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ALPINE-HIMALAYAN BLUESCHISTS

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INTRODUCTION

Blueschists gained a special tectonic significance after the advent of plate tectonics as marking the location of former subduction zones. Eskola (1939) proposed a high-pressure/low-temperature origin for the blueschists based on the greater density and hydrous nature of their minerals compared with the minerals of the neighboring greenschist and amphibolite facies. During the 1940s and 1950s a high-pressure/low temperature (HP/LT) origin for the blueschists was upheld in Europe (e.g. Brouwer & Egeler 1952, de Roever 1955), whereas a metasomatic origin was generally favored among geologists studying the North American Franciscan complex (e.g. Taliaferro 1943, Turner 1948, Brothers 1954). During the 1960s an HP/LT metamorphic origin for the blueschists became the predominant view (e.g. Coleman & Lee, 1963, Ernst 1965), although a satisfactory tectonic mechanism for creating the abnormal conditions required for blueschist generation was still missing. This mechanism was provided eventually by plate tectonics; Hamilton (1969) ascribed the origin of the Franciscan blueschists to processes in subduction zones, where the cold oceanic lithosphere is carried down into the hot asthenosphere. The very low heat conductivity of rocks is responsible for the generation of HP/LT conditions in the downgoing slab.

The southern part of the Cordilleran orogen in North America, where the Franciscan complex is located, is a noncollisional orogen, and the Franciscan blueschists represent rocks carried down to the subduction zone and subsequently uplifted and accreted into the active continental margin. Thus, the protoliths of the blueschists are either basic volcanic rocks and pelagic sediments representing the upper levels of the downgoing oceanic crust or voluminous graywackes carried down into the trench by turbidity currents. By contrast, the Alpine-Himalayan orogenic system is

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a major collisional orogenic belt formed through the collision of two major continents, Laurasia and Gondwana-Land (Figure 1). Although blueschists were first described and studied in the Alpine orogenic belt, their origin poses more problems in this region than in the Circum-Pacific blueschist complexes, where the dynamic, steady-state plate movements

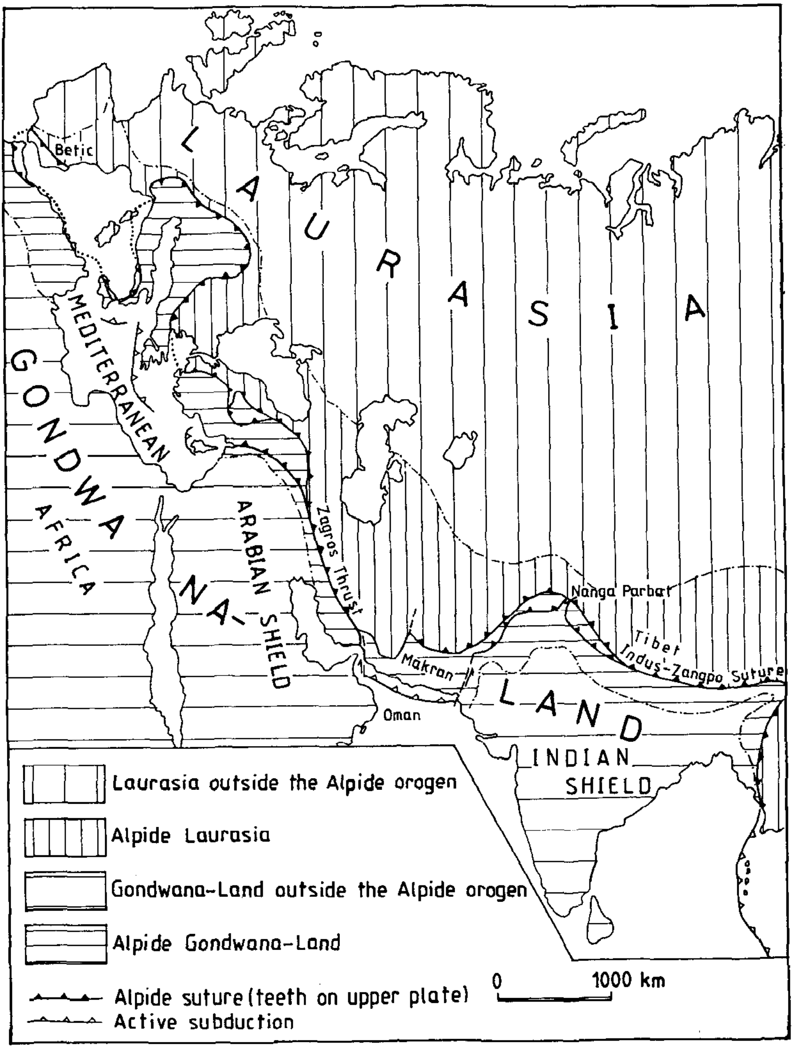

Figure 1 Tectonic map of Eurasia showing the major Alpide sutures and the Alpine/Himalayan orogen. Modified from Şengör (1987).

provide a plausible mechanism for their origin. Are the Alpine-Himalayan blueschists the result of a similar steady-state subduction process, or have they formed during continental collision or ophiolite obduction? What type of precollisional paleogeographic setting do the blueschist protoliths represent: oceanic crust, continental margin, or thinned or normal continental crust? In what part of the orogenic tectonic stack do the blueschists occur? What controls their distribution along the orogen? This review describes the major Alpine-Himalayan blueschist complexes in the framework of the above questions.

BLUESCHIST FACIES

The blueschist facies is characterized by a unique series of dense minerals comprising sodic amphibole, lawsonite, jadeite-rich sodic pyroxene, and aragonite. The more exotic blueschist minerals include deerite, ferro- and magnesiocarpholite, sudoite, etc. Apart from these major diagnostic blueschist minerals, garnet, pumpellyite, and greenschist facies minerals (with the exception of biotite and calcic amphibole) also frequently occur in this facies.

The blueschist facies is defined, in common with the recent general usage, in terms of mineral parageneses in the metabasic rocks characterized by the persistent and abundant presence of sodic amphibole with lawsonite or epidote or pumpellyite. Contrary to earlier views, recent experimental work (Maruyama et al 1986) has confirmed that glaucophane is stable only at high pressures of above 7 kbar. The only exception made to this definition of the blueschist facies is to include incipiently metamorphosed rocks, which largely preserve their igneous or sedimentary textures and minerals but also contain variable amounts of HP/LT minerals into the realm of the blueschist facies. Such incipiently metamorphosed blueschists are relatively widespread and have been described from the Franciscan complex (Coleman & Lee 1963), western Turkey (Okay 1982), Makran (McCall 1985), and the Himalaya (Honegger et al 1989).

In blueschist metabasites, sodic amphibole and Ca-Al silicate generally make up the bulk of the rock. Sodic amphibole in metabasites is generally of crossite and glaucophane composition (cf. Leake 1978), and jadeite does not normally occur in the metabasic rock compositions. Metamorphic terranes characterized by the presence of cogenetic calcic and sodic amphiboles in the metabasic rocks are regarded as transitional between the blueschist and greenschist facies. With increasing temperature, blueschists grade isobarically into eclogites involving continuous reactions between sodic amphibole, sodic pyroxene, garnet, and clinozoisite (Ridley 1984a, Schliestedt 1986, Okay 1989).

Blueschist facies minerals are best developed in metabasic rocks, metacherts, and metagraywackes and are only rarely found in metapelites, metaarenites, calc-silicates, and marbles. The bulk-rock chemistry of the latter group prevents the growth of the common blueschist minerals, and these lithologies exhibit generally low-grade greenschist facies mineral assemblages under blueschist facies conditions. Therefore, the recognition of blueschist facies conditions in metamorphic terranes made up of metapelites and marbles is difficult in the absence of metabasic interlayers. This is a serious problem in the Alpine-Himalayan orogen, where such metamorphosed passive continental margin sequences are common.

ALPINE-HIMALAYAN OROGEN

Permian paleogeographic constructions of the world indicate two megacontinents, Laurasia and Gondwana-Land, partly separated by a wedge-shaped westward-narrowing ocean, the Tethys. The Alpine-Himalayan orogen arose as a result of the continental collisions leading to the demise of the Tethys. A complexity in the evolution of the Alpine-Himalayan orogen was the rifting of continental blocks from Gondwana-Land and their collision with Laurasia prior to the terminal Gondwana-Land–Laurasia collision during the Tertiary. Such a major rifting event occurred during the Permo-Triassic, when a major continental fragment or fragments rifted off from Gondwana-Land and moved across the Tethys, creating new oceanic crust behind it (Neo-Tethys). Eventually, the fragment(s) collided with Laurasia during the early Mesozoic, causing Cimmeride orogenic movements in a zone extending from Turkey to China (Şengör 1987). The Indian subcontinent is another example of an independently moving microcontinent within the Tethys that rifted off from Gondwana-Land during the Jurassic; traveled across the Tethys, creating the Indian Ocean behind it; and finally collided with Laurasia in the Eocene, giving rise to the Himalayan orogeny (Figure 1). As is apparent from these rifting and collision events, the Laurasian margin of the Tethys was throughout its history mostly an active margin, whereas the Gondwana-Land margin was largely a passive continental margin, with the notable exception of the area of the Western Alps.

