A case of obduction related high P low T metamorphism in upper crustal nappes, Arabian continental margin, Oman: P–T paths and kinematic interpretation

ARTICLE *in* TECTONOPHYSICS · SEPTEMBER 1988

Impact Factor: 2.87 · DOI: 10.1016/0040-1951(88)90253-3

4 AUTHORS, INCLUDING:

Bruno Goffé
Centre Européen de Recherche et d'E…

André Michard
Université Paris-Sud 11

Tectonophysics, 151 (1988) 363–386
Elsevier Science Publishers B.V., Amsterdam – Printed in The Netherlands

A case of obduction-related high-pressure, low-temperature metamorphism in upper crustal nappes, Arabian continental margin, Oman: *P–T* paths and kinematic interpretation

BRUNO GOFFÉ [1], ANDRÉ MICHARD [2*], JEAN ROBERT KIENAST [3] and OLIVIER LE MER [2**]

[1] *ER 224 du CNRS, Département de Géologie, Ecole Normale Supérieure, 46 rue d'Ulm, 75230 Paris Cedex (France)*
[2] *ERA 887 du CNRS, Laboratoire de Géologie Structurale, 1 rue Blessig, 67084 Strasbourg Cedex (France)*
[3] *ERA 727 du CNRS, Laboratoire de Pétrologie Métamorphique, Université Pierre et Marie Curie, 4 place Jussieu, 75230 Paris Cedex 05 (France)*

(Received October 14, 1987; revised version accepted December 11, 1987)

Abstract

Goffé, B., Michard, A., Kienast, J.R. and Le Mer, O., 1988. A case of obduction-related high-pressure, low-temperature metamorphism in upper crustal nappes, Arabian continental margin, Oman: *P–T* paths and kinematic interpretation. In: F. Boudier and A. Nicolas (Editors), The Ophiolites of Oman. *Tectonophysics*, 151: 363–386.

In the southeastern Oman Mountains, the Saih Hatat tectonic window reveals Arabian continental material beneath the oceanic (sensu. lato) nappes, i.e., the Hawasina sedimentary units and the huge Sumail ophiolite. The northeastern (internal) half of the window suffered an intense deformation associated with a high-pressure, low-temperature (H*P*–L*T*) metamorphism, approximately dated at 80–75 Ma according to the available stratigraphic data. After the deformation, due to the absence of collision, the obduction-inherited structural and metamorphic pattern was virtually unmodified.

A metamorphic map is established based on mineral assemblages in metapelites and metabasites. Low-grade assemblages occur in the highest units (Muscat nappes, Quryat unit). Characteristic minerals are Fe/Mg carpholite ($\pm$lawsonite) associated with kaolinite or pyrophyllite and locally sudoite in the metapelites while lawsonite and blue amphiboles are found in the metabasites. High-grade rocks outcrop in the deepest and easternmost units (As Sifah antiform) and are typified by an epidote–glaucophane-bearing eclogite assemblage. Intermediate units (Hulw) show assemblages of intermediate grade (chloritoid schists and glaucophane-bearing metabasites). Higher grade assemblages such as talc–chloritoid were not found. Inverted gradients are observed locally in the Muscat nappes (pyrophyllite-bearing units above kaolinite–quartz-bearing units) and these are ascribed to late metamorphic thrust displacements which are demonstrated by microstructures.

A complex *P–T* evolution is fairly well documented by the mineral assemblages. Prograde evolution in the subophiolitic (continental rocks) began with a rather "hot" gradient of about 30°–35°C/km (with chloritoid growth in the higher units). Thereafter pressure increased with the temperature remaining constant or even decreasing (chloritoid replaced by kaolinite + oxides and carpholite growth) in synkinematic conditions. The peak *P–T* conditions are close to 8 kbar, 270°C for the Muscat nappes and 11 kbar, 400°C for the As Sifah unit. The corresponding surface-related gradient is low (10°–15°C/km). During the retrograde evolution, pressure and temperature generally decreased concomitantly, but in some of the upper units temperature remained constant or increased slightly for some time, at which point sudoite growth from carpholite took place.

