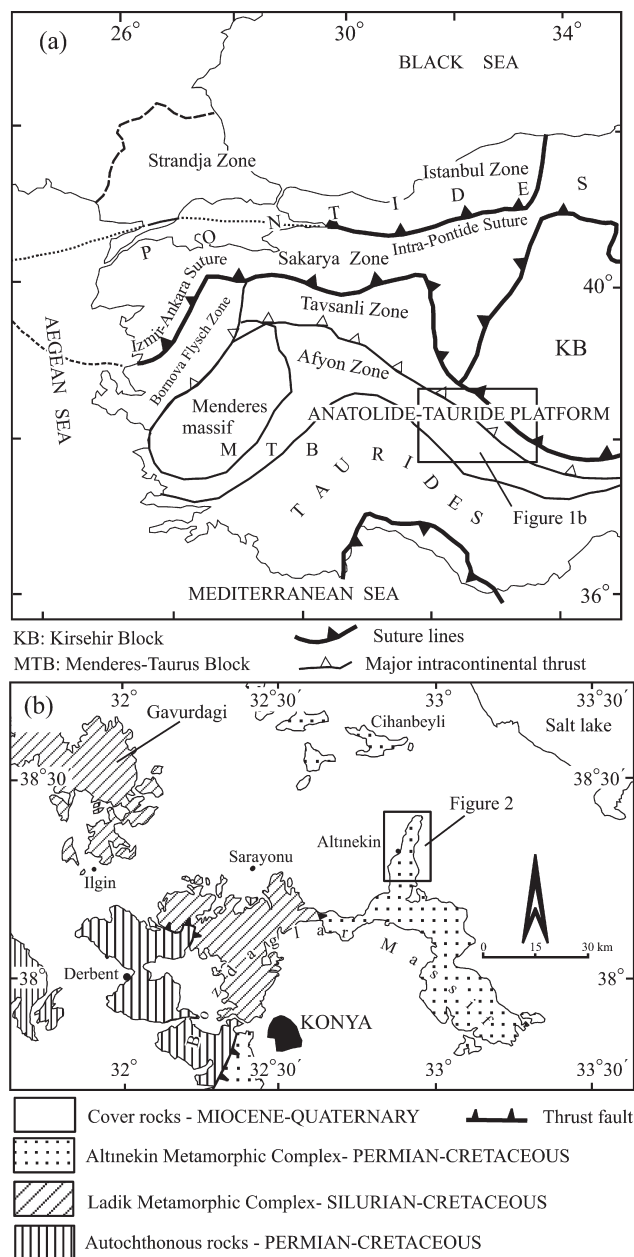


Figure 1. Location of the study area in the context of the geology of western Turkey: (a) major tectonic units of western Turkey; (b) location of the Altınekin Complex.

1986; Okay *et al.* 1996). The Pontides comprise the Sakarya, Istanbul and Strandja zones and belong to Laurasia. South of the suture is the Anatolide–Tauride Platform, which belongs to the Gondwanaland realm. The enigmatic Inner Tauride Suture Zone subdivides the platform into two continental fragments, the Menderes–Tauride and Kırşehir blocks. The Menderes–Tauride Block comprises the Tavşanlı, Afyon and Bornova Flysch zones and the Menderes Massif.

Within Turkey there are several high-pressure, low-temperature (HP–LT) metamorphic belts which reflect Neotethyan ocean closure (Okay 1984, 1986, 1989; Okay and Monié 1997; Okay *et al.* 1996; Sherlock *et al.* 1999). The distribution of glaucophane-bearing rocks in Turkey was first described by Kaaden (1966). Okay (1986) assigned these occurrences to five major HP–LT provinces on the basis of tectonic setting, age, and regional distribution: these are the Tavşanlı Zone, the Alanya blueschists, the Bitlis Massif, the Pontide blueschists and other occurrences such as the Sarkoy and Marmara blueschists.

The Tavşanlı Zone, representing the northward-subducted north-facing continental margin of the Menderes–Tauride Block, is the largest and most important HP–LT metamorphic belt in Turkey. It is thrust over the Afyon Zone, which is composed of Devonian to Palaeocene sedimentary rocks metamorphosed to greenschist facies (Okay 1984, 1986). The Tavşanlı Zone has a NW–SE-trending orientation and is 350 km in length and 50–100 km in width. It extends from Bursa to east of Konya and comprises volcanic and sedimentary rocks metamorphosed to blueschist facies during the mid-Cretaceous (Okay 1984, 1986). It is surrounded on all sides by sutures, to the north by the İzmir–Ankara Suture, and elsewhere by the Inner Tauride Suture. Its north-western part is well known (Okay 1984, 1986; Okay *et al.* 1996), but its south-eastern continuation has received relatively little attention, the few published studies generally being concerned with the stratigraphy and structural features of the region. This paper is a preliminary attempt to redress the balance.

The Altınekin Complex, NW of Konya, is considered to be part of the south-eastern continuation of the Tavşanlı Zone (Eren 1993). It is located in the Kutahya–Bolkardag belt (Özcan *et al.* 1988, 1990), which includes passive continental margin sequences of the Tauride–Anatolide Platform together with allochthonous, ophiolite-bearing assemblages (Özgül and Göncüoğlu 1998). It comprises HP–LT metamorphic rocks ranging in age from Permian to Cretaceous (Wiesner 1968; Göğer and Kiral 1969; Özcan *et al.* 1988; Eren 1993, 1996a) and ranging in grade from the epimetamorphic zone (Çelik and Karakaya 1999) to the blueschist facies.

The aim of this paper is to study the metamorphism of higher-grade units in the Altınekin Complex, and compare the results with those obtained from other parts of the Tavşanlı Zone. Specific objectives include (i) establishing the relative timing of deformation and blueschist–facies metamorphism, (ii) constraining the P – T conditions of metamorphism, hence geothermal gradients, within this segment of the subduction zone, and (iii) determining, as far as possible, P – T paths taken by the rocks. The results will be used to provide constraints on the mechanism by which the high-pressure rocks were exhumed. The absolute timing of the HP–LT metamorphism is the subject of a separate study.

2. GEOLOGY OF THE KONYA AREA

Metamorphic rocks crop out to the north of Konya in an arcuate massif collectively named the Bozdaglar Massif (Eren 1993; Figure 1(b)). Blueschists within this massif were originally thought to have formed during the Variscan Orogeny (Kaaden 1966), but the demonstration of Permian to Cretaceous protolith ages for rocks north-east of Konya (Karaman 1986) established a Cretaceous or Tertiary age for the metamorphism. Eren (1993, 1996a, 1996b, 2001) and Karakaya *et al.* (2002) described the stratigraphy and structural features of the Bozdaglar Massif and demonstrated that it consists of several tectonic slices.

The Bozdaglar Massif includes three main structural units, which are in tectonic contact with each other. These are, from west to the east, the autochthonous Upper Permian–Cretaceous Gökçeyurt Group, the allochthonous Silurian–Cretaceous Ladik Metamorphic Complex, and the U. Permian–Cretaceous Altınekin Metamorphic Complex (Figure 1(b)). The rocks of the Gökçeyurt Group, which forms part of the Afyon Zone of Okay (1984), have undergone very low-grade to low-grade greenschist–facies metamorphism. The Ladik Metamorphic Complex was metamorphosed under HP–LT conditions (Kaaden 1966; Bayıç 1968; Özcan *et al.* 1988; Kurt unpublished PhD thesis; Hekimbaşı unpublished PhD thesis) and was subjected to at least four deformational phases during the Alpine Orogeny (Eren 1996b, 2001). The Altınekin Metamorphic Complex was also subjected to HP–LT metamorphism, and is the subject of this paper.

2.1. The Altnekin Metamorphic Complex

The Permian–Cretaceous Altnekin Metamorphic Complex crops out northwest of Konya and tectonically overlies the Ladik Metamorphic Complex (Figure 1(b)). The Altnekin Complex comprises three structural units. These are, from bottom to top, an U. Permian–Cretaceous metasedimentary succession (which correlates with the autochthonous Gökçeyurt Group and parts of the allochthonous Ladik Complex), a Cretaceous accretionary complex and a Mesozoic ophiolite (Figure 2; Eren 2003).

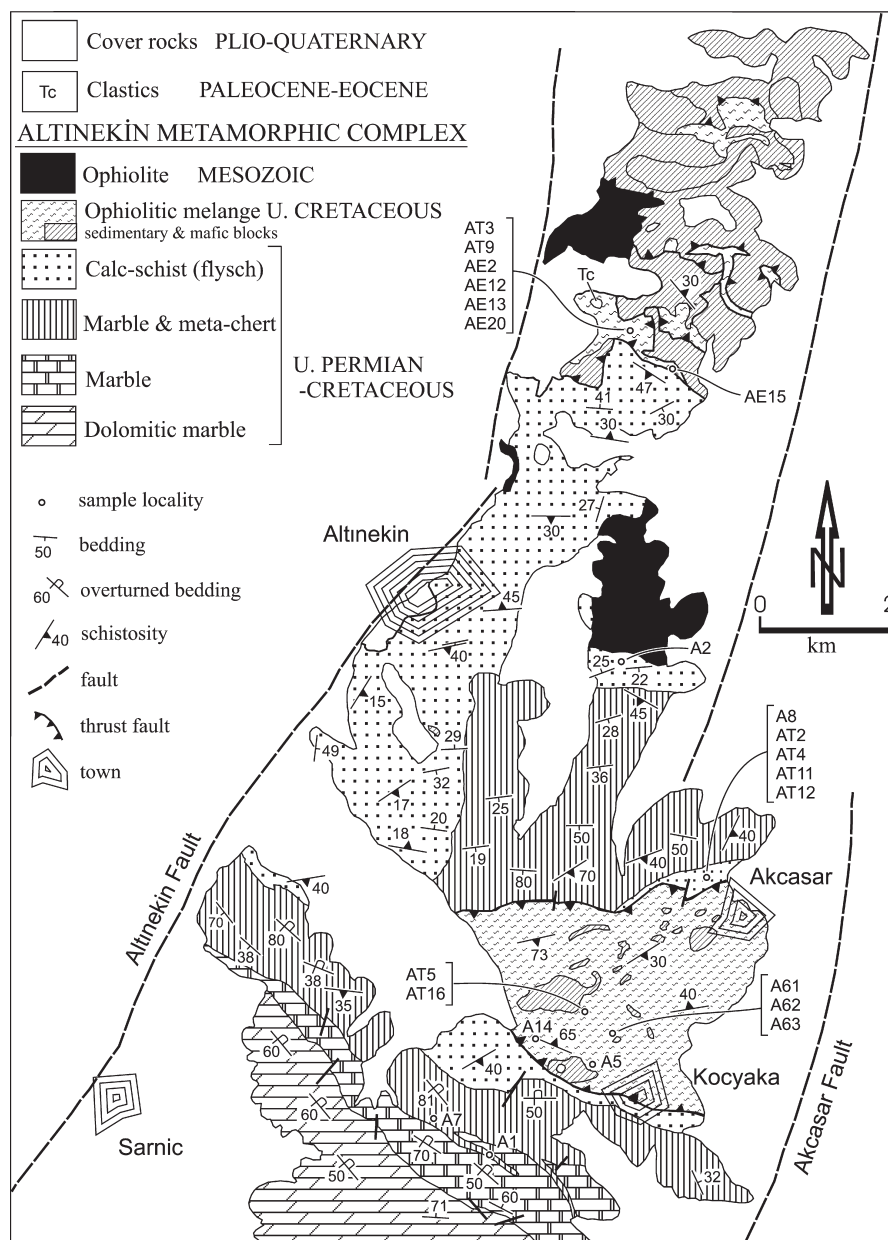


Figure 2. Geological sketch map of the Altnekin Complex, showing sample localities. All blank areas are outcrops of cover.