BLUESCHIST COMPLEXES OF THE ALPINE-HIMALAYAN OROGEN

Blueschists are not evenly distributed along the Alpine-Himalayan orogen. In terms of the outcrop area, over 80% of the known blueschists occur between Spain and western Turkey. This is partly a result of the better-known geology of this region compared with the much larger segment of the orogen between eastern Turkey and Indochina. However, recent detailed studies in the Himalaya suggest that the scarcity of the blueschists in this eastern segment is a real feature. The temporal distribution of the blueschists in the Alpine-Himalayan orogen is also uneven, with few blueschists associated with the early Mesozoic Cimmeride orogeny; most blueschists are related to the evolution of the Neo-Tethys.

Here we examine blueschists in four major areas along the Alpine-Himalayan chain (Figure 1): the western Apulia comprising the western Mediterranean area, the eastern Apulia of Greece and western Turkey, the Oman-Makran region, and the Himalaya.

Western Apulia

In the western Mediterranean region, a belt of blueschists up to 75 km wide can be followed along the internal zones of the strongly curved Alpide orogen from the Betic zone in Spain through Calabria and Corsica to the Alps s.s. (Figure 2). The continuity of this blueschist belt was strongly disrupted by later events, such as the opening of the Tyrrhenian oceanic basin, and by major strike-slip faulting and the extensive cover of the nappes or postorogenic deposits. However, individual blueschist complexes in this region share many common features, such as the age of HP/LT metamorphism, tectonic setting, and type of protolith, and they clearly form a single metamorphic province, here called the western Apulian blueschist belt.

The western Mediterranean sea is characterized by a Hercynian (late Paleozoic) metamorphic basement with abundant granitic rocks. The Alpine cycle started with the Triassic transgression leading to the deposition of thick platform carbonates in the central Mediterranean, while in the surrounding regions thinner red bed, limestone, and evaporite sequences were laid down. The rifting of the carbonate platform began in the Early to Middle Jurassic and was related to the opening of the North Atlantic. It led to the differentiation of a northern Laurasian continental margin, corresponding with the Helvetic zone of the Western Alps, a central basin with two oceanic seaways (Valais and Piemont) separated by a continental platform (Briançonnais) and a southern carbonate platform–type continental margin (Austroalpine zone and the Southern Alps; see Figure 3), which was initially contiguous to Gondwana-Land and formed its northern passive continental margin (Trümpy 1975, Frisch 1979). During the Jurassic a large continental fragment, called Apulia or Adria (Channel et al 1979), split off from Gondwana-Land with the creation of a narrow oceanic seaway, whose remnant is the present day eastern

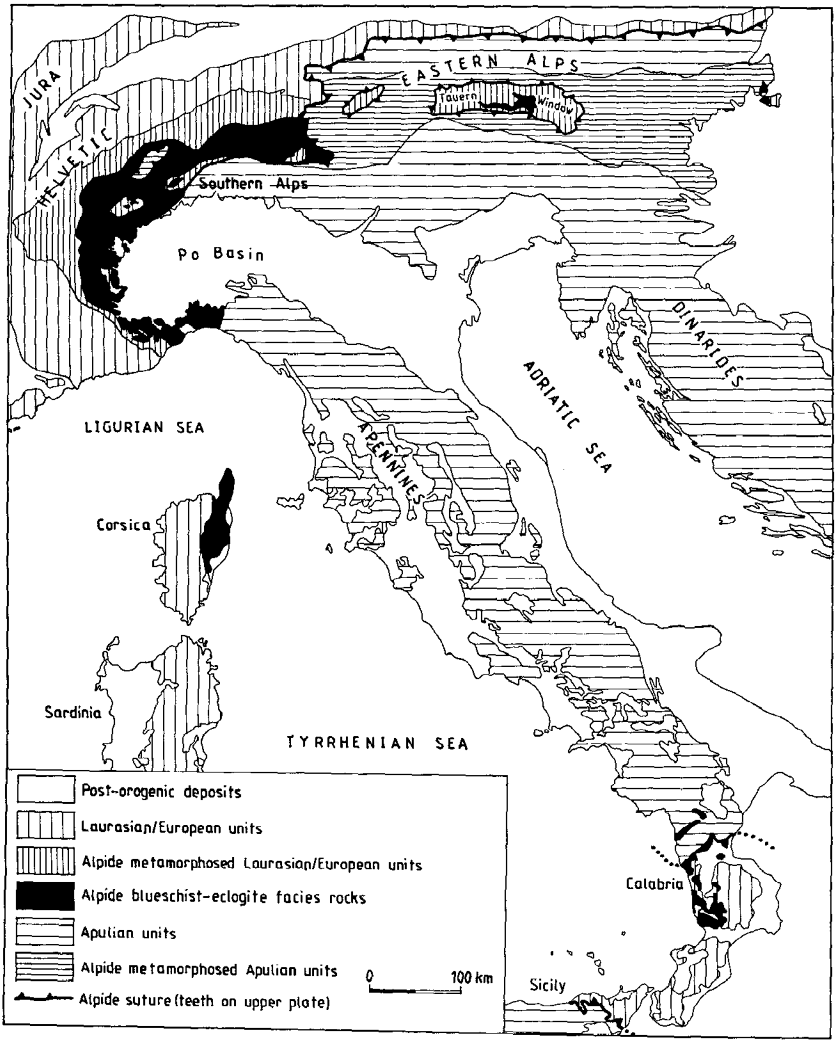

Figure 2 Simplified tectonic map of southcentral Europe, showing the western Apulian blueschist belt. Regions shown in black indicate blueschist/eclogite outcrops irrespective of the degree of the later, lower P/T overprint.

Mediterranean (Figure 3; Dewey et al 1973, Biju-Duval et al 1977). The Alpide orogeny between Italy and western Turkey is the result of the collision of this Apulian platform with Laurasia.

The western Apulian blueschists can be grouped into three regions:

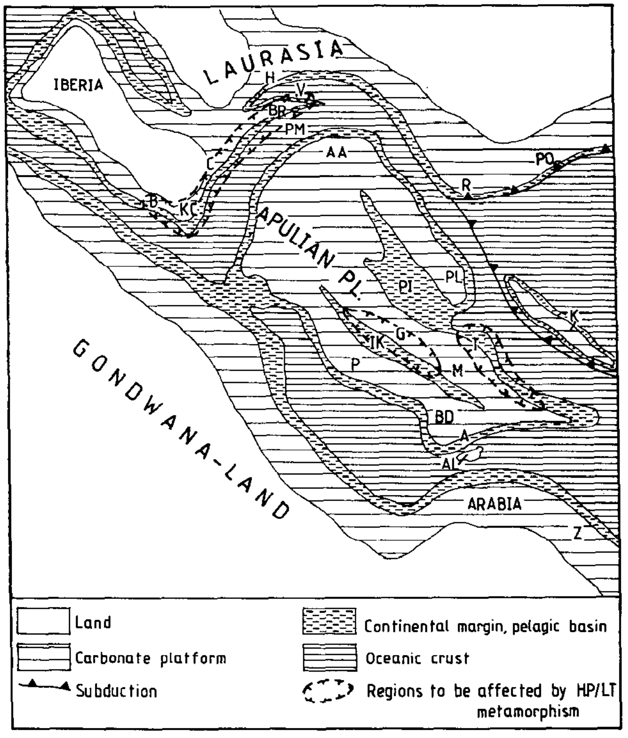

Figure 3 Schematized paleogeographic map of the Apulian platform for the Late Jurassic–Early Cretaceous. Modified from Dercourt et al (1985) and Şengör et al (1984b). A–Antalya, AA–Austroalpine, AL–Alanya, B–Betic, BD–Beydağları, BR–Briançonnais, C–Corsica, G–Gavrova-Tripolitza, H–Helvetic, IK–Ionian-Kızılca, K–Kırşehir, KC–Kabyl-Calabria, M–Menderes, P–Paxos, PI–Pindos, PL–Pelagonian, PM–Piemont, PO–Pontid, R–Rhodope, T–Tavşanlı, V-Valais, Z–Zagros.

Alpine (s.s.) blueschists (including those of the Eastern Alps and Corsica), Calabrian blueschists, and the Betic blueschists.

ALPINE (S.S.) BLUESCHISTS Most of the Penninic zone of the Alps, comprising paleogeographically the Valais and Piemont oceans and the Briançonnais platform, has undergone an HP/LT metamorphism (Figure 2). The Penninic zone consists of a number of north- and northwest-verging nappes that are tectonically overlain by the Austroalpine units representing rocks of the Apulian platform. In the Eastern Alps the Penninic zone outcrops in three large tectonic windows—Tauern, Engadine, and Wechsel-Semmering—underneath Austroalpine units that do not show blueschist facies metamorphism, while in the Western Alps most of the

Austroalpine cover has been removed by erosion (Figure 2). The Penninic nappes are thrust northward and westward over the Helvetic zone or on the Hercynian granitic basement.

