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Tectonophysics 279 (1997) 215–226

TECTONOPHYSICS

The conundrum of samail: explaining the metamorphic history

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Accepted 1 May 1997

Abstract

Recent thermochronology and thermobarometry place new constraints on the tectonic evolution of the Samail ophiolite. Zircon and radiolarian ages indicate formation of the Samail ophiolite crust at 97–94 Ma. Hornblende ages demonstrate that the metamorphic sole beneath the ophiolite formed and, only 1–4 m.y. later, cooled to 550°C. Thermobarometry dictates that the sole formed at peak temperatures of 775–875°C and unexpectedly high pressures of 1.1 GPa. These data imply extremely rapid (~200 km/m.y.) subduction beneath a very young ophiolite for a few m.y. to depths of 30–40 km, followed by equally rapid subduction of cold lithosphere beneath the ophiolite. High-pressure rocks beneath the ophiolite show subduction of the Arabian continental margin to depths as great as 70–80 km, but the temporal and genetic relationships between the metamorphic sole and the high-pressure rocks are poorly known. Key questions remaining are the magnitude, style, and age of extension of the ophiolite, and the timing of high-pressure metamorphic event(s).

Keywords: ophiolite; subduction; metamorphic sole; obduction

1. Introduction

The tectonic process by which enormous (>10,000 km^2), dense ophiolite sheets are emplaced from ocean basins onto low-density, subaerial continents has long interested geologists. Accepting that the emplacement of an ophiolite onto a continental margin results from subduction of the continental margin beneath the ophiolite (e.g., Nicolas, 1989), significant questions still remain. (1) Where in the ocean basin did intraoceanic thrusting begin? (2) How old was the ophiolite at the time of intraoceanic thrusting? (3) How far beneath the ophiolite was oceanic lithosphere subducted? (4) At what rate did the intraoceanic thrusting occur? (5) How old was the ophiolite at the time of continental subduction? (6) How far beneath the ophiolite was the continental margin subducted? (7) At what rate was the continental margin subducted? One might expect that oceanic lithosphere of any age could be emplaced as an ophiolite and that ophiolites are slivers of only the upper 10–15 km of oceanic lithosphere. Recently acquired petrologic and geochronologic data indicate that the Samail ophiolite, the best-exposed, largest, least-deformed, and perhaps most-studied ophiolite in the world (Fig. 1), was less than a few m.y. old when emplacement began, and rests upon emplacement-related high-pressure rocks exhumed from depths of 70–80 km.

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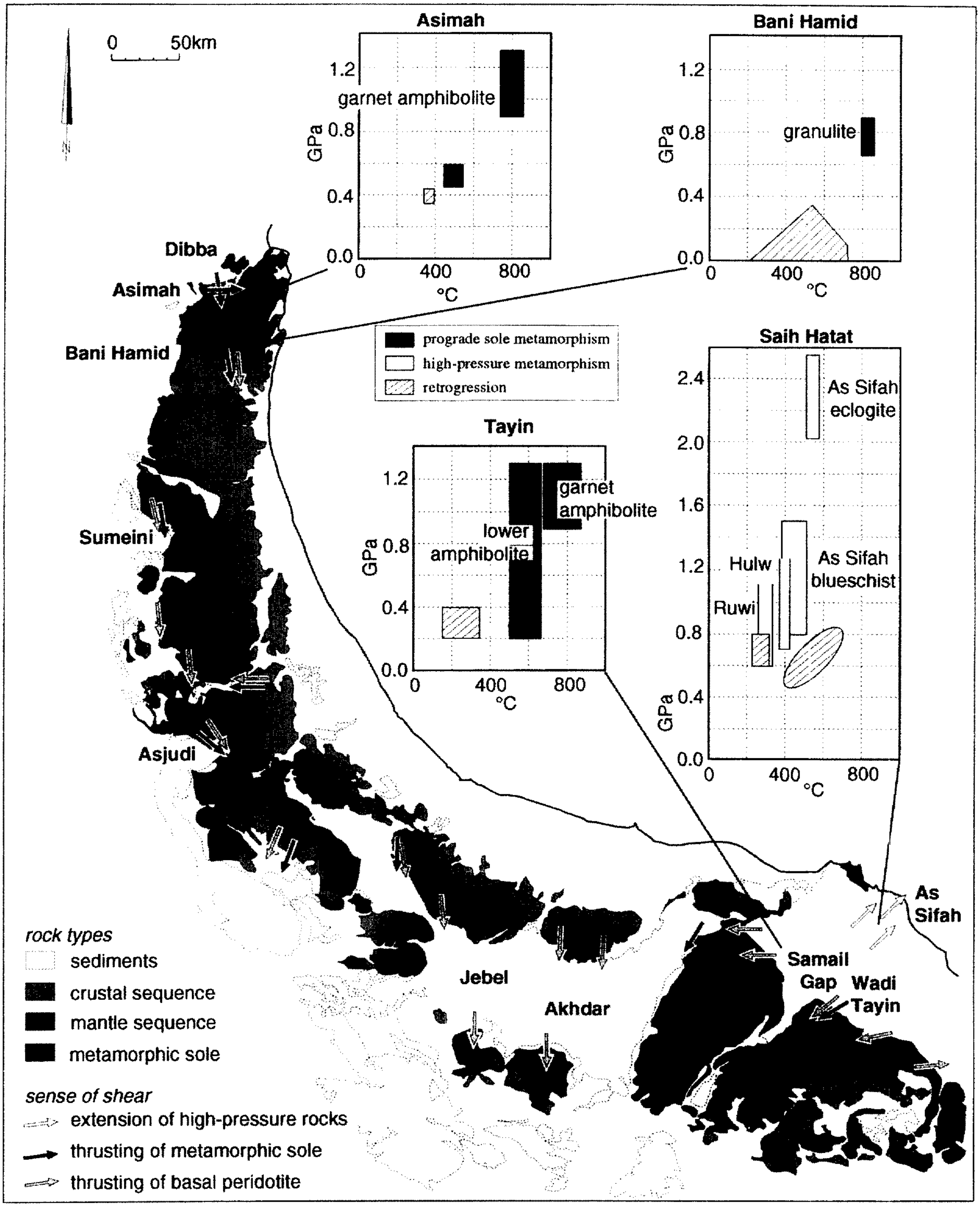
PII S0040-1951(97)00114-5


