Neotethyan ophiolites: formation and obduction within the life cycle of the host basins

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Abstract: To understand the Neotethyan ophiolites better, their place in the history of the host basins is explored, using the Jurassic Hellenic–Dinaric and some Cretaceous (mainly peri-Arabian) ophiolites as examples. These formed in mature (*c.* 60 Ma and 100 Ma old) seaways by spreading at rates that apparently were too high to persist for more than a fraction of the basin history. Each ophiolite group formed in a short time interval, about 10 Ma, soon after changes in the motions of plates in the host basins. Therefore, these ophiolites do not seem to have formed by normal long-term spreading along mid-ocean ridges, but their formation signifies special 'ophiolite events'. This fits well the widely accepted origin in a supra-subduction zone (SSZ) setting that was inferred from geochemical data. The ophiolites studied here are thus interpreted as having formed in new subduction zones that originated during changes in plate motions. The spreading during their accretion was driven by fast retreat (roll-back) of the subducting slabs. The western ophiolites of the Hellenic–Dinaric belt, dominated by mid-ocean ridge basalt (MORB)-like rocks but invaded by SSZ magmas, could have formed along ridges just before they failed (collapsed), but the age data fit better formation in a proto-back-arc setting alongside the more eastern ophiolites with the typical SSZ signature. The construction of the ophiolites examined here ended when they were detached from their substrate and pushed over the adjacent basins. At that stage they were underplated by metamorphic soles, but how the latter were emplaced still needs clarification. Continuing retreat of the subducting slabs consumed the host basins and pushed the ophiolites hundreds of kilometres until they were obducted over the nearby margins 15–20 Ma after formation. This framework seems to apply to many ophiolites and allows us to interpret them in terms of known processes, but also highlights problematic issues that still need to be resolved.

Ophiolites are a conspicuous though minor component of Phanerozoic and Neoproterozoic orogens (Dilek 2003*a*). Following Hess (1965) and Gass (1968) they are recognized as fragments of oceanic crust and uppermost mantle. Being important indicators of the past operation of the Wilson cycle and the existence of former oceanic basins between components of orogens, ophiolites were extensively studied. Early workers favoured an origin along mid-ocean ridges (MORs) (e.g. Gass 1968; Moores & Vine 1971; Coleman 1977), but later geochemical studies pointed at a supra-subduction zone (SSZ) origin of many ophiolites, probably in a young pre-arc (nascent forearc) setting similar to what is found in the forearcs in the western Pacific (Casey & Dewey 1984; Hawkins *et al.* 1984; Leich 1984; Pearce *et al.* 1984; Stern & Bloomer, 1992). Moores (1982) recognized the diversity of ophiolites and distinguished between 'Tethyan (Mediterranean)' ophiolites that are obducted (thrust) over passive continental margins and 'Cordilleran (Pacific)' ophiolites that are incorporated into accretionary complexes, but Shervais (2001) stressed that the magmatic history of ophiolites of both types followed similar evolutionary trends, although they differ in other respects (Beccaluva *et al.* 2004). Other ophiolite types were also recognized (Dilek 2003*b*). Studies of Mediterranean and Middle East ophiolites, especially the Troodos (Cyprus) and Semail (Oman) ophiolites, contributed much to these ideas (e.g. review by Robertson 2002). Although ophiolites were studied from various points of view, many questions regarding their origin and emplacement are still debated (e.g. Shervais 2001; Flower & Dilek 2003; Robertson 2004, and references therein).

One way to advance the understanding of ophiolites is to examine how their history relates to the evolution of the host basins. The present work follows this approach, but in view of the ophiolite diversity the discussion is limited to a few groups of Tethyan ophiolites: Jurassic Hellenic–Dinaric ophiolites and some Cretaceous ophiolites (Fig. 1), whereas other ophiolites in that region that were emplaced along active margins (Robertson 2002, 2004) are not considered. The present work builds on previous discussions, but stresses plate kinematic considerations and features of subduction zones. Below, some

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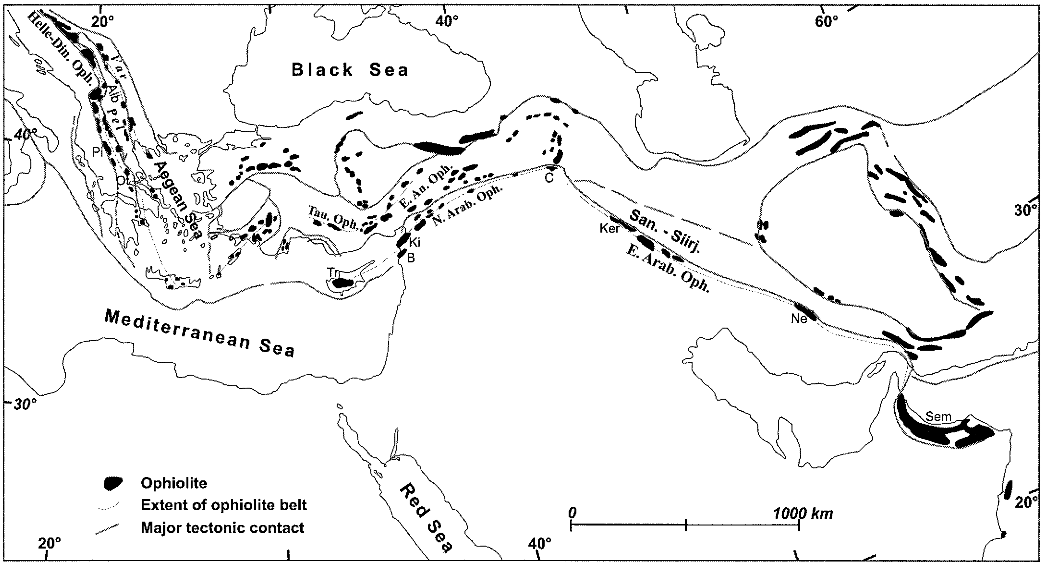

Fig. 1. Distribution of ophiolites in the studied region. Alb., Albanian ophiolites; B, Baer–Bassit ophiolite; C, Cilo ophiolite; Ker., Kermanshah ophiolite; Kızıldağ ophiolite; Ne, Neyriz ophiolite; Pel., Pelagonian–Korab block; Pi, Pindos ophiolite; Ot, Othris ophiolite; San. -Sirj., Sanandaj–Sirjan block; Sem., Semail ophiolite; Tr., Troodos ophiolite; V, Vourinos ophiolite; Var., Vardar zone.

general considerations regarding ophiolite origin are first briefly presented, and then features of the ophiolites examined here that are relevant for the subsequent discussion are summarized. Based on these data the relations between these ophiolites and the histories of the host basins are discussed, paying attention to problems that need further study. Here stratigraphic ages deduced from fossils and from radiometric dating are compared using the time scale of Gradstein *et al.* (2004).

Neotethyan examples

General setting and features

The ophiolites considered here formed in Mesozoic seaways or basins that existed in the realm between Eurasia and Gondwanaland. The history of this realm, whose closure gave rise to the Alpine orogenic chain, has been interpreted in somewhat different ways (e.g. LePichon *et al.* 1988; Robertson *et al.* 1991, 1996; Stampfli *et al.* 2001), but all interpretations agree that early in the Mesozoic several blocks (micro-continents) rifted away from the major continents, especially from Gondwanaland, and that their subsequent drifting led to the growth of new (Neotethyan) seaways. The ophiolites considered here originated in these seaways and were obducted on their passive margins. Now these ophiolites occur in belts that may be 1000–2000 km long that mark the sutures between blocks that bordered these basins.

The ophiolites examined here are of two major types (Boudier & Nicolas 1985; Nicolas 1989). Most of them have a mantle part that consists largely of harzburgitic tectonites and an overlying crustal part that comprises the complete classic Penrose pseudo-stratigraphy (Anonymous 1972), i.e. (1) a plutonic unit of ultramafic and mafic rocks, often with some high-level plagiogranites, (2) a sheeted dyke complex, and (3) an extrusive unit, often overlain by (or interbedded with) abyssal sediments. The total thickness, where all units are preserved, may reach 12 km or more. Less common are ophiolites whose mantle part consists mainly of lherozlitic tectonites, and the crustal part comprises plutonic and volcanic units, whereas a sheeted dyke complex is absent or poorly developed; their total thickness does not exceed 3–4 km.

The ophiolites are associated with allochthonous units that originated in the same basins and provide crucial information about these basins (e.g. Moores 1982; Robertson 2004, and references therein). They include: (1) strongly deformed metamorphic rocks, with inverted metamorphic gradients, that form soles up to several hundred metres thick at the base of the ophiolite bodies; their protoliths consist of basinal sediments and volcanic rocks that differ from the volcanic rocks in the ophiolites; (2) accretion–subduction complexes (mélanges) that formed during ophiolite obduction; they consist of imbricated slices of pre-ophiolite rocks that were scraped off the floors of the host basins and the passive margins on which the ophiolites were obducted, but mélanges derived from the ophiolites themselves are often also present, and serpentinite bodies are sometimes also found.

Ophiolites are generally accepted to represent new crust that formed by some type of sea-floor spreading, i.e. along zones of plate separation. Formation in an intra-oceanic setting, some distance from any continental margin, is also accepted, based on the nature of the overlying sediments (Robertson 2004). The metamorphic soles formed when their protoliths were overthrust by hot material, generally thought to be the ophiolites themselves (Spray 1984; Woodcock & Robertson 1977; but see below).

On the other hand, it is still debated whether ophiolites formed along MORs or in a SSZ setting. The first interpretation is appealing because the structure of ophiolites is readily explained as formed by sea-floor spreading. Nicolas (1989) interpreted the harzburgitic and lherzolitic ophiolites as having formed by fast and slow spreading, respectively. However, the geochemical signature of the igneous rocks of the harzburgitic ophiolites differs from mid-ocean ridge basalt (MORB) but resembles rocks found in SSZ settings: low-Ti basalts, island arc tholeiites (IAT), or very low-Ti depleted magmas with a boninitic affinity. The crystallization of plagioclase after pyroxenes and the occurrence of orthopyroxene in cumulates also differs from MORB. Thus formation of the harzburgitic ophiolites in an SSZ setting was advocated (e.g. Cameron *et al.* 1980; Laurent *et al.* 1980; Pearce *et al.* 1984) and is widely accepted, although typical island arcs or sediments derived therefrom are most often lacking. In contrast, the volcanic rocks in lherzolitic ophiolites are high-Ti MORB-like rocks that could have formed along MORs. However, in some ophiolites rocks of both types occur together (see below).

Ophiolites of the Hellenic–Dinaric orogen

An almost 1000 km long belt of Jurassic ophiolites extends through Greece, Albania and former Yugoslavia along the Hellenic–Dinaridic orogen (Fig. 1). They are thought to have formed in the Pindos Ocean, which existed between the Apulian and the Pelagonian (Korab) blocks (micro-continents) (Robertson *et al.* 1991; Doutsos *et al.* 1993; Smith 1993; Robertson & Shallo 2000; Robertson 2002). Another ophiolite belt with a different history, not treated here, extends east of the Pelagonian block and is considered to have formed in the distinct Vardar Ocean (Fig. 1; Robertson 2002).