The Penninic zone comprises Hercynian basement rocks, their detached Mesozoic sedimentary cover, and volcanic and pelagic sedimentary rocks (schistes lustres) of the Penninic oceanic basins and dismembered ophiolites. The Hercynian granitic and metamorphic rocks, which form large ductilely deformed basement nappes, are interpreted as relicts of continental fragments within the Penninic oceanic basins. Both the basement and cover nappes and the ophiolites of the Penninic zone are affected by an HP/LT metamorphism overprinted by a lower P/T metamorphism.

The Cretaceous HP/LT metamorphism in the Alps predated the major Eocene deformation that produced the nappe structure of the Penninic zone (e.g. Milnes 1978), and thus the HP/LT metamorphic rocks in the Alps occur as a collage of coherent tectonic units of differing metamorphic grades and with independent metamorphic histories. It is not clear whether these HP/LT metamorphic units initially constituted parts of a single large metamorphic terrain or whether they were metamorphosed piecemeal; the presence of some extremely high-pressure metamorphic rocks (Chopin 1987) and significant variation in isotopic ages suggest the latter possibility. The delineation of these HP/LT megablocks is largely hindered by the strong Eocene deformation and metamorphism.

In the Western Alps the HP/LT metamorphism shows a general decrease in grade from the internal parts, where eclogitic paragenesis is common, toward the external parts, which only show a blueschist facies metamorphism (e.g. Niggli 1973, Ernst 1973, Frey et al 1974). The essential ecologite paragenesis of omphacite + garnet is described from both the Penninic basement and cover nappes and the Penninic ophiolites (e.g. Bearth 1973, Evans et al 1979, Heinrich 1986, Pognante & Kienast 1987). In this 500-km-long belt of eclogite facies metamorphism there are Penninic tectonic units, such as the Grand Combin zone in the Zermatt-Saas area, which although having appropriate basic volcanic rocks do not show any evidence for early eclogite facies metamorphism (Dal Piaz & Ernst 1978).

The highest grade HP/LT metamorphic rocks in the Alps are found in the Dora Maira massif, which is one of the internal Hercynian basement complexes. A few-meters-thick but laterally several-kilometers-continuous metaquartzite band within the pre-Alpine basement of the Dora Maira massif contains very pyrope-rich garnet (90–98 mol % pyrope) with coesite inclusions, coexisting in textural equilibrium with phengite, kyanite, talc, and rutile (Chopin 1984, 1987). The mineral assemblage indicates pressures of formation of above 28 kbar and temperatures of about 700°C. The metaquartzite layer is in contact with thick monotonous albite- and biotite-bearing gneisses with marble layers and kyanite-eclogite lenses. Chopin (1984, 1987) believes that the coesite-bearing band represents an evaporitic sediment and that the ultra-high-pressure Alpine metamorphism has also affected the surrounding rocks, which implies burial of upper crustal rocks of at least several tens of square kilometers in area to depths of over 90 km.

In contrast to the internal parts, the external parts of the Penninic zone show only blueschist facies metamorphism characterized by the general absence of almadine-rich garnet and omphacite. There is little data on the type of transition from the eclogite facies to blueschist facies in the Western Alps. However, this transition is most likely marked by major thrusts, normal faults, or shear zones rather than being represented by an isograd. Blueschist facies assemblages are widely described from the ophiolitic rocks in the schistes lustres from the western part of the external Penninic zone, where the greenschist metamorphic overprint has been slight (e.g. Steen 1975, Pognante & Kienast 1987), and from Corsica (e.g. Brouwer & Egeler 1952, Caron et al 1981, Gibbons et al 1986).

The HP/LT metamorphic rocks of the Western Alps either tectonically overlie the Helvetic zone or directly overlie the European continental basement. In Corsica the basal contact of the blueschist unit over the Hercynian basement is marked by a 1000-m-thick major ductile shear zone developed within the Hercynian granitic rocks (Gibbons & Horak 1984). The upper parts of the shear zone contain sodic amphibole, while calcic amphibole is stable in the lower parts. This can be interpreted as a cooling effect at the base of the blueschist unit as it overrides on its uplift path the warmer European continental basement. The metamorphism imposed on the granitic basement related to this thrusting has been dated at 90–113 Ma (Cohen et al 1981).

In the Eastern Alps, HP/LT metamorphic rocks occur mainly in the Tauren tectonic window, which exposes Penninic units underneath the Austroalpine crystalline basement (Figure 2). The Penninic units in the Tauern tectonic window comprise a pre-Alpine crystalline basement and an allochthonous cover of Triassic carbonates and schistes lustres. As in the Western Alps, only part of the Penninic units in the Tauern window are apparently affected by the HP/LT metamorphism (Figure 2; Frank et al 1987). However, despite recent detailed petrological work, the limit and nature of these blueschist units are poorly known. High-grade kyanite-eclogites, estimated to have formed at pressures close to 20 kbar, occur only in a 3000-m-thick zone in the central Tauern window (Miller 1974, Holland 1979). This eclogite zone and the surrounding rocks are affected by a second glaucophane-lawsonite–type HP/LT metamorphism, which has a wider distribution but has apparently not affected the northern and eastern parts of the Tauern window (Frank et al 1987). A ubiquitous mid-Tertiary lower P/T metamorphism has widely overprinted the early HP/LT mineral assemblages and produced widespread pseudomorphs after lawsonite (e.g. Droop 1985, Selverstone & Spear 1985).

The Sesia-Lanzo zone, located in the internal parts of the Western Alps, has attracted considerable recent attention because it contains the only large eclogitic metamorphic terrane in the Alps with no lower P/T overprint and thus provides a clear example of the eclogite facies metamorphism in the continental crust. Unlike the other Alpine units affected by the HP/LT metamorphism, the Sesia-Lanzo zone is regarded as part of the Austroalpine, and thus Apulian, pre-Triassic continental basement. The Sesia-Lanzo zone consists of Hercynian high-grade schists, amphibolites, marbles, and ultramafic rocks, all intruded by late Hercynian granitoids and leucocratic dikes. This pre-Alpine basement was deformed and metamorphosed in the eclogite facies, which produced glaucophane-eclogites and metagranitoids with the mineral assemblage of quartz+jadeite-rich sodic pyroxene+garnet+phengite+paragonite (Compagnoni & Maffeo 1973, Compagnoni et al 1977, Koons et al 1987).

The isotopic dates for the Alpine HP/LT mineral assemblages show a relatively wide scatter but generally fall between 70 and 110 Ma (e.g Hunziker 1974, Bocquet et al 1974, Bonhomme et al 1980, Deutsch 1983, Chopin & Monie 1984). The presence of glaucophane and lawsonite detritus in a late Turonian Penninic flysch (Winker & Bernoulli 1986) indicates that some of the blueschists were exposed as early as 90 Ma. The HP/LT metamorphic rocks in the Alps are widely overprinted by a lower P/T metamorphism dated at around 48 Ma (e.g. Frey et al 1974) that partially and in some cases probably completely destroyed the early HP/LT mineral assemblages. In the Lepontine region (central Alps) this second Meso-Alpine metamorphism reached sillimanite grade. Naturally, no relics of blueschist mineral assemblages remain in this region; however, the presence of these assemblages on both sides of the Lepontine region suggests that this region had also undergone the early HP/LT metamorphism (Figure 2).

CALABRIAN BLUESCHISTS Calabria represents the southern continuation of the Alpine suture (Figure 1). Before the Oligocene, Calabria was probably positioned southeast of the present Sardinian coast, much nearer to the Alpine (s.s.) belt (Figure 3; Alvarez 1976). Calabria is made up of three major tectonic complexes superposed between the late Eocene and early Miocene (Ogniben 1973, Carrara & Zuffa 1976). The first is a carbonate-clastic unit of Triassic to Late Cretaceous/Paleocene age of Apulian affinity showing low-grade metamorphism ascribed to the greenschist facies. It is overlain tectonically by an ophiolite-flysch unit, which is in turn overthrust by a composite Hercynian crystalline basement nappe of high-grade gneiss, amphibolite, and granitoid. The ophiolite-flysch unit consists mainly of basalt, tuff, and rarer diabase and gabbro; the lower parts of the ophiolite stratigraphy are generally missing. The mafic volcanics are stratigraphically overlain by radiolarian cherts, limestones, and volcanoclastic rocks of Late Jurassic–Early Cretaceous age. This ophiolite-flysch sequence, exposed across an area of over 100 km in length (Figure 2), has undergone a blueschist facies metamorphism (Hoffmann 1970, de Roever 1972). There is a slight apparent increase in grade northward, with a lawsonite $\pm$ pumpellyite + albite + chlorite $\pm$ magnesioriebeckite assemblage in the south passing to a lawsonite + pumpellyite + glaucophane/crossite $\pm$ Na-pyroxene $\pm$ chlorite assemblage in the north. A greenschist facies overprint is also described farther south. The HP/LT metamorphism in Calabria is probably of mid- to Late Cretaceous age, similar to that in the Western Alps.

BETIC BLUESCHISTS The Betic Cordillera in southern Spain forms the westernmost part of the Alpide orogenic belt. It is a north-vergent orogen similar to the Western Alps, with an external Subbetic zone in the north made up of unmetamorphosed Mesozoic cover nappes and an internal Betic zone (s.s.) consisting of three main superposed tectonic units variably affected by Alpide metamorphism. Only the lowermost tectonic unit, the Nevado-Filabride complex, shows an early HP/LT metamorphism strongly overprinted by a lower P/T metamorphism; the two upper tectonic units, the Alpujarride and the Malaguide complexes, show only low- to medium-pressure metamorphism, which decreases in grade upward in the sequence (Egeler & Simon 1969, Torres-Roldan 1979).