Asimah
Bani Hamid
0 50km
1.2
garnet amphibolite
GPa 0.8
0.4
0.0
0 400 800
°C
granulite
Dibba
Asimah
Bani Hamid
prograde sole metamorphism
high-pressure metamorphism
retrogression
Saih Hatat
2.4
As Sifah eclogite
2.0
1.6
Tayin
garnet amphibolite
lower amphibolite
Hulw
As Sifah blueschist
Ruwi
Sumeini
Asjudi
As Sifah
Samail Gap
Wadi Tayin
Jebel
Akhdar
rock types
sediments
crustal sequence
mantle sequence
metamorphic sole
sense of shear
extension of high-pressure rocks
thrusting of metamorphic sole
thrusting of basal peridotite

2. Igneous crustal rocks

The tectonic setting of the Samail ophiolite is somewhat uncertain. Lower (V1) and middle (V2) lavas have been interpreted as transitional between mid-ocean-ridge and intraoceanic arc lavas (Pearce et al., 1981; Lippard et al., 1986), or, alternatively, as the products of normal mid-ocean-ridge spreading and subsequent melt extraction during intraoceanic thrusting (Ernewein et al., 1988; Pflumio, 1991). Plutonic rocks comprise an early gabbroic series and a late wehrlitic series. The early high-level gabbro–diorite–plagiogranite suite is interpreted as the source of the sheeted dike complex and the lower extrusive unit (Lippard et al., 1986; Nicolas and Boudier, 1991). The wehrlitic series is dominantly wehrlite–dunite–troctolite, with lesser amounts of gabbro, plagiogranite, and granite. The late intrusive complexes are genetically related to the middle extrusive sequence, V2, and are inferred to have intruded during the initial intraoceanic thrusting stage of ophiolite emplacement (Juteau et al., 1988).

Crystallization ages of the Samail crustal section are represented by thirteen near-concordant U/Pb zircon ages (Tilton et al., 1981) from plagiogranites. Most of the plagiogranites were interpreted as late-stage differentiates of the early gabbroic series, but one was inferred to be part of the later wehrlitic suite. Zircons from both suites overlap in age, supporting the suggestion of Juteau et al. (1988) that the wehrlitic series intruded prior to complete solidification of the gabbroic series. The zircon ages range from 97.3 to 93.5 $\pm$ 0.25 Ma (1 σ) (Fig. 2). Ten of them are restricted to a rather narrow interval of 95.4 to 94.5 Ma, with a mean age of 94.8 $\pm$ 0.1 Ma. Radiolaria in the lower V1 volcanics are Cenomanian, and foraminifera in the middle V2 volcanics are Cenomanian to Turonian (Tippit et al., 1981; Schaaf and Thomas, 1986; Beurrier et al., 1987). In the most recent rendition of the Cretaceous timescale (Gradstein et al., 1994) the Cenomanian and Turonian span 98.9–93.5 Ma and 93.5–89.0 Ma, respectively. Thus, the U/Pb plagiogranite ages are concordant, within uncertainty, with the fossil ages of the lower and middle volcanic rocks of the ophiolite.

Recent $^{40}Ar/^{39}Ar$ dating (Hacker, 1994; Hacker et al., 1997) demonstrated that the crustal section cooled at a rate normal for oceanic crust. Hornblende ages from crustal plagiogranites, gabbros, and from veins in peridotite, range from 93.6 $\pm$ 0.5 Ma to 96.3 $\pm$ 1.3 Ma with a mean age of 94.4 $\pm$ 0.3 Ma (Fig. 2). Granitic to dioritic stocks and dikes of the wehrlitic series that intruded upper mantle rocks, yielded hornblende ages of 93.3 to 94.1 $\pm$ 0.5 Ma with a mean of 93.8 $\pm$ 0.3 Ma. If the plagiogranites crystallized at ~700°C at 94.8 Ma and the hornblendes in the crustal section closed to Ar diffusion at ~550°C at 93.8 Ma, the cooling rate in the crust was ~150°C/y. This is compatible with the thermal model of Morton and Sleep (1985) of ocean ridges that considered convective hydrothermal cooling; they calculated that the upper levels of oceanic magma chambers in an East Pacific Rise-type spreading center cool below 550°C in 0.3–1.0 m.y.