The underlying allochthonous units (Smith *et al.* 1979; Jones & Robertson 1991; Jones *et al.* 1991; Bortolotti *et al.* 1996, 2004; Robertson & Shallo 2000; Saccani *et al.* 2003) record rifting and formation of the Pindos oceanic basin on the western side of the Pelagonian block. Deep-water sediments accumulated in the basin until the Oxfordian (Danelian & Robertson 2001). Triassic rifting is also indicated by widespread magmatism of that age on the Pelagonian block (Robertson *et al.* 1991; Pe Piper 1998; Robertson & Shallo 2000).

The ophiolites of the Hellenic–Dinaric belt were thrust eastward onto the Pelagonian (Korab) block (micro-continent) in Late Jurassic times (Smith 1993; Bortolotti *et al.* 1996; Rassios & Smith 2000; Robertson & Shallo 2000). However, the part of the Pindos basin west of them survived until the Eocene (Clift & Robertson 1989; Degnan & Robertson 1998) and it was only then that the area was deformed into a pile of SW-vergent thrust sheets. To represent this belt the better studied ophiolites in northern Greece and in Albania are considered in some detail.

The ophiolites of northern Greece. The Vourinos, Pindos, and Othrys ophiolites are well preserved (Fig. 1), whereas more southern ophiolites are disrupted. The Vourinos ophiolite comprises harzburgitic tectonites and displays the classic Penrose pseudo-stratigraphy, although the volcanic unit is reduced (eroded?) (Moores 1969; Smith 1993; Rassios & Smith 2000). The sheeted dykes and extrusive rocks consist mainly of low-Ti mafic rocks and minor andesites and dacites with an IAT affinity that are cut by dykes and small intrusions with a boninitic affinity (Beccaluva *et al.* 1984). The Pindos ophiolite comprises two thrust sheets (Capedri *et al.* 1980; Jones & Robertson 1991; Jones *et al.* 1991; Saccani & Photiades 2004; Saccani *et al.* 2004). The upper sheet consists only of cumulates and of harzburgitic tectonites. The latter generally occur in ophiolites with SSZ affinity, but the cumulates comprise also rocks that crystallized from MORB-like magmas. The lower sheet, although rather thin and disrupted, contains the entire pseudo-stratigraphy. The cumulates and some extrusive rocks have MOR affinities, whereas the volcanic rocks comprise alternating MORB, IAT and boninitic rocks, the last tending to be the younger ones. Dykes with boninitic affinity cross all units. Further south the Othris ophiolite preserves slices of harzburgites as well as lherzolitic tectonites (Smith *et al.* 1975, 1979; Saccani *et al.* 2004). Thus, these ophiolites display a range of rock types, some comprising only rocks with SSZ affinities whereas others include such rocks as well as rocks with MORB affinities, the latter being the older ones.

Dating of zircons by the U/Pb method (Liati *et al.* 2004) revealed that the Vourinos and Pindos ophiolites have the same ages within analytical uncertainty. A gabbro from the Pindos ophiolite yielded an age of 171 ± 3 Ma, whereas a gabbro and a plagiogranite from the Vourinos ophiolite yielded ages of 168.5 ± 2.4 Ma and 172.9 ± 3.1 Ma, respectively. Radiolarites overlying the Vourinos complex contain latest Bajocian–Early Bathonian (*c.* 168 Ma) and somewhat younger fossils (Chiari *et al.* 2003). Thus there is no clear age distinction between them.

The metamorphic soles beneath these ophiolites yielded $^{40}Ar/^{39}Ar$ ages (using updated decay constants; Spray *et al.* 1984) of 171 ± 4 Ma to 165 ± 3 Ma, close to the ages obtained by Liati *et al.* (2004). These ages and the occurrence of ophiolite-derived clasts in radiolarites of Early Bathonian to Early Callovian age (*c.* 166–162 Ma) in a mélange beneath the Pindos ophiolite (Jones *et al.* 1992) indicate tectonization shortly after ophiolite formation. Flexure in front of the approaching ophiolites caused the Pelagonian block to subside in Oxfordian–Kimmeridgian times (Danelian & Robertson 2001). Final uplifting and obduction are recorded by Tithonian–Beriassian (*c.* 148–142 Ma) sediments overlying the Vourinos ophiolite (Smith 1993).

Further south the ophiolites are disrupted but are similar, comprising rocks with both MOR and SSZ affinities, and have similar ages and similar emplacement histories (Robertson 2002; Liati *et al.* 2004; Saccani *et al.* 2004).

The Albanian ophiolites. These ophiolites are well preserved. They mark the northward prolongation of the Greek ophiolite belt and had a similar history (Fig. 1; Beccaluva *et al.* 1994; Shallo 1994; Bortolotti *et al.* 1996; Robertson & Shallo 2000; Saccani *et al.* 2004). The ophiolites along the eastern side of the belt comprise harzburgitic tectonites, are thick (≥ 8 km), and show a complete pseudo-stratigraphy, although in the south their upper parts were eroded. The sheeted dykes and volcanic rocks consist mainly of low-Ti basalts, basaltic andesites and dacites with an IAT character. Rocks with a boninitic affinity occur as dykes and as extrusive rocks in the higher parts of the volcanic units. The ophiolites along the western side of the belt are only up to 3–4 km thick, and comprise lherzolitic tectonites and plutonic and volcanic units that have variable thicknesses because of faulting; sheeted dykes are generally absent. The volcanic rocks consist of MORB-like high-Ti tholeiites. However, in many places the western type ophiolites contain rocks of both MOR and SSZ affinity; either these alternate in time or the SSZ-type rocks are the younger ones (Bébien *et al.* 1998, 2000; Bortolotti *et al.* 2002; Hoeck *et al.* 2002). These occurrences show that originally the two ophiolite types were not widely separated and that at some times the two magma types were available in the same places.

Dimo-Lahitte *et al.* (2001) obtained $^{40}Ar/^{39}Ar$ ages of 163.8 ± 1.8 Ma (Callovian) for a plagiogranite in the NE and 172.6 ± 1.7 Ma (close to the Aalenian–Bajocian transition) for a dyke in an ophiolite further south. Fossils in overlying cherts have Bathonian to Callovian or Oxfordian ages (Bortolotti *et al.* 1996). The metamorphic soles record P–T conditions that range from 800–860 °C, 0.9–1.2 GPa (equivalent 25–35 km depth) through 550–700 °C, 0.4–0.6 GPa (12–18 km) to greenschist-facies conditions (Carosi *et al.* 1996; Dimo-Lahitte *et al.* 2001). They yielded $^{40}Ar/^{39}Ar$ ages of 173–169 Ma (Aalenian–Bajocian) in the south and about 164 Ma (Callovian) in the north, with few intermediate ages between these. There is no difference between the ages on the eastern and western sides of the ophiolite belt (Dimo-Lahitte *et al.* 2001).

Ophiolites of all types are overlain by a Tithonian mélange and Barremian–Lower Valanginian turbidites, both consisting of components derived from the two ophiolite types, from the adjacent basin floor, and from the Korab block (notably quartzitic sandstones) (Bortolotti *et al.* 1996). Thus, at *c.* 150–140 Ma the ophiolites were already close to the Korab micro-continent, but still under the sea. These sediments occur also as slices in the sub-ophiolitic allochthons, which confirms that they formed during ophiolite obduction over the Korab block. After uplift the ophiolites were covered by shallow-water sediments from the Barremian. As further south, remnants of the Pindos Ocean survived west of the ophiolites until they were tectonized in the early Tertiary.

The Hellenic–Dinaric ophiolites extend northward into former Yugoslavia (Pamić *et al.* 2002; Fig. 1), but these have been less studied. They are generally strongly deformed. Here the tectonites are predominantly lherzolitic, and volcanic rocks transitional between IAT and MORB are present. They had a similar history to the more southern ophiolites.

Summary. The foregoing account shows that the Hellenic–Dinaric ophiolites formed in a narrow time interval between *c.* 170 Ma and *c.* 165 Ma within an oceanic basin that originated in the Triassic, i.e. this basin was 60–70 Ma old when the ophiolites formed. These ophiolites comprise bodies of the harzburgitic type with a complete classic pseudo-stratigraphy and magmas with a SSZ character, and also bodies of the lherzolitic type with an incomplete pseudo-statigraphy and magmas with a MOR character, but many of the latter contain also magmas with a SSZ character. Thus the two ophiolite types appear to have formed in close proximity in time and space. The ages of the metamorphic soles (173–164 Ma) are close to or overlap the ages of ophiolite formation. Obduction over the Pelagonian–Korab block continued 15–20 Ma after ophiolite formation. The part of the Pindos basin west of the ophiolites survived some 100 Ma after ophiolite obduction, until its final closure in the Tertiary.

Ophiolites of the Anatolian domain and peri-Arabian belt

In the Anatolian domain, east of the Aegean Sea, several belts of Cretaceous ophiolites are present (Fig. 1). They originated in different seaways that formed in the Early Mesozoic (probably Triassic) times and persisted into the Palaeocene (Şengör *et al.* 1988; Yazgan & Chessex 1991; Yılmaz *et al.* 1993; Okay & Tüysüz 1999; Robertson 2002). Here the focus is on the southern belt that also extends along the periphery of Arabia, whereas the more northern belts are only mentioned briefly.

A conspicuous belt of ophiolites extends along the Ankara–Erzincan suture (Fig. 1; Okay & Tüysüz 1999; Yaliniz *et al.* 2000; Robertson 2002; Önen 2003). They are incompletely preserved, but various components show a subduction-related character. These ophiolites were obducted southward in Campanian–Maastrichtian times while a volcanic arc was active on the north side of the host basin. Metamorphic soles yielded $^{40}Ar/^{39}Ar$ ages of 101 ± 4 Ma and 93 ± 4 Ma (Önen 2003). A noteworthy feature is that in western Turkey ophiolites were emplaced over blueschists that formed at *c.* 20 kbar (55–60 km depth) in the Campanian (Okay *et al.* 1998).

Further south ophiolites were obducted over the Tauride block (Fig. 1; Whitechurch *et al.* 1984; Dilek *et al.* 1999). They comprise harzburgitic tectonites and cumulates, whereas higher units are present in a few places only, probably because of syn-emplacement erosion (they may be represented by clasts in the associated mélanges). In these ophiolites the cumulates and higher units, where present, show an SSZ fingerprint (Parlak *et al.* 1996, 2000). Metamorphic soles of the Taurus ophiolites yield K/Ar ages scattered mainly between *c.* 100 Ma and *c.* 85 Ma, with an age concentration around 95–90 Ma, and a few $^{40}Ar/^{39}Ar$ ages of 94–90 Ma (Dilek *et al.* 1999; Parlak & Delaloye 1999). A special feature is that these ophiolites and their metamorphic soles are crossed by numerous diabase dykes that formed from evolved IAT magmas with an SSZ signature that sometimes were less refractory than the magmas that from which the host cumulates crystallized (Whitechurch *et al.* 1984; Parlak & Delaloye 1996; Dilek *et al.* 1999; Çelik & Delaloye 2003; Vergili & Parlak 2005). The dykes are not deformed, which shows that they were emplaced after the deformation of the metamorphic soles, but they do not extend into the underlying mélanges. Two $^{40}Ar/^{39}Ar$ ages of 90–91 Ma and three ages of 90–87 Ma were obtained, and one sample gave an age of *c.* 63 Ma (Parlak & Delaloye 1996; Dilek *et al.* 1999), which shows that most dykes were emplaced shortly after formation of the metamorphic soles.