The Nevado-Filabride complex is made up of a lower series of over 4-km-thick monotonous graphitic micaschists with quartzite intercalations believed to be pre-Permian in age overlain by a heterogeneous upper series of micaschists with metabasic, ultramafic, calc-schist, and marble horizons considered to be of Permo-Triassic age. The presence of ultramafic rocks in the sequence, likely relics of Alpide ophiolites, suggests that the Nevado-Filabride complex is a composite tectonic unit. It is not clear, mainly because of the lack of metabasic rocks in the lower series, whether the whole sequence has undergone the HP/LT metamorphism (cf. Gomez-Pugnaire 1984). The Nevado-Filabride complex could be similar to the Alanya nappes in southern Turkey, where units with differing metamorphic and structural histories were juxtaposed as nappes prior to the last deformational and metamorphic event (Okay 1989).

The upper series of the Nevado-Filabride complex shows a typical plurifacial metamorphism and, in fact, is the type locality where this term was coined (de Roever & Nijhuis 1964). The earliest eclogitic paragenesis in the metabasic rocks is omphacite + garnet + paragonite ± calcic amphibole ± kyanite. This eclogite facies metamorphism occurred, at least locally, under zero-strain conditions, resulting in the formation of coronatic eclogites where the igneous textures and some of the igneous clinopyroxene are preserved (Gomez-Pugnaire & Fernandez-Soler 1987). Sodic amphibole developed later, probably at the expense of garnet and omphacite, and was associated with the penetrative deformation. The HP/LT metamorphic rocks are variably overprinted by a lower P/T metamorphism, as evidenced by the formation of albite, barroisite, and chlorite in the metabasic rocks (de Roever & Nijhuis 1964).

The HP/LT metamorphism in the Betic zone is post-Triassic and probably Cretaceous in age, similar to the age of the blueschist metamorphism in the Calabria and Alps s.s. In fact, the tectonic line between the internal Betic zone and the external Betic is interpreted as a major Tertiary dextral transcurrent fault (Paquet 1974), which implies that the Betic zone was situated during the Cretaceous several hundred kilometers farther east, probably in direct continuity with the Alpide (s.s) orogenic belt (Figure 3).

Eastern Apulia

The Apulian microplate can be traced from the Austroalpine domain through the Dinarides, Carpathians, and Hellenides into the Anatolide/Taurides in Turkey, where its eastward continuation is terminated by a suture knot (Figure 1). The suture defining the northern boundary of the Apulian plate is followed from the northern margin of the Eastern Alps through the outer arc of the Carpathians and the Vardar zone in Greece to the Izmir-Ankara-Erzincan suture in Turkey. The Serbo-Macedonian and Rhodope massifs in Greece and the Pontides in Turkey represent the Laurasian continental margin (Figure 4).

Alpide metamorphic rock including blueschist outcrops are scarce in the Dinarides and Carpathians, which are dominantly made up of Mesozoic and Tertiary sedimentary rocks (cf. Metamorphic Map of Europe 1973). However, these rocks may be present under the sedimentary nappes of the Carpathians, similar to the case of the Eastern Alps. In contrast, blueschists occur in great abundance in the eastern Apulia region of Greece and Turkey (Figure 4).

There are several differences in the tectonic evolution of the western and eastern Apulia (cf. Robertson & Dixon 1984). The Neo-Tethys opened in the eastern Apulia during the Triassic–Early Jurassic rather than the mid-

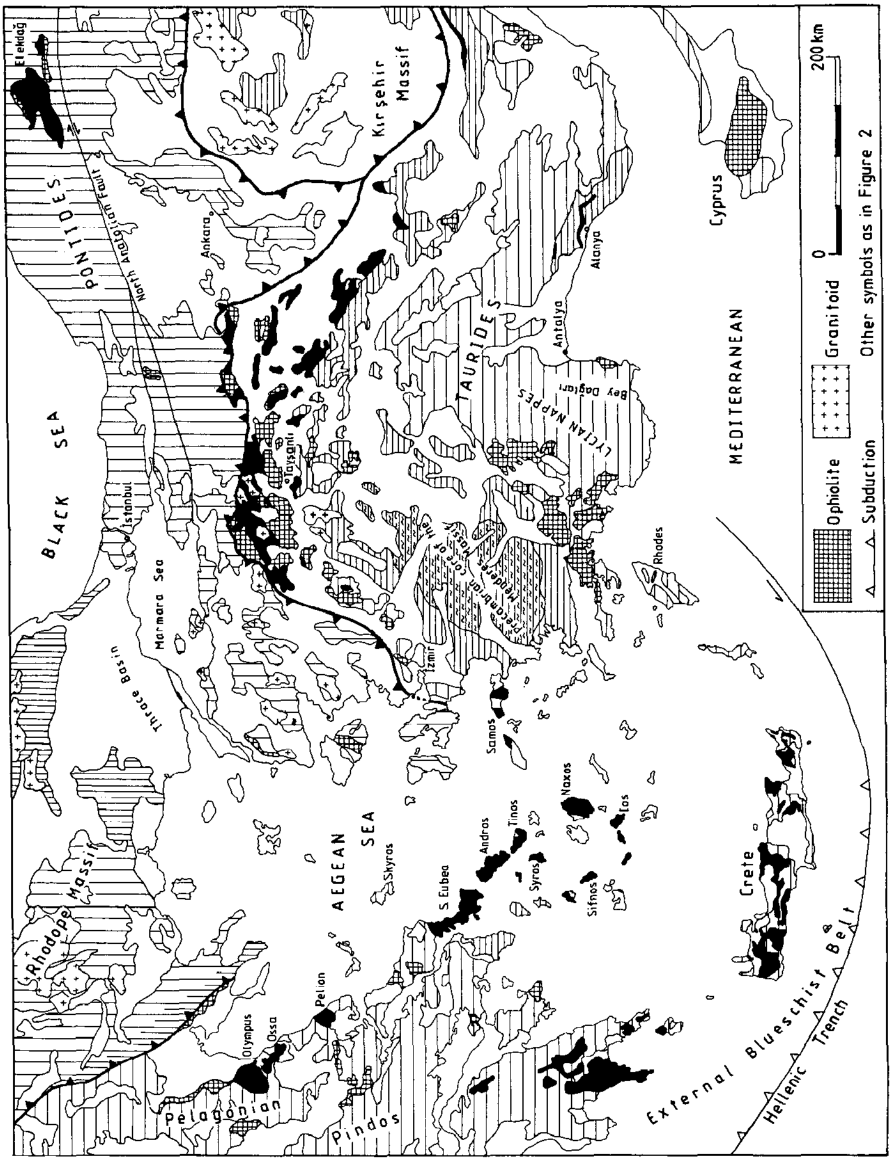

Figure 4 Tectonic map of Greece and western Turkey showing the eastern Apulian blueschist belts.

Jurassic, and it was consumed by a northward-dipping subduction zone; thus, unlike the Western Alps, the vergence of the main Alpide orogen in Turkey and in Greece is to the south and southwest. A major consequence of this was that, in contrast to the situation in the Alps s.s., the northern margin of Apulia was strongly affected by deformation and metamorphism (Figure 3), whereas the active Laurasian continental margin was relatively little involved in the Alpide orogeny.

The structural style in the Apulian units of Greece and Turkey is south- or southwest-vergent far-traveled nappes generally corresponding to specific paleogeographic domains. The nappes near the Neo-Tethyan suture (the internal zones) generally exhibit Alpide metamorphism, whereas those farther away (the external zones) commonly do not. In Greece, the Apulian platform consisted paleogeographically of alternating Mesozoic-Tertiary carbonate platform and basinal units, which are (from southwest to northeast) the Paxos carbonate platform, the Ionian/Plattenkalk basin, the Gavrovo-Tripolitza carbonate platform, the Pindos basin, and the Parnassos and Pelagonian carbonate platforms (Figure 3); to the northeast of the Pelagonian carbonate platform lay the Vardar zone, representing the northern continental margin of the Apulian plate in Greece (Bernoulli & Laubscher 1972, Jacobshagen 1986). These various paleogeographic domains now occur as immense nappe stacks. The major Alpide orogenic episodes in Greece were during the latest Jurassic and earliest Cretaceous, when ophiolites were obducted on the Pelagonian zone; during the mid- to late Eocene, when the major nappe structures of the Hellenides were produced probably as a result of continental collision; and during the early to middle Miocene, when there was major thrusting and nappe emplacement in the southern (external) parts of Greece.