3. Metamorphic sole

The Samail ophiolite was thrust over adjacent oceanic lithosphere and then onto the Arabian craton along a several-hundred-meters thick shear zone, or 'metamorphic sole', composed of peridotite mylonite overlying partially melted granulite-facies mafic rock, amphibolite, and greenschist facies sedimentary and basaltic rock (Searle and Malpas, 1980; Ghent and Stout, 1981; Boudier et al., 1988; Bucher, 1991). Immediately underlying are relatively unmetamorphosed sedimentary and volcanic rocks interpreted as pelagic, slope, rise, and shelf deposits from a Late Permian to Late Cretaceous passive margin and ocean basin that lay northeast of the Arabian craton (e.g., Glennie et al., 1974; Lippard et al., 1986; Robertson et al., 1990). The amphibolite-facies rocks are believed to represent oceanic

Fig. 1. Map of the Samail ophiolite. Arrows show motion of upper plate during deformation of basal portion of peridotite tectonite, metamorphic sole, and high-pressure rocks. P–T diagrams show inferred pressures and temperatures of metamorphism from the Asimah and Tayin localities of the metamorphic sole, from the metamorphic rocks included within peridotite at Bani Hamid, and from the Saih Hatat high-pressure area.

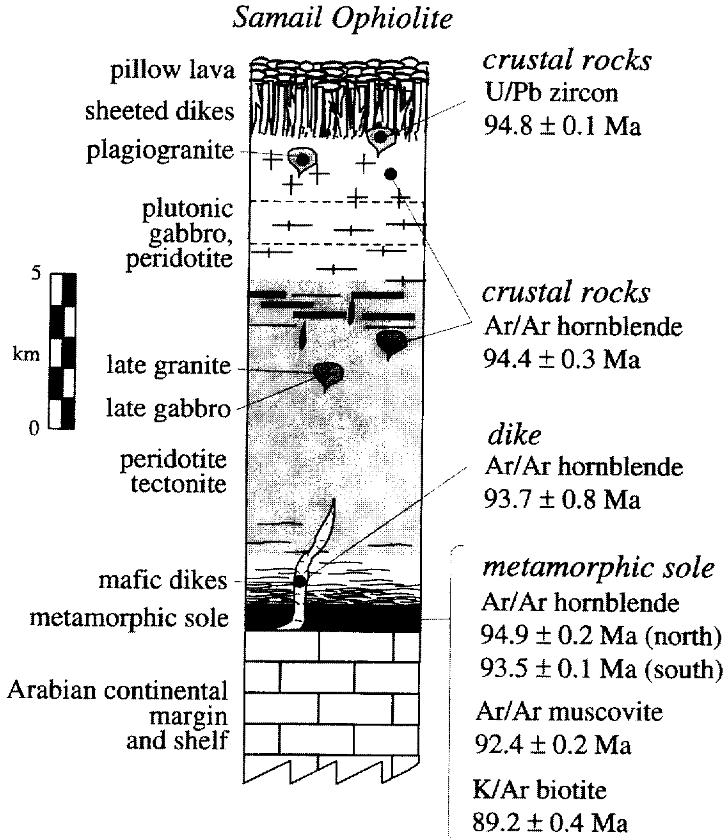


Fig. 2. Pseudostratigraphy of the Samail ophiolite showing average crystallization and cooling ages for the igneous crust and metamorphic sole. Zircon ages are U/Pb ages from Tilton et al. (1981), hornblende and muscovite ages are $^{40}Ar/^{39}Ar$ ages from Hacker et al. (1997), and biotite ages are K/Ar ages with uncertainties <1 m.y. from Gnos and Peters (1993).

crust overridden during the early, intraoceanic thrusting stage of emplacement, whereas the lower-grade rocks are inferred to be basalt, clastic sediment, and chert overridden at a later stage (Searle and Malpas, 1980).

The metamorphic sole has been extensively investigated in three, now classic, areas: Tayin, Sumeini, and Asimah (Fig. 1). The bulk of each section is composed of gneissic to schistose amphibolites that are dominated by intermediate-composition plagioclase + tschermakite to magnesio-hornblende. High structural levels at Tayin and Sumeini contain andesine–labradorite + diopside + pargasitic–hastingsitic hornblende, with rare pods of quenched partial melt; almandine-rich garnet is also present within a few meters of the overlying peridotite. Common to all the amphibolite-facies rocks is a high-temperature fabric defined by recrystallized hornblende and plagioclase (Boudier et al., 1985; Hacker and Mosenfelder, 1997).