Farther east, north of the Arabian platform, two ophiolite belts are present (Fig. 1): (1) the north Arabian belt, which is thrust over the Arabian platform, to be discussed below; (2) the SE Anatolian belt, preserved in higher tectonic slices, which also include sediments, mélanges and volcanic rocks, all being thrust over the neo-autochthonous cover of the north Arabian ophiolites (Aktaş & Robertson 1990; Yazgan & Chessex 1991; Yılmaz 1993; Yılmaz *et al.* 1993; Yiğitbaş & Yılmaz 1996; Robertson 2002; Beyarslan & Bingöl 2000; Parlak *et al.* 2004). The sediments in these tectonic units show that the host basin persisted until the Late Eocene, but was tectonized already at the end of the Cretaceous. The special features of the SE Anatolian ophiolites that are important in the present context are their SSZ affinity and the fact that in Coniacian–Maastrichtian times calc-alkaline volcanic–plutonic magmas were emplaced on some of them, whereas at the same time a magmatic arc along which granitoids were emplaced formed on the southern side of the Tauride block (Yazgan & Chessex 1991; Parlak & Rizaoğlu 2004; Parlak *et al.* 2004). Thus the SE Anatolian ophiolites record a direct link between SSZ ophiolites and a calc-alkaline magmatic arc.

The north Arabian ophiolite belt. This ophiolite belt, *c.* 1000 km long, originated in a basin next to the northern Arabian margin (Robertson 2002, 2004; Garfunkel 1998, 2004). It includes the

Troodos (Cyprus), Baer–Bassit (Syria), and Kizildağ (Turkey) ophiolites, and smaller less well-studied bodies as far east as the Cilo ophiolite (Fig. 1). The continuity of this belt as far west as Cyprus is indicated by the similarity of the allochthonous units in Cyprus and in NW Syria (Robertson 2000) and by distinct magnetic anomalies over the intervening sea floor (Woodside 1977). This belt forms the northern part of the peri-Arabian ophiolite crescent defined by Ricou (1971) (see below).

The allochthonous units associated with the Baer–Bassit ophiolite and similar units further east in Turkey as well as the Mamonia complex in Cyprus contain Triassic deep-water sediments, which indicates that the north Arabian margin and adjacent deep basin were shaped already in the Triassic (235–225 Ma), when MORB-like magmatism took place (Robertson & Woodcock 1979; Robertson 1990; Malpas *et al.* 1992; Yılmaz 1993; Al-Riyami *et al.* 2000; Garfunkel 2004). The margin and adjacent part of the deep basin persisted to mid-Cretaceous times at least.

The little deformed Troodos ophiolite (Gass 1980; Robertson & Xenophontos 1993, and references therein), whose base is not exposed, consists of two parts. The main northern part shows a rather regular arrangement of all units of the classic pseudo-stratigraphy (actually it inspired the Penrose definition). The smaller southern part has a complex structure and is separated from the northern part by the east–west-trending Arakapas fault zone, interpreted as a fossil transform (MacLeod & Murton 1993). The tectonites in the main part comprise a block of medium depleted lherzolites next to highly depleted harzburgites (Batanova & Sobolev 2000). The overlying units show an SSZ affinity: cumulates formed from wet magmas that crystallized plagioclase after pyroxenes (Hébert & Laurent 1990); sheeted dykes and extrusive units consisting of a basalt–andesite–dacite–rhyodacite suite and a younger Mg-rich picrite–basalt–basaltic andesite suite with boninitic affinities, both derived from different depleted IAT (Baragar *et al.* 1990; Robinson & Malpas 1990; Bednarz & Schminke 1994; Portnyagin *et al.* 1997). Boninitic rocks occur in the Arakapas fault zone (Robinson & Malpas 1990; MacLeod & Murton 1993). In places the cumulates are cross-cut by later, much less deformed, ultramafic and mafic cumulate bodies (Malpas 1990) that record a relatively late-stage magmatic activity, perhaps related to the younger volcanic rocks. Some plagiogranites yielded U/Pb zircon ages of 90–93 Ma (Turonian), compatible with the ages of fossils in sediments overlying the volcanic rocks (Mukasa & Ludden 1987). Structural studies of the sheeted dykes and extrusive units revealed important normal faulting and block tilting indicative of extension perpendicular to the dykes (now approximately east–west), but the overall pseudo-stratigraphy was not disrupted (Varga & Moores 1985; Allerton & Vine 1991). Metamorphic soles are not known, as the base of the Troodos is not exposed, but amphibolite-grade slices resembling metamorphic soles of other ophiolites in the region are exposed next to the ophiolite (Malpas *et al.* 1992) and they yield similar $^{40}Ar/^{39}Ar$ ages of 90–83 Ma (Spray & Roddick 1981).

The Kızıldağ (Hatay) ophiolite (Delaloye & Wagner 1984; Lytwyn & Casey, 1993; Dilek & Thy 1998; Dilek *et al.* 1999; Bağci *et al.* 2005) also shows all units of the classic pseudo-stratigraphy, but they are often in fault contact. The faulting was interpreted by Dilek & Thy (1998) and Dilek *et al.* (1999) as recording extension during accretion of the ophiolite. The tectonites consist largely of harzburgites. The sheeted dykes and volcanic rocks are similar to those in Troodos, ranging from IAT and basaltic andesites to rocks with boninitic affinity, indicating an SSZ affinity.

The Baer–Bassit ophiolite, unlike the Troodos and Kızıldağ ophiolites, is broken into slices that are imbricated with the sub-ophiolitic allochthonous units, but all the units of the pseudo-stratigraphy are present (Al Riyami *et al.* 2000, 2002). The tectonites consist predominantly of harzburgites. The dykes and extrusive rocks consist of depleted IAT and rocks with a boninitic affinity, resembling the higher lavas of the Troodos ophiolite. The metamorphic sole yielded K/Ar ages of 85–95 Ma (Delaloye & Wagner 1984; Al Riyami *et al.* 2002). The presence of lavas with both normal and reverse magnetization (Morris *et al.* 2002) records a somewhat younger age, but this magnetization may have been acquired during late hydrothermal alteration.

Obduction of the ophiolites downflexed the north Arabian margin, which produced a foreland basin in the Campanian (Yılmaz 1993). The Kızıldağ and Baer–Bassit ophiolites are overlain by ophiolite-derived clastic rocks and shallow-water sediments of Late Maastrichtian and younger ages (Delaloye & Wagner 1984; Dilek & Thy 1998; Al-Riyami *et al.* 2000), which indicates exposure after obduction over the continental north Arabian platform. Farther east similar relations are documented (Yılmaz 1993). The Troodos ophiolite had a different history (Robertson 2000, 2002; Lord *et al.* 2002). In the Campanian it was covered mainly by thin basinal sediments, except in western Cyprus where fine-grained calc-alkaline volcaniclastic deposits (the

Kannaviou Formation) accumulated. At the same time the Mamonia complex was tectonized and contributed to a mélange (the Moni mélange, now south of the ophiolites) that does not contain material derived from the ophiolite. The two units were brought together in the Maastrichtian and then contributed clasts to a common cover, which in turn is overlain by lower Tertiary pelagic sediments. This records an underwater position and emplacement on the thin crust of the Levant basin.

Palaeomagnetic data (Morris *et al.* 1998) show that the assembly of these units in Cyprus involved considerable lateral motions. Palaeomagnetic data also indicate that the main part of the Troodos ophiolite rotated *c.* 90° counter-clockwise on a vertical axis, mostly in Campanian–Maastrichtian times (Morris 1996), various parts of the Baer–Bassit ophiolite rotated from *c.* 100° to 220° counter-clockwise during obduction (Morris *et al.* 2002). While these ophiolites were obducted, a basin with a complex history persisted farther north until Late Eocene times and there the SE Anatolian ophiolites formed (see above).

The east Arabian (Zagros) ophiolite belt. The eastern part of the peri-Arabian ophiolite crescent originated in a basin between Arabia and the Sanandaj–Sirjan block (Fig. 1). Because the latter may have been distinct from the Tauride block, which controlled the seaway in which the north Arabian ophiolites originated, the two parts of the peri-Arabian belt are here considered separately. The east Arabian margin was shaped by late Permian and Triassic rifting (Lippard *et al.* 1986; Rabu 1993). Palinspastic reconstructions (e.g. Le Pichon *et al.* 1988; Stampfli *et al.* 2001) show that afterwards a wide oceanic area developed east of the east Arabian margin.

The east Arabian belt is best represented by the well-studied Semail ophiolite (Fig. 1). The sub-ophiolite allochthonous units represent a complex 300–500 km wide basinal area, the Hawasina basin, that existed off the Oman passive margin and received basinal sediments until Cenomanian–Turonian times. A similar feature has not been documented further north. The Semail ophiolite, *c.* 500 km long and 50–100 km wide, shows the classic pseudo-stratigraphy (Lippard *et al.* 1986; Rabu 1993). The tectonites, predominantly harzburgites, record high-temperature penetrative deformation that was interpreted as recording flow during and after melt extraction under a somewhat irregular spreading centre (Nicolas 1989). However, some chromites in harzburgites contain inclusions of basalt with an SSZ character (Schiano *et al.* 1997). The cumulates mostly show a crystallization order different from that of MORB (Lippard *et al.* 1986). The older and most voluminous part of the volcanic rocks (Geotimes series) consists of basalts and basaltic andesites transitional between MORB and IAT with a weak SSZ signature; these are considered to be related to much of the sheeted dyke complex and the cumulates (Alabaster *et al.* 1982; Lippard *et al.* 1986; Ernewein *et al.* 1988). These are overlain by basalts, andesites and some dacites (Alley series) with a more pronounced SSZ signature (Alabaster *et al.* 1982; Lippard *et al.* 1986; Ernewein *et al.* 1988; Umino *et al.* 1990). Locally the latter series contains boninitic flows, and such magmas also form dykes in the underlying units (Ishikawa *et al.* 2002). The cumulates and tectonites are often crossed by late-stage discordant and little deformed minor intrusions of wehrlites and pyroxenites, as well as gabbros, diorites and trondhjemites that are related to the younger volcanic rocks (Lippard *et al.* 1986; Juteau *et al.* 1988; Umino *et al.* 1990). Some of the more acid rocks sometimes intrude the sheeted dykes. In places the late intrusions show a calc-alkaline differentiation trend (Lachize *et al.* 1996). The ultramafic tectonites are crossed by numerous dykes (Python & Ceuleneer 2003). These include a group of troctolites and olivine gabbros that crystallized from MORB-like magmas that precipitated plagioclase before pyroxene (unlike the cumulate unit) and intruded the harzburgites while they were still hot (>1000 °C). A second, more widespread group consists of gabbro–norites and pyroxenites that were precipitated from highly depleted magmas with SSZ affinity and were emplaced while the host rocks were cooling.