In Turkey, the Anatolides represent the metamorphosed and the Taurides generally the unmetamorphosed, more external parts of the Apulian plate. As in Greece, various paleogeographic domains can be distinguished within the Anatolides and the Taurides (Şengör & Yılmaz 1981, Okay 1986, Poisson 1986): southward from the Izmir-Ankara suture, these are the Tavşanlı zone representing the northern passive continental margin of the Apulian plate; the Menderes massif carbonate platform and the Kızılca basin, which are probably the lateral equivalents of the Gavrovo-Tripolitza platform and the Ionian basin in Greece, respectively; the Beydağları carbonate platform and the Antalya basin, which probably developed oceanic crust during the Cretaceous; and the Alanya platform (Figure 3). The deformation history in Turkey was more complex than that in Greece, so that these paleogeographic domains cannot be observed as regular continuous zones. The two major Alpide orogenic episodes in western Turkey occurred during the mid- to Late Cretaceous, when ophiolites were obducted on the Anatolides and the Taurides, and during the mid- to late Eocene, when the Anatolides and the Taurides were tectonically sliced and formed south-vergent nappe piles (Şengör & Yılmaz 1981). During the early Miocene there was renewed southeasterly thrusting restricted to the southwestern Taurides (Brunn et al 1971).

Blueschists of various ages and grades are well represented in the eastern Apulian area of Greece and western Turkey (Figure 4). The earliest known blueschists in this area are the pre-Alpide Triassic Karakaya and Elekdağ blueschists in the Pontides of northern Turkey, which are related to the Cimmeride orogeny (Şengör et al 1984b, Okay 1986). In Greece the earliest blueschists seem to be small scattered occurrences associated with the latest Jurassic–earliest Cretaceous ophiolite obduction over the Pelagonian carbonate platform (e.g. Jacobshagen & Wallbrecher 1984). The only isotopic ages for these blueschists and associated high-pressure greenschists come from Crete and Gavdos, where phengite K/Ar ages are around 150 Ma (Late Jurassic; Seidel et al 1977). The tectonic setting, stratigraphy, and petrology of these early blueschist complexes are poorly known, and they are not further considered here.

The four major, relatively well described Alpide blueschist complexes of the region are the mid-Cretaceous Tavşanlı zone, the Alanya nappes, the mid-Eocene Cycladic belt, and the Oligocene–early Miocene External Hellenide belt. All four of these HP/LT belts lie within the Apulian platform (Figure 3).

THE TAVŞANLI ZONE BLUESCHISTS The Tavşanlı zone is situated directly south of the major Tethyan suture in western Turkey (Figure 4). It is made up of two main tectonic units: a more extensive and widespread coherent lower unit (Orhaneli unit) with a well-characterized laterally traceable stratigraphy and with a ductile metamorphic fabric, and an upper unit (Ovacık Unit) consisting of an imbricated volcano-sedimentary complex showing only brittle deformation and an incipient blueschist metamorphism (Çoğulu 1967, Servais 1981, Okay 1984). The Tavşanlı zone is tectonically overlain by an ophiolite represented now largely by ultramafic rocks. Mineral assemblages of the rare gabbros and dikes in the ultramafic rocks indicate that the ophiolite nappe has not undergone HP/LT metamorphism (Okay 1984). The Tavşanlı zone is thrust southward on the Late Cretaceous–Paleocene wildflysch of a Mesozoic carbonate platform unit (Afyon zone), which only shows greenschist facies metamorphism (Figure 4).

The lower unit of the Tavşanlı zone is characterized by a regular tripartite stratigraphy consisting of a lower phyllite unit passing upward into a several-kilometers-thick white massive marble unit, overlain by a thick section made up of intercalated metabasite, metachert, and metashale. The contacts between these units are gradational, and the whole sequence has undergone a uniform HP/LT metamorphism and ductile penetrative deformation. The lower unit of the Tavşanlı zone shows a typical prograde glaucophane-lawsonite zone metamorphism. The mineral assemblage of the metabasic rocks indicates a two-stage prograde metamorphism: an early static stage characterized by a lawsonite + sodic pyroxene + chlorite + sphene assemblage where sodic pyroxene is pseudomorphous after augite, followed by a sodic amphibole–forming event leading to the observed glaucophane-lawsonite zone mineral assemblage of sodic amphibole + lawsonite + chlorite + phengite + sodic pyroxene + sphene (Okay 1980a,b). Sodic amphibole formation is associated with the onset of penetrative deformation; rocks directly underlying the ophiolite nappe or the upper Tavşanlı unit still retain the lawsonite zone mineral assemblages, presumably because penetrative deformation of these rocks was inhibited by the thick and rigid ophiolite lid (Okay 1980b, 1986).

The upper unit in the Tavşanlı zone consists of complexly imbricated slices of spilitized basalt and agglomerate, radiolarian chert, red and green pelagic shale, graywacke, and tectonic lenses of serpentinite and talc. Although the rocks appear unmetamorphosed in the field, and the primary igneous textures and minerals are largely retained, typical blueschist minerals such as lawsonite, aragonite, and sodic pyroxene can locally be observed to have grown in the amygdales and veins of spilites, and pelagic limestones have been transformed into aragonite marbles comprising several-centimeters-large aragonite crystals (Okay 1982). Texturally and mineralogically, these rocks are akin to the Franciscan greenstones and jadeite-bearing graywackes (Coleman & Lee 1963, Ernst 1965). Associated with this incipient blueschist metamorphism, there has been a regional metasomatism involving topotactic replacement of augite in spilites by sodic pyroxene, resulting in rocks with 7–8% Na_2O (Okay 1982).

Data on the depositional age of the Tavşanlı zone are scarce. Triassic and Late Jurassic–Early Cretaceous paleontological ages are obtained from the upper unit (Servais 1981; Ş. Genç, personal communication, 1987). A speculative late Paleozoic–Mesozoic depositional age can be assigned to the lower unit based on stratigraphical correlation with similar but slightly metamorphosed or unmetamorphosed units farther south. K/Ar phengite ages show a predominant range of 80 to 115 Ma, while the less reliable sodic amphibole K/Ar ages are around 125 Ma (Çoğulu & Krummenacher 1967, Kulaksız & Phillips 1985).

ALANYA BLUESCHISTS The Alanya metamorphic complex, located in the Taurides north of Cyprus (Figure 4), illustrates many of the key features of the Alpide blueschist units, such as the passive continental margin–type protoliths, juxtaposition of metamorphic units with differing metamorphic grades, and complex metamorphic histories. The Alanya metamorphic complex comprises three superimposed flat-lying nappes of platform-type shallow-water lithologies that tectonically overlie the largely unmetamorphosed Mesozoic continental margin sediments of the Antalya complex (Figures 3 and 4; Okay & Özgül 1984). Of the three nappes, only the intermediate Sugözü nappe has undergone a mid-Cretaceous HP/LT metamorphism. The Sugözü nappe is only about 600 m thick but extends for over 40 km (Figure 4); it consists predominantly of garnet-micaschists, with rare bands and lenses of eclogites that show variable effects of a Barrovian-style greenschist facies overprint (Okay 1989). Unlike the HP/LT metamorphism, this late Barrovian metamorphism and associated deformation have affected all three nappes and produced a single dominant schistosity such that the nappes can only be distinguished on the basis of the metamorphic mineral assemblages; rocks overlying and underlying the garnet-micaschists of the Sugözü nappe lie within the chlorite zone of the Barrovian metamorphism and do not contain garnet (Okay 1989). The origin of the Alanya blueschists is problematic; they are either transported southward from the major Neo-Tethyan suture or are related to the destruction of a small oceanic basin south of the Apulian platform (Figure 4).

CYCLADIC BLUESCHIST BELT A discontinuous outcrop of blueschists extending for 500 km in a NNW-SSE direction from Mount Olympus in mainland Greece to the Cycladic islands west of the Menderes massif have been collectively called the Cycladic blueschist belt (Figure 4; Blake et al 1981). Information on the tectonic position of the blueschists comes mainly from mainland Greece, where blueschists are exposed in several tectonic windows extending from Olympus to southern Eubea (Figure 4). The structural sequence as exposed in these tectonic windows consists of three major units: a carbonate platform unit of Triassic to early/middle Eocene age (Olympus unit) at the base, a passive continental margin–type sequence (Ossa unit) in the middle, and a thick composite tectonic sheet (Pelagonian zone) at the top (Dürr et al 1978).

The Olympus unit, as exposed in the Olympus-Ossa regions and southern Eubea, is made up of over 2-km-thick carbonates of mid-Triassic to Eocene age with a major unconformity comprising the whole of the Jurassic, overlain by an Eocene flysch (Godfriaux 1970, Schmitt 1983, Katsikatsos 1970, Dubois & Bignot 1979). The Olympus unit shows a stratigraphy somewhat similar to that of the Menderes massif in western Turkey (Figure 4) and is correlated in Greece with the external carbonate platform unit, the Gavrova-Tripolitza zone. The Olympus unit is generally described as having undergone low-grade greenschist facies metamorphism (Godfriaux 1970, Blake et al 1981, Jacobshagen 1986). However, lawsonite, along with chlorite, quartz, and albite, is described from the Eocene flysch of the Olympus unit in the Olympus area (Schmitt 1983), which suggests that the Olympus unit has undergone a low-grade blueschist facies metamorphism, probably along with the overlying Ossa unit. This makes more sense tectonically, considering that the age of the HP/LT metamorphism in the Cycladic blueschist belt is generally regarded as mid-Eocene.

