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Exhumation of blueschists along a Tethyan suture in northwest Turkey

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Abstract

A blueschist belt formed during the mid-Cretaceous (~88 Ma) high-pressure metamorphism of a passive continental margin sequence of shale, siltstone, quartzite and limestone occurs immediately south of the Tethyan suture in northwest Turkey. A transect across the Tethyan suture was mapped to constrain the exhumation of blueschist-facies assemblages. The presence of jadeite, glaucophane and lawsonite in the blueschists indicates *P–T* conditions of 20 kbar and 430°C. The blueschists are tectonically overlain by a Cretaceous oceanic accretionary complex showing an incipient blueschist metamorphism. A peridotite slab with subvertical layering lies along a low-angle fault contact over the blueschists and the accretionary complex. Isolated slices of Early Cretaceous (101 ± 4 Ma) garnet-amphibolites occur at the base of the peridotite. The garnet-amphibolites, which show an incipient blueschist-facies overprint, have probably formed during the oceanic subduction in the foot-wall of the subduction zone and later underplated the hanging-wall. Blueschists and peridotites are abruptly truncated in the north by the steeply dipping suture fault, that juxtaposes unmetamorphosed and slightly deformed Triassic to Jurassic sediments of the northern continent against the peridotite and the accretionary complex. Tectonic juxtaposition of the lower-pressure accretionary complex and peridotites over the higher-pressure blueschists implies the removal of a 45-km-thick oceanic mantle sequence above the blueschists. Regional geological and geometric constraints suggest that the exhumation of the blueschists was achieved by a combination of two simultaneously acting mechanisms. The first was the detachment of the upper crustal blueschist sequence from its basement in the subduction zone and its incorporation to the hanging-wall followed by buoyant ascent. The second mechanism is the progressive shallowing of the subducted continental lithosphere resulting in the thinning of the mantle wedge above the subducted slab. The blueschists were partly exposed prior to the Paleocene continent–continent collision. Post-collisional Eocene (52–48 Ma) calc-alkaline plutons, that were emplaced in the blueschists and the overlying peridotite at ~10 km depth, are interpreted to have formed by crustal anatexis from advective heating by mantle melts. These have probably been generated during the upwelling of the asthenosphere under the shallowing continental lithosphere.

Keywords: blueschists; exhumation; Ar/Ar dating; suture; Turkey; metamorphic soles; ophiolite

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1. Introduction

In Phanerozoic orogenic belts, large regions of blueschist and eclogite facies rocks are present that are characterised by pressures of metamorphism corresponding to depths exceeding the normal thickness of the continental crust. The exhumation of such regional high-pressure/low-temperature (HP/LT) metamorphic belts has been the subject of several models including thrusting and concomitant erosion in an oceanic subduction zone (Hsu, 1991), underplating and extension (Platt, 1986), delamination of the upper crust followed by buoyancy-driven exhumation (e.g., England and Holland, 1979; Wijbrans et al., 1993), subduction shallowing (Krueger and Jones, 1989) and extrusion between two colliding plates (e.g., Chopin et al., 1991). These models predict different structural settings and different tectonic evolutionary paths for the HP/LT metamorphic rocks. For example, in the thrusting–erosion model the highest-pressure metamorphic rocks are expected to occur in the top of the thrust stack. In the underplating–extension model the structural sequence should resemble that of an accretionary complex, extension and thrusting are expected to occur contemporaneously in the different parts of the accretionary prism, and the exhumation of the HP/LT metamorphic rocks should be a steady-state process. On the other hand, the extrusion model predicts that the exhumation of the HP/LT metamorphic rocks should be a singular event rather than a steady-state process and should be contemporaneous with the continent–continent collision. We aim to test these models in a regional Tethyan blueschist belt by placing stratigraphic, structural, petrological, geochemical and geochronological constraints on the exhumation of the blueschists. The area chosen in northwest Turkey is suitable for such a study as a late Tertiary updoming and erosion have exposed several kilometres of structural thickness. Furthermore, structural relationships are generally unambiguous, and the granodiorites that intrude the blueschists provide robust P–T points and thermal data for the exhumation history. Moreover, detailed petrological, geochemical and geochronological data on the blueschists and related rocks have recently been published (Okay and Kelley, 1994; Harris et al., 1994).

A second aim of the paper is to describe an Alpide suture, along which a major part of the Tethyan ocean was consumed. Suture zones are important tectonic structures as they record the history of subduction, accretion and collision that leads to the formation of an orogen. The area studied includes a well-exposed section of the 1300-km-long Izmir–Ankara–Erzincan suture, that separates Laurasian units to the north from Gondwanian units to the south.

2. Tectonic setting

Western Turkey comprises several continental fragments with distinctive stratigraphic, structural and metamorphic features that were amalgamated by continent–continent collision during the Alpide orogeny in the Early Tertiary, following the obliteration of the intervening Tethyan oceanic lithospheres by northward-dipping subduction zones (Sengör and Yılmaz, 1981; Okay, 1989). One of these continental fragments is the Sakarya Zone (Fig. 1) characterised by a clastic and volcanic basement, that was deformed and metamorphosed in the latest Triassic and unconformably overlain by a Jurassic–Cretaceous sedimentary sequence. It is bounded to the south by the Izmir–Ankara suture, generally regarded as the main Neo-Tethyan suture in Turkey. South of the suture, the Anatolide–Tauride platform is typically characterised by the general absence of the Late Triassic deformation and metamorphism, and by well-developed, thick Triassic to Cretaceous platform carbonate sequences.

Within the Anatolide–Tauride platform of western Turkey major faults define the boundaries of several units (Fig. 1). Immediately south of the Izmir–Ankara suture between Bursa and Nallıhan, the Tavsanlı Zone forms a 50-km-wide coherent blueschist belt, representing the northward subducted passive continental margin of the Anatolide–Tauride platform. Its western margin is in contact with a latest Cretaceous–Paleocene flysch zone (the Bornova Flysch Zone) containing Mesozoic limestone blocks, several kilometres large, in a Maastrichtian–Paleocene greywacke–shale matrix (Okay and Siyako, 1993). Farther south is the Afyon Zone, a Devonian to Paleocene sedimentary sequence metamorphosed in greenschist facies. In the Afyon Zone, the Mesozoic carbonate sedimen-

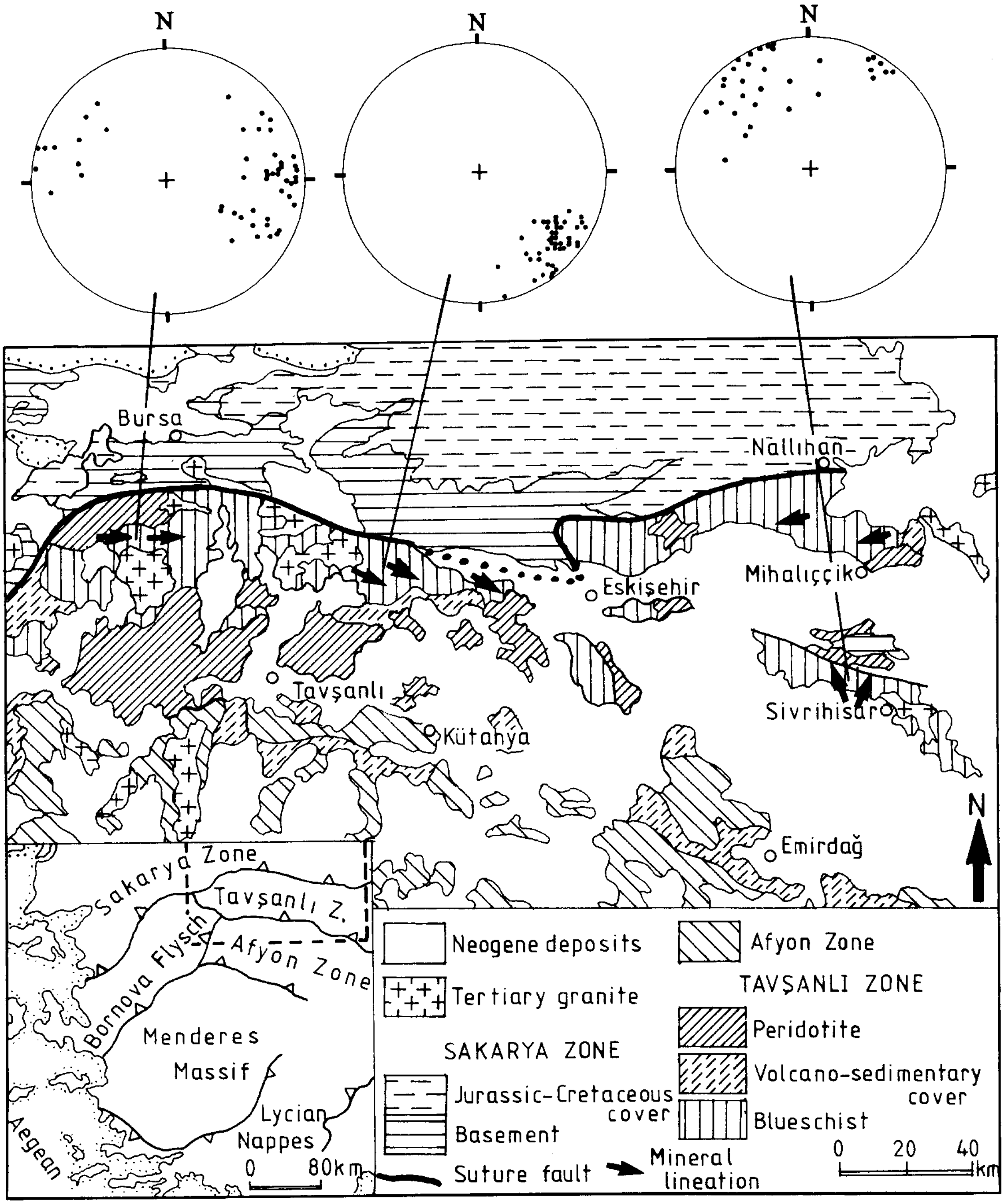


Fig. 1. Tectonic map of northwestern Turkey. Inset in lower left corner shows major tectonic zones. The mineral lineation data are from Lisenbee (1971) for south of Bursa, from Okay (1981) for northeast of Tavsanlı, and from Gautier (1984) for the Sivrihisar area.

tation continues up to the early Maastrichtian. It is succeeded by upper Maastrichtian–Lower Paleocene olistostromal wild-flysch with limestone, serpentinite, mafic volcanic rocks and blueschist blocks (Fig. 2; Özcan et al., 1988; Göncüoğlu et al., 1992). The Menderes Massif, to the south of the Afyon Zone, is a major Oligocene dome that exposes a Precambrian gneissic basement overlain by a Palaeozoic to Eocene sedimentary sequence metamorphosed in greenschist/amphibolite facies (Sengör et al., 1984). It is tectonically overlain to the southeast by the Lycian Nappes largely comprising Mesozoic continental slope and rise sequences and an overlying large peridotite slab. The Lycian nappes, including the peridotite, are generally believed to have been derived from the region of the Izmir–Ankara suture (e.g., Ricou et al., 1975).