High-level plagiogranites yielded Cenomanian U/Pb zircon ages of 93.5–97.9 Ma clustering around 95 Ma (Tilton *et al.* 1981), whereas $^{40}Ar/^{39}Ar$ ages from plagiogranites, gabbros and veins range from *c.* 96 Ma to *c.* 93 Ma (mean 94.4 ± 0.5 Ma), and granitic and dioritic rocks of the wehrlitic series yielded a mean age of 93.8 ± 0.3 Ma (Hacker *et al.* 1996). These ages are compatible with the presence of late Cenomanian to early Turonian (*c.* 95–91 Ma) fossils in sediments within the volcanic rocks, but pelagic deposition continued into the Santonian (Tippit *et al.* 1981). The latest volcanic rocks (Salahi unit), associated with Coniacian (89–86 Ma) fossiliferous sediments, consist of alkali basalts with a within-plate character (Alabaster *et al.* 1982; Lippard *et al.* 1986; Umino *et al.* 1990).

Metamorphic soles underlie the entire ophiolite and record peak metamorphic conditions that vary from *c.* 13 kbar, 800 °C to *c.* 7 kbar, 700–800 °C, corresponding to depths that considerably exceed the thickness of the ophiolite (Hacker & Gnos 1997; Searle & Cox 1999). They yielded $^{40}Ar/^{39}Ar$ hornblende ages that cluster in narrow ranges of 93.5 ± 0.1 Ma and 94.9 ± 0.2 Ma in the north and south of the ophiolite (Hacker *et al.* 1996; K–Ar ages show a much wider scatter) whereas micas from the soles yielded slightly younger ages, recording their cooling within a few million years (Hacker *et al.* 1996; Hacker & Gnos 1997). Some soles are crossed by undeformed dykes. A hornblende in one dyke gave a $^{40}Ar/^{39}Ar$ age of 93.7 ± 0.8 Ma (Hacker & Gnos 1997). In the north the tectonites and the lower cumulates are cut by dykes and lenses of peraluminous granites that gave $^{40}Ar/^{39}Ar$ ages of 89–90 Ma and a K/Ar age of 85 ± 3 Ma (Hacker *et al.* 1996; Searle & Cox 1999).

Palaeomagnetic data (Perrin *et al.* 2000; Weiler 2000) show that that the southern part of the ophiolite rotated *c.* 20° counter-clockwise whereas its northern part rotated *c.* 120° clockwise after the end of magmatism, but differential rotations between portions of the northern part had taken place already during the last stage of volcanic activity.

The Hawasina basin and the nearby Oman margin were considerably deformed in Late Cenomanian–Turonian to Santonian times, which led to deposition of mélanges (the Muti Formation) that contain rock fragments derived from the margin and from within the basin, but not from the ophiolite (Lippard *et al.* 1986; Robertson 1987; Rabu 1993). This could arise if the ophiolite was still far from the deformed area, or if it was still low-lying and little deformed, and so did not contribute clastic deposits. The nearby platform edge was fractured and eroded at the beginning of this period, and locally minor volcanism took place, but in the Coniacian (87–86 Ma) the platform edge was downflexed and by the Campanian a foreland basin of >100 km width developed and was filled by up to 4 km of sediments in front of the advancing nappes (Patton & O'Connor 1988; Boote *et al.* 1990; Warburton *et al.* 1990; Rabu 1993). Clastic deposits derived from the ophiolite and underlying allochthon appear in the Late Campanian (at *c.* 75 Ma), which shows that then the ophiolite advanced already over the continental margin and was uplifted enough to be eroded. The end of obduction is marked by erosion and subaerial weathering of the ophiolite and subsequent deposition of Late Maastrichtian marine sediments over the ophiolite. The oceanic basin NE of Oman persisted after emplacement of the Semail ophiolite, and a relict still survives, but in the Tertiary additional deformation took place (Coleman 1981; Lippard *et al.* 1986; Rabu 1993).

A special feature of the Semail ophiolite is that it overlies strongly deformed slices of rocks that were metamorphosed at high-pressure conditions, reaching eclogite-facies conditions that were variously estimated at 500–580 °C and 2.0–2.4 GPa (Searle & Cox 1999) and 550–580 °C and 1.2–1.6 GPa (El-Shazly 2001), which correspond respectively to depths of 70–80 km and 40–50 km. The protoliths consist of continental margin rocks, indicating subduction of the edge of the Arabian continent, but the history of these rocks is debated (Gray & Gregory 2003; Searle *et al.* 2003). In the present context the important point is that the Hawasina basin was still undeformed when the ophiolite and its metamorphic sole formed, so the latter were still far (hundreds of kilometres) from the continental margin, and so their evolution until then could not have been directly related to the subduction of the edge of Arabia.

The more northern east Arabian ophiolites have been little studied. The Neyriz ophiolite (Fig. 1) is rather faulted but comprises all units of the pseudo-stratigraphy (Sarkarinejad 1994). The tectonites consist of harzburgites and subordinate lherzolites, whereas the dykes and volcanic rocks have a similar geochemical character to the main part of the lavas in the Semail ophiolite, and differ only little from MORB. The Kermanshah ophiolite comprises harzburgitic tectonics, and its dykes have SSZ chemistry (Desmonds & Beccaluva 1983). These ophiolites were obducted at the same time as the Semail ophiolite and then a basin still remained NE of them (Ricou 1971; Ghazi *et al.* 2004).

Summary. The foregoing outline shows that the Cretaceous ophiolites of the Anatolian domain and the peri-Arabian belt formed within a basin that originated in Triassic or Permian times, and thus was *c.* 100 Ma old when the ophiolites formed. These ophiolites are constrained to have formed and evolved in a narrow time interval (often *c.* 10 Ma), although they originated in distinct seaways. Where metamorphic soles have been dated they mostly yield Cenomanian–Turonian ages, very close to, or even overlapping, the age of the ophiolite rocks. The sheeted dykes and extrusive rocks in these ophiolites have an SSZ geochemical signature, which is strongly expressed in the Anatolian and North Arabian ophiolites, but weakly in most of the lavas of the Semail and probably Neyriz ophiolites, although the younger lavas and dykes in the Semail ophiolite have a well-expressed SSZ signature. In the latter, some rocks that display a calc-alkaline differentiation trend are present. In the east Anatolian ophiolites this trend is well displayed. Also noteworthy is the presence of undeformed dykes cutting the metamorphic soles of the Tauride ophiolite and the Semail ophiolite. Final obduction took place some 20 Ma after ophiolite formation, but the host basins persisted for variable periods (often 30–40 Ma, whereas the basin next to the Semail ophiolite has not closed yet).

Place of ophiolite development in the history of the basins

To highlight the evolution of ophiolites in relation to the host basins it is convenient to divide their history into several stages. Shervais (2001) divided the history of SSZ ophiolites, mainly based on their petrological evolution, into the following stages: birth, formation of the bulk of the ophiolite; youth, formation of components derived from depleted sources; maturity, formation of arc-like components; death, end of magmatism and thrusting over the nearby basin; resurrection, emplacement of the ophiolite. Here the emphasis is on the ophiolite–host basin relations rather than on the petrological evolution, so a somewhat different division is used: (1) the pre-ophiolite stage, which produced the palaeogeographical setting in which the ophiolites originated; (2) the ophiolite formation stage; (3) the ophiolite tectonization and obduction stage, which may overlap the previous stage; (4) the final, post-obduction stage, ending with the closure of the basin. Stage (2) includes the first three stages of Shervais (2001), which he noted may overlap to some extent in time and space, and stage (3) overlaps his two subsequent stages.

Pre-ophiolite stage

The foregoing outline shows that the ophiolites considered here formed when the host basins were already well developed. Although these basins were destroyed, it is possible to constrain their widths and the spreading rates of ridges that existed in them.

It is expected that after initial rifting the basins widened by nearly symmetrical sea-floor spreading along ridges, similar to the present ridges. As spreading progressed the ridges moved away from the passive margins and remained in the middle of the basins as long as there was no subduction (Fig. 2). If subduction began on one side of a basin, the ridge could move farther away from its other margin and could even be subducted. Thus if a ridge survives some time after basin initiation, its average long-term spreading rate is constrained by the width of a basin at that time (Fig. 2). If faster spreading ridges existed, they could not persist, because that would imply production of an area wider than the basin. Such fast-spreading ridges would either be subducted or their spreading will have stopped or slowed down.

Palinspastic plate reconstruction at *c.*105 Ma, shortly before the ophiolites formed (Fig. 3), allows us to apply these considerations to the basins that hosted the Cretaceous ophiolites. It is seen that wide oceanic seaways existed then in the region, implying considerable sea-floor spreading since their inception. Along any transect the total width of the seaways is given approximately by the convergence between Eurasia and Arabia–Africa (taking into account that a part of the convergence (100–200 km(?)) was taken up by shortening of continental areas). The convergence across the Anatolian domain was *c.* 2400 km since 105 Ma ago and *c.* 1800 km since 92 Ma ago, using the plate kinematics of Müller & Roest (1992). Thus when the north Arabian ophiolites formed the host basin could have been 500–800 km wide, leaving enough room for similar seaways farther north. If subduction had not occurred in this basin, then the average rate of its north–south opening was 0.5–0.8 cm a^{-1} (the actual opening could have been in a somewhat different direction). To accommodate this opening (and the opening of more northern seaways), together with the eastward motion of Africa–Arabia relative to Europe, the subduction must have been oblique somewhere in this region before 105 Ma (Fig. 3).

Farther east the motion of Africa–Arabia relative to Eurasia since Mid-Jurassic times reduced the oceanic area east of Arabia (Fig. 3; Le Pichon *et al.* 1988; Müller & Roest 1992; Stampfli *et al.* 2001). This requires (oblique) subduction somewhere on the NE side of this oceanic area, perhaps along the Sanandaj–Sirjan block, where Late Jurassic and Cretaceous igneous activity took place (Berberian & Berberian 1981; Berberian & King 1981). The total convergence between Oman and Eurasia was *c.* 3200 km since 105 Ma ago and *c.* 2300 km since 92 Ma ago. Thus, taking into account the presence of other seaways in this area (McCall 1997), the basin hosting the Semail ophiolite could have been up to 2000–2200 km wide when this ophiolite formed. Thus, if a ridge survived in this basin for 120 Ma (since the Late Triassic) its average spreading rate must have been less than *c.* 3.5 cm a^{-1}, or else it would have been subducted.