The nummulite-bearing Eocene carbonates and the overlying synorogenic flysch in the Olympus unit (Godfriaux 1970, Dubois & Bignot 1979) indicate that the thrusting of the Ossa unit over the Olympus Unit occurred during the Eocene. This is corroborated through the isotopic dating of the mylonites in the thrust zone in the Olympus region (Barton 1976). The Ossa unit, generally regarded as the only Cycladic blueschist unit in mainland Greece, is exposed in the tectonic windows of Olympus, Ossa, Pelion, and southern Eubea (Figure 4), where it forms several-kilometers-thick passive continental margin–type sequences of limestone, shale, quartzite with minor basic volcanic rock, keratophyre, and ophiolitic olistostrome. Owing to the intensive deformation and lack of fossils, a reliable stratigraphy cannot be constructed; however, the blueschist sequence most probably records stratigraphically the development of a carbonate platform–type passive continental margin and its eventual demise through ophiolite obduction. Blueschists have been given different names in different tectonic windows [Ossa or Ambelakia units in Olympus and Ossa (Godfriaux 1970, Katsikatsos et al 1982, Schmitt 1983), the Makrinitsa unit in Pelion (Ferriere 1982), and the Styra-Ochi unit in southern Eubea (Guernet 1978, Bavay & Romain-Bavay 1980, Dürr 1986)], although they most probably form a single continuous tectonic sheet between Olympus and southern Eubea.

Blueschist metamorphism is difficult to document in large areas in mainland Greece owing to the scarcity of metabasic rocks. However, sodic amphibole and lawsonite are described from metabasic rocks in the Olympus-Ossa (Schmitt 1983), Pelion, (Ferriere 1982) and southern Eubea (Bavay & Romain-Bavay 1980, Maluski et al 1981, Dürr 1986) regions. In the Olympus-Ossa region, the blueschist metamorphism of the Ossa unit is of incipient type; in the metabasites, the igneous texture and mineralogy are largely retained, and sodic amphibole and lawsonite occur sporadically (Derycke et al 1974, Schmitt 1983). The interfolded tectonic contacts between the Olympus and Ossa units in the Olympus-Ossa region (Schmitt 1983) and the similarity in the grade and type of metamorphism in both units suggest that the Olympus and Ossa units underwent jointly a syn- to posttectonic HP/LT metamorphism.

The Eocene limestone and flysch in the Olympus unit give a lower age limit for the HP/LT metamorphism. An Eocene age of blueschist metamorphism in the Ossa unit in mainland Greece is also indicated by $^{39}Ar/^{40}Ar$ dating from southern Eubea. Maluski et al (1981) obtained homogeneous $^{39}Ar/^{40}Ar$ age spectra from phengites, ranging from 33 to 50 Ma (mid-Eocene–early Oligocene). The pre-Maastrichtian ages given by the highly heterogeneous age spectra of sodic amphibole (Maluski et al 1981) are probably due to excess argon, a common phenomenon in this mineral (cf. Altherr et al 1979).

Blueschists are well exposed in the Cycladic islands, which include spectacular high-grade blueschist outcrops in Sifnos and Syros, the type locality of glaucophane (Hausmann 1845). However, because of the general lack of fossils, strong deformation, and disrupted outcrop pattern, the tectonic position of the blueschists in the Cyclades is less clear than that of the mainland blueschists. No Olympus-type carbonate platform is recognized in the Cyclades, probably because the Olympus carbonate platform is strongly deformed and metamorphosed along with the overlying Ossa unit so that, owing to the general lack of fossils, these two tectonic units cannot be distinguished (cf. Papanikolaou 1987). The various units of unmetamorphosed ophiolite and sedimentary rocks, long thought to be small thrust slices overlying the Greek blueschists, are now interpreted as downfaulted blocks along major listric normal faults (Ridley 1984b, Lister et al 1984); thus a considerable rock succession originally overlying the blueschists is now missing.

The blueschist sequence in the Cyclades consists of a pre-Alpide metamorphic basement of augen-gneiss and garnet-micaschist exposed on Ios and Naxos (Jansen & Schuiling 1976, van der Maar & Jansen 1983), overlain by a thick variegated sequence of limestone, dolomite, shale, quartzite, basic volcanic rock, and ophiolitic olistostrome. The schist-gneiss basement, which shows evidence of a major pre-Permian metamorphism (Henjes-Kunst & Kreuzer 1982), may be correlated with the Precambrian metamorphic core of the Menderes massif (Figure 4; Şengör et al 1984a) and probably constitutes the pre-Alpide metamorphic basement of the Olympus platform. The overlying sediment-dominated cover probably represents the Olympus and Ossa units.

In the Cyclades, both the pre-Alpide metamorphic basement and the overlying rocks have undergone a high-grade eclogitic blueschist facies metamorphism variably overprinted by a later, lower P/T metamorphism. There are several good petrological descriptions of the blueschists on the Cycladic islands (e.g. for Sifnos, see Okrusch et al 1978, Schliestedt 1986, Schliestedt & Matthews 1987; for Syros, see Ridley 1984a; for Ios, see van der Maar & Jansen 1983, Henjes-Kunst 1980). In areas where the high-temperature overprint is weak or absent (e.g. on Syros or Sifnos), the metabasic rocks comprise the eclogitic paragenesis of garnet+omphacite+glaucophane+zoisite (Ridley 1984a, Schliestedt 1986). Jadeite+quartz paragenesis occurs in the metakeratophyres on Sifnos (Okrusch et al 1978) and on Syros (Ridley 1984a).

K/Ar, Rb/Sr, and $^{39}Ar/^{40}Ar$ white mica ages from the well-preserved blueschists of Sifnos (Altherr et al 1979) and Syros (Maluski et al 1987) range between 35 and 50 Ma and indicate an Eocene HP/LT metamorphism, an age the same as that on mainland Greece. The Eocene blueschists were variably overprinted by a late Oligocene–early Miocene (20–25 Ma) lower P/T metamorphism (Andriessen et al 1979, Altherr et al 1979), which locally resulted in partial melting (for example, on Naxos; Jansen & Schuiling 1976).

THE EXTERNAL BLUESCHIST BELT OF THE HELLENIDES This blueschist belt is of interest in terms of its very recent HP/LT metamorphism (late Oligocene–early Miocene) and its tectonic setting (immediately north of the active Hellenic trench). It forms an arcuate belt, roughly parallel with the Hellenic trench, extending from the Peloponnesus to Crete (Figure 4).

At the base of the nappe pile that makes up the external Hellenides is a carbonate-rich Late Permian–Oligocene sequence (Plattenkalk unit) distinctive for its pelagic cherty limestones of Jurassic to lowermost Oligocene age (Bonneau 1984) overlain by a syntectonic Oligocene flysch. The Plattenkalk unit, correlated with the Ionian zone, is overthrust by the Phyllite-Quartzite unit made up of pre-Alpide metamorphic rocks and Permo-Triassic shale, sandstone, conglomerate, carbonate, basalt, and evaporite with gypsum. The Phyllite-Quartzite unit is tectonically overlain by the Gavrovo-Tripolitza and Pindos units and by a composite nappe of ophiolite and sedimentary and metamorphic rocks. Out of the five units that make up the nappe pile of the external Hellenides, only the two lowermost units, the Plattenkalk and the Phyllite-Quartzite units, have undergone low-grade blueschist facies metamorphism. Sodic amphibole and lawsonite are described from the Phyllite-Quartzite unit in Crete and in the Peloponnesus (Seidel et al 1982, Thiebault & Triboulet 1984). The metamorphism of the Plattenkalk unit is more difficult to document because of the absence of metabasic rocks in the sequence, and some workers ascribe only a greenschist facies metamorphism to the Plattenkalk unit (e.g. Hall et al 1984). However, the presence of magnesiocarpholite along with diaspor and pyrophyllite in the metabauxites from the Plattenkalk unit (Seidel et al 1982) suggests that this unit has also undergone HP/LT metamorphism.

The late Oligocene–early Miocene age of the HP/LT metamorphism in the external Hellenides is based on the presence of Oligocene flysch in the Plattenkalk unit and on K/Ar dating in the Phyllite-Quartzite unit, which predominantly yields early Miocene ages (Seidel et al 1982).

Oman-Makran

Regionally important blueschists are not reported along the 2800-km-long Alpide belt between western Turkey and Makran. This is partly a result of the comparative scarcity of petrological data from this region; however, it is most likely that extensive blueschist terranes, such as those that occur in the Apulian area, are not present in this major orogenic segment. The minor blueschist occurrences of central and eastern Turkey are reviewed by Okay (1986). In Iran, blueschists are not described along the 1600-km-long Main Zagros thrust, which forms the main Neo-Tethyan suture between Gondwana-Land and Laurasia (Figure 1). The continent-continent collision in this segment was of Miocene age (Dewey et al 1973), compared with the Eocene and older collisions in the Apulian area, and thus the equivalents of the Penninic and Helvetic units have not yet formed. This explains the absence of collision-related metamorphism south of the Main Zagros thrust. However, ophiolites were obducted during the Late Cretaceous southward over the Arabian platform (Ricou 1971, Berberian & King 1981). Regional blueschist metamorphism, if it exists, must be sought in the ophiolitic mélanges and in the continental margin sediments underlying the ophiolites. In central Iran there are several narrow ophiolitic mélange zones that are reported to include blueschists (Berberian & King 1981). These blueschists form small fault slivers associated with Upper Cretaceous pelagic limestone and with dismembered ophiolite and turbidite, and they represent exotic tectonic blocks within accretionary complexes (Sabzehei 1974, Tirrul et al 1983).