Garnet–clinopyroxene thermometry indicates that the highest temperatures attained were 775–875°C (Ghent and Stout, 1981; Searle and Malpas, 1982; Hacker and Mosenfelder, 1997). The dearth of aluminum silicates and quartz has hampered estimates of metamorphic pressure in these high-variance assemblages. Early barometric calculations based on the jadeite component in clinopyroxene (Ghent and Stout, 1981) yielded imprecise estimates of 0.5–0.6 GPa. The recent discovery of kyanite in garnet amphibolites in the Asimah area (Gnos, 1997) provides an exciting new indication that pressures were twice that high. Multi-equilibrium calculations and element partitioning between clinopyroxene and garnet or orthopyroxene indicate peak temperatures of 800 ± 50°C for two-pyroxene rocks and garnet amphibolite at Asimah (Gnos and Kurz, 1994; Gnos, 1997). At such temperatures, the presence of kyanite constrains pressures to have been $\geq 1.1 \pm 0.2$ GPa (e.g., Bohlen et al., 1991); multi-equilibrium calculations based on coexisting phases in garnet amphibolites from the Asimah sole indicate equivalent metamorphic pressures of 1.0–1.2 GPa. Pressures of 1.1–1.5 GPa were calculated by Gnos (1997) for garnet amphibolites from Tayin, using compositions published by Ghent and Stout (1981). The absence of kyanite from sole localities other than Asimah is thus likely the result of a lack of appropriately pelitic layers. Prograde greenschist-facies rocks crop out beneath amphibolite in several areas (Searle, 1980). In the Asimah area, prograde greenschist-facies rocks comprise 1–1.5 km of metamorphic gabbro, alkaline volcanic rocks, quartz-rich metasedimentary rocks, and minor marble. Peak metamorphic temperatures of 450–550°C are indicated by magnesio-hornblende + albite assemblages in the metagabbroic rocks (Bucher, 1991).

An unusual enclave of metamorphic rocks crops out in the Bani Hamid area (Fig. 1). These rocks are entirely enclosed within peridotite, perhaps as a thrust imbricate, and lie 1.5–2 km structurally above the Asimah section of the metamorphic sole (Gnos and Nicolas, 1996). A 2-km-thick section consists dominantly of quartzose granulite-facies rocks with less mafic and carbonate rock (Gnos and Kurz, 1994). The occurrence of ferrian sapphirine + quartz, spinel + quartz, cordierite, and sillimanite, combined with two-pyroxene thermometry, suggests peak metamorphic conditions of 800–850°C and 0.65–0.9 GPa (Gnos and Kurz, 1994). Metamorphic pressures are lower at Bani Hamid than at Asimah, as expected from the structurally higher position of the former.

All these metamorphic sole localities show a weak greenschist-facies or lower-grade metamorphic overprint. For example, the metagabbroic low-rank amphibolite facies rocks of the Asimah area (450–550°C) are overprinted by actinolite–actinolitic hornblende + albite + epidote grown at 340–380°C, 0.3–0.5 GPa (Bucher, 1991). The common greenschist-facies assemblage in the mafic rocks at Wadi Tayin is albite + actinolite + K-feldspar + epidote + K-white mica + chlorite + sphene ± hematite ± pumpellyite ± calcite, formed at pressures of $\geq$0.3 GPa and temperatures of 150–350°C (Hacker and Mosenfelder, 1997).

Early estimates of the time of metamorphic sole formation were based on K/Ar dating (Allemann and Peters, 1972; Searle et al., 1980; Lanphere, 1981; Boudier et al., 1985; Lippard et al., 1986; Montigny et al., 1988; Gnos and Peters, 1993), which produced a 12 m.y. spread in hornblende ages, including ages older than the zircon ages of the oceanic crust. Recent $^{40}Ar/^{39}Ar$ dating of the metamorphic sole (Hacker, 1994; Hacker et al., 1997) has demonstrated a much narrower range of more precise ages. Most significantly, the hornblende $^{40}Ar/^{39}Ar$ ages postdate crustal zircon ages by only 0–4 m.y. (Fig. 2). Hornblendes from Tayin yielded $^{40}Ar/^{39}Ar$ ages that range from 92.6±0.6 to 94.9±0.5 Ma, with a mean of 93.5±0.10 Ma. Hornblendes from sole localities in the northern part of the ophiolite (chiefly Sumeini and Asimah) span 92.6 ± 0.6 to 95.7 ± 0.3 Ma, with a mean of 94.9±0.2 Ma. The $^{40}Ar/^{39}Ar$ hornblende ages suggest that K/Ar hornblende ages younger than ~92.5 Ma reflect Ar loss and that K/Ar ages older than ~95.5 Ma result from excess ^{40}Ar. Muscovite $^{40}Ar/^{39}Ar$ ages from the metamorphic sole range from 90.9 to 93.4 ± 0.3 Ma, and have a weighted mean age of 92.4 ± 0.2 Ma. Published K/Ar muscovite ages (Allemann and Peters, 1972; Gnos and Peters, 1993) are concordant with the $^{40}Ar/^{39}Ar$ ages, except for a few outliers. K/Ar biotite ages span 85.5±5 to 92.5±4 Ma (Allemann and Peters, 1972; Searle et al., 1980; Gnos and Peters, 1993) with a weighted mean of 89.2 ± 0.4 Ma. The hornblende, muscovite, and biotite ages from the metamorphic sole are successively younger, as expected from their closure temperatures of 550°C, 475°C, and 450°C, respectively (Hacker et al., 1997). Hacker et al. (1997) also measured a $^{40}Ar/^{39}Ar$ hornblende age of 93.7±0.8 Ma for a diabase dike that cuts amphibolite in the Tayin section, postdates the amphibolite-facies metamorphism and deformation of the sole, and predates minor greenschist-facies metamorphism and deformation.