The history of the Pindos basin is not well constrained, but it could not have been much wider than 500–800 km, given the space between the

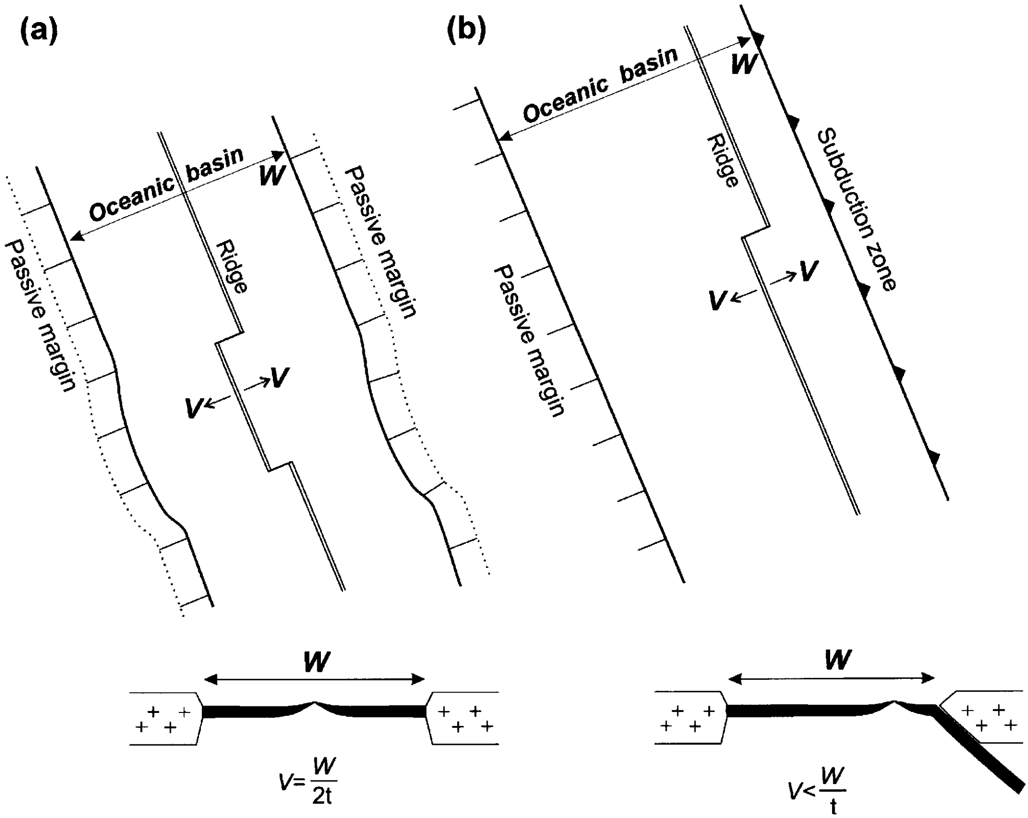

Fig. 2. Constraint on spreading rate of ridges. (**a**) Situation in ocean in which there is no subduction. (**b**) Situation in oceanic basin with a subduction zone on one side: ridge does not remain in middle of basin. Top, map view; bottom, cross-section. w, width of the basin at time t; v, half-spreading rate.

major continents in that area. Thus, if a ridge persisted in this basin for 60 Ma its average spreading rate could not have exceeded 1.6–2.7 cm a^{-1}, if a subduction zone existed in this basin before ophiolite formation (otherwise long-term average spreading rate would have been half these rates).

In summary, these considerations show that the ophiolites considered here formed in 60–100 Ma old basins that were 500 km to a couple of thousand kilometres wide, which amplifies the conclusion of Shervais (2001) that ophiolites form in wide basins. The ridges that produced these basins must have been slow spreading if they persisted to the time of ophiolite formation. Otherwise they would have produced areas wider than the basins (alluded to by Robertson & Woodcock 1979), so either they would have been subducted or their activity would have been intermittent. The latter option, although not impossible, is difficult to evaluate in basins that no longer exist, but there do not seem to be any firm arguments for the general applicability of such a scenario.

Ophiolite formation stage

The setting in which the ophiolites discussed above formed is constrained by their regional framework, their structure, and their chemical signature.

The regional framework. The regional framework of the basins hosting the Cretaceous ophiolites considered here is constrained, as discussed above, by the motions of Africa–Arabia relative to Eurasia. In the present context the important feature, accepted in all plate knematic models (e.g. LePichon *et al.* 1988; Müller & Roest 1992), is that these motions changed considerably *c.* 105 Ma ago or somewhat later (Fig. 3), i.e. slightly before the formation of the Anatolian

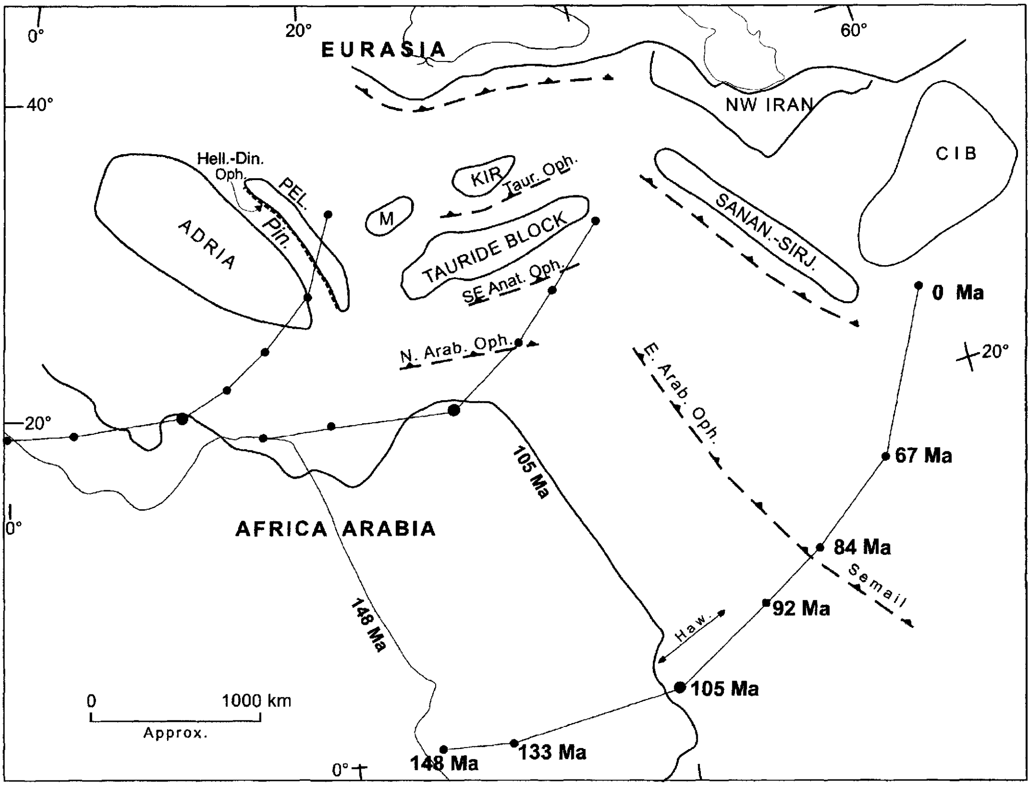

Fig. 3. Palaeogeographical situation about 105 Ma ago. Position of Africa–Arabia relative to Eurasia at various times is after Müller & Roest (1992). Coordinates relative to Eurasia are shown for reference. Positions of micro-continents are approximate, shown to outline the Neotethyan seaways (Iranian blocks are shown schematically based on Şengör *et al.* (1988) and McCall (1997)). Also shown are possible positions of new subduction zones along which the ophiolite belts mentioned in the text formed. Older subduction zone are shown by thicker lines.

and peri-Arabian ophiolites. Figure 3 shows that in the Anatolian domain the motion of Africa–Arabia changed from nearly parallel to its northern margin to convergence at an angle of *c.* 45° to this margin at rate of *c.* 4 cm a^{-1}. The convergence was probably partitioned between the seaways in this region, so new subduction zones should have formed in them. East of Arabia convergence at an overall rate of >6 cm a^{-1} continued, but its direction changed by *c.* 30°, so that it became almost perpendicular to the Arabian–Oman continental margin. The plate motions across the Pindos Ocean are not known, but the initiation of opening of the Ligurian seaway on the western side of Apulia shortly before *c.* 170 Ma (Lombardo *et al.* 2002) would probably change the motion across the Pindos Ocean close to the time of formation of the Hellenic–Dinaric ophiolites. In a wider context this may be related to the opening of the central Atlantic at that time (Smith 2004). Thus, the ophiolites considered here formed close to the time of changes in the motions of the plates or micro-continents bordering the host basins.

Internal structure of the ophiolites. The interpretation that ophiolites represent new crust that formed by some type of sea-floor spreading implies that their width perpendicular to the sheeted dykes measures the amount of spreading during their formation. Thus the largest bodies (Troodos, Semail, and Vourinos–Pindos) record spreading of ≥100 km, whereas smaller ophiolites record spreading of tens of kilometres.

The spreading rates are difficult to determine, however, because dating is not precise enough. Therefore indirect indications were used. Nicolas (1989), who favoured ophiolite origin along MORs, proposed that differences in spreading rates controlled the structure of various ophiolite types. As the factors that control generation of new crust in both MOR and SSZ settings are likely to be similar, it is instructive to compare the structure of ophiolites with that of oceanic crust, regardless of the setting envisaged for their production. Detailed studies of the structure of oceanic crust (Dilek *et al.* 1998; Karson 1998) show that the crust that formed along slow-spreading ridges (e.g. in the Atlantic Ocean, spreading rates 2–4 cm a^{-1}) has an irregular structure, often as a result of faulting, so that its components have variable thicknesses. This is interpreted as resulting from limited and intermittent magma supply, which does not allow fast enough construction of new crust, so a part of the spreading is accommodated by slip on normal faults and shear zones (amagmatic spreading). The lherzolitic ophiolites that have such a structure (e.g. the west Albanian ophiolites) could therefore have formed by slow spreading. It is noteworthy that similar features were observed also in the Mariana back-arc basin (Ohara *et al.* 2002), supporting the idea that the same factors control spreading in SSZ and MOR settings. On the other hand, the layered structure of the harzburgitic ophiolites with a complete Penrose pseudo-stratigraphy resembles the structure of crust formed along the fast-spreading East Pacific Rise (spreading rates 8–12 cm a^{-1}). Thus the Semail ophiolite is interpreted to have formed by fast spreading, perhaps 10 cm a^{-1} (Nicolas 1989; Dilek *et al.* 1998). The Troodos and Kızıldağ ophiolites are faulted but still show the Penrose pseudo-stratigraphy better than lherzolitic ophiolites, suggesting formation by intermediate spreading, perhaps 4–6 cm a^{-1}. These are, however, only approximate estimes.

The important point in the present context is that such spreading rates could be maintained only during time intervals that are considerably shorter than the life span of the basins hosting these ophiolites, for otherwise areas wider than these basins would be generated. Thus a spreading rate of 10 cm a^{-1} could generate the entire width of the basin hosting the Semail ophiolite in *c.* 20–25 Ma, whereas a spreading rate of 2–3 cm a^{-1} could generate the entire width of the basin hosting the Troodos and Kızıldağ ophiolite in *c.* 20–30 Ma. Thus, ophiolite generation appears to mark short episodes of fast crustal accretion, considerably faster than the above estimates of the long-term average spreading rates in the host basins.

The sites of ophiolite generation. The sediments directly overlying the ophiolites record formation in an intra-oceanic setting, far from continental margins (Robertson 2004). The nature of the protoliths of the metamorphic soles (basinal sediments and volcanic rocks, including MORB-like rocks) and the presence of such components in the sub-ophiolitic allochthons (Robertson 2002, 2004) also indicate that the ophiolites were thrust over intra-oceanic areas.