The area studied is located south of Bursa and straddles the Izmir–Ankara suture between the Sakarya Zone and the blueschist belt (Fig. 1). It is one of the rare localities in western Turkey, where the suture is not covered by Neogene terrigenous deposits. The large-scale structure in the southern part of the area studied is characterised by an Oligocene dome centred around the circular Orhaneli granodiorite, which exposes regionally metamorphosed blueschists in a tectonic window under the peridotite (Fig. 1).

3. Southern continent

3.1. Blueschist sequence

The blueschists in the Orhaneli area make up a regionally metamorphosed clastic-limestone sequence overlain tectonically by an accretionary complex and a peridotite slab. The base of the blueschists is not exposed. The blueschists are intruded by Eocene granodiorite plutons and are unconformably overlain by Miocene continental deposits (Figs. 3 and 4).

The blueschist sequence comprises an over 1-km-thick basal metaclastic unit overlain by >300-m-thick massive white marbles. The metaclastics are medium-grained, strongly foliated micaschists representing metamorphosed siltstones and shales. The petrology of the micaschists was studied in detail by Okay and Kelley (1994) along the Kocasu valley in the southwestern part of the area investigated (Fig. 3). The critical mineral assemblage of the micaschists is quartz + phengite + glaucophane + jadeite ± chloritoid ± paragonite ± chlorite. The micaschists contain metaquartzites and jadeite-schist horizons, that are several metres thick. The jadeite-schists assemblage consists of sodic pyroxene with over 95% end-member jadeite, coexisting with quartz and phengite. Rare metabasic intercalations with sodic amphibole + lawsonite + chlorite + sodic pyroxene paragenesis occur in the transition zone with the overlying marbles. Similar mineral assemblages exist eastwards around Orhaneli (Lisenbee, 1972).

The metamorphic foliation in the Orhaneli region dips gently northwards as a result of the Oligocene updoming. The micaschists exhibit tight to isoclinal folds varying in amplitude from microscopic to over 100 m, resulting in partial transposition of the bedding. The axial planes dip gently to the north, and fold axes, that lie in the foliation plane, show a considerable scatter around a preferred orientation in a N75°E direction, which is subparallel to the variably but moderately plunging calcite elongation lineation in blueschist marble (Fig. 1; Lisenbee, 1972). The mineral elongation lineation, essentially defined by preferred orientations of calcite in marbles and of sodic amphibole in metacherts, is a ubiquitous feature of the blueschists in northwest Turkey (Çoğulu, 1967; Okay, 1981; Servais, 1982; Gautier, 1984; Monod et al., 1991). In all investigated regions in the Tavsanlı Zone, except the Sivrihisar region, the mineral lineation trends east or southeast, subparallel to the trace of the suture (Fig. 1). The widespread early isoclinal to tight folds in the Orhaneli area are locally refolded by east- and south-vergent asymmetric open folds (Lisenbee, 1972). A steeply dipping crenulation cleavage is well developed in the Kocasu region.

The assemblage jadeite + paragonite + quartz in the micaschists constrains the metamorphic pressure between 12 and 25 kbar. Pressures are further constrained by the equilibrium assemblage chloritoid, glaucophane, paragonite, chlorite and quartz to 20 ± 2 kbar at 430°C (Okay and Kelley, 1994). The absence of garnet in both the metabasites and micaschists indicates maximum temperatures of about 450°C, while the chloritoid–glaucophane equilibria indicate minimum temperatures of 420°C at 20 kbar

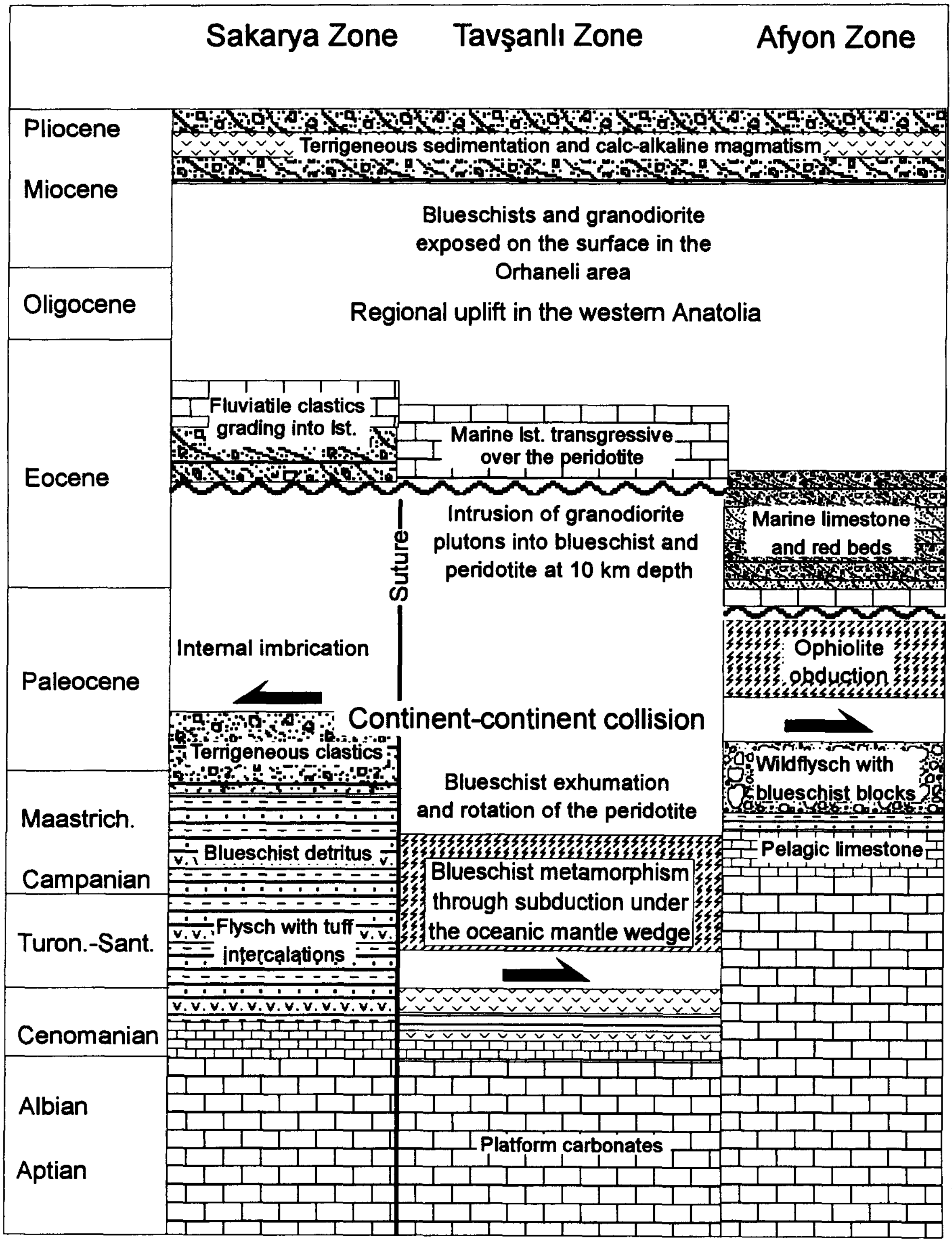


Fig. 2. Depositional and tectonic events in northwest Turkey.

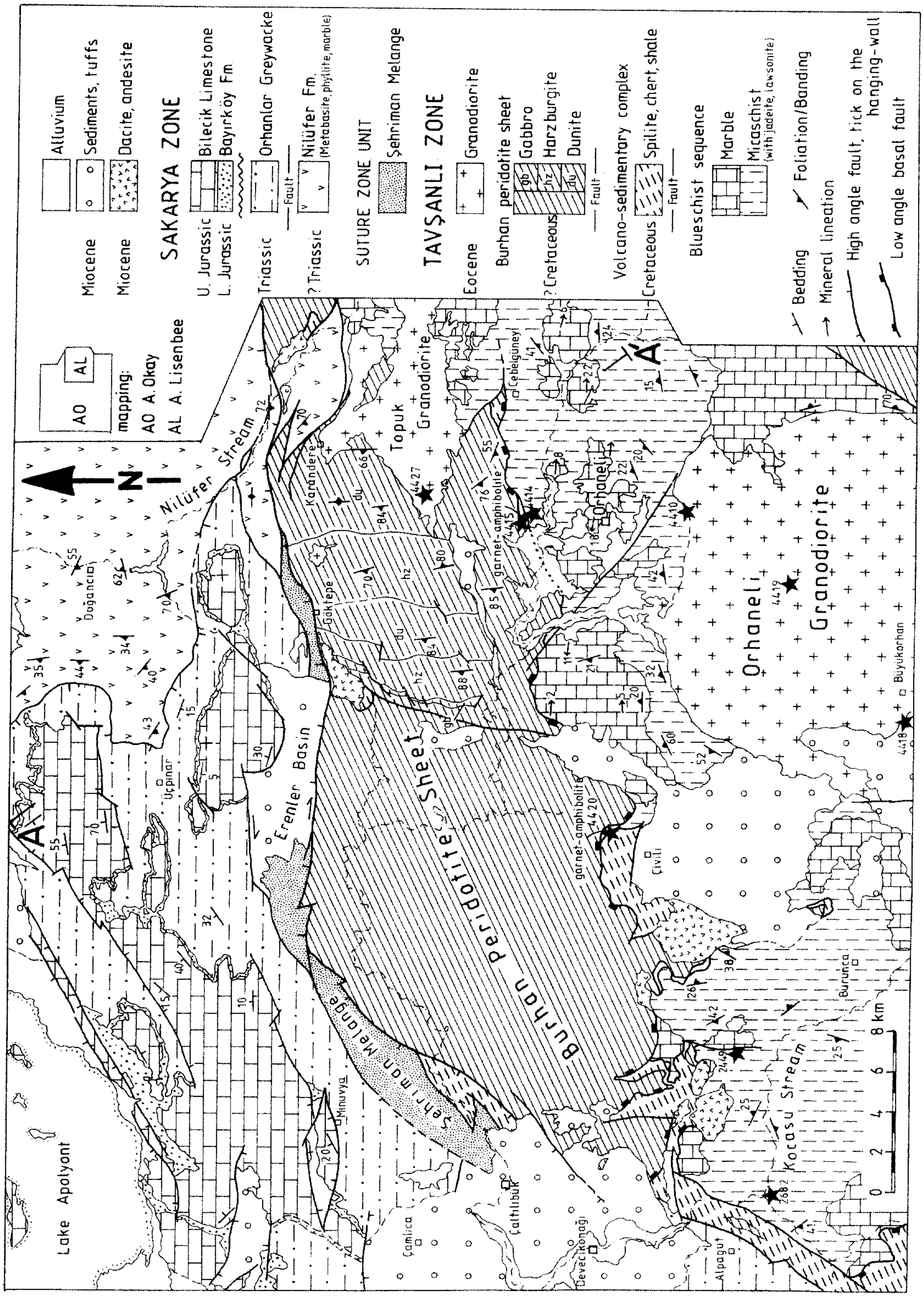


Fig. 3. Geological map of the Orhaneli area. Stars with adjoining numbers show the location of isotopically dated samples. See Fig. 1 for location.