The U. Permian–Cretaceous metasedimentary succession consists of four different mappable units, which all interfinger. Black and grey dolomitic marbles of U. Permian–L. Jurassic age are present at the base. This unit grades laterally and vertically into a thick sequence of light coloured marbles of U. Triassic–Cretaceous age. In places, metacherts can be observed as intercalations within the marble. A Cretaceous unit of pelagic, thinly-bedded metacherts, calc-schists and marbles conformably overlies the marbles. The youngest unit consists of U. Cretaceous turbiditic metaclastics (flysch).

The Permian–Cretaceous metasedimentary succession represents the passive continental margin of the Menderes–Tauride Block (Özgül and Göncüoğlu 1999) and has been overthrust by the Cretaceous accretionary complex (the Koçyaka Ophiolitic Mélange; Karaman 1986; Eren 2003). The latter consists of metamorphosed mafic volcanics, dolerites, pyroclastic rocks, marbles, cherts, metaclastic rocks, serpentinites and blocks of marble and schist. It consists generally of imbricate tectonic slices between overlying metasedimentary rocks and underlying ophiolite slabs. Many of these rocks are strongly sheared and show penetrative deformation (Eren 2003). Özgül and Göncüoğlu (1999) describe the unit as an olistostrome. Metabasites in the mélange generally range in composition from Island Arc Tholeiite (IAT) to Mid-Oceanic Ridge Basalt (MORB) (Yılmaz *et al.* 1998), suggesting that their protoliths formed in a back-arc basin environment (Floyd *et al.* 2003). The mélange is overlain tectonically by meta-ophiolite (the Aktepe Ophiolite; Eren 2003), small outliers of which are scattered throughout the study area.

The intensity of metamorphism and deformation increases from the lowest structural unit to the uppermost unit, and from south to the north in the Bozdağlar Massif. Clay fractions of rocks of the U. Permian–Cretaceous metasedimentary series are composed of illite, chlorite and locally pyrophyllite, and illite crystallinity is in the epimetamorphic zone. Pumpellyite, epidote, albite, tremolite, glaucophane and prehnite also occur in these rocks (Karakaya unpublished MSc thesis 1992). Metabasites within the Koçyaka Ophiolitic Mélange contain sodic pyroxene, sodic amphibole, actinolite, chlorite, phengite, albite, epidote, garnet, lawsonite and pumpellyite, as well as relict igneous augite, hornblende and plagioclase (Özgül and Göncüoğlu 1999). No detailed information is yet available on the metamorphic conditions recorded by these rocks.

In the following sections, the petrography, mineral chemistry and phase relations of selected rocks from the northern part of the Altinekin Metamorphic Complex are described. These data are used to establish the time relations of deformation and metamorphism, deduce metamorphic reactions and calculate P – T conditions of metamorphism, the results of which enable us to quantify depths of subduction and geothermal gradients. Reaction textures and P – T data are used together to constrain the retrograde P – T paths taken by rocks in different units within the Altinekin Complex.

3. PETROGRAPHY OF THE ALTINEKİN METAMORPHIC COMPLEX

In this section we describe the mineralogy and textures of selected rocks from the northern part of the Altinekin Metamorphic Complex. Our observations overlap to some extent with those of Özgül and Göncüoğlu (1999), but concentrate on petrogenetically significant assemblages and textures. The mineral assemblages are listed in Table 1. Sample localities are shown in Figure 2.

3.1. Meta-igneous rocks of the Koçyaka Ophiolitic Mélange

The rocks described in this section are typical of the meta-igneous blocks within the Ophiolitic Mélange and flysch units.

Epidote blueschists. Metabasalts (e.g. samples AE13, AE20, AT5) are fine-grained granofelses typically containing sodic amphibole, chlorite, albite, epidote, phengite, quartz and titanite. The sodic amphibole, epidote and phengite commonly form subhedral to euhedral crystals up to 0.5 mm (exceptionally 1 mm) long with a decussate texture, associated with small (0.5–1 mm diameter) simple-twinned albite poikiloblasts and interstitial chlorite and quartz (Figure 3(a)). Calcite is locally present as either interstitial grains or porphyroblasts up to 2 mm across. The sodic amphiboles are strongly pleochroic and typically unzoned. The epidotes are yellow in plane-polarized light,

Table 1. Mineral assemblages of selected rocks of the Altınekin Complex. Mineral name abbreviations from Kretz (1983). Na-Amp, sodic amphibole (glaucofane or crossite). R, relict matrix phase; I, relict inclusion in major matrix phase; A, retrograde alteration product

<i>Metabasites (blocks in Koçyaka Ophiolitic Mélange):</i>	
AE13	Na-Amp + Chl + Ab + Qtz + Cal + Phe + Ttn (+ Ep + Ap)
AE20	Na-Amp + Ep + Chl + Ab + Qtz + Phe + Ttn
AT4	Na-Amp + Ep + Chl + Ab + Cal + Phe + Ttn (+ Act ^A)
AT5	Na-Amp + Ep + Chl + Ab + Qtz + Phe + Ttn
AT11	Na-Amp + Ep + Chl + Ab + Qtz + Cal + Ttn
AT16	Act + Na-Amp ^I + Ep + Chl + Ab + Cal + Ms + Ttn
A5	Na-Amp + Lws + Ab + Cal + Ttn (+ Chl ^A + Pmp ^A + Ep ^A)
A62	Pmp + Ep + Chl + Ab + Cal + Ttn (+ Omp ^I)
A63	Pmp + Ep + Chl + Lws ^I + Zo ^R + Ms + Ttn
<i>Phengite-Epidosites (blocks in Koçyaka Ophiolitic Mélange):</i>	
AT3	Ep + Chl + Phe + Ttn
AT9	Ep + Chl + Phe + Ttn
<i>Metachromitites (block in Koçyaka Ophiolitic Mélange):</i>	
A14	Chrom + Cr-Pmp + Cr-Phe
<i>Marbles, calc-schists & mica-schists (matrix of Koçyaka Ophiolitic Mélange):</i>	
AE2	Cal + Ank + Qtz + Mg-Chl + Cr-Phe
AE12	Tr + Talc + Cal + Mg-Chl
A61	Ep + Chl + Phe + Cal + Ab + Qtz (+ Na-Amp ^I + Trm + Ttn + Ore)
AE15	Na-Amp + Grt + Ep + Qtz (+ Phe + Tur + Ap)
<i>Calc-schists & mica-schists (U. Permian–Cretaceous metasediments):</i>	
A2	Na-Amp + Ep + Chl + Ab + Qtz + Phe + Cal (+ Ttn + Ap + Pyt)
A8	Na-Amp + Qtz + Phe + Cal (+ Ep + Ap)
AT2 (i)	Na-Amp + Cal + Qtz + Ttn + Chl ^A + Phe
(ii)	Na-Amp + Ep + Phe + Qtz + Ttn + Chl ^A
AT12	Chl + Cal + Qtz (+ Ab + Phe + Tur + Ore + Ap + Zrn)
<i>Micaceous quartzites (U. Permian–Cretaceous metasediments):</i>	
A1	Na-Amp + Phe + Qtz (+ Tur + Chl ^A + Ap)
A7	Na-Amp + Phe + Qtz (+ Tur + Chl ^A + Ap)

indicating appreciable Fe³⁺ contents. In metagabbros (e.g. AT11) the sodic amphibole forms 5 mm × 10 mm aggregates of aligned euhedral prisms and subordinate chlorite, presumably pseudomorphing a ferromagnesian mineral in the gabbroic protolith. Here, the amphibole prisms show colour zoning from dark cores to paler rims.

The epidote-blueschists are generally devoid of retrograde alteration, though in some (e.g. AT4) there is minor replacement of sodic amphibole by actinolite and of epidote by chlorite. In sample AT16, the greenschist-facies retrogression is dominant; here sodic amphibole is preserved mainly as randomly oriented inclusions within albite poikiloblasts, and the matrix consists largely of fine-grained, aligned actinolite and chlorite. Retrogression in this rock was clearly associated with penetrative deformation. The unretrograded metabasic epidote-blueschists have no tectonic fabric.

Lawsonite blueschist sample A5 is a medium-grained granofels containing randomly oriented, unzoned, euhedral prisms of sodic amphibole up to 3 mm × 0.5 mm, separated by poikiloblasts of albite and calcite enclosing clusters of small (<0.1 mm) euhedral lawsonite crystals (Figure 3(b)). Aggregates of titanite granules are also abundant. The albite + lawsonite domains presumably represent the plagioclase sites in the igneous protolith. Tiny lawsonite inclusions occur within some of the amphiboles. Sodic amphibole shows local replacement by retrograde chlorite, epidote and dark green pumpellyite. Sodic pyroxene and aragonite are absent.

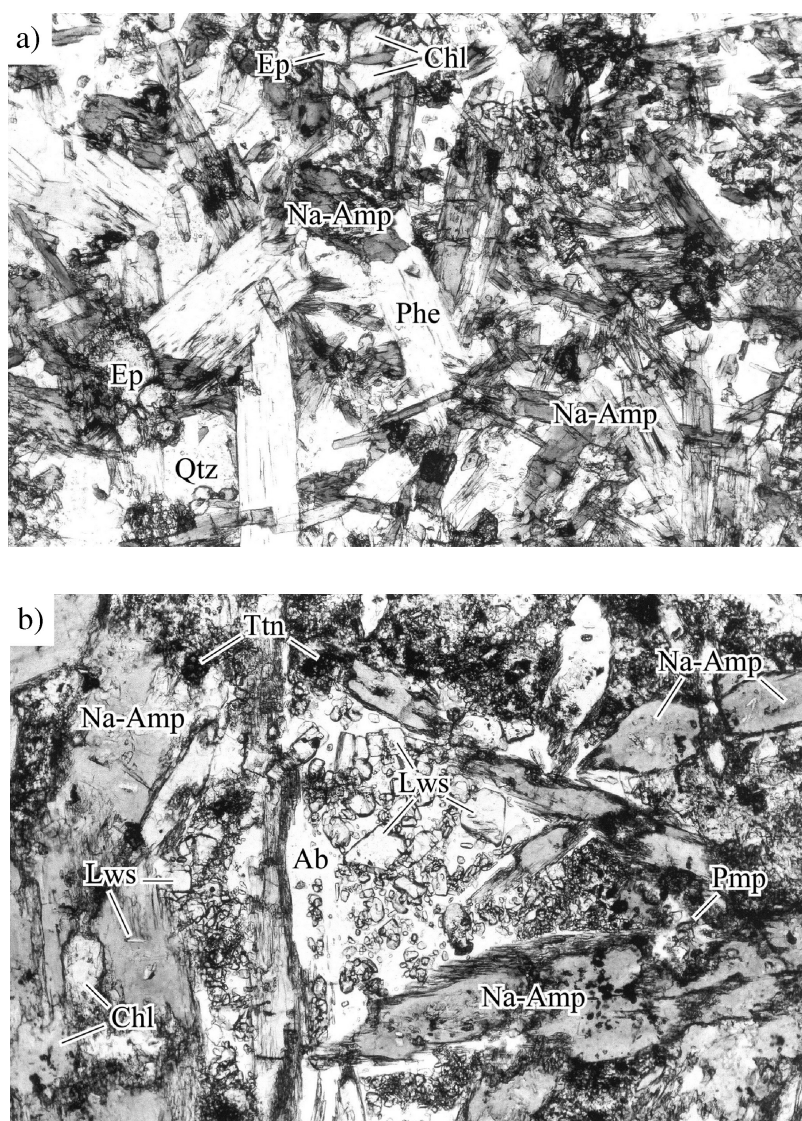


Figure 3. Photomicrographs showing petrographic features of selected rocks of the Altinekin Complex. Fields of view $1.6\text{ mm} \times 1.07\text{ mm}$. Mineral name abbreviations from Kretz (1983). (a) Randomly oriented sodic amphibole (Na-Amp), epidote (Ep), chlorite (Chl), phengite (Phe) and quartz (Qtz). Epidote blueschist AT5. Plane-polarized light. (b) Lawsonite crystals (Lws) in an albite poikiloblast (Ab) between prisms of sodic amphibole (Na-Amp) and granules of titanite (Ttn). Sodic amphiboles, which locally contain lawsonite inclusions, are partially altered to chlorite (Chl) and pumpellyite (Pmp). Lawsonite blueschist A5. Plane-polarized light. (c) Inclusion of omphacite (Omp) in epidote (Ep) in a matrix of chlorite (Chl), pumpellyite (Pmp) and calcite (Cal). Pumpellyite greenstone A62. Plane-polarized light. (d) Inclusions of lawsonite (Lws) in micropoikiloblasts of pumpellyite (Pmp). Pumpellyite greenstone A63. Crossed polars. (e) Aggregate of euhedral epidote (Ep) and pumpellyite (Pmp) in chlorite (Chl) in a matrix of pumpellyite and relict zoisite (Zo). Pumpellyite greenstone A63. Plane-polarized light. (f) Boudinaged sodic amphibole prism (Na-Amp) and granules of quartz (Qtz) in a matrix of strained calcite (Cal). Calc-schist A8. Plane-polarized light.