Geochemical signature. The foregoing review shows that the geochemical signature of igneous rocks in the harzburgitic ophiolites resembles that of rocks in SSZ rather than in MOR settings. The strength of the SSZ fingerprint is variable, however. At one extreme, in the Semail ophiolite, a considerable part of the lavas and cumulates formed from magmas transitional between MORB and IAT, whereas only the less voluminous younger volcanic rocks and intrusions have a well-expressed SSZ signature. In other ophiolites (e.g. in the Hellenic-Dinaric belt), rocks with both MOR-like and SSZ fingerprints are present, sometimes alternating in time. Generally, rocks with a boninitic affinity typical of an SSZ setting tend to appear late in the ophiolite history.

It is the SSZ geochemical fingerprint of many ophiolites (which seems to characterize most ophiolites: Shervais 2001) that is the main argument for linking their formation with young subduction zones. If such ophiolites formed along MORs it is difficult to understand why the rocks normally found along the present-day ridge do not dominate, or at least are common, in ophiolites.

Models of formation

The foregoing considerations show that the Semail, Troodos and other ophiolites considered here formed in special short-lived events (unless the spreading rates during their formation were much overestimated). Moreover, the fact that both the Jurassic and Cretaceous ophiolites considered here formed within short time intervals in belts that are often *c.* 1000 km long or even longer also points to special events, which, given the regional framework, took place close to times of changes in plate motions. This makes it difficult to consider ophiolite formation as a result of normal long-term spreading of ridges in the host basins. Thus, interpreting ophiolites as originating along MORs is not as simple a model as appears at first sight.

These features fit, however, the alternative interpretation of the harzburgitic ophiolites as originating in an SSZ setting that was based primarily on their geochemical fingerprint. The common occurrences of IAT-like rocks, which are characteristic of SSZ settings but are uncommon in MORs, and especially the occurrence of rocks with a boninitic affinity, which are known

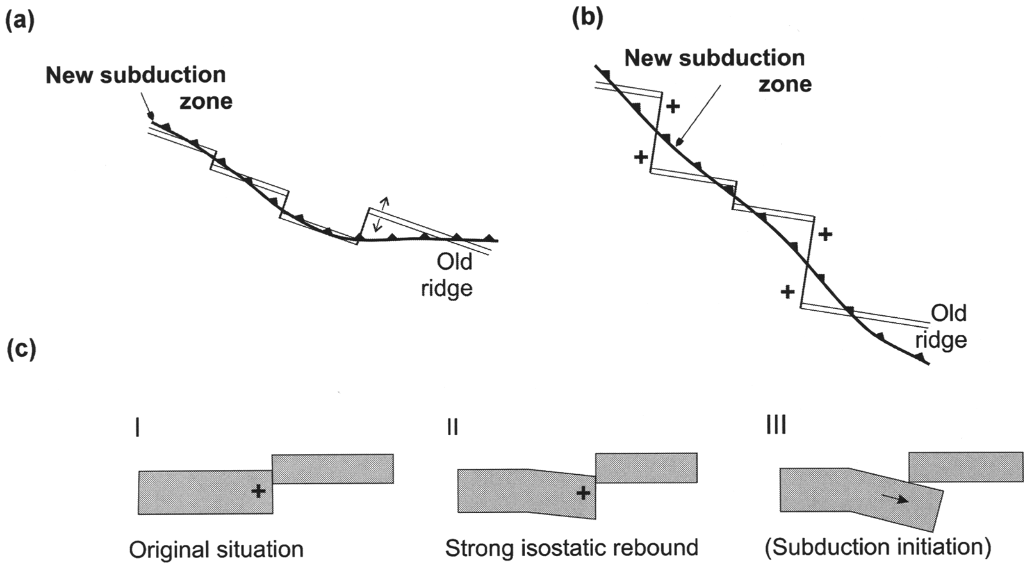

Fig. 4. Development of ophiolites in an SSZ setting, from inception of a new subduction zone to obduction. (See text for discussion).

only in young intra-oceanic forearcs, point at an origin in an SSZ setting (Pearce *et al.* 1984), which is now widely accepted. Analogy with western Pacific forearcs (Casey & Dewey 1984; Hawkins *et al.* 1984; Leich 1984) inspired later interpretations (e.g. Stern & Bloomer 1992), and is generally followed here. Hence the formation of ophiolites in short time intervals (*c.* 10 Ma) along long belts is considered to express distinct 'ophiolite events' in which strips of crust were produced along new subduction zones that formed in response to changes in plate motions (which was suggested already by Moores (1982), from a different point of view). These considerations provide the framework for the following attempt at interpreting the features of the harzburgitic ophiolites considered here (the lherzolitic ophiolites will be discussed later).

The polarity of the subduction is not directly constrained, but it can be assumed, following Coleman (1981) and Moores (1982), that during emplacement of Tethyan ophiolites the descending slabs dipped away from the passive margins on which the ophiolites were obducted. Intra-oceanic subduction with such a polarity, but not the opposite one, will consume the basinal areas between these margins and the site of ophiolite formation. Assuming such a polarity during ophiolite formation allows the same subduction zone to serve for both ophiolite formation and obduction. This interpretation fits the ophiolites considered here, as there is no record of subduction along the passive margins on which they were obducted (but does not necessarily apply to Cordilleran-type ophiolites). It should be noted that if subduction along an active margin brought an ophiolite to this margin, then another subduction zone should be invoked for the intra-oceanic formation of the ophiolite.

The mechanism by which new subduction zones are initiated is not known, and perhaps they arise in several different ways (Casey & Dewey 1984; Stern 2004). Given the original intra-oceanic position of the ophiolites considered here, the subduction zones over which they formed must have originated within the host basins. Possibly an intact part of a basin failed under the influence of externally applied forces, as in the central Indian Ocean (Bull & Scrutton 1992; Krishna *et al.* 1998). At first, a system of thrusts develops, but later thrusting is localized to form a single subduction zone (Fig. 4a). Another possibility is that density differences along transforms may influence the process (Casey & Dewey 1984). However, as transforms are expected to be in isostatic equilibrium, as is the case today, there is no force to cause spontaneous foundering of the older and heavier sides. If such foundering occurred, then in the early stages the isostatic equilibrium would be greatly disturbed in a manner that will oppose the foundering (Fig. 5). Thus, failure induced by externally applied force, such as may arise during changes in plate

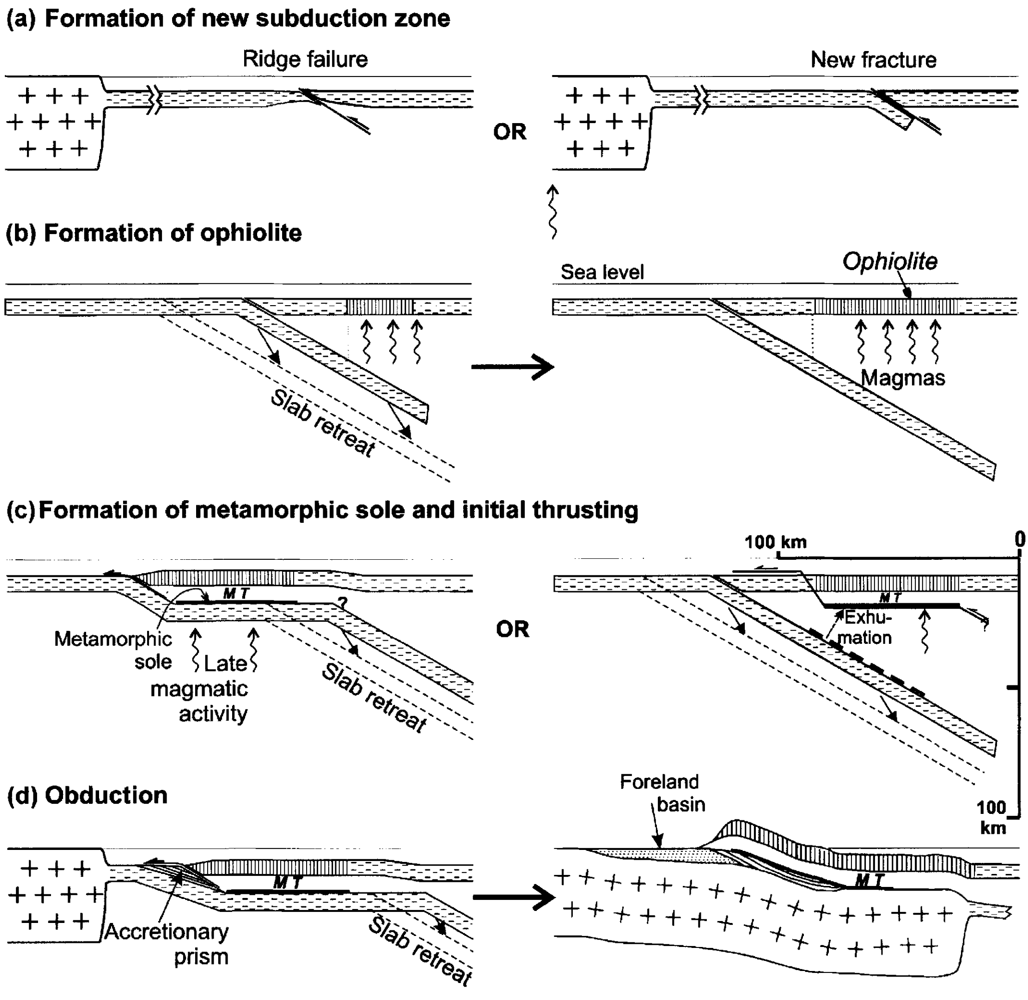

Fig. 5. Problems involved in the formation of new subduction zones along previous plate boundaries. (**a**) Formation along a ridge with small-offset transforms (map view): the subduction zone cannot follow only the weak zero-age crust along the ridge crest, but it must cut across transforms. (**b**) Formation along a ridge offset by long transforms (map view): the subduction zone cannot follow zero-age crust; it is also noteworthy that the lithosphere is older and heavier on alternating sides of each transform. (**c**) Formation of a new subduction zone at the expense of a long transform (cross-section): the older and colder side (+) may tend to sink (stage I), but in order to begin to subduct (stage III) it must pass through stage II, in which isostatic rebound will occur and prevent further sinking of the older lithosphere unless there is an externally driving force. MT, metamorphic sole.

motions, is more probable than spontaneous foundering. It should also be noted that the formation of new subduction zones of 1000 km length or more (the length of some ophiolite belts) requires the existence of much longer transforms, because the polarity of density reversals changes along the transforms. It is not clear how to place such long transforms in the basins hosting the ophiolites considered here, but this requires further study. A change in plate motions may also cause failure along a ridge because this is the weakest place within the basin (e.g. Coleman 1981; called 'ridge collapse' by Robertson 2004). However, ridges are usually not simple zones of weakness but consist of segments separated by transforms, so their failure is more complicated than simply disrupting a continuous weak ridge crest, unless transform offsets are small (Fig. 5). It also involves to difficulty of subducting hot and buoyant ridge crests.