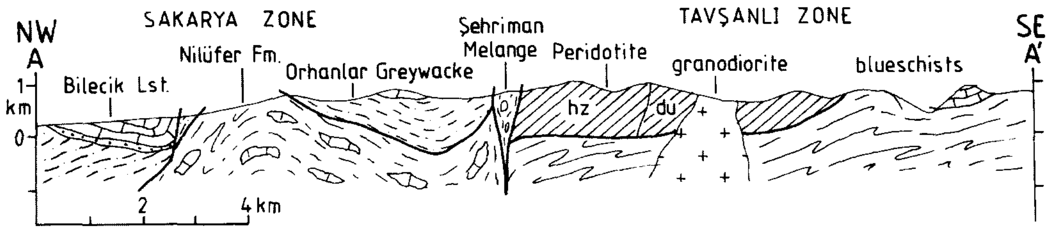


Fig. 4. Geological cross-section of the Orhaneli area. See Fig. 3 for location.

(Okay and Kelley, 1994). The estimate for the *P–T* conditions of the blueschist metamorphism in the area studied is therefore 430 ± 30°C and 20 ± 2 kbar (Fig. 5). The general absence of garnet in metabasic and metapelitic rocks, the lack of greenschist facies overprint textures, such as actinolite rims around sodic amphibole, or replacement of lawsonite by epidote, common in most other blueschist regions, impose tight constraints on the *P–T* paths followed by the blueschists and indicates that the *P–T* path has stayed largely within the blueschist facies field (Fig. 5).

Three blueschist samples from the Orhaneli area have been analysed using a high-resolution Ar/Ar laser probe technique (Okay and Kelley, 1994; Harris et al., 1994). Phengites from a jadeite-schist from the Kocasu valley (sample 2449, Fig. 3) yielded a weighted mean age of 88.5 ± 1 Ma (2 sigma errors), though another jadeite-schist from the same region (2682) yielded phengite ages ranging from 83.5 to 70 Ma (Okay and Kelley, 1994). Sodic amphibole from a metachert (4414a) north of Orhaneli 32 km farther east (Fig. 3) yielded an Ar/Ar isochron age of 108 ± 8 Ma (Harris et al., 1994). This sample contained a significant excess argon component and it is not clear how much significance can be attached to the age difference of 20 m.y. The absence of a greenschist overprint in the blueschists implies that HP/LT metamorphism persisted over a protracted period. However, a spread of Cretaceous K/Ar ages (120–65 Ma) was also reported from the blueschists in the eastern part of the Tavsanlı Zone (Çoğulu and Krummenacher, 1967; Kulaksız and Phillip, 1985). A similar age disparity is reported from Oman, where the isotopic ages of blueschist and eclogite assemblages range from 131 to 68 Ma (Lippard, 1983; Montigny et al., 1988; El-Shazly and Lanphere, 1992; Searle et al., 1994), many of which predate the age of the metamorphic soles (101 to 95 Ma, Montigny et al., 1988) and the Maastrichtian (75 Ma) ophiolite obduction. The interpretation of this spread in blueschist ages, whether apparent (Goffé et al., 1988; Searle et al., 1994) or real (Montigny et al., 1988; El-Shazly and Lanphere, 1992) is controversial. A spread of Ar/Ar ages for individual phengite grains (from 42 to 30 Ma) from the HP/LT metamorphic rocks on Sifnos in the Cyclades has been interpreted as being due to sequential uplift and cooling of the HP/LT metamorphic sequence with the older ages located near the top of the sequence (Wijbrans et al., 1990), which in the Orhaneli area is the basal fault plane above the blueschists (Fig. 3). However, in the present case there is only a maximum vertical distance of 500 m between assemblages from which the 101 Ma and 88 Ma ages have been determined; such a small vertical separation is unlikely to be responsible for the 20 m.y. age difference. On the other hand, it is becoming increasingly apparent that Ar/Ar mineral data from HP/LT metamorphic rocks often yield ages considerably older than the 'real' age of the metamorphism. Ar–Ar ages from the Dora Maira Massif in the Western Alps, for example, fall in the range of 90–105 Ma (Monié and Chopin, 1991), though the age of the UHP metamorphism is probably around 38 Ma (Tilton et al., 1991). Recent UV laser analysis indicated widespread excess argon, which is now thought to be related to fluid poor conditions during the HP/LT metamorphism (Kelley et al., 1994a; Arnaud and Kelley, 1995). To test for possible excess argon in the Turkish phengite, laser traverses along two phengites of 0.5–0.7 mm were made from the pristine jadeite-schist (sample 2449, Fig. 6). A similar study in the Western Alps revealed very large age variations within individual grains (Arnaud and Kelley, 1995). The ultra-violet laser technique used is detailed in Kelley et al. (1994b)

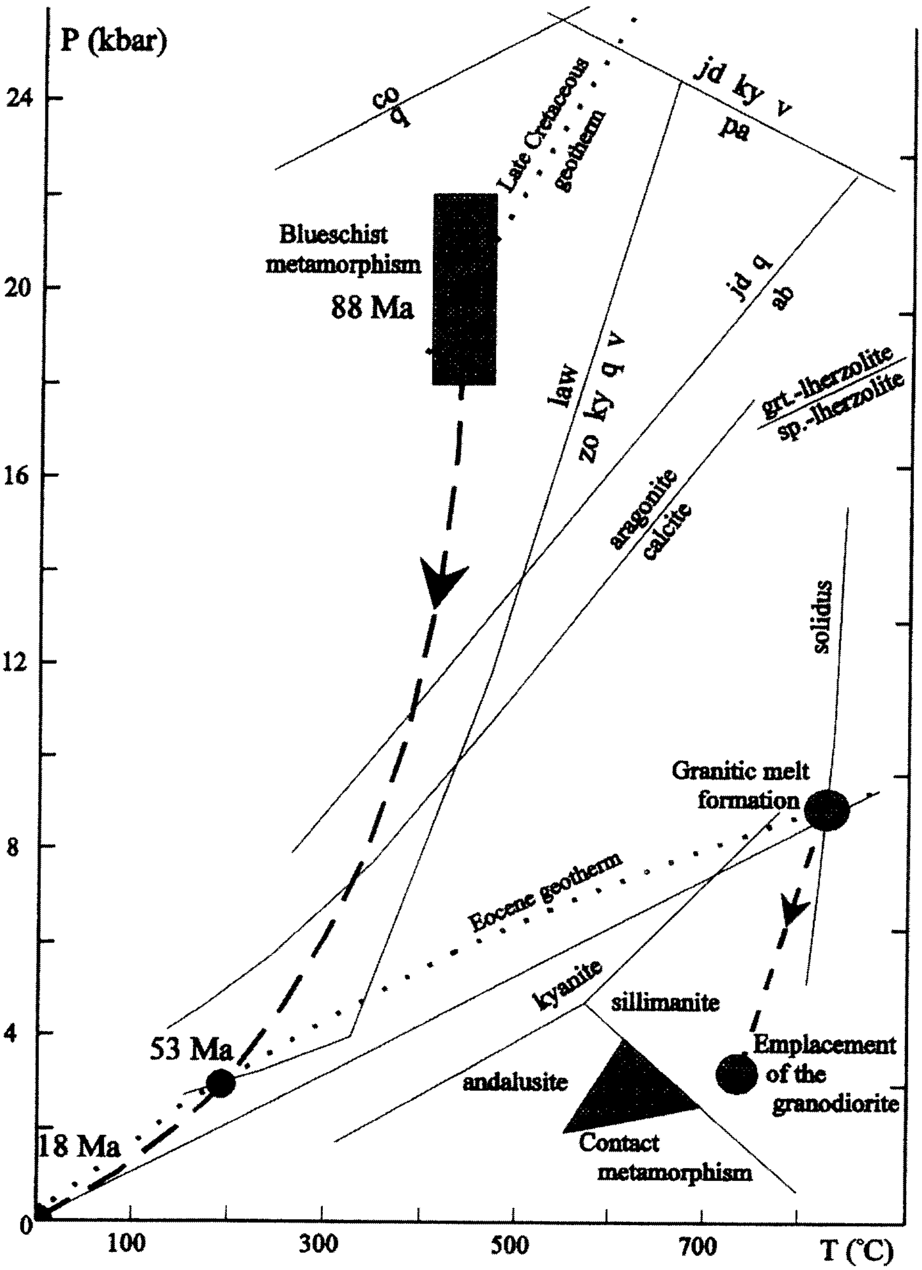


Fig. 5. Pressure–temperature diagram showing P–T–t path followed by the blueschists in the Orhaneli area (heavy dashed line), conditions of emplacement of the Orhaneli granodiorite and relevant equilibria. The two dotted lines show possible isotherms in the continental crust of the Orhaneli area for 88 Ma and 53 Ma. The 88 Ma geotherm is adapted from Peacock (1992) for steady-state P–T conditions in a subducted slab undergoing blueschist-facies metamorphism. The 53 Ma geotherm is constrained by the inferred Eocene partial melting at <10 kbar and by the absence of greenschist facies overprint in blueschists. Solidus for the dehydration melting of a biotite- and hornblende-bearing tonalite is from Huang and Wyllie (1986) and Rutter and Wyllie (1988). Depth scale on the right is for oceanic lithosphere. The lower-pressure stability of lawsonite is after Liou (1971); all other reactions are calculated using THERMOCALC of Holland and Powell (1990). Abbreviations: *ab* = albite; *ar* = aragonite; *co* = coesite; *jd* = jadeite; *ky* = kyanite; *law* = lawsonite; *pa* = paragonite; *q* = quartz; *v* = H_2O; *zo* = zoisite.

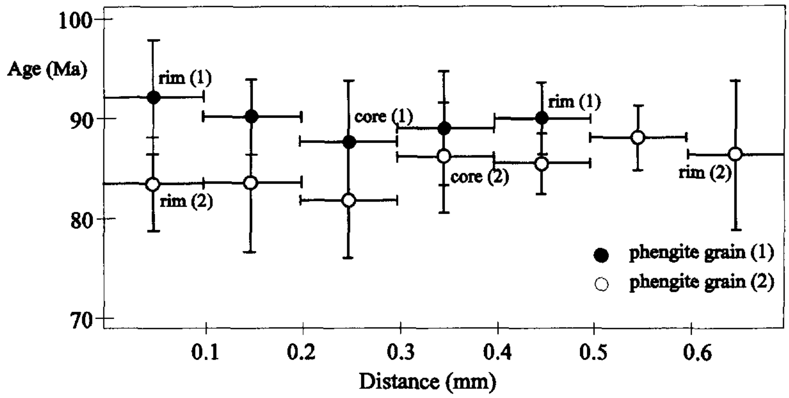


Fig. 6. Argon laser ages obtained from traverses along two phengite grains from jadeite-schist (sample 2449). Analytical data are given in Table 1.

and Arnaud and Kelley (1995). Although the age data are of lower precision due to the small volume of sample analysed at each stage, traverses show no age gradient within the grains (Fig. 6, Table 1) and thus no late stage argon loss or influx of excess argon. A weighted mean age of 87 ± 3 Ma confirms the earlier spot ages from this sample. Moreover, in-situ UV laser analyses of quartz in the same sample indicate very small amounts of argon with near atmospheric ratios (unlike the Alpine samples, cf. Arnaud and Kelley, 1995) further indicating no excess argon in the system (Table 1). Thus, we regard 88 Ma (Turonian/Coniacian) as a reasonable age for the blueschist metamorphism in the western part of the Tavsanlı Zone.