Pumpellyite greenstones. These rocks are fine-grained epidote–chlorite–pumpellyite granofelsels that lack a tectonic fabric and contain relict high-pressure minerals. Sample A62 contains abundant euhedral prisms up to 2 mm long in a fine-grained matrix of chlorite, green pleochroic pumpellyite and titanite. Albite and calcite form small poikiloblasts. The epidote contains rare, partially altered inclusions of relict green sodic pyroxene (Figure 3(c)). Sample A63 is extremely rich in pumpellyite and is totally devoid of albite; its Ca,Al-rich composition suggests that the protolith was some sort of calc-silicate rock, possibly a rodingite. The bulk of the rock consists of

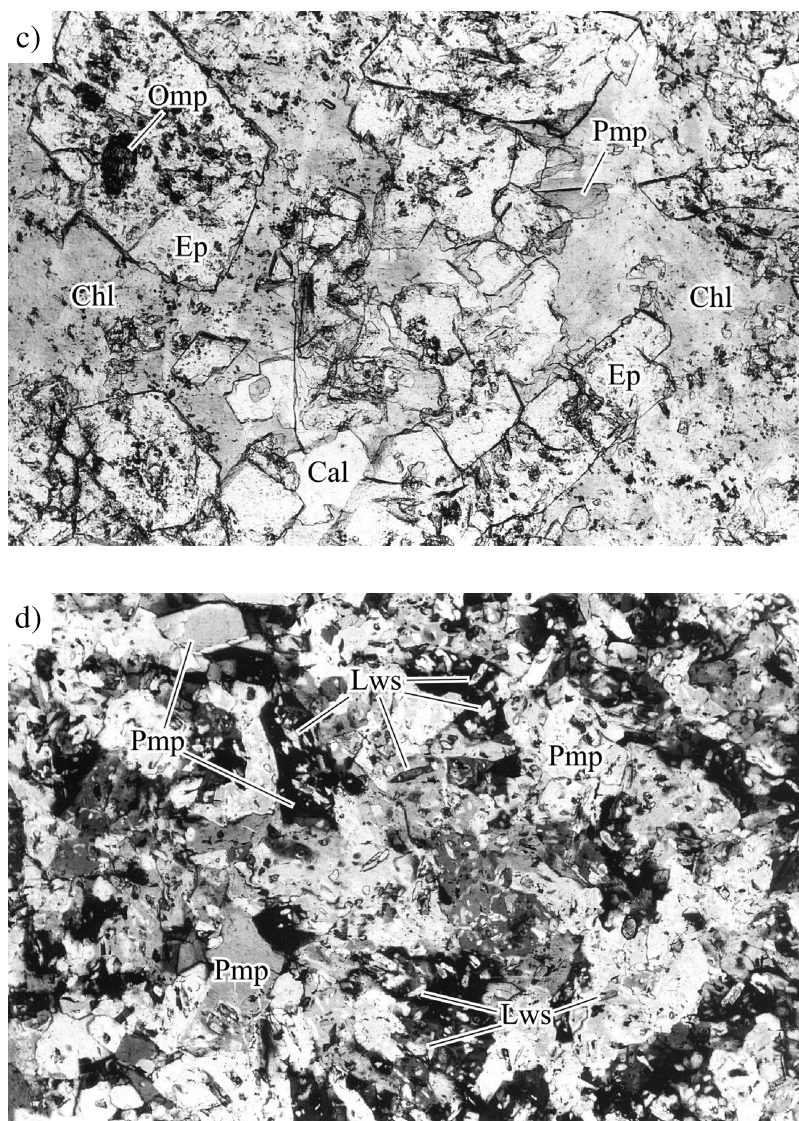


Figure 3. Continued.

interlocking pumpellyite micro-poikiloblasts, up to 0.5 mm long, enclosing tiny inclusions of lawsonite (Figure 3(d)). Between the pumpellyites are small, late muscovites and ragged relict prisms of zoisite. Chlorite occurs as scattered aggregates in which pumpellyite and epidote are euhedral (Figure 3(e)).

Phengite epidosites (e.g. AT3, AT9) are granofelses containing randomly oriented, pale yellow, subhedral epidote prisms up to 2 mm × 1 mm in a fine-grained matrix of olive-green phengite, chlorite and minor titanite. The protoliths were probably metasomatized basalts.

Meta-chromitite sample A14 consists mainly of angular relict grains of chromite up to 1 mm across, commonly in clusters up to 4 mm across which represent the remains of original igneous chromites. Relict grains have dark orange–red cores (presumably igneous compositions) and irregular opaque rims. The chromites are fringed by tiny (<0.1 mm long) prisms of pink pleochroic Cr-pumpellyite ('shuiskite') and separated by patches and veins of randomly oriented, fine-grained, pale green chromian phengite.

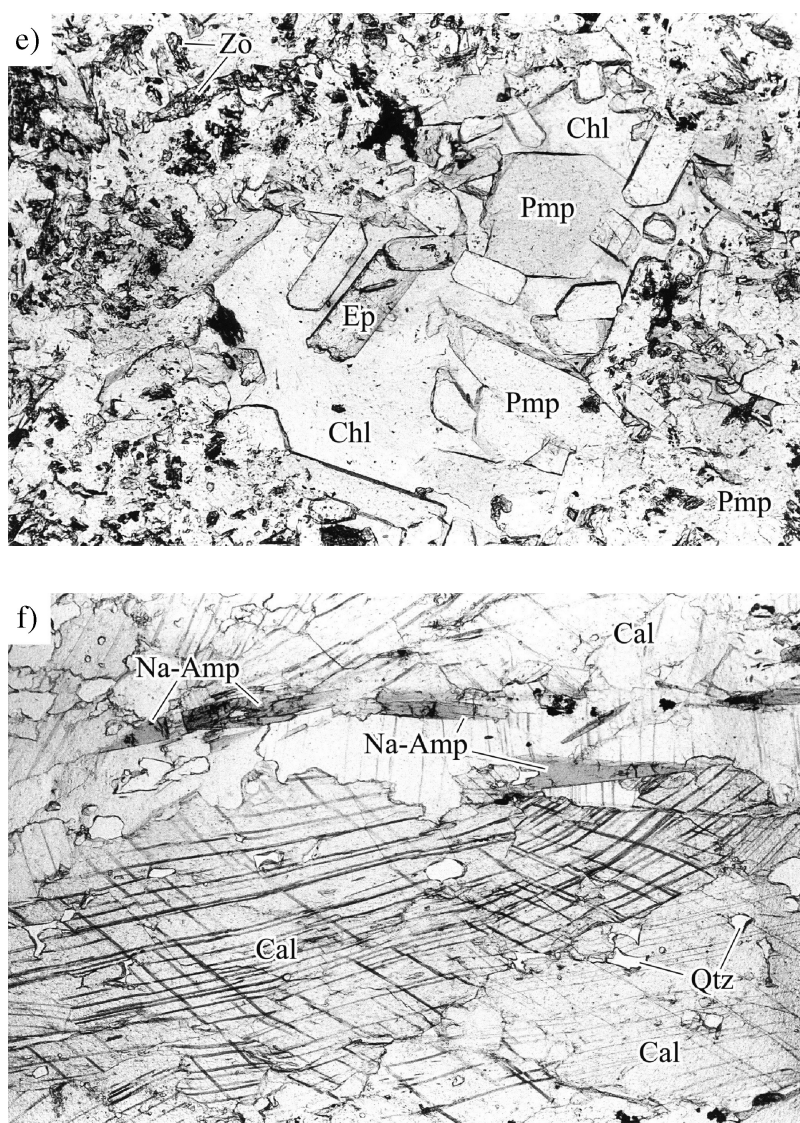


Figure 3. Continued.

3.2. Metasediments

The rocks described in this section are typical of the metamorphosed Cretaceous sediments of the metasedimentary series and the matrix of the Koçyaka Ophiolitic Mélange.

Calc-schists. Sodic-amphibole-bearing calc-schists of the Cretaceous flysch (e.g. A2, A8) consist of aligned prisms of strongly pleochroic sodic amphibole and, locally, epidote in a matrix of granoblastic calcite and quartz and, in some varieties, aligned flakes of chlorite and/or phengite of high aspect ratio. Albite, where present, forms post-tectonic porphyroblasts up to 1 mm across. The amphibole shows slight colour zoning (with paler rims) and is locally boudinaged (Figure 3(f)). The zoning is cut by the boudinage fractures indicating that extension postdated amphibole growth (Figure 3(f)). The epidote is partially replaced by chlorite and calcite. Calcite and quartz both show strongly undulose extinction indicative of penetrative retrograde strain. Calc-schist sample A6, from the Ophiolitic Mélange, is similar but shows more pronounced late-metamorphic strain and lacks matrix sodic

amphibole, the latter occurring only as small relict inclusions within albite porphyroblasts. Other samples (e.g. AT12) are amphibole free and contain quartz and calcite with much intracrystalline strain.

Sample AT2, from the flysch, contains both calcareous and non-calcareous layers. The calcareous layer (AT2i) contains radiating sprays of sodic amphibole in a granoblastic matrix of calcite and quartz. The non-calcareous layer (AT2ii) is a sodic amphibole epidote mica schist in which amphibole porphyroblasts and titanite and epidote microporphyroblasts cut across a penetrative, fine-grained phengite schistosity, indicating that the peak of blueschist facies metamorphism postdated at least one phase of plastic deformation. In the same layer, the amphibole has been boudinaged and altered to retrograde chlorite in extensional fractures, and the penetrative schistosity is locally transected by an oblique extensional crenulation cleavage. These observations indicate that the rock underwent a second phase of deformation after the peak of metamorphism.

Marbles. Impure marbles are common within the Ophiolitic Mélange. Tremolite marble sample AE12 contains partly aligned euhedral tremolite prisms up to 3 mm long and scattered flakes of talc and colourless Mg-chlorite in a matrix of granoblastic calcite. Quartz marble sample AE2 consists of brown subhedral rhombs of ankerite up to 3 mm across in a matrix of granoblastic calcite and quartz containing aligned flakes of Mg-chlorite and chromian phengite. The granoblastic minerals in both samples show undulose extinction indicative of late-stage deformation.

Quartzites. Metacherts of the metasedimentary series are schistose sodic-amphibole-bearing micaceous quartzites. In many samples (e.g. A1 and A7), the sodic amphibole forms euhedral, unzoned, strongly pleochroic prisms up to 5 mm long set in a fine-grained schistose matrix of strained and sutured granoblastic quartz and subordinate aligned phengite. The long axes of the amphibole porphyroblasts lie parallel to, but locally appear to cut, the phengite schistosity, but are randomly oriented within that plane. In other samples, the amphibole shows some alignment and boudinage, indicating post-metamorphic strain.

Metacherts also occur within the Ophiolitic Mélange. Sample AE15 is similar to the above except that phengite is absent and garnet and epidote are present. The amphibole is aligned and boudinaged, and the epidote forms yellow prisms up to 1 mm long parallel to the amphibole lineation. The garnet, probably spessartine, occurs as tiny (<0.2 mm) euhedral crystals.