The Hellenic–Dinaric ophiolites may have formed by failure (collapse) of a ridge in the Pindos Ocean where there is no indication of former subduction. This allows the western part of the belt to be a remnant of a ridge, but it does not fit the absence of a clear east–west age difference across the belt, suggesting a different origin (see below). In the seaway between Arabia and the Tauride block, where the north Arabian and SE Anatolian ophiolite belts originated, two subduction zones may have formed: one off the Arabian margin along which Troodos–Kızıldağ ophiolites formed, and one farther basinwards, along which the island arc next to the Tauride block formed (Fig. 3), so perhaps only one of them initiated by ridge failure, but they could have formed differently. The palaeomagnetic data from Troodos indicate that the southern subduction zone was oriented close to east–west. Other subduction zones are assumed to have formed farther north of the Tauride block to account for the Tauride and the more northern Anatolian ophiolites. In the basin east of Arabia, which was shrinking since Jurassic time, a subduction zone probably existed already along the Sanandaj–Sirjan block (Fig. 4). However, the continuation of magmatic activity along this block into the Late Cretaceous (Berberian & Berberian 1981; Berberian & King 1981) suggests that this subduction zone remained active until that time, so it was not available as the site of formation of the east Arabian ophiolites. Therefore another subduction zone is assumed to have formed inside the basin (Fig. 3). In all these cases the new subduction zones could form by ridge failure, provided that ridges still survived in the basin, which is uncertain, and provided that this is a viable mechanism; otherwise, failure of the basin floor should be assumed. Hacker & Gnos (1997) suggested that the Semail ophiolite formed along a transform fault. This must have trended closer to east–west than to north–south, based on the palaeomagnetic data, and should have been comparable in length with the Semail ophiolite. It is not clear how such a fault fits into the history of the host basin.

The production of ophiolites by some form of sea-floor spreading in an SSZ setting is interpreted as a consequence of retreat (roll-back) of the descending slabs (Fig. 4b; Elsasser 1971 (who called this behavior 'retrograde subduction'); Chase 1978). This is a common behaviour of subducted lithospheric slabs (Garfunkel *et al.* 1986), being a result of their negative buoyancy, which adds a downward component of motion to their dip-parallel descent, so that the slabs descend at a steeper angle than their dips (the dip need not change). This causes the hinges where the lithosphere bends into the mantle to retreat, i.e. to move opposite to the slab dip, at rates of a few centimetres per year (Garfunkel *et al.* 1986; updated models of plate motion do not change this conclusion). Mantle flow may also promote slab retreat (Flower & Dilek 2003). In the present context it is important to note that slabs in young subduction zones (e.g. next to the Scotia Sea) also retreat at fast rates, and this also occurred along the western Pacific subduction zones immediately after their formation (Stern & Bloomer 1992). Thus if in a basin that shrinks at a rate of say 1–2 cm a^{-1}, slab retreat was faster, say 3 cm a^{-1}, then sea-floor spreading at a rate of 2–1 cm a^{-1} will occur behind the trench, so there a 100 km wide strip of new crust will form in 5–10 Ma. In the case of the Semail ophiolite, considered to have formed by spreading at a rate of *c.* 10 cm a^{-1}, even faster slab retreat must be envisaged.

Magma generation along the ophiolite-related subduction zones can be related to models of the dynamics and thermal state in the mantle wedge of subduction zones. These show that viscous drag of the descending slabs induces a corner flow in the wedges and that this flow is much enhanced by slab retreat, because material is sucked into the wedge to fill the space left behind the slabs (Fig. 6; Garfunkel *et al.* 1986; Davies & Stevenson 1992; Kincaid & Sacks 1997), and it is further modified when spreading behind the trench diverts some material towards the site of spreading (Ribe 1989). This flow replenishes the mantle wedge with hot material, which counteracts the cooling effect of the descending slab and thus allows continuing magma generation. However, the region close to the trench remains stagnant and cold, and there magmas are not generated (Davis & Stevenson 1992; Kincaid & Sacks 1997). Next to modern trenches this cold amagmatic region is 150–200 km wide (Gill 1981) and igneous activity occurs only where the slab depth exceeds 65–130 km (England *et al.* 2004). In nascent subduction zones this cold region can be smaller, especially if subduction begins in young sea floor (e.g. near ridges) where the mantle is hot (Kincaid & Sacks 1997). In such cases igneous activity may occur closer (50 km or less?) to the trench above shallow (30 km?) parts of the slabs.

Magmas with an SSZ fingerprint are considered to be derived from mantle sources that had already been depleted to varying degrees by previous partial melting events (e.g. MORB production), and were remelted as a result of addition of slab-derived aqueous fluids that carry a 'subduction component' of various trace elements (Gill 1981; McCulloch & Gamble 1991;

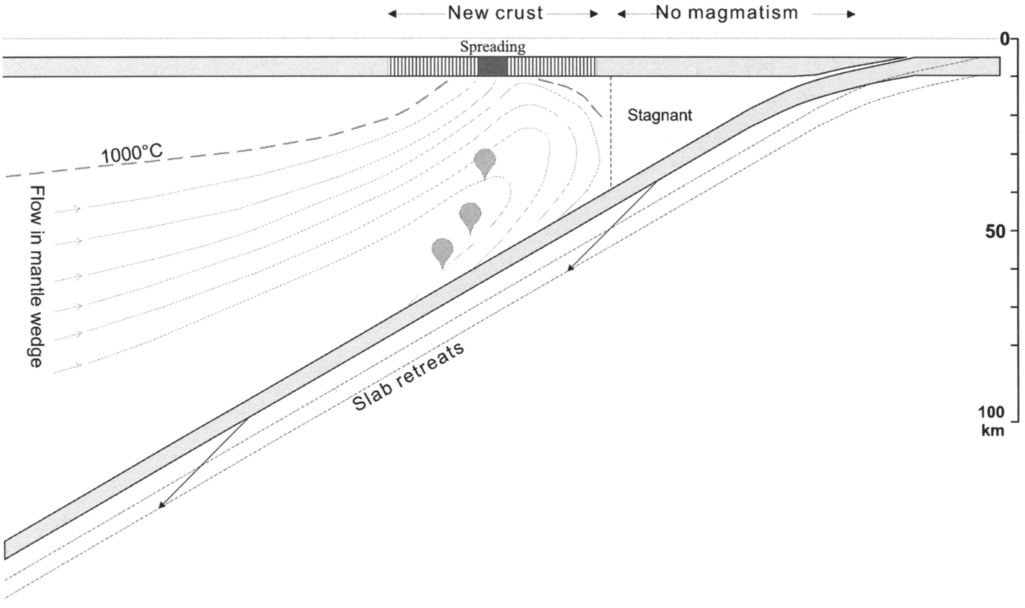

Fig. 6. Model of flow and magma generation in the mantle wedge in subduction zones. (See text for discussion.)

Hawkesworth *et al.* 1993; Pearce & Peate 1995). Boninitic-like lavas are derived from sources that are more depleted than the IAT sources (Pearce *et al.* 1992). In all cases, the addition of water lowers the melting temperature and is essential for promoting magma generation from otherwise refractory sources.

The flow system in the mantle wedge can entrain depleted peridotites from beneath the overriding plate (Fig. 6), where they were left behind after having contributed partial melts to generate the overlying crust. However, to be hot enough to flow and to melt again, such material must be derived from some depth beneath the Moho. More fertile mantle coming from greater depth can also be entrained by the flow and brought into the mantle wedges. Alternatively, spreading behind the subduction zone may also occur in a proto-back-arc basin (before a real arc develops) into which normal mantle material rises and melts to form magmas similar to MORB (similar to present-day back-arcs: Hawkins 1995, 2003), and then the still hot and ductile residue can be swept towards the narrower part of the wedge, where it is hydrated and remelted. This implies that the ophiolites represent only a part (that was obducted) of the zone of magma generation next to newly formed subduction zones.

This picture is probably oversimplified and requires further elaboration. The occurrence side by side of various magmas derived from sources that were depleted to different extents (e.g. Troodos, Semail, Pindos) strongly suggests that the mantle wedge is a heterogeneous region, probably on the scale of a few kilometres to several tens of kilometres, which results from the flow being more complex than suggested by simple models. The tendency of magmas with a strong SSZ character, and especially with a boninitic affinity, to appear in advanced stages of construction of the ophiolites considered here (and in many others as well: Shervais 2001) raises the possibility that diapiric upwellings that promote decompression melting of hydrated refractory mantle become more important as time passes (Pearce *et al.* 1992).

These considerations allow us to assess the position of the lherzolitic ophiolites that occur along the Hellenic–Dinaric ophiolite belt. Their structure, which differs from that of the harzburgitic ophiolites, and the MORB-like character of their extrusive rocks raise the possibility that they formed along a slow-spreading MOR, perhaps just before it failed to produce a new subduction (Robertson & Shallo 2000), and when they became situated behind the newly generated trench they were invaded by SSZ magmas (Fig. 5b). Inserguieux-Filippi *et al.* (2000) suggested that a MORB source may continue to rise and melt some time after ridge failure and inception of a new subduction zone. Although this model does not consider slab retreat and assumes a deeply rooted hot rising column beneath the ridge, which may not be realistic, the process envisaged deserves further evaluation. Still, the question arises of why such situations do not occur elsewhere. A more difficult problem is the absence of a resolvable age difference between the western and eastern ophiolites. These questions can be resolved if the western ophiolites formed in a proto-back-arc, i.e. by spreading some distance west of the subduction zone in a setting where now MORB-like magmas sometimes form (Hawkins 1995, 2003), whereas the eastern ones formed closer to the site of subduction. It remains to be seen whether the geochemical data allow such a model. If they do, then it is likely that such a distal part of the subduction system existed in other cases as well, but usually this not preserved in the obducted ophiolites.

The Semail ophiolite reveals additional complexities. On the one hand, the chemical fingerprint of the sheeted dykes and extrusive rocks, especially the younger ones, points to formation in an SSZ setting, and this is supported by the presence of boninitic volcanic rocks and by dykes with a pronounced SSZ character in the mantle tectonites. On the other hand some dykes that intruded the tectonites while they were still hot (1100–1200 °C) were derived from MORB-like magmas, and unlike the cumulates display the crystallization order of MORB (Python & Ceuleneer 2003). Moreover, the Semail ophiolite somewhat resembles the crust of the Hess Deep in the Pacific Ocean (Francheteau *et al.* 1990). These observations raise the question of the temporal and genetic relations between tectonites and the overlying parts of ophiolites, and whether all the parts of ophiolites formed simultaneously. Such a question was also raised regarding the Troodos ophiolite (Thy & Esbensen 1993). Alternatively, one may envisage that the position of the ophiolites relative to the subduction zones changed during their construction.

In summary, an origin of ophiolites in an SSZ setting next to newly formed subduction zones fits their composition and petrography, and provides a framework for their interpretation. In particular, their formation can be related to the geometry and thermal-mechanical processes in the mantle wedges above the subducting slabs, although more needs to be known about the complexities of magma formation in such settings.

Tectonization and obduction stage

The end of magmatic construction marks the transition to a new stage in ophiolite development in which tectonic rather than igneous processes dominated their evolution as they approached the site of future obduction. This involves detachment from the original substrate, underplating by metamorphic soles, and internal deformation. Of these, the metamorphic sole formation is the best known, and probably among the earliest events.