There is no direct evidence for the depositional

Table 1
Ar-laser isotopic data from two phengite traverses and quartz grains from jadeite-schist (sample 2449) from the Orhaneli region, northwest Turkey

Grain	$^{40}Ar/^{39}Ar$	$^{38}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$	^{39}Ar	$^{40}Ar^{*}/^{39}Ar$	Age (Ma)	Distance ±50 μm
Phengite							
11 (edge)	5.216	0.023	0.0022	10.03	4.55±0.26	83±5	50
12	5.064	0.017	0.0017	11.59	4.56±0.38	84±7	150
13	5.248	0.018	0.0027	10.49	4.46±0.32	82±6	250
14	5.443	0.019	0.0025	11.18	4.70±0.31	86±5	350
15	5.262	0.019	0.0020	11.31	4.67±0.16	86±3	450
16	5.148	0.017	0.0011	10.51	4.82±0.18	88±3	550
17 (edge)	5.148	0.017	0.0015	8.05	4.72±0.42	86±8	650
21 (edge)	5.314	0.017	0.0009	9.57	5.03±0.32	92±6	50
22	5.035	0.016	0.0004	10.07	4.93±0.21	90±4	150
23	4.891	0.017	0.0003	10.93	4.79±0.34	88±6	250
24	4.965	0.016	0.0003	10.88	4.87±0.32	89±6	350
25	4.923	0.016	0.0000	10.38	4.92±0.20	90±4	450
Undeformed quartz							
1	23.300	0.058	0.0674	0.68	3.4±4.8	60±100	
2	40.693	0.086	0.1194	0.52	5.4±4.9	100±90	
3	22.011	0.093	0.0903	0.48	4.7±5.3	−90±100	
4	21.328	0.065	0.0887	0.52	4.9±3.5	−90±70	
Deformed quartz							
1	21.421	0.105	0.0476	0.45	7.4±5.6	130±100	

^{39}Ar amounts in $\times 10^{-12}$ cc STP.

age of the pre-metamorphic blueschist sequence. However, the normal stratigraphic sequence of the Tauride–Anatolide platform comprises Palaeozoic clastics overlain by thick, massive Mesozoic platform carbonates. If analogous with the Tavsanlı Zone, the metaclastic part of the blueschist sequence may have a Palaeozoic and Lower Triassic, and the marble section a Mesozoic depositional age.

3.2. *Eocene plutons*

Two large granodiorite plutons intrude the blueschists and the overlying Burhan peridotite slab and are unconformably covered by the Miocene terrigenous deposits (Figs. 3 and 4). They are part of four plutons that intrude the blueschist sequence in the western part of the Tavsanlı Zone (Fig. 1). Both the Topuk pluton to the north and the Orhaneli pluton to the south are undeformed. These equigranular granodiorites are characterised by plagioclase, quartz, biotite, alkali feldspar and hornblende. The geochemistry of both granodiorites has been studied by Harris et al. (1994). Both plutons have a uniform meta-aluminous composition with silica contents of 63–69%. Selective LIL trace element enrichment in the granodiorites suggests a typical calc-alkaline affinity. The absence of Y and HREE depletion precludes garnet as a significant restite phase, thus indicating melt formation at pressures less than 10 kbar. Mantle-derived melts are required in the petrogenesis of these magmas either as parental magmas for the granodiorites to evolve by fractionation processes in shallow magma chambers or as a convective heat source for crustal anatexis.

Both plutons have contact aureoles marked by the destruction of the blueschist mineral assemblages and by the formation of andalusite and cordierite in the inner contact aureole. The contact aureole mineral assemblages, as well as the Al-content of the hornblende in the granodiorite, indicate that the Orhaneli granodiorite was emplaced at a pressure of 3 ± 1 kbar (Harris et al., 1994).

Ar/Ar laser spot analysis on biotite from the Orhaneli granodiorite (4419A) yielded a good isochron age of 52.6 ± 0.4 Ma, which is compatible with a biotite isochron of 52.4 ± 1.4 Ma from a hornfels (4410) 2 m from the granodiorite contact (Fig. 3). These ages suggest rapid cooling after intrusion at a high level in the crust in the Middle Eocene (Harris et al., 1994). A slightly younger isochron age of 47.8 ± 0.4 Ma is obtained from biotite and hornblende grains from the higher level Topuk granodiorite (4427A).

4. Oceanic assemblages

The blueschist sequence is tectonically overlain by units that were part of the Tethys ocean. From bottom to top these units comprise the volcano-sedimentary complex, the amphibolite metamorphic soles and the Burhan peridotite slab (Figs. 3 and 4).

4.1. *Volcano-sedimentary complex*

The volcano-sedimentary complex is made up of spilitised mafic volcanic, pyroclastic and tuffaceous rocks with minor radiolarian chert, pelagic shale, pelagic limestone and serpentinite. It occurs generally as imbricate tectonic slices along the base of the Burhan peridotite slab. The total structural thickness of the volcano-sedimentary complex may reach up to several hundred metres. The mafic volcanic and pyroclastic rocks, which constitute over 80% of the volcano-sedimentary complex, comprise blocks of radiolarite and limestone. The volcano-sedimentary complex is transected by a large number of shear zones with tens to hundreds of metres spacing. A lens of red, micritic pelagic limestone, a few metres thick, is intercalated with radiolarian cherts in the volcano-sedimentary complex east of Çaltılıbük. It contains the microfossil *Pithonella ovalis* indicative of a Cenomanian to Maastrichtian age (Okay and Kelley, 1994). Late Cretaceous–Paleocene pollens (Lisenbee, 1972) are also reported from the volcano-sedimentary complex north of the village of Cebelgüney (Fig. 3).

The rocks of the volcano-sedimentary complex are free of penetrative deformation. The common mineral assemblage in the mafic volcanic and pyroclastic rocks is Ti-augite, albite and chlorite, while pumpellyite and aragonite commonly occur in the amygdales. Lawsonite was found in one specimen replacing igneous plagioclase phenocrysts (Fig. 7a). Sodic pyroxene and lawsonite frequently occur in veins and amygdales in mafic volcanic rocks from a similar volcano-sedimentary complex northeast of

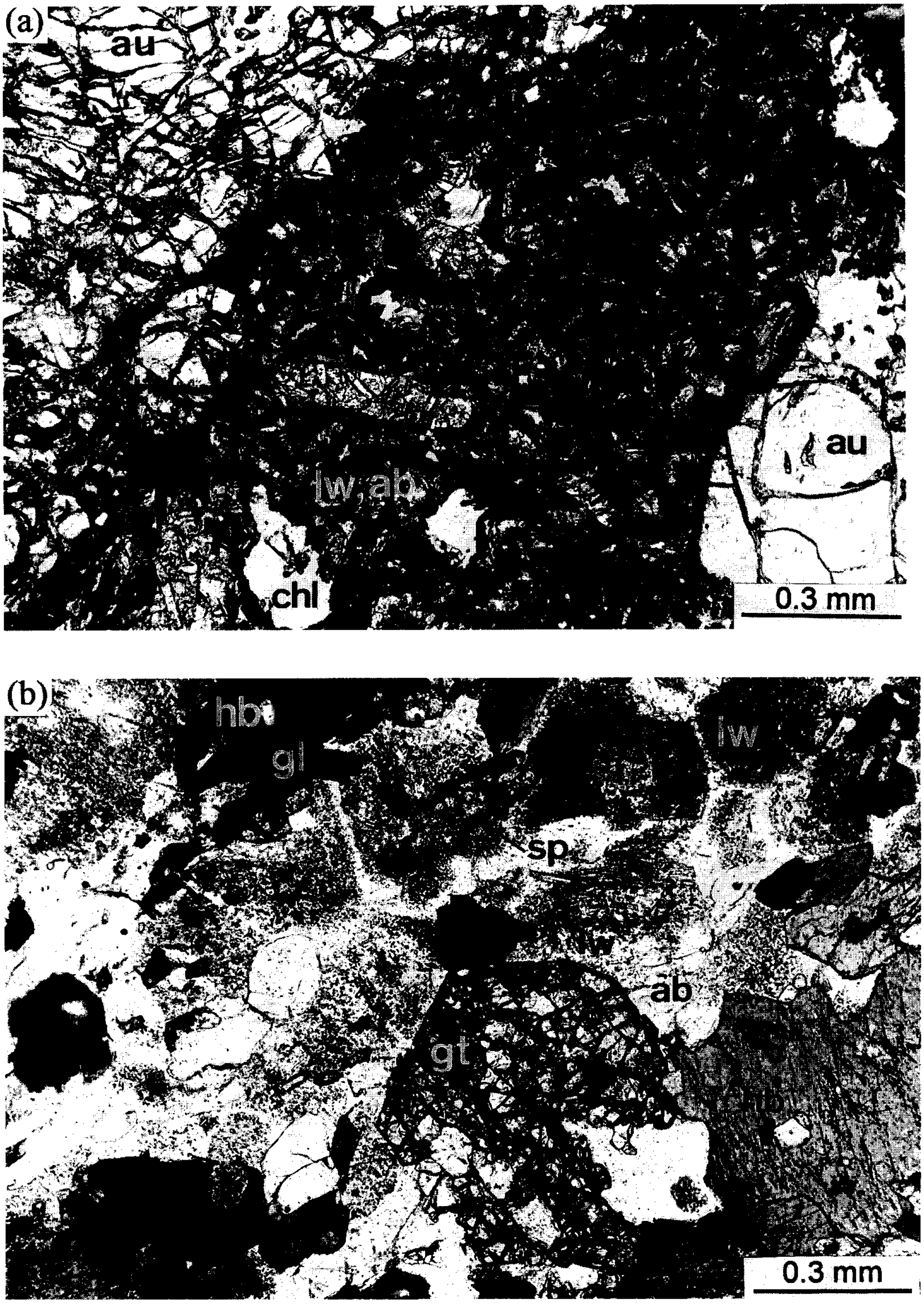


Fig. 7. Photomicrographs (plane polarised light). (a) Incipient blueschist-facies metamorphism in basalt (4424A) from the accretionary complex with lawsonite (*lw*) and albite (*ab*) pseudomorphs after plagioclase coexisting with chlorite (*chl*) and igneous augite (*au*). (b) Fine-grained lawsonite (*lw*) aggregates in albite (*ab*) and sodic amphibole rims (*gl*) around hornblende (*hbl*) illustrating incipient blueschist metamorphism of the high-grade garnet-amphibolite (4420); *gt* = garnet.

Tavsanlı (Fig. 1; Okay, 1984). The lack of recrystallisation, the occurrence of aragonite and lawsonite, and absence of jadeite in the mafic volcanic rocks indicate metamorphic conditions of 6 ± 2 kbar and $200 \pm 50°C$.

Similar volcano-sedimentary complexes are widespread in northwestern Turkey and are often referred to as ophiolitic melanges, although they lack a well-defined matrix. They represent sediment-starved Cretaceous accretionary complexes, made up of accreted upper levels of oceanic crust.