4. High-pressure rocks

In the Saih Hatat area (Fig. 1), units of the Paleozoic passive margin sequence and Cretaceous synorogenic sediments were metamorphosed at relatively high pressures and low temperatures. From structurally highest to lowest they are: the Ruwi unit, which contains lawsonite and carpholite formed at 250–320°C and >0.6 GPa; the Yiti unit, with crossite + epidote and carpholite formed at transitional greenschist–blueschist facies conditions of 250–400°C and >0.6 GPa; the Wadi Mayh and Hulw units, containing carpholite and crossite + epidote grown at 360–420°C and >0.7 GPa; and the As Sifah unit, with glaucophane + garnet schists recrystallized at 380–500°C and 0.8–1.5 GPa, enclosing eclogite pods that reached peak temperatures of 500–580°C (Goffé et al., 1988; El-Shazly et al., 1990; Searle et al., 1994). Peak pressures experienced by the eclogites have been estimated as 2.3 ± 0.25 GPa from multi-equilibrium calculations (Searle et al., 1994) and modeling of cracking of garnets with quartz inclusions (Wendt et al., 1993).

The Wadi Mayh and Hulw units are overprinted by greenschist-facies albite + epidote assemblages grown at 250–330°C and 0.6–0.8 GPa, and the As Sifah unit shows post-kinematic albite + actinolite + epidote grown at 550 ± 100°C, 0.65 ± 0.15 GPa during N–S extension (Searle et al., 1994). These pressures and temperatures are similar to those of the greenschist-facies overprint of the metamorphic sole (Fig. 1). There is considerable disagreement about the P–T path followed by the high-pressure rocks; Goffé et al. (1988) show counterclockwise paths for all the units except one, while El-Shazly et al. (1990) and Michard et al. (1994) show clockwise paths for the same units. Searle et al. (1994) also show a clockwise evolution for the eclogite-facies rocks, while Wendt et al. (1993) argued that the decompression path deviated only slightly from the compression path.

There are indications that the amphibolite-facies metamorphism of the sole thrust preceded high-pressure metamorphism. The lawsonite- and carpholite-bearing Ruwi unit is a mélange that contains amphibolite blocks that may have been derived from the metamorphic sole (Searle et al., 1994). A few scattered localities in the metamorphic sole contain minerals that may have formed during high-pressure metamorphism, but also are compatible with recrystallization at the modest pressures that would have been produced by the overlying ophiolite nappe. For instance, magnesio-riebeckite and stilpnomelane were reported from the Dibba zone (Fig. 1) (Allemann and Peters, 1972; Gnos and Kurz, 1994), winchite and crossite was found at Samail Gap (Rabu, 1987), and pumpellyite was found at Tayin (Hacker and Mosenfelder, 1997).

The age of the high-pressure metamorphism has been a matter of debate for years, with various workers preferring one or two metamorphic events at roughly 80 or 110 Ma. Most white-mica ages on the high-pressure rocks span an enormous range from 112 to 72 Ma, but there are additional more extreme outliers. The $^{40}Ar/^{39}Ar$ white-mica spectra reported by Montigny et al. (1988) are hump-shaped and, by analogy with other studies of high-pressure rocks that revealed Ar spectra influenced by excess ^{40}Ar or incompletely degassed grains (Hacker and Wang, 1995; Hannula and McWilliams, 1995; Ruffet et al., 1995), may not represent an actual geologic event. The internal discordance of Montigny et al.'s $^{40}Ar/^{39}Ar$ spectra indicates that their K/Ar ages, as well as those of Lippard (1983), also cannot be relied upon to indicate the time of the high-pressure metamorphism. Likewise, none of the spectra produced by El-Shazly and Lanphere (1992) yielded plateau ages that include more than two steps and >50% of the released ^{39}Ar; therefore we also reject them. Searle et al. (1994) have published the only $^{40}Ar/^{39}Ar$ white-mica age from eclogite that yielded a plateau, of 96 ± 2 Ma. Because every other sample of these high-pressure rocks has yielded uninterpretable $^{40}Ar/^{39}Ar$ spectra (including the other two published by Searle et al., 1994), we consider this age suspect as well.

Metamorphism of the high-pressure units must have occurred before 65 Ma when Paleocene limestones were deposited on top (Bechennec et al., 1990), and good constraints can be placed on the end of the emplacement process, which is signaled by the first appearance of igneous detritus on the Arabian craton in the mid- to late-Campanian (78–71 Ma) Juweiza Formation (Warburton et al., 1990; Rabu, 1993). Goffé et al. (1988) and Searle et al. (1994) correlated the lowest-grade high-pressure rocks, the Ruwi unit, with the Coniacian–Santonian (89.0–83.5 Ma, Gradstein et al., 1994) Muti Formation (Rabu et al., 1990) exposed farther west around Jebel Akhdar. This correlation requires that metamorphism of the lowest-grade units was no earlier than 89 Ma. M.P. Searle (pers. commun., 1996) later proposed instead that the Ruwi is more likely the equivalent of the Triassic–Cretaceous Haybi mélange. Zircon fission-track dating (Saddiqi et al., 1995) showed that the low-grade units (e.g., Ruwi) cooled below ~210°C at 80 ± 8 Ma, whereas the As Sifah unit remained above this temperature until 68 ± 2 Ma.