In most cases the metamorphic soles are not crossed by igneous rocks, indicating emplacement after the end of magma addition to the ophiolites. In some cases, such as the Semail and Tauride ophiolites, dykes intrude the sole and the overlying tectonites, but they account for limited and last SSZ magmatic additions to the ophiolites. Because the protoliths of the soles originated in the host basins, their emplacement beneath the ophiolites implies that the latter were detached from their substrate. This could allow displacement of the ophiolites away from the SSZ of magma generation and supply, and thus should follow closely the end of this supply. To extend under the entire width of the up to 10–15 km thick ophiolites, the detachment surfaces should be rather flat (Fig. 4c), but cases such as the Semail ophiolite whose base cuts across the pseudo-stratigraphy (Boudier *et al.* 1988) show that they need not be horizontal. The detachments are thus distinct from the dipping tops of the descending slabs over which the ophiolites originated but are new fractures behind the trench (Fig. 4c), as recognized by Jones *et al.* (1991).

To explain the timing of metamorphism close to the end of ophiolite construction and the inverted metamorphic gradients of the soles, they were interpreted to have formed when the still hot ophiolites were thrust over the adjacent basins where the sole protoliths originated. This is easily envisaged for ophiolites that originated along MORs, provided ridge failure and thrusting begin immediately after ophiolite production (Coleman 1981; Spray 1984). However, the finding that the igneous protoliths of the soles differ from the ophiolite rocks is incompatible with this model, because ridge spreading is expected to be symmetrical. This is an additional strong argument against a MOR origin of the ophiolites considered here (Searle & Cox 1999; Robertson 2002, 2004).

Such a scenario can be envisaged also for ophiolites that originated in an SSZ setting, explaining their displacement away from the area of SSZ magma supply, but it encounters difficulties (e.g. Wakabayashi & Dilek 2000). To metamorphose the basin floor that they override the ophiolites must remain sufficiently hot, which requires very fast thrusting. Hacker *et al.* (1996) and Hacker & Gnos (1997) inferred 20 cm a^{-1} for the Semail ophiolite, which considerably exceeds plate velocities of Neotethys (and most plate velocities in general) and thus is problematic. Another problem is that some soles record pressures corresponding to depths that considerably exceed the thickness of ophiolites. In Albania and Oman pressures reaching 0.9–1.2 GPa and 0.7–1.3 GPa were found, corresponding to depths of 28–37 km and 22–40 km, respectively (Dimo-Lahitte *et al.* 2001; Searle & Cox, 1999). A sole of a Tauride ophiolite records $T > 560$ °C, $P > 8$ kbar, corresponding to a depth of *c.* 30 km or more (Dilek & Whitney 1997). An origin at such depths implies considerable thinning of the ophiolite after sole formation (Wakabayashi & Dilek 2000), but this has not been documented.

Alternatively, the soles were metamorphosed in the subduction zone and then were exhumed and underplated the ophiolites (Fig. 4c), which explains the depth of metamorphism and also dyke intrusion after sole formation. Only material that was subducted during or shortly after ophiolite formation is expected to be exhumed, but not material that subducted earlier and descended to considerable depth, which explains why the metamorphism of the soles occurred later than ophiolite generation (see Wakabayashi & Dilek 2000). The exhumation mechanism remains obscure, however. Thus, both scenarios face dificulties, although the second is perhaps less problematic. Shervais (2001) proposed that the soles formed when the ophiolites overrode a ridge, but in the cases considered here this remains unproven. It is also unlikely that ridges are generally available to be subducted shortly after ophiolite formation, although perhaps this may happen in some places.

In any case, after detachment and sole emplacement the basins between the ophiolites and the margins on which they were obducted were eliminated, which is generally ascribed to continuing subduction and slab retreat (Fig. 4). In the time intervals between the formation of the ophiolite considered here and their obduction (15–20 Ma) subduction at rates of several centimetres per year can consume basinal areas that are a few hundred kilometres wide. Probably, the ophiolites were internally deformed during this stage, as recorded in Oman (Boudier *et al.* 1988; Yanai *et al.* 1990), but insufficient studies in other cases do not allow us to outline a general picture. There is no clear record of the expected magmatism related to the subduction in this stage, though the late calc-alkaline rocks in the east Anatolia belt and the Kannaviou Formation of Cyprus may have originated in this setting. Apparently, the zone behind the ophiolites where such activity could occur is not obducted and its record is lost. During this period the ophiolites should have formed the leading edges of the overriding plates, although no longer above sites of magma generation. As the ophiolites advanced they bulldozed the basin floor and the continental slope, and shed debris into the adjacent basin, all of which became stacked into the allochthonous subduction–accretion complexes (accretionary prisms) at their base (Fig. 4d).

Palaeomagnetic data from the Semail, Troodos and Baer–Bassit ophiolites revealed large (90° and more) rotations on vertical axes. A portion of the Semail ophiolite began to rotate already during the last stages of volcanism, but in the other cases it can only be established that at least in part the rotations took place during emplacement, indicating severe disruption. The causes and mechanism of the rotations may be related to oblique convergence, but further study is required to fully elucidate the process. Because of the lack of palaeomagnatic data from other ophiolites it is not clear whether this is a widespread feature of ophiolite obduction.

As the ophiolites and the underlying allochthonous slices were obducted, their weight is expected to downflex the regions in front of them, producing foreland basins (Fig. 4d). Such foreland basins were documented in front of the Greek and Albanian ophiolites and in front of the north Arabian ophiolites, and over the Oman margin. In the first case only deepening was observed, indicating starved foreland basins, whereas in the latter cases the flexural lows were filled with sediments. As such basins extend some 100 km in front of the tectonic loads, depending on the strength of the flexed plate (Garfunkel & Greiling 2002), their formation over the continental margins records the final approach of the ophiolites. It is noteworthy that the top of a continuous flexed plate slopes down beneath the tectonic load and is not depressed in front of the latter. In such cases the ophiolites have to climb up this slope during obduction, which suggests pushing from behind. However, in reality the situation may be more complex, e.g. along the Oman margin where considerable faulting and deformation took place (Robertson 1987). In addition, there and in western Turkey exhumed high-P metamorphic rocks were emplaced not long before ophiolite obduction, although the timing of exhumation is not well constrained (Okay *et al.* 1998; Searle & Cox 1999; Gray & Gregory 2003; Searle *et al.* 2003). In these cases the development of the continental margins was much more complex than just being overrun by ophiolites and it cannot be described by simple flexure of continuous plates. The question then arises as to whether similar processes operated in other places as well, but are hidden at depth. If this is the case, the obduction process may be considerably more complicated than envisaged in Figure 4d.

In summary, although the scenario outlined here describes the end of ophiolite construction and its subsequent displacement toward the sites of obduction in terms of known processes, the above considerations also reveal several problems. This shows that in some cases at least the processes may be much more complicated than shown in Figure 4.

Post-obduction stage

The ophiolites examined here were obducted over passive margins some time before the final closure of the host basins, which may persist for tens of millions of years. This stage of the host basin history is beyond the scope of the present work. It is only noted that the final continental collision that eliminated their remnants may deform the ophiolites and lead to their thrusting over the colliding blocks, which determines the final structural relations.

Discussion and conclusions

The foregoing considerations highlight some general features of the ophiolites examined here and their relations with the histories of the host basins.

(1) The ophiolites originated in mature basins that probably formed by slow spreading (0.5–3.0 cm a^{-1}). If faster-spreading ridges existed in the basins, then they must have been subducted or were no longer active by the time the ophiolites formed, or else such ridges would have produced areas wider than the basins. Changes in spreading rate could also occur, but there are no direct constraints on this possibility.

(2) The ophiolites formed in intra-oceanic positions by some form of sea-floor spreading. Spreading rates of the ophiolites with harzburgitic tectonites, which are inferred from comparison with oceanic crust, could produce the entire width of the host basins in 10–30 Ma, so they could be maintained for limited periods only. Production of these ophiolites thus signified changes in the evolution of the host basins and could not have been a long-lived process.

(3) The ophiolites examined here occur in belts, and those in each belt formed in short time intervals (up to 5–10 Ma), ophiolite events, shortly after changes of the direction and/or rate of plate motions.

(4) The geochemical signature of the ophiolites with harzburgitic tectonites points at an origin in an SSZ setting. In view of the previous inferences the ophiolite events are therefore thought to signify the formation of new subduction zones following changes of plate motions. These ophiolites formed by spreading behind new trenches as a result of slab retreat, which also influences the dynamic and thermal structure of magma generation in the mantle wedges.

(5) The less common ophiolites with lherzolitic tectonites and with a MORB-like geochemical signature formed near those with an SSZ signature and were invaded by magmas with an SSZ character. They may have formed along ridges just before the latter were disrupted. Alternatively they may have formed alongside the ophiolites with an SSZ signature in a setting analogous to back-arc basins (although there were still no true arcs).

(6) Shortly after formation, the ophiolites were detached from their substrate and displaced from the zone of magma SSZ magma supply. Metamorphic soles were emplaced beneath the detachments, and in some cases were intruded by dykes. The way in which this took place remains problematic. Later the basins between the ophiolites and the margins on which the latter were eventually obducted were consumed by continuing subduction and slab retreat; the ophiolites became the leading edges of the overriding plates, and eventually were thrust over accretionary prisms built of materials scraped off the floor of the consumed basins and margins on which the ophiolites were obducted.

(7) The weight of the approaching ophiolites and the underlying allochthonous units should downflex the margins on which they were emplaced. In simple flexure models the underlying plate slopes basinward, requiring up-slope pushing of the obducted ophiolites. The real situation may be much more complex, involving fracturing of the margin, and in some cases also emplacement of high-*P* metamorphic rocks, which limits the applicability of simple flexure models.

The crucial point in the above considerations is that the harzburgitic ophiolites originated in an SSZ setting, soon after formation of the subduction zones. This is based primarily on petrological evidence from many such ophiolites worldwide (Shervais 2001) and is strongly supported by the analogy with western Pacific forearcs. The above considerations regarding the place of ophiolite formation in the history of the host basins support and fit well into this interpretation, and also explain the remarkable similarity in age of large groups of ophiolites. Such distinct ophiolite events have been recognized also in other parts of the world (Ishiwatari 1994).

This framework allows interpretation of the history and main features of the ophiolites of the type considered here in terms of known processes. However, the above discussion also reveals the need to further clarify many aspects of this history. The way in which new subduction zones form is still not clear. Why is ophiolite construction a short-lived process, probably lasting ≤ 10 Ma? Why are they detached from their substratum and displaced away from the zone of SSZ magma generation, whereas other subduction zones (e.g. western Pacific) follow a different evolutionary trend that leads to construction of island arcs? If the distinctive ophiolite rock-associations normally form in new subduction zones, why are they not commonly found associated with old island arcs? Various aspects of metamorphic sole formation and of the obduction mechanism also need further clarification.

In conclusion, the foregoing considerations show that examination of ophiolites from the standpoint of the histories of the host basins provides significant insights regarding their origin and history, supplementing the wealth of other data regarding ophiolites. This allows us to explain their formation and history in terms of known processes, but also reveals and highlights significant problems that must be clarified before ophiolite histories can be understood.

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