4.2. Amphibolite soles

Two amphibolite slices outcrop at the base of the Burhan peridotite slab in the Orhaneli area. They are similar to the metamorphic soles described from the base of other ophiolite complexes in northwest Turkey (Gautier, 1984; Önen and Hall, 1993), in Oman and in Newfoundland (e.g., Williams and Smyth, 1973; Spray et al., 1984). Such metamorphic soles are generally considered to form during the early stages of intra-oceanic thrusting by the combined effects of residual heat from the overlying hot ophiolite body and the frictional heat generated along the thrust plane. Both amphibolite slices, separated by 16 km, are about 150 m thick. They consist of medium-grained, strongly banded amphibolites that dip 45° north under the silicified base of the peridotite (Fig. 3) and show sharp boundaries with the underlying blueschist and volcano-sedimentary complex. Mesoscopic red garnets are restricted to the upper 50 m of the amphibolite sequences.

The amphibolite-facies mineral assemblage in both slices is hornblende + plagioclase + epidote + quartz + rutile ± garnet ± opaque. Garnet forms several millimetre large, subidoblastic poikilitic porphyroblasts associated with smaller dark green hornblende and plagioclase (Fig. 7b). Epidote is a common matrix phase in the slice north of Orhaneli, whereas it occurs mainly as inclusions in garnet in the slice north of Çivili. Titanite occurs as a late mineral forming wide rims around rutile.

Garnet-amphibolites from both slices show an incipient blueschist-facies overprint, with the development of very fine-grained (<0.03 mm) turbid aggregates of lawsonite in plagioclase (Fig. 7b). Lawsonite aggregates are better developed in the Çivili slice, which also shows thin dark-blue sodic amphibole rims around some of the green hornblende grains. A similar incipient blueschist overprint in garnet-amphibolites from the base of an ultramafic sheet is reported from north of Kütahya (Önen and Hall, 1993).

One representative garnet-amphibolite from each slice was analysed by electron microprobe. The analytical conditions for the electron microprobe analysis were the same as those described in Okay and Kelley (1994). Garnets from both samples are similar in composition and are almandine-grossular solid solutions with minor pyrope and spessartine contents (Table 2). They show typical growth zoning profiles with a decrease in spessartine and increase in almandine and pyrope towards the rim. The calcic amphiboles are pargasites and pargasitic hornblendes. Epidotes have $Fe^{3+}/(Fe^{3+} + Al)$ ratios of 0.29–0.35 (Table 2). Plagioclase is albite (an_{1-3}) in both samples, although originally it must have been more calcic in composition. Lawsonite is rich in ferric iron (Table 2), characteristic of low-temperature lawsonites.

The garnet–hornblende geothermometer of Graham and Powell (1984) suggests metamorphic temperatures of $700 \pm 90°C$ and $688 \pm 40°C$ for the samples 4415B and 4420, respectively (Fig. 8). The Ti content of amphiboles coexisting with rutile is known to increase with increasing temperature (e.g., Raase, 1974). The Ti contents (0.14–0.17 per formula unit) of the analysed hornblendes indicate temperatures of 650–750°C (Schumacher et al., 1990). In experimental studies, clinopyroxene first appears in metabasic rock compositions at 700–750°C depending on pressure and oxygen fugacity (Spear, 1981) providing an upper temperature limit for the garnet-amphibolites without pyroxene. A temperature of $700 \pm 50°C$ is regarded as a reasonable estimate for the analysed garnet-amphibolites.

The hornblende compositions plot in or near the high-pressure amphibole fields of Laird and Albee (1981). Pressures can be quantitatively estimated using equilibria between garnet, hornblende and quartz. The following reactions, calculated using THERMOCALC (Holland and Powell, 1990), have shallow slopes and are relevant for the pressure estimates

Table 2
Representative mineral compositions from the garnet-amphibolites at the base of the peridotite (4415B and 4420) and from the incipiently metamorphosed mafic pyroclastic rock (4424A)

	Garnet			Hornblende		Epidote	Lawsonite	
	4415B rim	4420 core	4420 rim	4415B	4420	4415B	4420	4424A
SiO_2	37.99	37.30	37.53	42.28	40.39	37.74	37.55	37.92
TiO_2	0.09	0.24	0.12	1.02	1.20	0.12	0.06	0.19
Al_2O_3	20.72	20.55	20.29	13.86	13.74	25.07	28.07	29.75
Cr_2O_3	0.00	0.00	0.00	0.03	0.05	0.04	0.00	0.02
FeO	26.78	29.71	28.56	18.00	21.67	10.24	5.35	2.98
MgO	3.04	0.75	2.18	8.39	6.35	0.00	0.00	0.06
MnO	1.34	4.45	1.85	0.06	0.12	0.08	0.06	0.02
CaO	10.44	8.47	10.10	9.13	8.94	22.86	16.94	17.30
Na_2O	0.00	0.00	0.00	3.58	3.40	0.00	0.00	0.05
K_2O	0.02	0.03	0.05	1.00	1.72	0.09	0.02	0.00
	100.42	101.50	100.68	97.35	97.58	96.24	88.05	88.29
Structural formula based on:								
	12 ox	12 ox	12 ox	23 ox	23 ox	8 cat	8 ox	8 ox
Si	2.998	2.982	2.990	6.316	6.162	3.000	2.036	2.023
Al^4	0.002	0.018	0.010	1.684	1.838			
Al^6	1.926	1.918	1.894	0.757	0.633	2.349	1.795	1.871
Ti	0.006	0.014	0.008	0.115	0.140	0.006	0.003	0.008
Cr	0.000	0.000	0.008	0.004	0.008	0.002	0.000	0.000
Fe^{3+}	0.068	0.068	0.090	0.477	0.603	0.680	0.243	0.133
Fe^{2+}	1.700	1.918	1.812	1.772	2.162			
Mg	0.358	0.088	0.260	1.868	1.443	0.000	0.000	0.005
Mn	0.090	0.302	0.124	0.007	0.015	0.006	0.003	0.001
Ca	0.884	0.726	0.862	1.478	1.484	1.947	0.984	0.989
Na	0.000	0.000	0.000	1.048	1.020	0.000	0.000	0.005
K	0.002	0.004	0.006	0.193	0.341	0.010	0.001	0.000
	8.034	8.038	8.064	15.719	15.849	8.000	5.065	5.035
alm	56.1	63.2	59.1					
pyr	11.8	2.9	8.5					
spess	3.0	10.0	4.1					
gr, and	29.1	23.9	28.3					

alm, almandine; pyr, pyrope; spess, spessartine; gr, grossular; and, andradite. Ferric iron in hornblende is calculated on the assumption of 23 oxygens and $Si + Ti + Al + Cr + Fe^{3+} + Fe^{2+} + Mg + Mn = 13.00$.

(Fig. 8):

$$\text{pargasite} + \text{hornblende} + \text{ferrohornblende} = \text{grossular} + \text{almandine} + \text{edenite} + \text{quartz} + H_2O$$

$$\text{pyrope} + \text{pargasite} + \text{ferrohornblende} = \text{grossular} + \text{almandine} + \text{edenite} + \text{quartz} + H_2O$$

The two equilibria indicate metamorphic pressures of 9.3 ± 3.7 kbar and 7.8 ± 3.5 kbar for a given temperature of 700°C for the samples from the Orhaneli and Çivili slices, respectively. The coexistence of garnet, epidote, quartz and rutile indicates minimum pressures of about 6 kbar by the amphibole-free equilibrium (Fig. 8):

$$\text{grossular} + \text{rutile} + \text{quartz} + H_2O = \text{titanite} + \text{clinozoisite}$$

The metamorphic conditions of the incipient blueschist metamorphism in the garnet-amphibolite, estimated from the stability of lawsonite + albite and absence of jadeite, are 6 ± 2 kbar and 200 ± 100°C

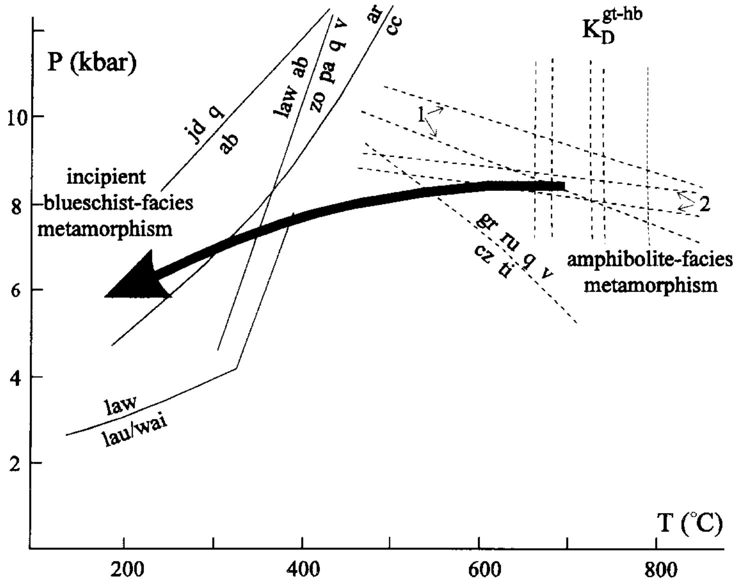


Fig. 8. Pressure–temperature–time path of garnet-amphibolites from the base of the peridotite. P–T conditions of the amphibolite facies metamorphism at 8.5 ± 3 kbar and 700°C are constrained by the garnet–hornblende geothermometer of Graham and Powell (1984) (K^{gt-hb}, vertical dashed lines) and by reactions (1) pargasite + hornblende + ferrohornblende = edenite + grossular + almandine + quartz + H_2O, and (2) pargasite + ferrohornblende + pyrope = edenite + grossular + almandine + quartz + H_2O for minerals in the two analysed samples. Conditions of incipient blueschist-facies metamorphism at 6 ± 2 kbar and 200 ± 100°C are given by the lawsonite + albite stability. Other reactions are calculated using THERMOCALC of Holland and Powell (1990). *ab* = albite; *ar* = aragonite; *cc* = calcite, *cz* = clinozoisite; *gr* = grossular; *jd* = jadeite; *lau* = laumantite; *law* = lawsonite; *pa* = paragonite; *sph* = titanite; *q* = quartz; *ru* = rutile; *v* = H_2O; *wai* = wairakite; *zo* = zoisite.

and indicate a virtually isobaric cooling path (Fig. 8). Such garnet-amphibolite slices are genetically different from those described from the base of ophiolites overlying unmetamorphosed continental margin sediments, which are exhumed soon after their generation and thus do not show a blueschist facies overprint.

Metamorphic pressures in the blueschists are about 14 kbar higher than those in the directly overlying volcano-sedimentary complex and garnet-amphibolites, suggesting the removal of about 45-km-thick strata between these units.

Nine laser-spot analyses of hornblende from the garnet-amphibolite from the Çivili slice (4415B) yielded an isochron age of 101.1 ± 3.8 Ma indicating an Albian cooling age for the metamorphic soles (Harris et al., 1994). This age is similar to that obtained from the metamorphic soles of the Lycian ophiolites (104 ± 4 Ma, Thuizat et al., 1981), which, however, do not show blueschist metamorphism and have been emplaced onto unmetamorphosed continental margin sediments.