4. MINERAL CHEMISTRY

Major-element analyses of minerals in selected rocks from the Altnekin Metamorphic Complex were obtained by EDS using the converted Geoscan electron microprobe at the University of Manchester, operating at 15 kV, 3 nA specimen current on Co metal, and 40 second count time per analysis. X-ray spectra were processed using Link Systems ZAF4/FLS software.

Representative analyses are listed in Tables 2–4.

4.1. Amphiboles

Fe^{3+} -corrected amphibole cation formulae were calculated assuming $\Sigma(\text{Si}, \text{Ti}, \text{Al}, \text{Cr}, \text{Fe}, \text{Mn}, \text{Mg}) = 13.00$ per 23(O) anhydrous (Droop 1987).

Sodic amphiboles are mainly glaucophane and ferroglaucophane in the IMA classification scheme (Leake 1997), and have virtually empty A sites and tetrahedral and M4 sites essentially full of Si and Na, respectively (Table 2). $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ values range from 0.33 to 0.71; values <0.5 characterize the amphiboles of the metasediments and lawsonite-blueschist, whilst values >0.5 characterize those of the epidote-blueschists (Figure 4). Calculated values of $\text{Al}/(\text{Al} + \text{Fe}^{3+})$ range from 0.4 to 1, with amphiboles of epidote-blueschists spanning the whole range, and amphiboles of the metasediments and lawsonite-blueschist mostly having values >0.75 (Figure 4). Within any one sample, there is considerable variation in $\text{Al}/(\text{Al} + \text{Fe}^{3+})$; part of this spread is probably due to uncertainties in the Fe^{3+} calculation method, but in some samples the variation reflects zoning, with high- Fe^{3+} analyses correlating with the dark-coloured cores.

Table 2. Representative electron-microprobe analyses of amphiboles, omphacite and albite

	Na-Amp A2 core	Na-Amp A2 rim	Na-Amp A5 mid	Na-Amp AE13 small prism	Na-Amp AT11 core	Na-Amp AT2 rim	Act AT4 replacing Na-Amp	Act AT16 matrix needle	Omp A62 inclusion in Ep	Albite A2 core of porph.	Albite A5 rim of poik.	Albite AE13 core of porph.
SiO ₂	55.79	56.17	56.06	58.14	56.29	56.45	51.32	55.39	55.16	67.98	68.33	68.79
TiO ₂	0.09	0.01	0.00	0.02	0.07	0.06	0.00	0.00	0.00	0.00	0.00	0.03
Al ₂ O ₃	9.03	9.16	10.57	10.33	5.69	8.76	5.49	2.02	8.63	19.19	19.26	19.59
Cr ₂ O ₃	0.02	0.00	0.05	0.00	0.00	0.00	0.33	0.00	0.00	0.10	0.00	0.00
Fe ₂ O ₃	3.10	2.38	2.21	2.63	9.09	4.45	4.06	1.60	4.36	—	—	—
FeO	15.24	15.32	15.93	8.79	8.75	12.71	7.19	8.61	6.65	0.06	0.14	0.01
MnO	0.36	0.25	0.18	0.14	0.45	0.18	0.00	0.00	0.03	0.00	0.00	0.02
MgO	5.82	6.25	5.18	9.85	9.45	7.11	15.56	16.55	5.50	0.00	0.00	0.00
CaO	0.33	0.23	0.12	0.30	1.00	0.11	10.68	10.92	11.92	0.29	0.05	0.03
Na ₂ O	7.19	7.60	7.28	7.33	6.53	7.18	1.87	1.02	7.38	11.74	11.67	11.91
K ₂ O	0.00	0.00	0.00	0.00	0.00	0.03	0.00	0.00	0.00	0.00	0.00	0.00
Total	96.97	97.37	97.58	97.53	97.32	97.04	96.50	96.11	99.63	99.36	99.45	100.38
Formula	23(O)	23(O)	23(O)	23(O)	23(O)	23(O)	23(O)	23(O)	6(O)	8(O)	8(O)	8(O)
Si	8.01	8.01	7.97	8.00	7.98	8.01	7.39	7.93	2.02	2.99	3.00	3.00
Ti	0.01	0.00	0.00	0.00	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.00
Al	1.53	1.54	1.77	1.67	0.95	1.47	0.93	0.34	0.37	1.00	1.00	1.00
Cr	0.00	0.00	0.01	0.00	0.00	0.00	0.04	0.00	0.00	0.00	0.00	0.00
Fe ³⁺	0.33	0.26	0.24	0.27	0.97	0.48	0.44	0.17	0.12	—	—	—
Fe ²⁺	1.83	1.83	1.89	1.01	1.04	1.51	0.87	1.03	0.20	0.00	0.01	0.00
Mn	0.04	0.03	0.02	0.02	0.05	0.02	0.00	0.00	0.00	0.00	0.00	0.00
Mg	1.25	1.33	1.10	2.02	2.00	1.51	3.34	3.53	0.30	0.00	0.00	0.00
Ca	0.05	0.04	0.02	0.04	0.15	0.02	1.65	1.67	0.47	0.01	0.00	0.00
Na	2.00	2.10	2.01	1.96	1.80	1.98	0.52	0.28	0.52	1.00	0.99	1.00
K	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00
Total	15.05	15.14	15.03	14.99	14.95	15.02	15.18	14.95	4.00	5.00	5.00	5.00
Mg/ (Mg+Fe ²⁺)	0.41	0.42	0.37	0.67	0.66	0.50	0.79	0.77	0.60	—	—	—

Table 3. Representative electron-microprobe analyses of micas and chlorites. All Fe assumed to be Fe²⁺, na: not analysed

	Ms A62	Ms A63	Phc AE13	Phc AT9	Phc AT5	Cr-Phc AT14	Chl A2	Chl A5	Chl A62	Chl A63	Chl AE13	Chl AT11	Chl AT5
SiO ₂	46.23	45.33	52.05	50.73	50.72	50.64	25.95	25.06	26.26	27.73	27.59	28.36	26.71
TiO ₂	0.04	0.27	0.09	0.05	0.00	0.20	0.05	0.04	0.06	0.00	0.00	0.00	0.08
Al ₂ O ₃	29.59	36.47	24.78	25.67	26.46	16.96	19.50	18.44	19.60	18.31	18.83	19.19	19.36
Cr ₂ O ₃	0.00	0.00	0.00	0.07	0.00	10.28	0.00	0.06	0.00	0.14	0.10	0.00	0.00
FeO	4.47	0.87	2.75	3.40	2.58	0.83	22.66	34.97	24.97	20.34	19.32	14.79	21.30
MnO	0.02	0.00	0.11	0.00	0.00	0.06	0.29	0.51	0.40	0.17	0.14	0.30	0.33
MgO	3.24	0.57	4.19	3.60	3.49	5.05	16.98	8.92	16.10	19.40	19.79	23.44	18.10
CaO	0.13	0.14	0.12	0.09	0.01	0.09	0.32	0.09	0.08	0.18	0.06	0.02	0.00
Na ₂ O	0.35	0.55	0.25	0.00	0.00	0.34	na	na	na	na	na	na	na
K ₂ O	10.73	10.81	10.76	10.45	10.84	9.92	na	na	na	na	na	na	na
Total	94.80	95.01	95.10	94.06	94.10	94.37	85.75	88.09	87.47	86.27	85.83	86.10	85.89
Formula	11(O)	11(O)	11(O)	11(O)	11(O)	11(O)	14(O)	14(O)	14(O)	14(O)	14(O)	14(O)	14(O)
Si	3.16	3.03	3.49	3.45	3.43	3.51	2.77	2.79	2.77	2.89	2.87	2.87	2.81
Ti	0.00	0.01	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.01
Al	2.39	2.87	1.96	2.05	2.11	1.38	2.45	2.42	2.44	2.25	2.31	2.29	2.40
Cr	0.00	0.00	0.00	0.00	0.00	0.56	0.00	0.01	0.00	0.01	0.01	0.00	0.00
Fe	0.26	0.05	0.15	0.19	0.15	0.05	2.02	3.25	2.20	1.77	1.68	1.25	1.88
Mn	0.00	0.00	0.01	0.00	0.00	0.00	0.03	0.05	0.04	0.01	0.01	0.03	0.03
Mg	0.33	0.06	0.42	0.36	0.35	0.52	2.70	1.48	2.53	3.02	3.07	3.54	2.84
Ca	0.01	0.01	0.01	0.01	0.00	0.01	0.04	0.01	0.01	0.02	0.01	0.00	0.00
Na	0.05	0.07	0.03	0.00	0.00	0.05	na	na	na	na	na	na	na
K	0.94	0.92	0.92	0.91	0.94	0.88	na	na	na	na	na	na	na
Total	7.14	7.02	6.99	6.97	6.98	6.97	10.01	9.99	9.99	9.97	9.96	9.98	9.97
Mg/(Mg + Fe)	0.56	0.54	0.73	0.65	0.71	0.92	0.57	0.31	0.53	0.63	0.65	0.74	0.60

Table 4. Representative electron-microprobe analyses of epidote-group and pumpellyite-group minerals and lawsonites

	Zoisite		Epidote		Epidote		Epidote		Epidote		Epidote		Pmp		Pmp		Shuiskite		Lws	
	A63	A2	A5	A62	A63	A62	A63	AT2	AT4	AT5	A62	A63	A5	A63	A5	A63	A14	A5	A63	A63
SiO ₂	39.21	36.87	38.05	37.92	38.30	37.92	38.30	37.92	37.52	37.62	37.15	36.88	37.10	34.60	37.79	36.96	34.60	37.79	36.96	36.96
TiO ₂	0.00	0.07	0.00	0.07	0.14	0.07	0.14	0.00	0.00	0.00	0.14	0.00	0.04	0.00	0.22	0.00	0.00	0.22	0.00	0.00
Al ₂ O ₃	32.89	21.80	24.35	25.05	26.90	25.05	26.90	24.63	23.69	24.23	25.05	25.01	24.83	10.42	31.29	31.67	10.42	31.29	31.67	31.67
Cr ₂ O ₃	0.05	0.11	0.07	0.05	0.06	0.05	0.06	0.00	0.09	0.00	0.02	0.08	0.10	20.59	0.00	0.03	20.59	0.00	0.03	0.03
Fe ₂ O ₃	0.25	14.54	12.60	11.64	8.43	11.64	8.43	11.85	12.31	12.13	1.71	1.78	2.56	1.09	—	—	1.09	—	—	—
FeO	—	—	—	—	—	—	—	—	—	—	2.40	1.66	3.26	0.68	0.45	0.16	0.68	0.45	0.16	0.16
MnO	0.00	1.26	0.12	0.05	0.02	0.05	0.02	0.07	0.20	0.27	0.56	0.33	0.38	0.63	0.00	0.12	0.63	0.00	0.12	0.12
MgO	0.00	0.02	0.08	0.18	0.09	0.18	0.09	0.00	0.00	0.00	3.06	3.08	2.22	3.25	0.18	0.00	3.25	0.18	0.00	0.00
CaO	24.69	22.00	23.30	23.31	23.76	23.31	23.76	23.04	23.29	22.79	22.43	22.84	22.92	19.99	17.70	17.29	19.99	17.70	17.29	17.29
Na ₂ O	0.00	0.00	0.00	0.05	0.00	0.05	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.38	0.10	0.00	0.38	0.10	0.00	0.00
K ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	97.09	96.67	98.57	98.32	97.70	98.32	97.70	97.51	97.10	97.04	92.52	91.66	93.41	91.63	87.73	86.23	91.63	87.73	86.23	86.23
Formula	12.5(O)	12.5(O)	12.5(O)	12.5(O)	12.5(O)	12.5(O)	12.5(O)	12.5(O)	12.5(O)	12.5(O)	24.5(O)	24.5 (O)	24.5(O)	24.5(O)	8(O)	8(O)	24.5(O)	8(O)	8(O)	8(O)
Si	3.00	3.00	3.00	2.99	3.00	2.99	3.00	3.01	3.01	3.01	5.99	5.98	5.98	6.02	2.00	1.99	6.02	2.00	1.99	1.99
Ti	0.00	0.00	0.00	0.00	0.01	0.00	0.01	0.00	0.00	0.00	0.02	0.00	0.01	0.00	0.01	0.00	0.00	0.01	0.00	0.00
Al	2.97	2.09	2.26	2.32	2.48	2.32	2.48	2.30	2.24	2.28	4.76	4.79	4.71	2.13	1.95	2.01	2.13	1.95	2.01	2.01
Cr	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.01	0.01	2.83	0.00	0.00	2.83	0.00	0.00	0.00
Fe ³⁺	0.01	0.89	0.75	0.69	0.50	0.69	0.50	0.71	0.74	0.73	0.21	0.22	0.31	0.14	—	—	0.14	—	—	—
Fe ²⁺	—	—	—	—	—	—	—	—	—	—	0.32	0.23	0.44	0.10	0.02	0.01	0.10	0.02	0.01	0.01
Mn	0.00	0.09	0.01	0.00	0.00	0.00	0.00	0.00	0.01	0.02	0.08	0.04	0.05	0.09	0.00	0.01	0.09	0.00	0.01	0.01
Mg	0.00	0.00	0.01	0.02	0.01	0.02	0.01	0.00	0.00	0.00	0.74	0.75	0.53	0.84	0.01	0.00	0.84	0.01	0.00	0.00
Ca	2.02	1.92	1.97	1.97	1.99	1.97	1.99	1.96	2.00	1.95	3.88	3.98	3.96	3.72	1.01	1.00	3.72	1.01	1.00	1.00
Na	0.00	0.00	0.00	0.01	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.13	0.01	0.00	0.13	0.01	0.00	0.00
K	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	8.00	8.00	8.00	8.00	7.99	8.00	7.99	7.98	8.01	7.99	16.00	16.00	16.00	16.00	5.01	5.02	16.00	5.01	5.02	5.02
Mg/(Mg + Fe ²⁺)	—	—	—	—	—	—	—	—	—	—	0.69	0.77	0.55	0.89	—	—	0.89	—	—	—
Al/(Al + Fe ³⁺ + Cr)	0.99	0.70	0.75	0.77	0.83	0.77	0.83	0.77	0.75	0.76	0.96	0.95	0.94	0.42	—	—	0.42	—	—	—