East of the Hormuz straits the Main Zagros thrust is transformed through a major transform fault into a northward-dipping active subduction zone where the ocean floor of the Gulf of Oman is being subducted beneath the southern margin of Asia (Figure 5; Farhoudi & Karig 1977). This Oman-Makran region, where continental collision has not yet occurred, forms a typical example of a precollisional orogenic belt. To the north of the subduction zone there is the several-hundred-kilometers-wide Makran accretionary complex consisting mainly of Late Cretaceous to Holocene trench-fill turbidites, with intercalated ophiolite and pelagic sediments in the north representing accreted oceanic crustal rocks (Farhoudi & Karig 1977).

On the south side of the Gulf of Oman a major mid-Cretaceous ophiolite (Semail ophiolite) was obducted southward during the Late Cretaceous

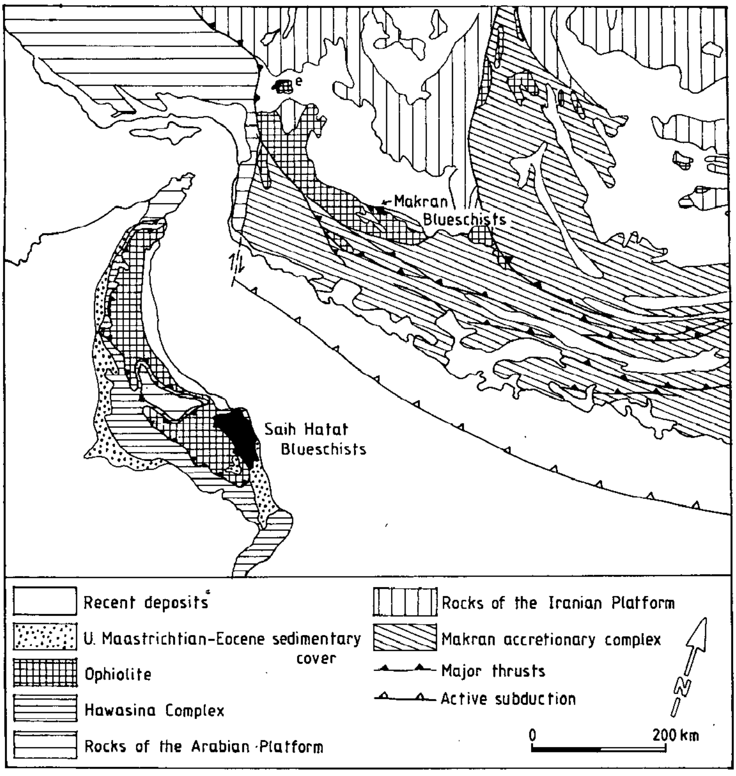

Figure 5 Tectonic map of the Oman-Makran region. Areas in black show blueschist outcrops.

over the Arabian passive continental margin and shelf sediments (Glennie et al 1974, Lippard et al 1986). The present structural succession in the Oman region consists of autochthonous sedimentary rocks reaching up to Campanian in age tectonically overlain by the Hawasina complex of deformed Mesozoic passive continental margin sediments that form complexly imbricated nappe stacks. The Hawasina complex is overthrust by a heterogeneous series of imbricated Mesozoic sedimentary, volcanic, and metamorphic rocks (Haybi complex), which is overlain by the Semail ophiolite. All these tectonic units are unconformably covered by the Maastrichtian limestones (Figure 5).

BLUESCHISTS OF THE OMAN-MAKRAN REGION Rocks of the Makran accretionary complex are generally regarded as not metamorphosed; however, recently, incipiently metamorphosed HP/LT metavolcanic and metasedimentary rocks have been described from a small region of 15 km^2 in the northern part of the Makran (Figure 5; McCall 1985). These rocks form a strongly tectonized sequence of basic volcanic rock, turbidite, and Late Cretaceous pelagic limestone. Basic volcanic rocks generally retain their igneous textures and minerals; sodic amphibole (crossite and glaucophane) has formed around the magmatic augites or in the matrix and is accompanied by chlorite, albite, pumpellyite, and calcite (McCall 1985). The sedimentary rocks are more fully recrystallized and commonly exhibit a foliation; they consist of quartz, feldspar, calcite, muscovite, chlorite, pumpellyite, sodic amphibole, and sphene. These Makran blueschists are similar to the incipiently metamorphosed rocks from the Tavşanlı zone (Okay 1982), although lawsonite, aragonite, and sodic pyroxene are not reported from Makran. However, detailed petrological studies in the northern Makran will most probably reveal more such incipiently metamorphosed blueschists. The age of the blueschist metamorphism in the northern Makran is probably post-Late Cretaceous, as indicated by the presence of partially recrystallized globotruncana-bearing limestones in the incipiently metamorphosed sequence (McCall 1985). Makran blueschists represent material accreted to the underside of the accretionary prism and subsequently uplifted by isostatic rebound and erosion (cf. Platt et al 1985). Blueschist metamorphism is probably going on below the northern Makran, where the accretionary complex reaches a thickness of about 20 km through continuous underplating.

Blueschists have also been recently discovered in northeast Oman in the Saih Hatat tectonic window (Michard 1983), which exposes an extensive area of over 1200 km^2 of low- to medium-grade metamorphic rocks underneath the Semail ophiolite (Figure 5). These rocks, initially thought to be a pre-Permian sequence (Glennie et al 1974), are now interpreted as either metamorphosed equivalents of the Hawasina complex or metamorphosed authochthon (Michard 1983, El-Shazly et al 1989). The Saih Hatat sequence of pre-Permian to Late Cretaceous age consists of quartzites, shales, dolomites, and limestones with minor basic volcanic rocks. Although the entire sequence in the Saih Hatat tectonic window is apparently affected by the HP/LT metamorphism and associated penetrative deformation (El-Shazly et al 1989), evidence for this event is restricted to the minor isolated basic volcanic rocks and Na-Fe-rich pelitic bands. Petrography of these diagnostic rock types indicates that the blueschist metamorphism was not uniform throughout the area and was variably overprinted by a lower P/T metamorphism. The blueschist metamorphic grade shows a broad decrease upward in the sequence and toward the southwest. The highest grade blueschists, with the eclogitic mineral assemblage of glaucophane + epidote + phengite ± paragonite ± garnet ± Na-pyroxene in the metabasic rocks, are exposed in the As Sifah area to the east. Metabasic rocks from other areas do not contain garnet and are generally characterized by the sodic amphibole + epidote paragenesis with abundant albite and chlorite. Ca-rich metapelitic rocks in the top part of the sequence contain lawsonite + chlorite + quartz ± carpholite (Le Mer et al 1987, El-Shazly et al 1989). The Oman blueschists are variably overprinted by a lower P/T metamorphism, as shown by the common barroisite rims around sodic amphibole and by the formation of biotite and albite.

The transgressive Maastrichtian limestones over the Saih Hatat metamorphic rocks give an upper age limit for the HP/LT metamorphism. K/Ar ages of 100 ± 4 and 80 ± 4 Ma from phengites from the As Sifah blueschists (Lippard 1983) indicate that the blueschist metamorphism was broadly contemporaneous with the obduction of the Semail ophiolite. The Oman blueschists formed as a result of the subduction of the Arabian continental margin and shelf, concomitant with the obduction of the Semail ophiolite generated in the back-arc basin above the subduction zone (Lippard et al 1986). This situation is quite analogous to the Tavşanlı zone blueschists (Okay 1984).

Himalaya

East of Makran we come into the realm of the Himalaya, produced through the collision of the Indian plate with Laurasia (Le Fort 1975). The Indian subcontinent, which was originally part of Gondwana-Land, split off from it during the Jurassic, moved rapidly across the Tethys (creating the Indian Ocean behind it), and eventually collided with the active Andean-type Laurasian margin during the Eocene. The 2250-km-long east-west-orientated Indus-Zangpo suture, generally marked by a steeply dipping fault with dismembered ophiolite lenses, separates the Indian subcontinent from Laurasia. To the north of the Indus-Zangpo suture is the Transhimalaya, a Cretaceous to Paleocene Andean-type island arc at the southern margin of Laurasia. Rocks of the Indian subcontinent to the south of the Indus-Zangpo suture occur in three major thrust sheets: the Higher Himalaya, the Lower Himalaya, and the Sub-Himalaya. The Higher Himalaya consists of up to 10-km-thick Cambrian to mid-Eocene sediments (the Tethyan zone) underlain by the Proterozoic crystalline basement of the Indian shield (the Central Crystallines). The Mesozoic-Cenozoic sediments of the Tethyan zone were deposited on a continental shelf on the northern border of the Indian platform. In the Higher Himalaya there are also a number of allochthonous sheets of ophiolite and ophiolitic mélange thrust in the early Eocene southward from the Indus-Zangpo suture over the Tethyan sediments. The post-early Miocene Main Central thrust juxtaposes the Central Crystallines of the Higher Himalaya over the Phanerozoic sediments of the Lower Himalaya. Apart from the thick Paleozoic and thin Mesozoic sediments, the Lower Himalaya comprises major klippen of gneiss and granite derived from the Central Crystallines. The Lower Himalaya is thrust southward along the Main Boundary thrust over the Sub-Himalaya; the latter consists of thick, molassic late Tertiary sediments.