4.3. Burhan peridotite slab

The Burhan peridotite slab is a monotonous peridotite body covering an area of about 500 km^2 (Fig. 1). As the upper parts of the ophiolite sequences are not present in the Tavsanlı Zone, the ages of the Burhan peridotite slab and other peridotite bodies are not directly known. However, geochronological studies in the metamorphic soles of several ophiolite bodies, e.g. in the Semail ophiolite in Oman (Lanphere, 1981), in the Greek and Yugoslavian ophiolites (Spray et al., 1984), in the Tauric ophiolites in southern Turkey (B. Hacker, pers. commun., 1994) have shown that the age of the metamorphic sole is virtually indistinguishable from that of the overlying ophiolite, which by analogy suggests an Albian age for the Burhan ophiolite.

The Burhan peridotite slab consists predominantly of harzburgite and dunite (>90%) with minor gabbro, pyroxenite and chromite. Its northern contact with the Sakarya Zone is defined by the suture zone while in the south it overlies the volcano-sedimentary complex or the blueschists along a shallowly dipping sharp basal fault (Figs. 3 and 4). This contact and the Burhan peridotite slab are cut by the Eocene Topuk granodiorite. Towards the southwest, east of the Devecikonağı, the peridotite is truncated by a subhorizontal basal fault, that is transected by the steeply dipping suture fault (cf. Fig. 3). An extrapolated cross-section of the Burhan peridotite slab suggests a maximum thickness of about 1.5 km. This rough estimate is consistent with the detailed gravity profiles, which show a maximum gravity value within the peridotite body, that is only 32 mGal greater than is found at the southern contact, giving a maximum thickness of 1 km assuming a density contrast of 0.7 between the peridotite and the micaschists (Lisenbee, 1972).

The internal structure of the Burhan peridotite slab is well known in its eastern part (Lisenbee, 1971, 1972; Tankut, 1980; Kaya et al., 1989). It consists of several kilometres thick, north-trending, subvertical harzburgite and dunite, and thinner gabbro and pyroxenite layers (Figs. 3 and 4). The contacts between these units are gradational over distances as great as 50 m. The peridotites show a distinct tectonic foliation, parallel to the compositional layering, and defined by the preferred orientation of olivine grains and by elongate enstatite crystals (Lisenbee, 1972). The peridotite youngs to the west and is bounded by a poorly defined fault zone, marked by serpentinite lenses, small Neogene basins and Miocene dacite intrusions (Fig. 3). The total thickness of the Burhan peridotite slab in this area, as measured perpendicular to layering, is over 13 km, in stark contrast to the vertical thickness less than 2 km. The western part of the peridotite slab is poorly known. It is mainly made up of harzburgite that also shows the steep north-trending tectonite fabric.

The Burhan peridotite slab, like the other peridotite bodies in the Tavsanlı Zone, is cut by east-trending, subvertical microgabbro dykes showing chilled margins. The dykes are generally a few metres thick and vary in abundance from 1 dyke over several hundred metres to 10 dykes over 30 m. They do not extend down to the volcano-sedimentary complex and are cut by the basal fault. The chilled margins of the dykes indicate that they were injected into an already cold oceanic lithosphere. Similar dykes of low-K tholeiitic composition are present in the peridotites from the Lycian Nappes (Whitechurch et al., 1984) and central Taurides (Lytwyn and Casey, 1995). The mineral assemblage in the dykes is augite, partly replaced by igneous hornblende and saussuritised plagioclase. This observation is critical since it indicates that the dykes, and by inference the Burhan peridotite slab, have not undergone the regional blueschist metamorphism that is observed in the immediately underlying marbles and micaschists. The common mineral assemblage in the Burhan peridotite, olivine + orthopyroxene + clinopyroxene + Cr-spinel, is not stable above 14 kbar at 1000°C (Fig. 6; Perkins et al., 1981), when garnet starts to form, again indicating a pressure discontinuity between the peridotite and the underlying blueschists.

4.4. Basal fault system of the Burhan peridotite slab

The Burhan peridotite rests along a shallowly to medium-dipping, sharp basal fault over the blueschist sequence. Slivers of the volcano-sedimentary complex are intercalated along this contact in the west. Eastwards they pinch out so that the peridotite is in direct contact with the blueschists (Figs. 3 and 4).

North of the Kocasu valley the volcano-sedimentary complex overlies the blueschists along a subhorizontal contact; the contact is sharp with no shearing in either the underlying blueschists or in the volcano-sedimentary complex. The contact between the volcano-sedimentary complex and the overlying peridotites is similarly sharp and subhorizontal, with flat-lying small klippen of silicified or serpentinised peridotite preserved over the volcano-sedimentary complex. Farther east the basal contact of the peridotite can be observed north of Orhaneli (Fig. 3). Silicified massive serpentinite sharply overlies the blueschists along this eastward trending 6.5 km long contact. In this area the contact dips at 45–50° north (Lisenbee, 1972).

The juxtaposition of the lower-pressure volcano-sedimentary complex and peridotite to the blueschists metamorphosed at pressures corresponding to a depth of about 60 km suggests omission of strata along the basal fault. A similar relationship has been described from California, where extensional faulting is believed to be responsible for the observed discontinuity in metamorphic pressures between the blueschists and overlying ultramafic rocks (Jayko et al., 1987).

5. Suture zone

In the Orhaneli area, the suture zone appears as a narrow, well-defined discontinuity ranging from a single fault plane to a zone of tectonic melange up to a maximum of 3.5 km in width (Figs. 1 and 4). This Sehriman melange consists of recrystallised limestone, greywacke, mafic volcanic rocks and radiolarite blocks (ranging in scale from centimetre to kilometre) embedded in a highly sheared siltstone–shale matrix. The melange also includes slices of Orhanlar Greywacke, that are several hundred metres thick, and coherent flysch-type sandstone–shale sequences of unknown origin. Özkoçak (1969) reports poorly preserved *Globotruncana* sp. tests in the fine clastics, suggesting a post-Late Cretaceous age for the formation of the melange. The Sehriman melange thus includes lithologies from the Sakarya Zone and the volcano-sedimentary complex, and its formation is probably related to late fault activity along the suture.

The suture fault system can be followed for 60 km along strike in the area studied (Fig. 3). Individual faults dip steeply (70–80°) towards the north or northwest. They are upper-crustal brittle faults and the main fault branch cuts the Miocene conglomerates of the Erenler basin northwest of Göktepe. The Erenler basin is a small Miocene pull-apart basin, bounded to the north and to the south by two fault strands of the suture fault. It is filled by poorly sorted, terrigenous, monomict conglomerates, over 300 m thick, with microgabbro and peridotite pebbles (Fig. 3). This suggests that the suture zone acted as a sinistral strike-slip fault during and after the Miocene. Several small Miocene dacite and andesite plugs and sills intrude the suture zone. The zone was probably also active during the Eocene, as the Topuk granodiorite intrudes very close and parallel to an off-shoot of the main suture fault (Fig. 3). The suture that separates blueschists metamorphosed at 60 km depth during the Late Cretaceous from unmetamorphosed Jurassic shelf limestones may be analogous with the Insubric Line in the Alps (e.g., Schmid et al., 1989).

6. Northern continent (Sakarya Zone)

In the area studied, the Sakarya Zone consists of a deformed and partly metamorphosed Triassic basement overlain unconformably by a Jurassic sedimentary cover.

6.1. The basement of the Sakarya Zone

The basement of the Sakarya Zone consists of a metabasite–marble–phyllite unit (Nilüfer Formation) tectonically overlain by a monotonous greywacke series (Orhanlar Greywacke) (Figs. 3 and 4). These units form part of the Karakaya Complex, an assemblage of Permo–Triassic units deposited on the active margin of the Permo–Triassic Tethys (Tekeli, 1981; Okay et al., 1996). The Nilüfer Formation is characterised by metabasites (fine- to medium-grained mafic tuffs and minor pillow lavas) with intercalated marbles and phyllites. It shows a strong deformation, with the development of cleavage in phyllites and fine-grained metatuffs, and extensive boudinage of marbles and pyroclastic flows. The unit has undergone a low-grade, high-pressure greenschist facies metamorphism as indicated by the mineral assemblage in the metatuffs: actinolite + albite + epidote + chlorite + titanite. Sodic amphibole occurs in some laminated siliceous rocks and as incipient grains in the massive pyroclastic flows. In the area studied, there are no data on the age of deposition and metamorphism of the Nilüfer Formation, whereas farther west, northeast of Bergama, Middle Triassic conodonts have been described from the upper parts of this Formation (Kaya and Mostler, 1992).

The Orhanlar Greywacke consists of a monotonous, chaotically deformed greywacke sequence, over 1000 m thick, with minor shale, siltstone, conglomerate and thinly bedded black chert. It comprises rare limestone olistoliths, 0.3–2 m in size, with a Carboniferous microfauna of *Bradyina* sp. and *Archaediscus* sp. In the region south of Çaltılıbük (Fig. 3), the limestone blocks are more frequent, reach up to 20 m in size and are characterised by an Upper Permian fauna of *Pseudofuselina* sp., *Eopolydiexodina* sp., *Afghanella* sp., *Sumatrina* sp., *Tetrataxis* sp. , *Langella perforata langei*, *Eotuberitina reitlingerae*. The presence of exotic Upper

Permian blocks and the unconformably overlying Liassic Bayırköy Formation indicate that the Orhanlar Greywacke is of Triassic age.

The Orhanlar Greywacke overlies the Nilüfer Formation along a steeply dipping folded fault contact marked by a shear zone, several metres thick (Figs. 3 and 4). As the Nilüfer Formation is known to be stratigraphically overlain by clastic sequences similar to the Orhanlar Greywacke in the Bergama region, the movement along the fault separating the two units in the Orhaneli area is probably Cretaceous or younger in age. This fault strikes subparallel to the suture zone and probably joins it in the Nilüfer valley (Figs. 3 and 4), a feature which also suggests an Alpide age for this fault. Stratigraphic relationships across the fault (younger rocks are emplaced over the older rocks) suggest an extensional fault.

6.2. The Jurassic cover of the Sakarya Zone

The Orhanlar Greywacke is unconformably overlain by undeformed, 50–200-m-thick Jurassic clastics (Bayırköy Formation) corresponding to widespread fluviatile to shallow water molassic deposits formed at the end of the Triassic Karakaya orogeny. The base of the Bayırköy Formation is marked by a conglomerate or pebbly sandstone, which passes up into thickly bedded sandstones, pebbly sandstones, siltstones and shales. Molluscs in the siltstones (*Pleuromya alduini* and *P.* aff. *tellini*) indicate a Jurassic age (Özkoçak, 1969; Kaya et al., 1989). From regions farther west and east, the age of the Bayırköy Formation is well established as Sinemurian to Pliensbachian (Altıner et al., 1991). The Bayırköy Formation is overlain by neritic Jurassic limestones (Bilecik Limestone), that are over 800 m thick. Their age ranges from Callovian to Tithonian in this region (Altıner et al., 1991).