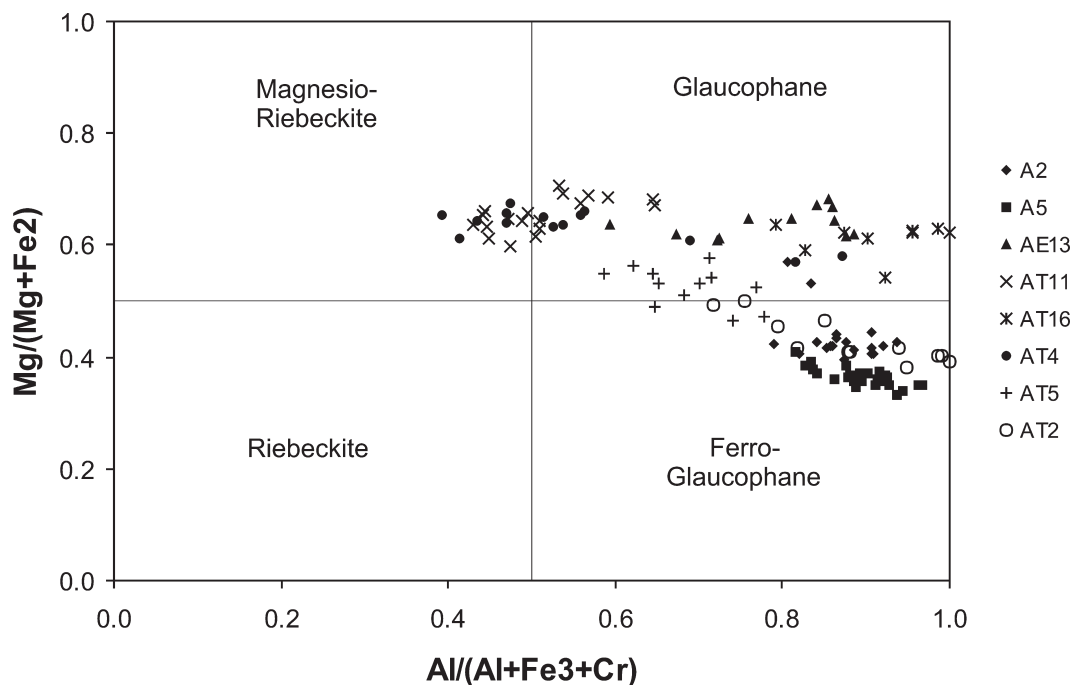


Figure 4. Compositions of analysed sodic amphiboles.

Actinolite overgrowing and replacing sodic amphiboles in retrograded epidote-blueschists AT4 and AT16 has appreciable Na in M4 sites (up to 0.54 a.p.f.u.). $\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$ values in both samples fall in the range 0.76–0.81.

4.2. Clinopyroxenes

Sodic pyroxene cation formulae were calculated by specifying a total of 4.00 cations per 6 oxygens (Droop 1987). The sodic pyroxene relicts in pumpellyite-greenstone A62 have compositions that plot in the omphacite field in the diagram of Essene and Fyfe (1967) (Figure 5). Na contents are fairly uniform (0.52–0.57 a.p.f.u. (atoms per formula unit)). $\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$ values range from 0.49 to 0.6 and Fe^{3+} contents range from 0.08 to 0.15 a.p.f.u.

4.3. Epidote-group minerals

Cation formulae of epidote-group minerals were calculated assuming all iron to be Fe^{3+} .

Zoisites in pumpellyite-greenstone A63 are virtually pure (Table 4), with Fe^{3+} contents generally less than 0.02 a.p.f.u.

Epidotes mostly have Fe^{3+} contents in the range 0.65–0.85 a.p.f.u. (Table 4). Exceptions are the epidotes in retrograded blueschist AT16 and pumpellyite-greenstone A63, which are more aluminous (0.45–0.52 and 0.48–0.53 Fe^{3+} a.p.f.u., respectively), and epidotes in mica-schist AT2ii, which are more ferrian (0.81–0.89 Fe^{3+} a.p.f.u.).

4.4. White micas

Mica formulae were calculated assuming all iron to be Fe^{2+} .

Phengites in epidiosites and unretrograded blueschists generally have Si contents in the range 3.4–3.5 per 11(O) formula unit (Table 3), indicative of moderately high amounts of celadonite substitution; such values are typical of phengites in many metabasic blueschists (e.g. Okay 1980a; El Shazly 1994) but are significantly lower than in some (e.g. Sorensen 1986). $\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$ values range from 0.55 to 0.75. Phengites in retrograded blueschists and in calc-schists possessing a late foliation have lower Si contents (e.g. 3.34–3.40 a.p.f.u. in A2). Na is

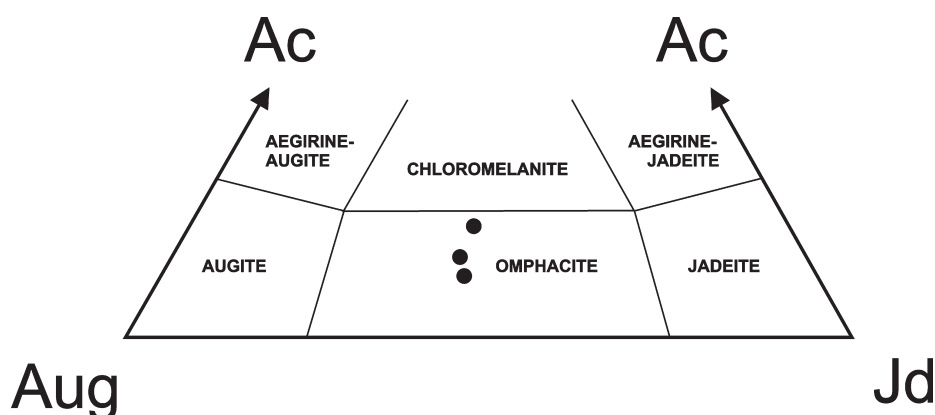


Figure 5. Compositions of relict sodic pyroxene inclusions in pumpellyite greenstone A62, on the jadeite–acmite–augite diagram of Essene and Fyfe (1967).

undetectable in most phengites, the exceptions being those of samples AE13 and A2 which contain 0.03–0.05 and 0.05–0.08 Na a.p.f.u., respectively.

Muscovites in pumpellyite–greenstones contain a maximum of 3.2 Si a.p.f.u. and have very variable $\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$ values (0.40–0.85 in A63), suggestive of disequilibrium.

Chromian phengites in metachromitite A14 contain 3.48–3.54 Si, 0.55–0.69 Cr and 0.01–0.05 Na a.p.f.u. and have $\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$ values in the range 0.89–0.92.

4.5. Chlorite

Chlorite formulae were calculated assuming all iron to be Fe^{2+} . Chlorites show a fairly narrow spread of Si contents, ranging from 2.75 to 2.96 a.p.f.u. (Figure 6, Table 3). In blueschists, $\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$ values of peak-metamorphic chlorite correlate with those of coexisting sodic amphibole, ranging from 0.57 to 0.76. The

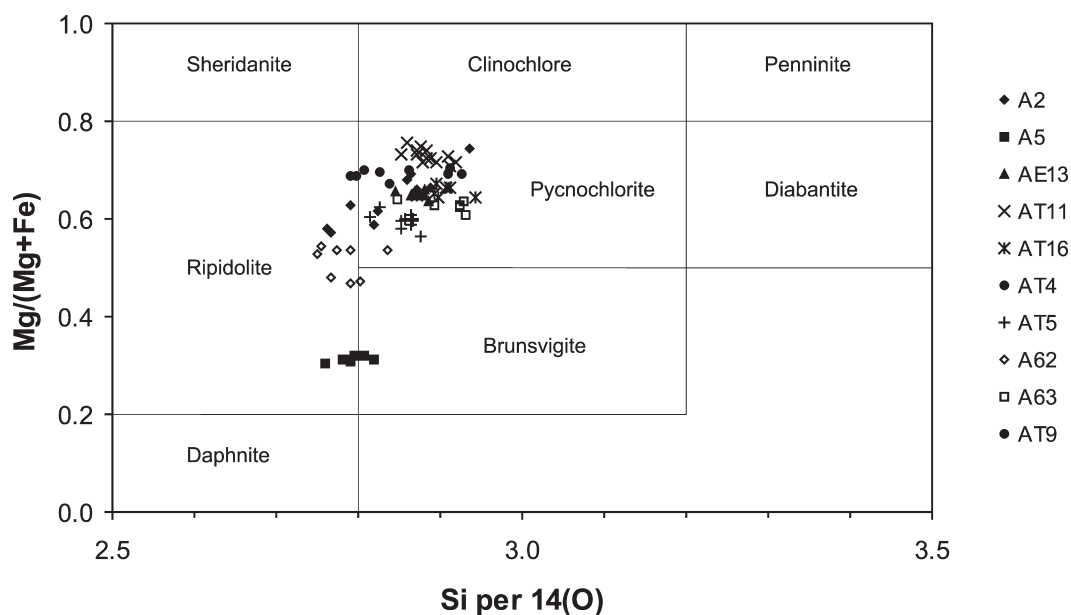


Figure 6. Compositions of analysed chlorites. Classification from Hey (1954).

retrograde chlorites in lawsonite blueschist A5 are relatively iron rich, with $\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$ values of 0.30–0.32. Chlorites in schist A2 show a large range in composition (Figure 6), with retrograde grains having lower Si and higher $\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$ than primary, schistosity-parallel flakes. $\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$ values of chlorites coexisting with pumpellyite in A62 and A63 range from 0.47 to 0.54 and 0.61 to 0.64, respectively.