Both the Higher and Lower Himalaya are extensively affected by a Tertiary Barrovian and lower P/T–type metamorphism that shows a general decrease upward and downward from the Main Central thrust, which produces an inverse metamorphic zonation in the Lower Himalaya (Windley 1983, Pecher & Le Fort 1986).

HIMALAYAN BLUESCHISTS The presently known blueschists in the Himalaya make up only a minute fraction of the exposed Himalayan metamorphic rocks, and as yet no real eclogites are reported from this major orogenic belt. All the known blueschists are located along the Indus-Zangpo suture in the western part of the Himalaya on both sides of the Nanga Parbat syntaxis (Figure 1; Virdi et al 1977, Shams et al 1980, Jan 1987).

Blueschists in the Ladakh area east of the Nanga Parbat syntaxis occur as up to a few-hundred-meters-thick, several-kilometers-long tectonic slices imbricated with other slices of ultramafic rock, gabbro, dolerite, pillow lava, and continental slope sediments (Honegger et al 1988). This narrow tectonic zone constitutes the Indus-Zangpo suture in the Ladakh area and separates rocks of the Indian subcontinent (Higher Himalaya) from those of Laurasia.

The protoliths of the blueschists are basic volcanic rocks and pyroclastics with minor intercalated radiolarian cherts and pelagic shales. The blueschist metamorphism has generally been of incipient character, although the degree of recrystallization varies from one tectonic slice to another (Honegger et al 1989). The basic volcanic rocks generally retain their primary textures and minerals. The igneous augites are partially replaced by sodic pyroxene, and hornblendes by sodic amphibole, and lawsonite has pseudomorphed the primary plagioclase (Honegger et al 1989). In comparison with the metabasic rocks, metasediments are fully recrystallized and metacherts contain the mineral assemblage of sodic amphibole + lawsonite + phengite + quartz + albite ± spessartine garnet ± sphene ± chlorite ± opaque (Honegger et al 1989).

There are no data on the depositional ages of the Himalayan blueschist protoliths. K/Ar and $^{39}Ar/^{40}Ar$ sodic amphibole, phengite, and whole-rock ages on the Nanga Parbat blueschists range from 70 to 100 Ma (Maluski & Matte 1984, Desio & Shams 1980, Honegger et al 1989).

These ages are considerably older than the Eocene continental collision in the Himalaya.

CONCLUSIONS

Blueschists are common rocks in the Mediterranean Alpides. In fact, as illustrated in Figures 2 and 4, they constitute areally over half of the rocks metamorphosed during the Alpide orogeny in the Apulian area. This is a peculiar feature of the Alpide orogeny, as compared with the Hercynian and other collisional orogenies where the regional metamorphism is generally of lower P/T type. This phenomenon, first stated by de Roever (1956), was ascribed to a decrease in geothermal gradient with time. However, the scarcity of blueschists in the Tertiary Himalayan orogen, where lower P/T metamorphic rocks are abundant (cf. Pecher & Le Fort 1986), indicates that other factors besides a possible decrease in geothermal gradient with time are important. One local factor was probably the presence of an isolated continental block, the Apulian platform, between two converging megacontinents, Laurasia and Gondwana-Land. The relatively narrow width of the Apulian platform and the presence of elongate regions of thinned continental crust (Figure 3) resulted in major continental subduction in the western Alpide orogeny compared with the Himalaya, where the larger and more rigid Indian subcontinent resisted subduction. Another factor is the presence of intraoceanic subduction zones in the central part of the Tethys; apparently, continental crust is more easily subducted in the intraoceanic arcs than in the Andean-type arcs.

Alpine/Himalayan blueschist lithologies range from evaporitic deposits and limestones to ophiolites reflecting various paleogeographic settings from shallow carbonate platform through passive and active continental margin to oceanic crust. Circum-Pacific-type blueschists, which represent metamorphosed accretionary complexes and minor accreted oceanic crust, are volumetrically insignificant and occur only in the eastern part of the Alpine/Himalayan orogen in Makran and the Himalaya. In its western part the Tethys was narrow, the volcanic arcs were poorly developed, and the trenches were bordered by extensive carbonate platforms that restricted the influx of clastic material (Figure 3). All of this explains the general absence of accretionary complex–type blueschists in the Apulian region. However, in its eastern part the Tethys became increasingly wider and over 8000 km of Tethyan oceanic crust is believed to have been subducted under the Transhimalaya; the absence of major accretionary complexes and related blueschists in this part of the Alpine/Himalayan orogen is probably due to their destruction during the continental collision and/or due to the compressional nature of the arc in this segment of the Tethys, which thus prevented the growth of the accretionary complexes (cf. Dewey 1980).

In the western part of the Alpine/Himalayan orogen, some carbonate platform and passive continental margin deposits have undergone blueschist metamorphism prior to the continental collision. In the Tavşanlı zone and in Oman, the HP/LT metamorphism is related to the Cretaceous ophiolite obduction; in both cases the passive continental margin was subducted and metamorphosed in a midoceanic subduction zone (Okay 1984, Lippard et al 1986) and a new subduction zone was initiated to the north of the clogged subduction zone to take up the plate convergence. The age similarity between ophiolite obduction and blueschist metamorphism, and the geochemistry of most of the Tethyan ophiolites, which indicates an above-subduction-zone setting (Pearce et al 1984), support such a model of ophiolite obduction and blueschist metamorphism. In the Taurides in Turkey and in the external parts of Oman, the ophiolites rest over the unmetamorphosed Mesozoic platform carbonates, which when traced toward the suture should show an increasing degree of blueschist metamorphism. The uplift of blueschists in the Tavşanlı zone and in Oman was achieved probably through normal faulting along the blueschist/ophiolite interface (cf. Platt 1986), so that at present essentially unmetamorphosed ophiolite overlies directly the blueschists in both regions. On the other hand, HP/LT metamorphism in the Cyclades and in the External Hellenide belt in Greece is related to progressive continental collision. This is indicated by the syn- to postcontinental collision ages of the blueschists in this segment and by the thick continental units overlying the blueschists.

In the Alpine/Himalayan realm, and probably also elsewhere, the initial HP/LT metamorphism seems to have occurred under zero strain conditions leading to static recrystallization and generally not involving progressive metamorphism. For example, some of the eclogitic rocks from the Sesia-Lanzo zone in the Western Alps (Koons et al 1987) or from the Betic zone (Gomez-Pugnaire & Fernandez-Soler 1987) retain the igneous textures of their protoliths but nevertheless show no mineral assemblages of the intermediate blueschist facies through which they must have passed. This static recrystallization stage is usually followed by a high-strain penetrative deformation that affects the mineral assemblage and mineral composition by overcoming nucleation difficulties and enlarging the equilibrium domain (Okay 1980b, 1986, Koons et al 1987). For example, in the Tavşanlı zone in northwest Turkey, sodic pyroxene has formed in the metabasites, probably metastably, during the static recrystallization stage and has pseudomorphed the preexisting augite, whereas lawsonite pseudomorphed the plagioclase. The onset of penetrative deformation eased the nucleation problems of sodic amphibole, which started forming at the expense of sodic pyroxene and chlorite (Okay 1982). In the Sesia-Lanzo zone pods of granitoids about 100 m in width have escaped the penetrative deformation and retain their primary igneous textures, with plagioclase pseudomorphed by jadeite + quartz + zoisite and biotite partially pseudomorphed by garnet + white mica (Compagnoni & Maffeo 1973). These metagranitoids pass laterally to the penetratively deformed orthogneisses, with a similar bulk-rock composition but one characterized by the presence of paragonite and omphacite rather than jadeite. These changes in the mineral assemblage and composition are attributed to the progressive enlargement of the equilibrium domain during the deformation, rather than to changes in pressure and temperature (Koons et al 1987). These indicate that the initial burial of the blueschists occurs rapidly and with no or little internal strain. Penetrative deformation starts at or near the maximum depth after the recrystallization has been completed and is probably associated with the uplift mechanism. A similar conclusion was reached by Platt & Lister (1985) after a detailed structural analysis of the blueschists in the French Western Alps.

A very common phenomenon observed in the Alpide blueschists is the greenschist facies overprinting of the blueschist and eclogite mineral assemblages. Recent work (Schliestedt & Matthews 1987, Okay 1989) has shown that fluid infiltration during the uplift, rather than an increase in temperature, is generally responsible for the observed greenschist facies overprint. For example, in the Alanya blueschists the temperature of the greenschist facies overprint was demonstrably less than that of the earlier HP/LT metamorphism (Okay 1989). The common occurrence of pristine and retrograded eclogites in close proximity within a single metamorphic terrain can also be explained through metamorphism caused by fluid infiltration. Considering that the uplift path of most ecologites and blueschists passes through the greenschist facies field, the main factor in the overprinting of the HP/LT mineral assemblages would be the access of fluids during the uplift.

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