5. Structural evolution

Structural studies (Boudier et al., 1985, 1988; Michard et al., 1989, 1994; Gnos and Nicolas, 1996) have revealed the kinematic histories of the metamorphic sole and high-pressure rocks (Fig. 1). Sense-of-shear indicators in the metamorphic sole indicate chiefly S-directed thrusting in the bulk of the ophiolite and W-directed thrusting in the southernmost massifs (Boudier et al., 1985). Boudier et al. (1988) interpreted these data to indicate southward (ridge-parallel) thrusting followed by westward (ridge-orthogonal) thrusting. $^{40}Ar/^{39}Ar$ measurements can be interpreted to show that the ridge-parallel thrusting was from 95.7 to 93.5 Ma and the ridge-orthogonal motion was from 94.9 to 92.6 Ma (Hacker et al., 1997). The structurally highest high-pressure units (Ruwi, Yiti, Wadi Mayh) also have limited microstructures or outcrop-scale fold–fault relationships indicating S-directed thrusting (Michard et al., 1984; Searle et al., 1994). Shear bands within and pressure gaps between the high-pressure units have been interpreted to reflect late down-to-the-north extension that accompanied and postdated the greenschist-facies overprint (Michard et al., 1994; Searle et al., 1994).

6. Discussion

The data and interpretations presented above powerfully constrain parts of the tectonic evolution of the Samail ophiolite yet leave other parts ambiguous. It has been known for 15 years that peak temperatures in the metamorphic sole reached ~800°C (Searle, 1980; Ghent and Stout, 1981), and thermomechanical simulations have shown that the ophiolite must have been less than a few million years old at the time intraoceanic thrusting began (Hacker, 1990, 1991). The only places in oceanic lithosphere where intraoceanic thrusting might have occurred at such high temperatures are spreading centers (Boudier et al., 1982), magmatic arcs (Pearce et al., 1981; Lippard et al., 1986) and hot spots. The compositions of the ophiolite extrusives rule out the possibility of a hot spot. Trace-element compositions of the middle lavas support genesis within an arc, and the lack of volcaniclastic rocks indicates that the arc must have been very immature. Trace-element concentrations suggest that the amphibolite-facies rocks of the metamorphic sole were not derived from Samail oceanic crust, which rules out a spreading center — where the upper and lower plates would be the same (Searle and Malpas, 1982). All these features can be reconciled if intraoceanic thrusting began at a transform fault separating <2-m.y.-old lithosphere (the Samail ophiolite) from older oceanic lithosphere, and if the middle lavas were produced when oceanic crust thrust beneath the ophiolite either melted or provided volatiles for melting of asthenosphere (Boudier et al., 1988). A transform fault is the most favorable place for thrusting in oceanic lithosphere because it juxtaposes lithosphere of different buoyancy, elevation, and temperature across a weak zone (Casey and Dewey, 1984). Thrusting across a transform fault could have produced the early ridge-parallel thrusting that Boudier et al. (1988) inferred. Structures within peridotite have been used to argue for a transform fault in the area of Wadi Tayin (Ceuleneer et al., 1988; Nicolas et al., 1988).

The recent discovery that rocks in the Asimah and Tayin localities of the metamorphic sole record pressures of 1.1 ± 0.2 GPa requires that the sole formed at 35–40 km depth (Gnos, 1997). $^{40}Ar/^{39}Ar$ hornblende ages from the metamorphic sole and a cross-cutting dike indicate that the characteristic amphibolite-facies deformation and metamorphism of the sole was complete by ~93.5 Ma (Hacker, 1994; Hacker et al., 1997). If the plagiogranite zircon ages of ~97.3 to 93.5 ± 0.25 Ma date the formation of Samail oceanic crust, the hornblendes in the sole cooled below 550°C in 0–4 m.y after the formation of the crust.

The only key piece of information missing is the timing of the high-pressure event(s). The most credible age for the eclogites is 96 ± 2 Ma on phengite (Searle et al., 1994), coeval with cooling of the metamorphic sole, but this remains unduplicated.

If the low-grade Ruwi unit is correlative with the Muti Formation (see discussion above), the age of the high-pressure metamorphism in that unit is constrained to postdate ~89 Ma (Rabu et al., 1990).

7. Tectonic model

Crustal zircon ages and radiolarian cherts overlying the ophiolite indicate that Samail ocean crust existed by the Cenomanian (98.9–93.5 Ma). Subsequently, intraoceanic thrusting must have underplated oceanic crust beneath the ophiolite to depths of 30–40 km (Fig. 3, (1)). All previous models have assumed subduction of the sole to depths of 15–20 km, consonant with the reconstructed thickness of the ophiolite. The record of 1.1 ± 0.2 GPa metamorphism in the sole requires that at 94 Ma (1) another 15–20 km thick thrust sheet lay on top of the ophiolite (Nicolas, 1989), (2) the ophiolite was a single thrust sheet 30–40 km thick and the sole rocks were exhumed by normal faulting to their present position beneath a thinner section of the ophiolite, or (3) the ophiolite was a single thrust sheet 30–40 km thick and has subsequently been thinned during extension.