4.6. Pumpellyite

All pumpellyite analyses were processed assuming an ideal formula $\text{W}_4\text{X}_2\text{Y}_4\text{Z}_6\text{O}_{21}(\text{OH})_7$ (Coombs *et al.* 1976). Fe^{3+} contents were calculated by the method of Cortesogno *et al.* (1984), which assumes a charge-balanced total of 16.00 cations per 24.5(O) anhydrous formula unit.

Pumpellyites have Si and Ca contents in the ranges 5.96–6.05 and 3.85–3.98 atoms per 24.5(O), respectively (Table 4), indicating that the W sites contain virtually their full complement of Ca, and that the Z sites are full of Si. Al contents range from 4.61 to 4.85 a.p.f.u., indicating that the X sites are *c.* one-third full of Al, the remainder being Fe^{3+} (0.15–0.37), Mg and Fe^{2+} . $\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$ values of pumpellyites in greenstones A62 and A63 lie mostly in the ranges 0.63–0.72 and 0.69–0.79, respectively, whilst those of retrograde pumpellyites in lawsonite-blueschist A5 range from 0.34 to 0.50.

Shuiskites in metachromitite A14 resemble the low-Cr pumpellyites described above, except that over half the Al is substituted by Cr, the latter totalling between 2.74 and 3.69 a.p.f.u.

4.7. Lawsonite

All analysed lawsonites are essentially pure (Table 4), the only significant impurity being Fe (up to 0.03 a.p.f.u.).

4.8. Albite

Albites in all analysed rocks have compositions very close to the end-member, with a maximum of 1% anorthite component (Table 2).

4.9. Chromite

Ferric iron contents of chromites were calculated assuming a total of 3.00 cations per 4.00 oxygens (Droop 1987). The reddish brown relict igneous chromites in metachromitite A14 have compositions averaging $\text{Spl}_{35}\text{Chr}_{30}\text{MgChr}_{31}\text{Galax}_{01}\text{Trev}_{01}\text{Mt}_{02}$, whilst the opaque rims have compositions averaging $\text{Spl}_{06}\text{Hc}_{18}\text{Chr}_{59}\text{Galax}_{13}\text{Mt}_{04}$.

5. PHASE RELATIONS AND METAMORPHIC REACTIONS

Phase relations in metabasites of the Altnekin Complex are displayed on ACF projections from albite, quartz and water in Figure 7 and on AFM projections from epidote, albite, quartz and water (Harte and Graham 1975) in Figure 8. Al and Fe^{3+} were combined into a single component for both types of projection.

The typical peak-metamorphic mineral assemblage in the epidote-blueschists is sodic amphibole + epidote + chlorite + albite + quartz + phengite + titanite (Table 1; Figure 7(a),(b); Figure 8(a),(b)). The most common type of retrogression in these rocks is the replacement of sodic amphibole by actinolite and of epidote by chlorite; taken with the observation that the actinolite–chlorite tie-line crosses the sodic amphibole–epidote tie-line on the ACF projection (Figure 7(c)); this indicates that the retrogression may be accounted for by the Fe,Mg-continuous reaction



This reaction is widely held to define the transition from epidote-blueschist to greenschist facies (see e.g. Liou *et al.* 1985; Maruyama *et al.* 1986; Evans 1990).

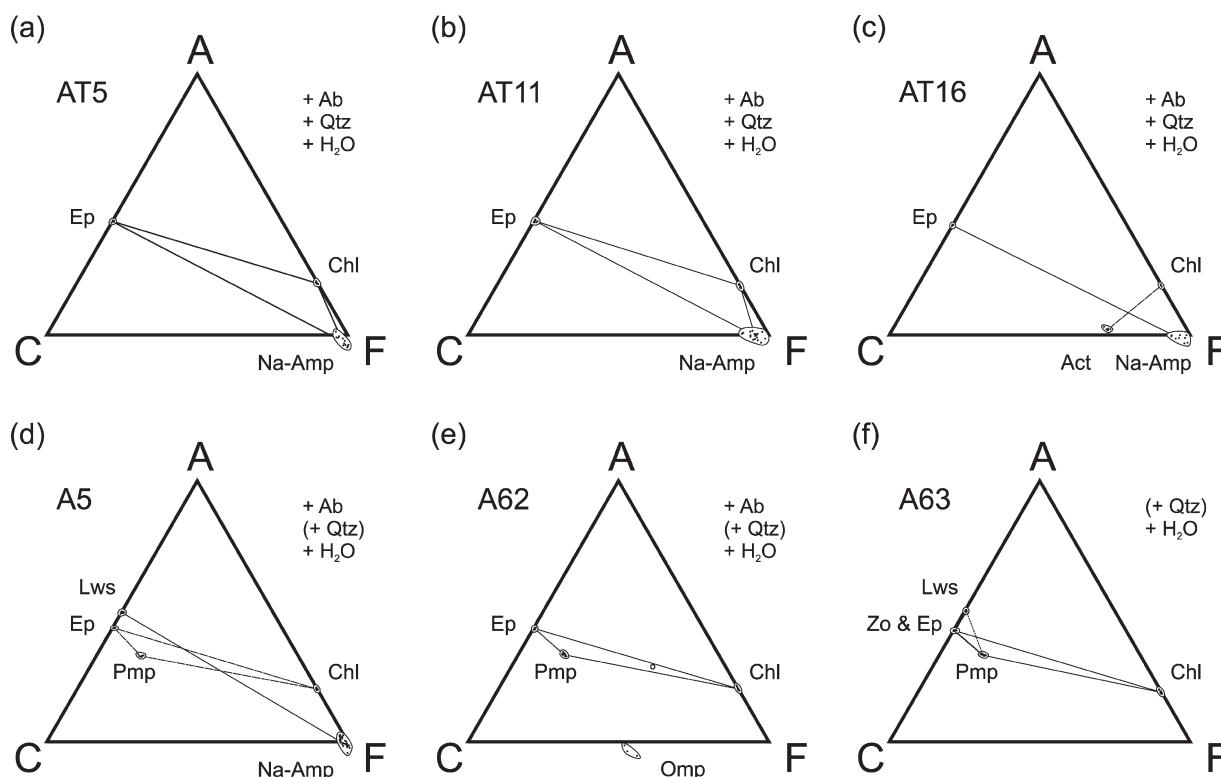
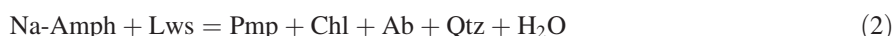


Figure 7. Phase relations of metabasites shown on Al_2O_3 – CaO – $(\text{Fe,Mg})\text{O}$ (ACF) projections from albite, quartz and water. Tie-lines relating to dominant assemblage, solid; tie-lines relating to relict assemblage in A62, dotted; tie-lines relating to retrograde assemblage in A63, dashed. (Qtz): no quartz actually present in rock. Small circle (A62): estimated bulk composition based on modal mineralogy.

The lawsonite-blueschist A5 contains the peak-metamorphic assemblage sodic amphibole + lawsonite + albite, and the amphibole is partially replaced by retrograde chlorite, pumpellyite and epidote. On the ACF diagram (Figure 7(d)), the sodic amphibole–lawsonite tie-line is crossed by both pumpellyite–chlorite and epidote–chlorite tie-lines, suggesting sequential operation of the following Fe,Mg-continuous reactions:



and



Quartz was not found in this rock, but the small volumes and grain sizes of reaction products involved may have prevented its detection. The reaction



may also have occurred locally in the rock, but not in isolation, as it fails to account for growth of retrograde pumpellyite.

The pumpellyite + epidote + chlorite assemblage, which is retrograde in the lawsonite-blueschist, forms the dominant, late mineral assemblage in the pumpellyite greenstones (A62 and A63), suggesting that the latter are strongly retrograded high-P rocks. Textural evidence for this is provided by the numerous tiny lawsonite inclusions within pumpellyite crystals in A63 and by the inclusions of relict omphacite within epidote in A62. The earliest mineral assemblage in A63 for which there is evidence is lawsonite + zoisite, with which pumpellyite probably

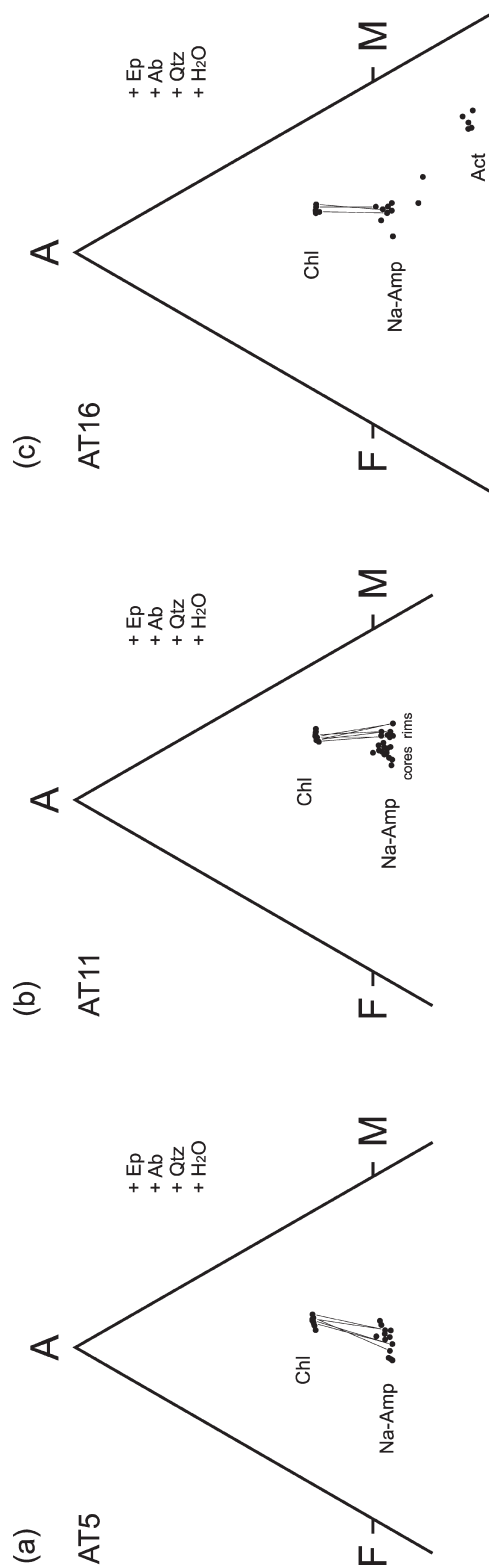
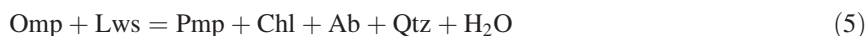


Figure 8. Phase relations of epidote-blueschists shown on $\text{Al}_2\text{O}_3\text{--FeO--MgO}$ (AFM) projections from epidote, albite, quartz and water.

also coexisted; growth of texturally late epidote and chlorite can be explained by Reaction (3) (Figure 7(f)). In sample A62, no direct evidence exists of the minerals that coexisted with the early omphacite, but the fact that the estimated bulk composition of this rock plots just below the epidote–chlorite join, closer to chlorite than epidote (Figure 7(e)), suggests that the coexisting minerals were chlorite and a low-Fe,Mg aluminous phase such as lawsonite. Early lawsonite + sodic pyroxene + chlorite assemblages, predating growth of sodic amphibole, have been described by Okay (1980b) and Okay and Kelley (1994) from the Tavşanlı Zone further north. Transformation of such an assemblage to the observed pumpellyite + epidote + albite + chlorite assemblage could be explained by the operation of the reaction



followed by Reaction (3). Alternative reaction sequences involving sodic amphibole-bearing intermediates are also possible. In either case, quartz would be expected as a reaction product; its absence from the rock may be due to removal of silica dissolved in percolating fluid.