The absence of Cretaceous to Early Tertiary rocks south of Bursa (Fig. 1) is probably due to Late Tertiary erosion, as such sequences are present both west and east (Saner, 1980; Altıner et al., 1991; Okay et al., 1996). In these regions, the Bilecik Limestone ranges up to the Valanginian and is unconformably overlain by pelagic argillaceous limestones of mid-Cretaceous age (Vezirhan Formation), and by more than 1000 m of Upper Cretaceous tuffaceous flysch with serpentinite, blueschist and Bilecik limestone olistoliths (Fig. 2). This flysch sequence shows a regressive development and passes upwards to fluviatile sediments of Paleocene age.

7. Post-tectonic deposits

Paleocene to Oligocene deposits are absent both in the Orhaneli area and more generally in the Tavsanlı Zone. Only a small outlier of Middle Eocene neritic limestone is exposed north of the town of Tavsanlı. It unconformably overlies the peridotites (Bas, 1986), indicating that the Tavsanlı Zone was below sea level around 50–42 Ma ago, when the Orhaneli and Topuk granodiorites were intruding the blueschist sequence at a depth of $\sim$10 km. The Oligocene was marked by a major uplift of most of western Turkey (Fig. 2). During the Miocene significant calc-alkaline magmatism occurred contemporaneously with continental sedimentation within small fault-controlled basins throughout northwestern Turkey. Several of these basins exist in the Orhaneli area (Fig. 3), filled by conglomerates, sandstones, marls, lacustrine limestone, dacitic and andesitic tuffs and containing large borate and lignite deposits. Several dacite and andesite plugs and sills occur within these basins (Fig. 3). We have dated individual biotite grains from a dacite tuff west of Büyükorhan (sample 4418, Fig. 3) using the Ar/Ar laser probe technique. The analytical procedures for isotopic dating are given in Okay and Kelley (1994). The dacite tuff lies directly over the Orhaneli granodiorite, and thus provides an upper limit for the age of exhumation of the pluton and adjoining blueschists. The biotites from the dacite gave a mean age of 17.6 ± 0.2 Ma, indicating that both the blueschist and the granodiorite were exposed by the Early Miocene.

8. Tectonic evolution

8.1. Pre-subduction stage (Liassic to Albian)

A feature of the study area is the lack of continental slope and rise sequences on both sides of the suture. Northwards of the suture the Bilecik Limestone approaches within 1.5 km of it (Figs. 3 and 4) with no sign of a change to a more pelagic slope facies. Jurassic–Cretaceous slope sequences are, however, known from farther east in the Sakarya Zone and indicate the establishment of a south-facing, passive continental margin during Early to Middle Jurassic times (Görür et al., 1983). Such sequences in the Bursa region were probably destroyed by thrusting and erosion, or alternatively displaced by strike-slip faulting.

South of the suture, the protoliths of the massive blueschist marbles are Mesozoic shelf-type limestones. Presumably, the corresponding continental slope and rise sequences have been either subducted or thrust southward prior to the blueschist metamorphism. By comparison with the Hawasina Nappes in Oman (e.g., Bechennec et al., 1988; Michard et al., 1994), the Lycian Nappes south of the Menderes Massif (Fig. 1) may represent the missing continental slope and rise deposits of the Anatolide–Tauride platform (Fig. 9).

8.2. Subduction and blueschist metamorphism (Albian to Coniacian)

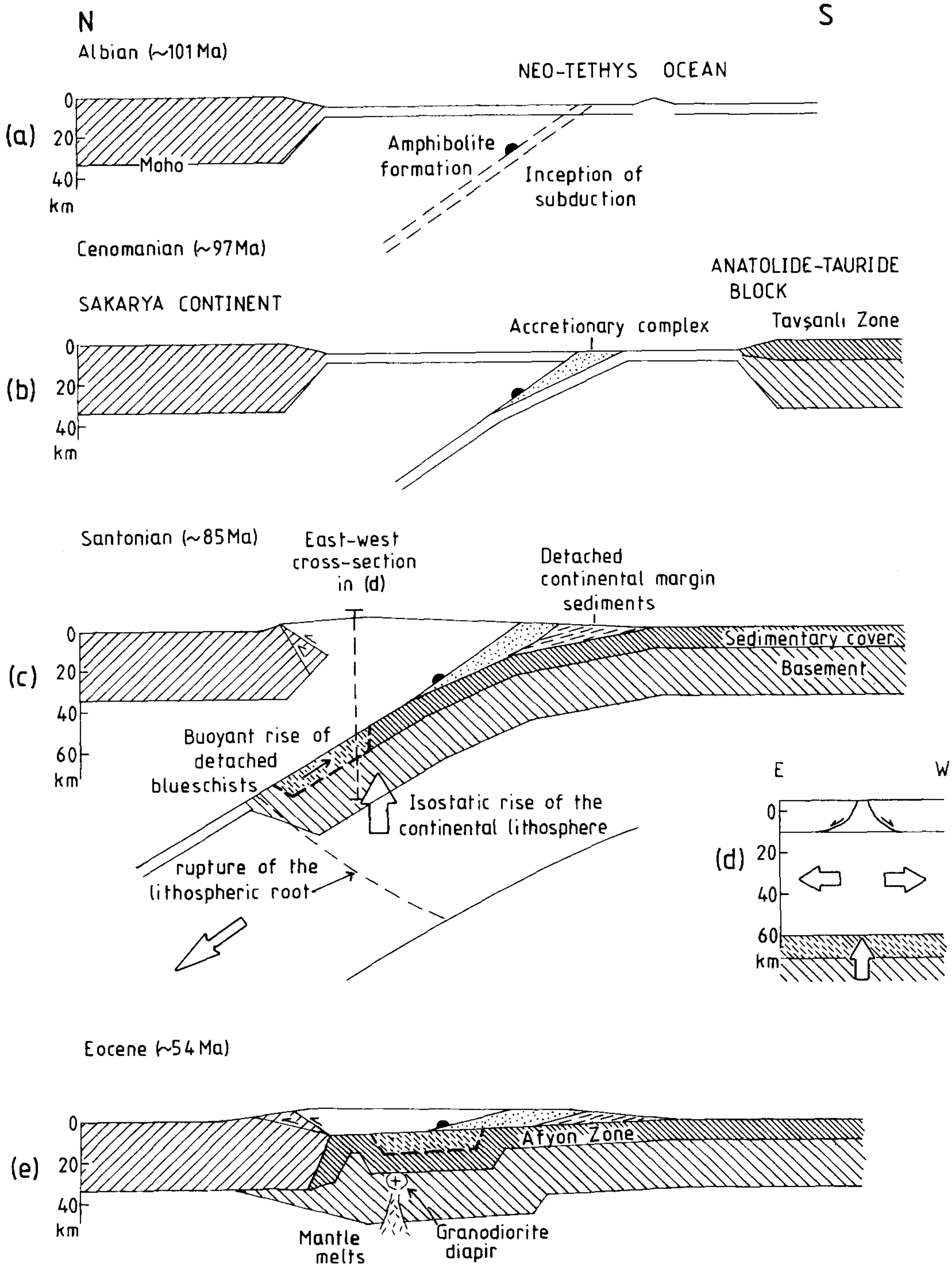
The initiation of northward-directed subduction of the Tethys is generally regarded as Albian based on the occurrence of mafic to intermediate tuffs within the Cenomanian–Maastrichtian deposits of the Sakarya Zone (Sengör and Yılmaz, 1981). However, the absence of Cretaceous intrusive rocks in the western part of the Sakarya Zone, including the region studied, suggests that either an Andean-type magmatic arc did not develop because the Tethys ocean was too narrow and/or the subduction zone was too shallow, or that the Neo-Tethyan ocean floor was being consumed along a northward-dipping intra-oceanic subduction zone. The extensive occurrence of peridotite slabs above the blueschists favours the latter hypothesis. The garnet-amphibolite metamorphic soles probably formed during the Albian inception of this intra-oceanic subduction zone through the residual heat of the overlying hot mantle wedge and subsequently underplated the hanging-wall (Fig. 9a). The estimated 8.5 ± 3.5 kbar pressure for the garnet-amphibolite soles from the Orhaneli area corresponds to a 25 ± 10-km-thick overburden of oceanic crust and mantle (Fig. 9a). The volcano-sedimentary complex that was accreted in the Albian–Cenomanian interval came into contact with a cooler hanging-wall and underwent incipient blueschist facies metamorphism along with the garnet-amphibolite (Fig. 9b). Similar high P/T conditions inferred for the volcano-sedimentary complex and for the incipient blueschist facies metamorphism in the garnet-amphibolite slices suggest that these two contrasting units resided jointly in an oceanic subduction zone. A similar scenario is envisaged for the garnet-amphibolites underlying an ultramafic slab in the Cascades in Washington, that have undergone a blueschist-facies overprint (Brown et al., 1982).

Following the subduction of the oceanic lithosphere, the northern passive margin of the Anatolide–Tauride platform was probably subducted in the intra-oceanic subduction zone resulting in the blueschist-facies metamorphism in the Tavsanlı Zone (Fig. 9c). The lithostatic pressure over the subducted continental margin was created by a 60-km-thick oceanic mantle wedge and by the underplated accretionary prism, as shown by the present position of the blueschists under the peridotite and the volcano-sedimentary complex.

8.3. Blueschist exhumation (Coniacian to Maastrichtian)

A major tectonic problem in the Tavsanlı Zone, and also in other blueschist regions like Oman or New Caledonia, is the mechanism of removal of the 60-km-thick wedge above the subducted continental slab. A model explaining the removal of this mantle

Fig. 9. Schematic cross-sections illustrating a possible model for formation and exhumation of blueschists in northwest Turkey. (a) Onset of north-dipping subduction zone and formation of garnet-amphibolites under a hot mantle wedge (black half dot). (b) Narrowing of the Tethys by continuous northward subduction and formation of an accretionary complex. (c) Subduction of the Anatolide–Tauride platform, and blueschist metamorphism in the continental crust. Blueschists are subsequently detached from their basement and begin their buoyancy-driven ascent along the subduction zone. The subduction zone begins to shallow after the rupture of the oceanic lithospheric root. (d) East–west cross-section illustrating ductile flow in the lower part, and extensional faulting in the upper part of the mantle wedge to accommodate the rising subducted continental lithosphere. (e) The mantle wedge is thinned and the blueschists, metamorphosed at 60 km depth, are juxtaposed with accretionary complex and garnet-amphibolites.


N
S
Albian (~101 Ma)
NEO-TETHYS OCEAN
(a)
0
20
40
km
Moho
Amphibolite formation
Inception of subduction
Cenomanian (~97 Ma)
ANATOLIDE-TAURIDE BLOCK
SAKARYA CONTINENT
Accretionary complex
Tavşanlı Zone
(b)
Santonian (~85 Ma)
East-west cross-section in (d)
Detached continental margin sediments
Sedimentary cover
Basement
(c)
60
Buoyant rise of detached blueschists
Isostatic rise of the continental lithosphere
rupture of the lithospheric root
E
W
(d)
Eocene (~54 Ma)
Afyon Zone
(e)
Mantle melts
Granodiorite diapir

wedge in northwest Turkey must be consistent with the following regional observations (Fig. 2):

(a) Blueschist detritus, including glaucophane-lawsonite schist pebbles, occurs in the upper Campanian–Paleocene clastics of the Sakarya Zone (Norman and Rad, 1971; Batman, 1978), suggesting that at least some blueschists in the eastern part of the Tavsanlı Zone were already exhumed by the Late Cretaceous before the Paleocene continent–continent collision.