All of these hypotheses have shortcomings. There is no evidence of burial of the crustal section of the ophiolite to 15–20 km, so the first possibility can be rejected. The second possibility is also not feasible because the amphibolite-facies rocks of the metamorphic sole and overlying basal harzburgites are intruded by mafic dikes, one of which yielded a hornblende $^{40}Ar/^{39}Ar$ age of 93.7 ± 0.8 Ma (Hacker et al., 1997). This indicates that the harzburgite/sole contact cannot have been the normal fault responsible for thinning. The third possibility implies that the mantle section has been thinned from 35–40 km to 8–13 km, or ~300%. Casey and Dewey (1984) suggested that the mantle section of an ophiolite may be thinned during intraoceanic subduction if deeper-level underplating causes extensional collapse of the overlying ophiolite; however, the continuity of mantle structures observed in Oman (Ceuleneer et al., 1988) may rule out intra-ophiolitic extension of this scale. Determining the age, style, and magnitude of extension in the ophiolite is thus a fertile area for research.

If thrusting began at 95 Ma at 200 km/m.y., the metamorphic sole could have been established across the 200-km base of the ophiolite by 94 Ma. By 93.8 Ma, the same rocks had cooled to <550°C. If thrusting of cold material beneath the sole was responsible for the cooling (Hacker, 1990), the thrusting must have occurred at 200 km/m.y. for the entire metamorphic sole to have cooled by 93 Ma (Fig. 3, (2)). Thrusting rates any slower than this do not meet the tight constraints provided by the zircon and hornblende ages (Hacker et al., 1997).

Searle et al. (1994) have written that "almost all the geochronology ... suggests that the HP metamorphism was occurring at ... 95–90 Ma". If the eclogite formation was coeval with cooling of the metamorphic sole — and the strongest data in favor of this is a muscovite $^{40}Ar/^{39}Ar$ age of 96 ± 2 Ma, which we indicated above should be considered suspect until more data are available — this is an important constraint. The peak temperatures of 500–580°C inferred for the As Sifah eclogites are equivalent to temperatures inferred for the metamorphic sole at 94 Ma, based on $^{40}Ar/^{39}Ar$ hornblende ages. However, the 2.3 GPa pressure inferred for the eclogites (Wendt et al., 1993; Searle et al., 1994), is double that inferred for the sole. If the eclogites formed at 70–80 km depth in the same subduction zone where the metamorphic sole was forming at 30–40 km they cannot have had the same peak temperature. Moreover, the eclogites indicate that the subducting plate was continental material at 70–80 km depth, so it is unlikely that the subducting plate was oceanic material at shallower depth. Perhaps there was a single subduction zone in which the metamorphic sole formed first and, after significant cooling of the subduction zone by subduction of colder material, the eclogites formed second (Fig. 3 (3A)). Or, possibly the metamorphic sole formed in an intraoceanic subduction zone and the eclogites formed in a second subduction zone closer to the Arabian craton (Fig. 3 (3B)). Both of these scenarios are compatible with existing data.

If, on the other hand, the lowest-grade high-pressure rocks had a Coniacian–Santonian protolith (Goffé et al., 1988; Searle et al., 1994), some or all of the high-pressure metamorphism can be no older than 83.5–89 Ma. Michard et al. (1994) proposed that all the high-pressure rocks formed at ~80 Ma by subduction of the continental margin beneath the ophiolite. To achieve the 70–80 km of burial re-