6. METAMORPHIC CONDITIONS AND P–T PATHS

In this section, mineral equilibria are used to estimate the peak and retrograde metamorphic conditions experienced by the U. Permian–Cretaceous metasediments and the Koçyaka Ophiolitic Mélange, and to constrain the retrograde *P–T* paths followed. Because many of the mineral assemblages have high variance, it is not generally possible to constrain *P–T* conditions using the intersection of isopleths of equilibria involving end-member components that are all present in the assemblages concerned; consequently, we have had to rely heavily on the use of equilibria as stability limits, with the activities of absent components set at unity or at partition-corrected values (Sutthirath *et al.* 2001). The *P–T* coordinates of equilibria were calculated using version 3.1 (2001) of the program THERMOCALC (Powell and Holland 1988; Holland and Powell 1998). The activity–composition relations used for solid-solution minerals are listed in Table 5. The abbreviations of Holland and Powell (1998) are used for the names of mineral end-member components in Table 5 and the reactions listed below.

6.1. Blueschists

The absence of jadeitic pyroxene and presence of albite as an apparently stable peak-metamorphic phase in both epidote and lawsonite blueschists means that the end-member equilibrium curve for the reaction



serves as an upper pressure limit for both types of blueschist.

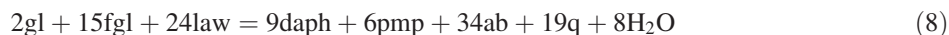
Table 5. Activity–composition relations in solid-solution minerals used for thermobarometry. End-member component abbreviations as in Holland and Powell (1998). ss.: solid solution. a_i : activity of end-member component i . X_i^j : mole fraction of element i on site j . $\square$: vacancy

<i>Sodic amphibole:</i>	Ideal mixing on sites, but with half the S_{mix} contribution from tetrahedral sites (Holland and Powell, 1998)	$a_{\text{gl}} = X_{\square}^{\text{A}} (X_{\text{Na}}^{\text{M4}})^2 (X_{\text{Mg}}^{\text{M1,3}})^3 (X_{\text{Al}}^{\text{M2}})^2 (X_{\text{Si}}^{\text{T2}})^2$
<i>Calcic amphibole:</i>	Regular solution model of Dale <i>et al.</i> (2000)	$a_{\text{fgl}} = X_{\square}^{\text{A}} (X_{\text{Na}}^{\text{M4}})^2 (X_{\text{Fe}^{2+}}^{\text{M1,3}})^3 (X_{\text{Al}}^{\text{M2}})^2 (X_{\text{Si}}^{\text{T2}})^2$
<i>Omphacite:</i>	Ideal coupled mixing	$a_{\text{jd}} = X_{\text{Al}}^{\text{M1}} \quad a_{\text{di}} = X_{\text{Mg}}^{\text{M1}}$
<i>Epidote-group:</i>	Ordering model of Holland and Powell (1998)	
<i>Chlorite:</i>	Ordering model of Holland <i>et al.</i> (1998)	
<i>Pumpellyite:</i>	Ideal mixing on sites, assuming Mg, Fe & Al on the X sites fully disordered (Frey <i>et al.</i> 1991)	$a_{\text{pmp}} = 4(X_{\text{Ca}}^{\text{W}})^4 (X_{\text{Mg}}^{\text{X}}) (X_{\text{Al}}^{\text{X}})$
<i>Garnet (py-alm ss.):</i>	Ideal mixing on sites	$a_{\text{alm}} = (X_{\text{Fe}}^{\text{VIII}})^3$

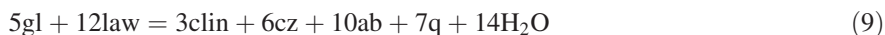
Lawsonite blueschist. For this rock (A5), an upper temperature limit is provided by the coexistence of lawsonite and albite, instead of zoisite, paragonite and quartz:



the end-member curve of which intersects that of reaction 6 at *c.* 430°C and 12.7 kbar (Figure 9(a)). Pressure minima for the peak assemblage are provided by activity-corrected component equilibria allied to Reactions (2) and (4), the most influential of which are



and



In the absence of sodic pyroxene, quartz and chlorite from the peak assemblage of A5, the partition-corrected equilibrium



with $a_{\text{jd}} = 0.5$ and $a_{\text{di}} = 0.17$ for the absent pyroxene, provides a poorly constrained lower-temperature limit. Together, these limits place the most likely peak-metamorphic *P–T* conditions of A5 at 375 $^{+60}_{-150}$ °C and 11 $^{+2}_{-4}$ kbar (Figure 9(a)).

Sample A5 contains abundant calcite, yet the deduced *P–T* conditions lie within the stability field of aragonite. Although there is no textural evidence for the former existence of aragonite, the latter was presumably present at the metamorphic peak and transformed entirely to calcite during decompression. The kinetics of the aragonite to calcite transformation are known to be rapid (Carlson and Rosenfeld 1981).

In calculating the equilibrium positions of dehydration reactions (e.g. (7) and (8)), the presence of a pure H₂O fluid phase was assumed. Any dilution of H₂O in the metamorphic fluid would shift these dehydration curves to lower temperatures and higher pressures, thus reducing the *P–T* field within which A5 could have crystallized. If the presence of a fluid during metamorphism is assumed, supporting evidence for an H₂O-rich composition is provided by the coexistence of lawsonite, CaCO₃ and titanite, the stability field of which is bounded, on an isobaric *T–X*_{CO₂} diagram, by the pure-phases equilibria



and



which, at 11 kbar and 375°C, constrain *X*_{CO₂} to values between 0.001 and 0.002.

The retrograde *P–T* path of the lawsonite blueschist evidently crossed to the low-pressure side of equilibrium curve (9), implying that decompression, rather than cooling, was the dominant effect. However, the absence of retrograde paragonite suggests that Reaction (7) was not crossed, implying that decompression was accompanied by some cooling. A retrograde path with a steep *dP/dT* (*c.* 6 kbar/100°C) is indicated.

Epidote blueschists. Most of the epidote blueschists and calc-schists contain the peak-metamorphic sub-assemblage sodic amphibole + chlorite + albite, to which the following pressure-sensitive end-member equilibrium may be applied:



In the absence of paragonite, the pressure-sensitive NFASH univariant reaction



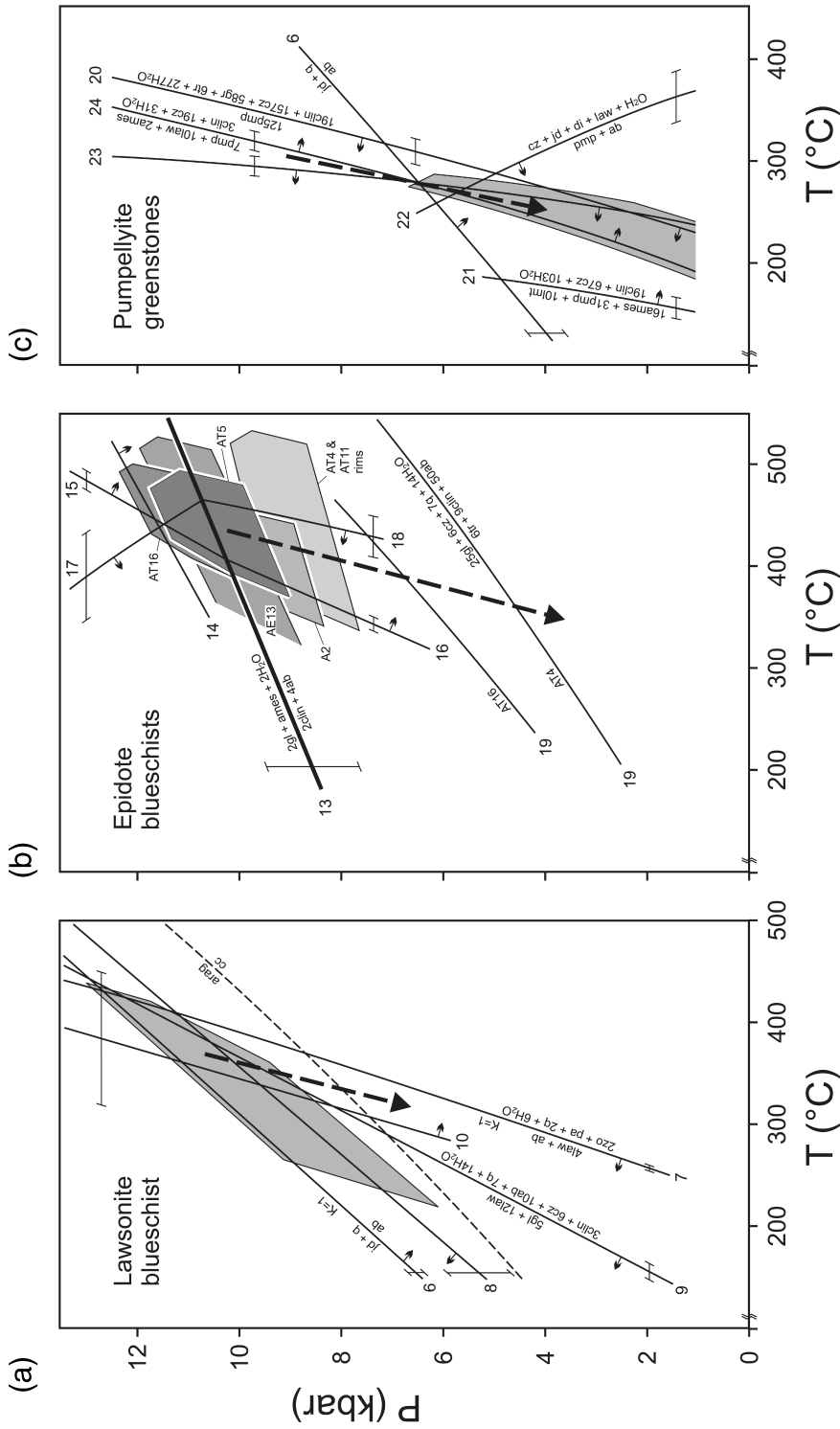
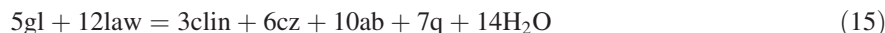


Figure 9. Metamorphic P - T conditions (shaded areas) deduced for dominant mineral assemblages in selected rocks from the Altnekin Complex: (a) lawsonite blueschist A5 (southernmost part of the southern outcrop of Ophiolitic Mélange), peak of metamorphism; (b) epidote blueschists (northern outcrop of Ophiolitic Mélange, plus northern part of southern outcrop of Ophiolitic Mélange and associated flysch), peak of metamorphism; (c) pumpellyite greenstones (southernmost part of the southern outcrop of Ophiolitic Mélange), retrograde metamorphism. Thick solid line, equilibrium curve used as a geobarometer; thin solid lines, equilibrium curves used as stability limits (arrow on side of stable assemblage), thin dashed line, equilibrium curve shown for information only. 1σ error bars shown where appropriate. ' $K = 1$ ': pure-phases curve. Numbers refer to equilibria mentioned in text. See the text for the identities of the unlabelled curves. Large arrows: deduced P - T paths. See the text for an explanation.

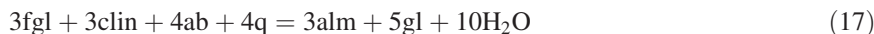
in the petrogenetic grid of Guiraud *et al.* (1990) provides an upper pressure limit. Lower temperature limits are provided by the coexistence of sodic amphibole, chlorite, epidote, albite and quartz and absence of lawsonite, via the equilibria



and



The absence of garnet in the epidote-blueschists means that the following equilibria provide upper temperature limits:



and



When calculated for the activities of components in their respective solid solutions (in the case of almandine in garnet, corrected for partitioning of Fe and Mg with chlorite), equilibria (13) and (15)–(18) yield the following P – T conditions of equilibration (Figure 9(b)): AT5, $440 \pm 50^\circ\text{C}$, 10.5 ± 1.7 kbar; AT16, $450 \pm 50^\circ\text{C}$, 11.3 ± 1 kbar; AT4, $440 \pm 100^\circ\text{C}$, 9 ± 1.2 kbar; AT11, $440 \pm 100^\circ\text{C}$, 9 ± 1.3 kbar; A2, $400 \pm 50^\circ\text{C}$, 9.4 ± 1 kbar. For AT11, the pale rims of the zoned amphiboles were used; the more riebeckitic cores yielded a pressure *c.* 0.6 kbar lower. The results listed above all fall within the low-pressure part of the ‘typical’ epidote blueschist field calculated by Evans (1990).