(b) Carbonate deposition continued in the Afyon Zone to the south of the Tavsanlı Zone until the early Maastrichtian, and was followed by the deposition of over 3-km-thick upper Maastrichtian–Lower Paleocene wildflysch with blueschist and ophiolite blocks (Fig. 2). Peridotite slices were thrust southward over the wild-flysch in Middle to Late Paleocene (Göncüoğlu et al., 1992). This suggests that southward thrusting had apparently no role in the blueschist exhumation until the middle Paleocene, by which time the blueschists were at shallow depths or even exposed.

(c) The absence of Cretaceous–Paleocene sedimentation in the Tavsanlı Zone suggests that it underwent erosion during the exhumation of the blueschists.

(d) The east–west trend of the suture and the north–south orthogonal convergence between Laurasia and Gondwana during the Late Cretaceous–Paleocene (e.g., Patriat et al., 1982) indicate northward orthogonal subduction during the Cretaceous. Thus, the mineral elongation lineation in the blueschists, which is generally subparallel to the trend of the suture (Fig. 1), was apparently formed or rotated under east–west extension at a high angle to the direction of subduction. The mineral stretching lineation in the blueschists is generally shown by elongate calcite grains in marble. As aragonite, and not calcite, was the stable $CaCO_3$ polymorph during the high P/T metamorphism (cf. Fig. 5), the stretching lineation must have formed during the exhumation of the blueschists.

(e) In obducted ophiolites, the compositional and tectonite fabrics in the ophiolite are generally parallel to the basal thrust. The almost perpendicular attitude of the layering relative to the basal thrust in the Burhan peridotite body indicates 80–90° rotation of the ophiolite along faults. Segments of such faults are probably preserved as the basal fault of the Burhan peridotite body. The attitude of the layering in the peridotite slab indicates rotation along a north-trending subhorizontal axis, suggesting E–W extension, which is normal to the assumed direction of subduction. Extension and block rotation must have occurred between about 88 Ma (the age of the HP/LT metamorphism) and 52–48 Ma (the age of intrusion of the granodiorites into the blueschist/peridotite contact). During this time interval the blueschists in the Orhaneli area rose from a depth of 60 km to 10 km.

(f) The contact between blueschists and overlying peridotite is a sharp fault, and not a thick mylonite zone, as described from many extensional detachment faults (e.g., Davis, 1980).

(g) Pre- or syn-metamorphic thrusting would have resulted in repetition of the stratigraphic sequence, and be indicative of underplating. Such structures are not recognised in the blueschist sequence, although they may be present below the exposed level of the blueschist micaschists in the Kocasu region.

(h) Palaeomagnetic data and kinematic reconstructions based on the opening of the Atlantic and Indian oceans (e.g., Patriat et al., 1982) indicate that Gondwana and Laurasia were on a collision course during the Senonian. This suggests that the blueschists were exhumed within an overall compressive regime.

(i) The Middle Eocene marine limestones, which rest unconformably on the peridotite in the Tavsanlı Zone northeast of Tavsanlı (Fig. 1), indicate a normal thickness of continental crust at this time.

The structural position of the blueschists below oceanic sequences that record lower metamorphic pressures is not consistent with a simple thrusting–erosion model (Hsu, 1991). A model involving extrusion between two colliding continental plates is also not applicable because the blueschists were at lower crustal levels or on the surface prior to the Paleocene continent–continent collision, as evidenced by the blueschist detritus in the Upper Cretaceous flysch in the Sakarya Zone. A steady-state model involving underplating and extension (Platt, 1986) could bring the blueschists nearer the surface, but it would essentially leave the subduction geometry unchanged and so cannot explain how the continental crust returned to a normal thickness before the Middle Eocene.

The model that best meets the regional observations is a combination of detachment in the upper crust and buoyancy of subducted continental lithosphere (Wijbrans et al., 1993), and of progressive shallowing of the subduction zone (Krueger and Jones, 1989). The first mechanism involves the detachment of the upper crust from its basement, incorporation into the hanging-wall of the subduction zone and subsequent buoyant ascent. In this way it is possible to juxtapose regional blueschists metamorphosed at 60 km depth with the accretionary complex at 25 km depth without requiring a thrust that reaches the surface (Fig. 9c). Mechanical arguments for this mechanism are discussed by England and Holland (1979). The buoyancy forces result from the density contrast between the crustal slice and the surrounding mantle, the thickness and the viscosity of the crustal slice, and the angle of subduction. Opposing forces result from the shear stress on the down-going slab, which depends on the velocity of subduction. Subduction of the continental crust will result in a major increase in the density contrast between the down-going slab and the surrounding mantle, and a decrease in subduction velocity, especially after the rupture of the attached oceanic lithospheric root. The density of the blueschists in the Orhaneli area, calculated from the average mineral modes of the micaschists (Okay and Kelley, 1994) is 2.95, substantially lower than that of the upper mantle (3.31). Minimum vertical and north–south structural dimensions of the blueschists of the Tavsanlı Zone suggest that the detached upper crustal slice was at least 5 km thick and 30 km long. For a subduction angle of 30° the detached crustal slice must have moved 70 km up in the subduction zone. Thermal relaxation within the ascending slice would have been impeded by the underplated, cold continental crustal material and the temperature of the overlying mantle wedge (Fig. 9c).

The buoyant ascent of the detached upper crustal slice along the subduction zone does not, however, lead to a removal of the mantle wedge above the descending continental slab. A mechanism that would remove this wedge is subduction shallowing, which has been invoked as a mechanism for the exhumation of the Franciscan blueschists (Krueger and Jones, 1989). Major changes in subduction dip have been suggested in western North America (Coney and Reynolds, 1977; Keith, 1978) and in the Andes (Jordan et al., 1983) based on the variation in the location of the magmatic arc. In North America the subduction dip is estimated to have decreased from 50° to 10° in 20 m.y. (Keith, 1978). In the case of subduction of continental lithosphere, the subduction dip would be expected to decrease rapidly following the rupture of the attached oceanic lithospheric root, due to the density contrast between continental and oceanic lithospheres. The shallowing of the continental slab would induce uplift of the mantle wedge, which presumably would thin by ductile flow in the viscous lower part and by extensional faulting (e.g., Dewey, 1988) in the upper brittle part (Fig. 9d). As north–south compression and continental subduction were probably continuing during the shallowing of the subducted continental slab in the Campanian–Maastrichtian interval, extension would be induced in an east–west direction, as observed in present-day Tibet, where north–south thrusting is occurring contemporaneously with east–west extension (e.g., Mercier et al., 1987). The north-south-striking normal faults in the Burhan peridotite slab and the east–west-trending mineral stretching lineation in the blueschists may have formed during this time interval.

8.4. Collision and post-collision compression (Paleocene to Miocene)

The Paleocene transition from flysch to molasse sedimentation in the Sakarya Zone and the Paleocene southward thrusting of the peridotite over the Afyon Zone (Fig. 2) may be taken as the onset of the continent–continent collision. The continuing compression in the post-continental-collision stage in western Turkey resulted in southward-younging thrusts from the Paleocene in the Afyon Zone to the Early Miocene in the Lycian nappes. A major event in this period in the Tavsanlı Zone was the Eocene calc-alkaline plutonism. The calc-alkaline magmatism may be related to the upwelling of the asthenosphere following the rapid isostatic rebound of the subducted continental slab (Fig. 9c). Such an upwelling would result in partial melting of the upwelling mantle and possible anatexis in the overlying crust (Fig. 9e). The location of the Eocene plutons in the northernmost part of the Tavsanlı Zone, where maximum asthenospheric upwelling is expected to have occurred, supports such a model.

The basement of the subducted continental crust in the Tavsanlı Zone presumably consisted of biotite-bearing Precambrian gneisses similar to those exposed in the core of the Menderes Massif to the south (e.g., Satır and Friedrichsen, 1986). The base of this crystalline basement must have undergone dehydration melting by convective heat provided by mantle-derived basaltic melts. The depth of dehydration melting, based on the granodiorite geochemistry (Harris et al., 1994) was less than 30 km. Assuming that the continental crust was of normal thickness in the Middle Eocene, as indicated by the nummulite-bearing marine limestones unconformably overlying the peridotite (Fig. 1), this places the melting zone at the base of the crust. Significant fractions of granite melt are likely to be transported extremely rapidly through the crust by fracture propagation (Clemens and Mawer, 1992). Hence, granitic melt production and emplacement of plutons are closely spaced in time, indicating high transient geotherms in the continental crust during the Middle Eocene (Fig. 5). The preservation of the blueschist assemblages under these conditions requires that fluid-absent conditions prevailed. It may be predicted, however, that the lower parts of the crust in the Tavsanlı Zone will show an increasingly higher temperature overprint on the blueschist/eclogite assemblages.

9. Conclusions

The regional blueschist terranes in northwest Turkey represent a subducted passive continental margin, which was initially overlain by a wedge of oceanic lithosphere, 60 km thick. The exhumation of the blueschists was achieved in a compressive regime by a combination of two mechanisms: (1) the detachment of blueschists from their lower-crustal basement in the subduction zone followed by buoyant ascent in the hanging-wall of the subduction zone (Wijbrans et al., 1993), and (2) the shallowing of the subducted continental lithospheric slab (Krueger and Jones, 1989). The underlying cause of both mechanisms was the buoyancy of the continental crust surrounded by mantle rocks. This model is applicable to regional blueschists that were metamorphosed under a mantle wedge, such as those from Oman (e.g., Goffé et al., 1988) or from New Caledonia (e.g., Brothers and Blake, 1973).

A second conclusion relates the preservation of blueschist facies mineral assemblages to their exhumation. Rapid exhumation is commonly invoked to preserve blueschist minerals, such as jadeite or lawsonite. In northwest Turkey, south of Bursa, blueschist facies rocks metamorphosed at 88 Ma were at 10 km depth at 53 Ma and probably did not reach the surface until 18 Ma, and yet unstable HP/LT metamorphic minerals such as jadeite and lawsonite are preserved. Similarly, Triassic ultrahigh-pressure metamorphic rocks with coesite in Dabie Shan in China are intruded by Cretaceous coarse-grained granitic rocks (Okay et al., 1993), indicating that they were still at considerable depth 60 Ma after the ultrahigh-pressure metamorphism. It seems that HP/LT metamorphic rocks can reside tens of million years in the upper parts of the crust and preserve their mineral assemblages provided they stay dry; mineral reaction kinetics at low temperatures are slow even on a geological time scale provided fluid-absent conditions prevail (cf. Hacker et al., 1992).

In well-studied orogenic belts such as in the Alps or in the Himalayas, sutures separating rocks deposited on opposing margins of oceans are recognised as steep faults or steeply dipping narrow shear zones. This is also the case in northwest Turkey, where the Tethyan suture is represented by a single fault plane or by a melange zone less than 3 km across. This suture zone in northwest Turkey separates unmetamorphosed shallow-marine Jurassic limestones from regional blueschists metamorphosed at 60 km depth during the Late Cretaceous. However, the exhumation of blueschists was largely completed prior to the continental collision, and the formation of the suture zone had a little role in their exhumation.

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