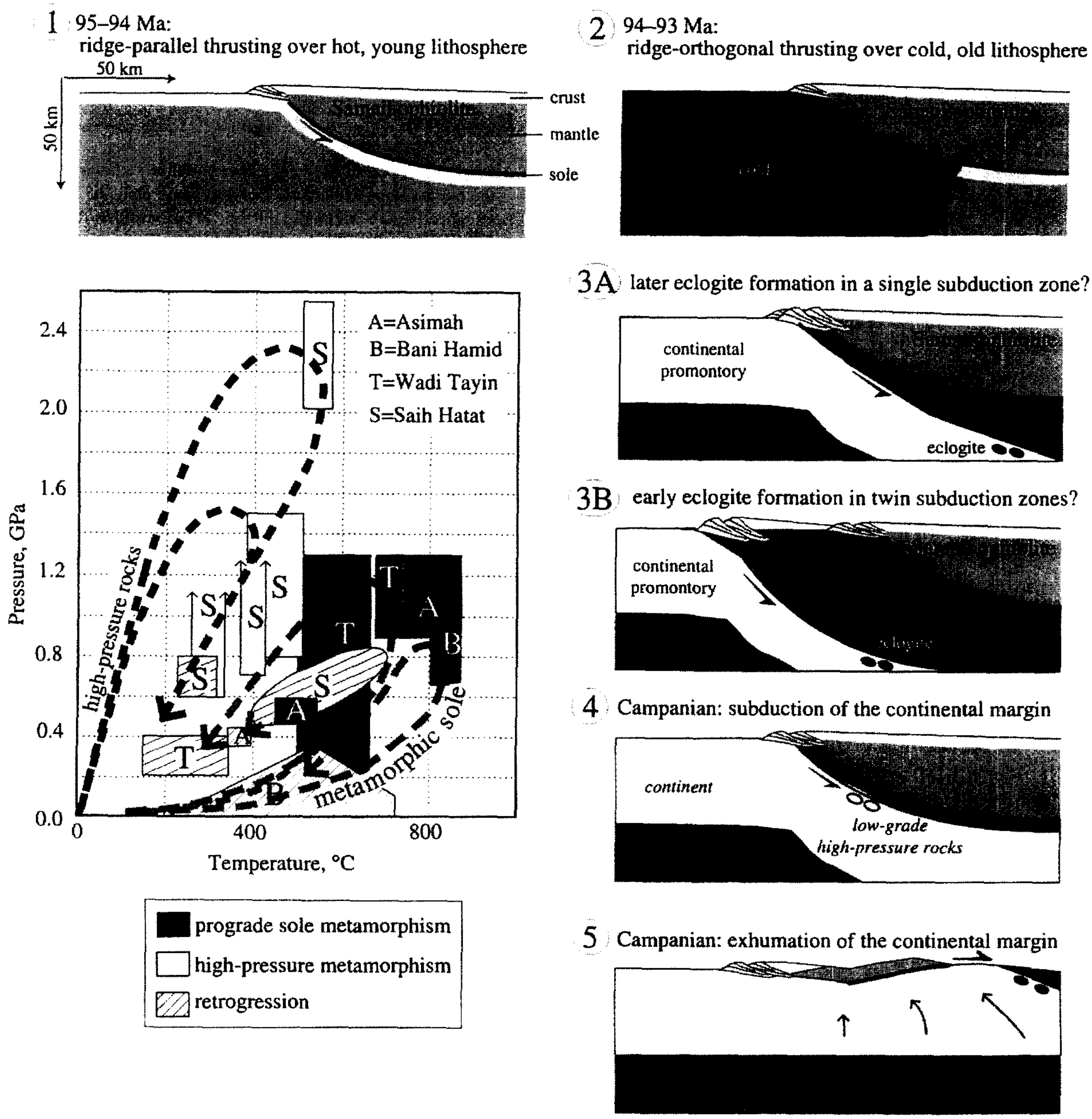


Fig. 3. Pressure–temperature paths and schematic tectonic history of the Samail ophiolite. Counterclockwise P–T paths for the metamorphic sole were calculated by Hacker (1991). (1) Thrusting of hot, young lithosphere beneath the ophiolite is required by the peak temperatures of the metamorphic sole. (2) Thrusting of cold, old lithosphere beneath the ophiolite is required by the $^{40}Ar/^{39}Ar$ hornblende ages of the metamorphic sole; the As Sifah eclogites may have formed at this time. (3A) Single subduction zone in which metamorphic sole formed first and, after significant cooling of the subduction zone by subduction of colder material, the eclogites formed second. (3B) Metamorphic sole formed in an intraoceanic subduction zone and the eclogites formed in a subduction zone closer to the Arabian craton. (4) The ophiolite arrived on the continent in the Campanian, as indicated by the sedimentary record. This may not be the first time of continental subduction beneath the ophiolite, but the lowest-grade high-pressure rocks, at least, formed at this time. (5) The subducted continental margin was exhumed.

quired by the eclogites they suggested that a second continental or oceanic thrust sheet was sandwiched between the upper, Samail ophiolite nappe and the subducted cratonal margin. Searle et al. (1994) proposed a similar model in which the eclogites were buried to 70–80 km by subduction of the continental margin along a second, deeper thrust beneath the ophiolite emplacement thrust. In either case, the shape of the continental margin may have been quite irregular; the material that formed the As Sifah eclogites may have been a Florida-like promontory subducted long before the rest of the continental margin.

Further progress discerning the tectonic history of the Samail ophiolite clearly rests upon determining the age(s) of the higher-grade high-pressure metamorphic units, the age(s) of the greenschist-facies overprint(s) of the metamorphic sole and the high-pressure rocks, and the style, magnitude and age of extension of the mantle section.

8. Conclusions

The Samail ophiolite originated by intraoceanic thrusting in oceanic lithosphere less than a few million years old. The intraoceanic thrust most likely was a transform fault juxtaposing Samail lithosphere with older lithosphere. Oceanic lithosphere was subducted to a depth of 30–40 km and accreted beneath the ophiolite as the metamorphic sole. How the metamorphic sole was exhumed to its present position beneath a much thinner section of oceanic lithosphere is undetermined, but could be addressed by a structural study of the extension of the Samail ophiolite. The Arabian continental margin was subducted 70–80 km beneath the ophiolite. Obtaining information about the timing of this high-pressure metamorphism relative to the intraoceanic thrusting will significantly constrain the tectonic process that placed ophiolite onto the Arabian continental margin.

Acknowledgements

Adolphe Nicolas and Françoise Boudier have been the premier force working for the last 15 years to understand the Samail ophiolite in the framework of an oceanic spreading center and it gives us great pleasure to make this small contribution in recognition of their achievements. Helpful reviews were given by André Michard and Rebecca Jamieson.

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