As with the lawsonite blueschist, the ubiquity of titanite as the stable Ti-rich phase in the epidote blueschists constrains, *via* Equilibrium (11), the maximum X_{CO_2} of any peak-metamorphic fluid phase in these rocks to *c.* 0.01, the remainder presumably being mostly H_2O .

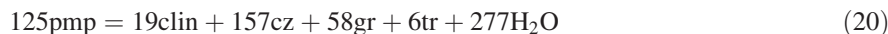
An upper pressure limit on the retrograde P – T conditions of the epidote blueschists is given by the equilibrium



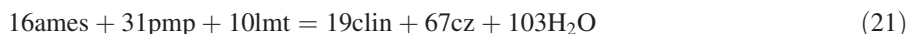
which is the end-member version of Reaction (1). Taking the glaucophane activity from sodic amphibole and the tremolite activity from actinolite, maximum pressures for the greenschist-facies retrogression in AT4 and AT16 are each *c.* 4 kbar lower than their respective peak-metamorphic pressures (Figure 9(b)). These findings imply that the retrograde P – T paths of the epidote-blueschists were strongly decompressive, but the P – T vectors are poorly constrained.

6.2. Pumpellyite greenstones

The dominant, late mineral assemblage in the pumpellyite greenstones (pumpellyite + epidote + chlorite + albite) has a stability field that is restricted to temperatures below 300°C and pressures below 7 kbar. The upper temperature limit is provided by the equilibrium defining the breakdown of Mg-pumpellyite to clinochlore, clinozoisite, grossular, tremolite and water:

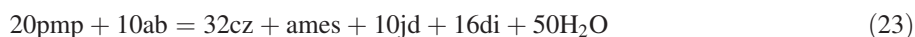


and, in the absence of laumontite, the lower temperature limit by the equilibrium

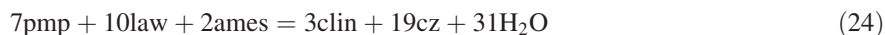


Further constraints are provided by the instability of omphacite (relicts of which are preserved inside epidote in sample A62—Figure 3(c) with respect to albite; when corrected for the estimated activity of jadeite in the omphacite (0.3), the curve for Equilibrium (6) is depressed by 2.5–3 kbar from its pure-phases position (Figure 9). The

instability of omphacite and/or lawsonite in the presence of pumpellyite, epidote, chlorite and albite also means that the following equilibria can be used as stability limits:



and



When corrected for component activities in solid-solution minerals, Equilibria (22) and (23) (applied to sample A62) and (24) (sample A63) define a wedge-shaped field with a very narrow temperature range for the late assemblage (250–280°C at 5 kbar, 200–250°C at 2 kbar—Figure 9(c)).

Equilibria (23) and (24) also place tight constraints on the slope of the P – T path for the evolution from omphacite- and lawsonite-bearing early assemblages to the dominant late one. The curve for (23) was evidently crossed in such a way as to stabilize its low-temperature assemblage (pmp + ab), whilst that of (24) was crossed in such a way as to stabilize its high-temperature assemblage (clin + cz + H₂O); because these curves intersect at a low angle, the retrograde P – T path is precisely located (Figure 9(c)), and a dP/dT slope of *c.* 10 kbar/100°C is indicated.

7. DISCUSSION

7.1. Metamorphic conditions in the SE Tavşanlı subduction zone

The peak-metamorphic P – T conditions recorded by a given rock map onto a single point on the peak-metamorphic geotherm. In the absence of an array of isochronous P – T points from a coherent metamorphic unit, no further information on the shape of that geotherm can be gained. Nevertheless, the peak-metamorphic P – T point provides an indication of the overall slope of the geotherm. In the discussion below, peak-metamorphic P – T points are converted to peak-metamorphic T/d values (where T = peak-metamorphic temperature, and d = depth of burial implied by pressure at peak of metamorphism) by assuming that d (in km) = $3.5 \times P$ (in kbar). The peak T/d value is commonly referred to as the ‘average geothermal gradient’.

The peak-metamorphic P – T conditions deduced in this paper confirm that the rocks of the northern Altinekin Complex underwent metamorphism in a régime of low peak T/d of *c.* 10–14°C/km, consistent with a subduction-zone setting, thus supporting the view that the complex once formed part of the Neo-Tethyan subduction-zone along the İzmir–Ankara Suture. Pressure estimates indicate that rocks of both the Cretaceous flysch unit and the U. Cretaceous Koçyaka Ophiolitic Mélange were tectonically buried to depths in excess of 30 km.

Intriguing differences in peak-metamorphic P – T conditions exist both between and within these units, which may be significant. Epidote blueschists of the flysch unit (A2, AT4 and AT11) yield pressures that are somewhat lower than those of the Koçyaka Ophiolitic Mélange (AE13, AT5, AT16) (Figure 9), indicating that the former were either subjected to higher T/d values (*c.* 12–14°C/km) than the latter (*c.* 11–12°C/km) during subduction, or underwent limited re-equilibration during post-peak decompression. The absence of a correlation between the amount of greenschist-facies alteration or late foliation development and calculated peak-metamorphic pressures supports the former explanation. Differences in peak T/d values are also recorded by rocks within the southern outcrop of the Ophiolitic Mélange (Figure 2). The lawsonite blueschist (A5), from near Koçyaka in the southernmost part of that outcrop, appears to have experienced a peak temperature (*c.* 375°C) and T/d value (*c.* 10°C/km) significantly lower than epidote blueschists from further north in the same unit. The slope of the retrograde P – T path of the pumpellyite greenstones (A62 and A63), also from the southernmost part of the outcrop, suggests that the peak temperatures experienced by these rocks were also relatively low. Together, these

findings suggest that the southernmost part of the Ophiolitic Mélange had a different metamorphic history from the rest of the unit. One possible explanation for this is that it belongs to a separate thrust slice.

An alternative explanation is given by Özgül and Göncüoğlu (1999), who maintain that the blocks within the Ophiolitic Mélange underwent HP–LT metamorphism in the subduction zone prior to being exhumed and incorporated in the sedimentary matrix, and that the whole mélange was then subducted again, resulting in a second episode of HP–LT metamorphism. If true, this model would help to account for metamorphic differences between blocks. However, we find no evidence to support a two-stage subduction model; our data indicate that the meta-sediments (including the matrix of the mélange) experienced the same number of HP–LT metamorphic cycles as the metabasic blocks (i.e. one) and that the overprinting relationships reflect evolving P – T conditions during a single cycle of subduction and exhumation.

The peak-metamorphic pressures recorded by rocks of the Altınekin Complex (9–11 kbar) correspond to burial depths of 31–38 km, equivalent to the base of a continental crust of normal thickness. These pressures are modest in comparison with the values of *c.* 19–22 kbar obtained by Okay and Kelley (1994) and Okay (2002) for the NW Tavşanlı Zone. Thus, rocks in the SE Tavşanlı Zone appear to have been subducted to only about half the depths of the rocks in the NW. Further studies may show the existence of higher-pressure rocks in the SE, but large differences in peak-metamorphic T/d value between the two areas point to a fundamental difference in style of subduction. The T/d value of *c.* 5°C/km deduced by Okay and Kelley (1994) and Okay (2002) for the NW Tavşanlı Zone is one of the lowest values recorded by metamorphic rocks anywhere on Earth, and is much lower than our values from the SE. The modelling studies of Peacock (1992) suggest that, if subducted rocks support moderate shear stresses (>0.2 kbar), as is likely, the geothermal gradient in a steady-state subduction zone will be minimized by minimizing convergence rate and maximizing subduction angle, other factors (basal heat flux, thermal conductivity etc.) being equal. This may indicate that the SE segment of the Tavşanlı subduction zone involved either a higher convergence rate or a shallower subduction angle than the NW segment.

7.2. Exhumation of blueschists in the SE Tavşanlı Zone

The mechanisms by which high-pressure metamorphic rocks at convergent plate margins are returned to the Earth's surface are hotly debated. Several different mechanisms have been proposed (see reviews by Platt 1993; Hynes 2002; Kurz and Froitzheim 2002), including (a) underplating and erosion (Rubie 1984), (b) corner flow, driven by subduction drag, in the hanging-wall wedge (Cowan and Silling 1978), (c) underplating and gravitational spreading of the hanging-wall wedge (Platt 1986), (d) detachment and buoyant uprise of fault-bounded slices (Wijbrans *et al.* 1993), potentially containing smaller masses of entrained dense rock (England and Holland 1979), (e) tectonically forced extrusion of fault-bounded masses up the subduction channel (Michard *et al.* 1993; Roselle and Engi 2002), (f) subduction zone shallowing due to buoyancy of subducted continental rocks beneath a mantle wedge (Krueger and Jones 1989), and (g) subduction zone shift (Hsu 1991). These mechanisms are not generally mutually exclusive, and may occur concurrently and/or sequentially; all may have some part to play in any given orogenic wedge, but their relative importance depends on local circumstances, particularly depth. Mechanisms (a), (b) and (c) are likely to be most influential for rocks at crustal depths, and (d) and (e) for rocks at mantle depths. In principle, some help in distinguishing between these exhumation mechanisms in any given HP–LT metamorphic complex can be provided by the P – T paths and deformation histories of the rocks.

The prograde P – T paths of rocks of the Altınekin Complex are largely unconstrained, but the amphibole zonation in sample AT11 points to prograde compression, implying that the rocks experienced $P_{\max}$ and $T_{\max}$ simultaneously. However, this conclusion is not secure, as it is based on the assumption that the chemical composition of the chlorite remained constant while that of the amphibole changed. The retrograde P – T paths of the lawsonite blueschist, epidote blueschists and pumpellyite greenstones are similar to one another, being strongly decompressive with subordinate cooling (Figure 9). Such paths, indicative of thermal relaxation towards higher geothermal gradient with time, are not compatible with exhumation mechanisms that involve forced extrusion or buoyant rise of rocks back up the subduction channel ((d) and (e), above), as the latter are likely to result in 'hairpin'-like P – T paths in which cooling and heating occur in the same thermal régime (e.g. Matsumoto *et al.* 2003). On the contrary,

the P – T paths deduced for the Altinekin Complex are more consistent with a retrograde evolution within the hanging-wall wedge, with exhumation driven by either corner flow, erosion, gravitational spreading or subduction zone shallowing. Further steps towards distinguishing between these possibilities may be made with reference to the deformation history of the rocks.

Metasediments of the Altinekin Complex preserve textural evidence for two phases of plastic deformation, one before and one after the thermal peak of blueschist-facies metamorphism. The first deformation presumably reflects the action of simple shear stresses as the rocks were being subducted. The second deformation resulted in intracrystalline deformation of calcite (Figure 3(f)), indicating that it occurred while the rocks were within the stability field of calcite, rather than aragonite. In deformed metabasites, such as AT16, the only tectonic fabric we see postdates the blueschist-facies metamorphism and is defined by alignment of the overprinting greenschist-facies minerals. Within the Altinekin Complex, therefore, the second of the two recognized deformations occurred while the rocks were undergoing decompression. This temporal association of deformation and decompression may be fortuitous; on the other hand, it may signify a causal connection between these two processes, in which case a purely erosive mechanism for the exhumation would be ruled out. Instead, the findings would be compatible with exhumation by a process that involved deformation of the hanging-wall wedge, such as corner flow, gravitational spreading or subduction zone shallowing. Okay *et al.* (1998) argue that the high- P lawsonite blueschists of the NW Tavşanlı Zone were exhumed by a two-stage process: (i) detachment and buoyant ascent up the subduction channel, and (ii) buoyancy-driven subduction zone shallowing. Our findings are consistent with the operation of the latter process in the SE Tavşanlı Zone also. Further analysis of this problem awaits detailed structural and geochronological studies.

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