The metamorphic pattern and mineral evolution are interpreted in the framework of the Late Cretaceous obduction. The already cold oceanic lithosphere was thrust upon a relatively hot thinned passive margin (with coeval basaltic

Present addresses:
* Laboratoire de Géologie Historique et Structurale, Bâtiment 504, Université Paris Sud, 91405 Orsay (France).
** Laboratoire de Géologie Structurale, Université de Bretagne Occidentale, avenue Le Gorgeu, 29283 Brest (France).

0040-1951/88/$03.50

volcanism). Pressure progressively increased in the passive margin due to the wedge shape of the advancing ophiolite, the maximum thickness of which was about 25 km in the As Sifah area. In this area, the overburden was also attributable to the accumulation of continental material (up to 5 km thick) by means of a complex set of synmetamorphic, often conjugated folds and thrusts. The basal peridotites of the obducted slab were probably not hotter than 300 °C during its thrusting upon the continental margin. The temperature of the tectonic pile below the ophiolite remained low because it was thrust above more external and still colder continental areas. Late thrusting in the same southerly direction controlled the retrograde evolution, together with erosional and gravity-driven tectonic unloading.

Introduction

Oman has been famous among geologists, since Reinhardt (1969), Alleman and Peters (1972) and Glennie et al. (1973, 1974) described the huge ophiolitic nappe there (the Sumail nappe) that was thrust upon the Arabian continent. It is a typical "obduction zone" as defined by Coleman (1971) who described the area as "an important variant of plate tectonics in the zones of compression" by comparison with the classical subduction zone. In these descriptions of Oman geology as well as in the somewhat more recent and detailed works by Searle et al. (1980), Pearce et al. (1981) and Hopson et al. (1981), the only metamorphic rocks described below the ophiolitic nappe are those of the metamorphic sole, i.e., the amphibolites and greenschists located under the Sumail nappe and formed during the intraoceanic detachment (shear) of the ophiolitic lithosphere.

High-pressure, low-temperature (H*P*–L*T*) metamorphism (blueschist facies) was first discussed with regard to Oman by Bailey (1981), Michard et al. (1981) and Boudier and Michard (1981). The occurrence of the characteristic mineral Fe/Mg carpholite was touched on by Michard et al. (1983), while Lippard (1983) described the occurrence of garnet and glaucophane-bearing schists. Finally, restricted eclogite outcrops were identified by Le Métour et al. (1986). The tectonic setting of the H*P*–L*T* metamorphism was outlined by Michard et al. (1984) and Michard (1983) who noted that the metamorphism affects deformed continental material overlain by the oceanic nappes. Hence, it characterizes the lower slab of the obduction zone.

The study of the Oman H*P*–L*T* metamorphism is of twofold interest. First, it offers an efficient tool for deciphering the kinematics and P–T conditions of the latest stages of the obduction process, during which the oceanic slab climbs upon the continental crust (Boudier et al., 1985; Michard et al., 1985; Lippard et al., 1986; Michard, 1987). Second, it offers a good opportunity to unravel an almost (compared to other chains) undisturbed H*P*–L*T* metamorphic zoning related to a single, major tectonic event: the Oman Mountains are the only part of the Tethyan belt where the obduction-related tectonic pile escaped the later collisional events (the "quiet obduction" of Boudier and Michard, 1981).

The purpose of this paper is to describe the H*P*–L*T* history of metamorphism of the southeastern Oman Mountains with the aim of gathering data for further geodynamic interpretation.

Geological setting

The Oman H*P*–L*T* metamorphic rocks outcrop in the southeastern Oman Mountains, along the Gulf of Oman shoreline within the innermost units of the obduction belt (Fig. 1). They constitute the northeastern part of a large tectonic window revealing continental material below the oceanic thrust units: the Saih Hatat anticline (Fig. 2).

The stratigraphy of the continental material is best shown in the southwestern part of the Saih Hatat and in its western equivalent, the Jebel Akhdar anticline: both areas escaped H*P* metamorphism and accompanying deformation. The constituent rocks can be divided into an Upper Permian–Upper Cretaceous cover sequence and a basement of Late Proterozoic–Ordovician age affected by a Paleozoic (Hercynian?) folding event (see below).

In the Saih Hatat window, the pre-Permian rocks consist of more than 1500 m of Upper Proterozoic metagraywackes and volcanics (mostly

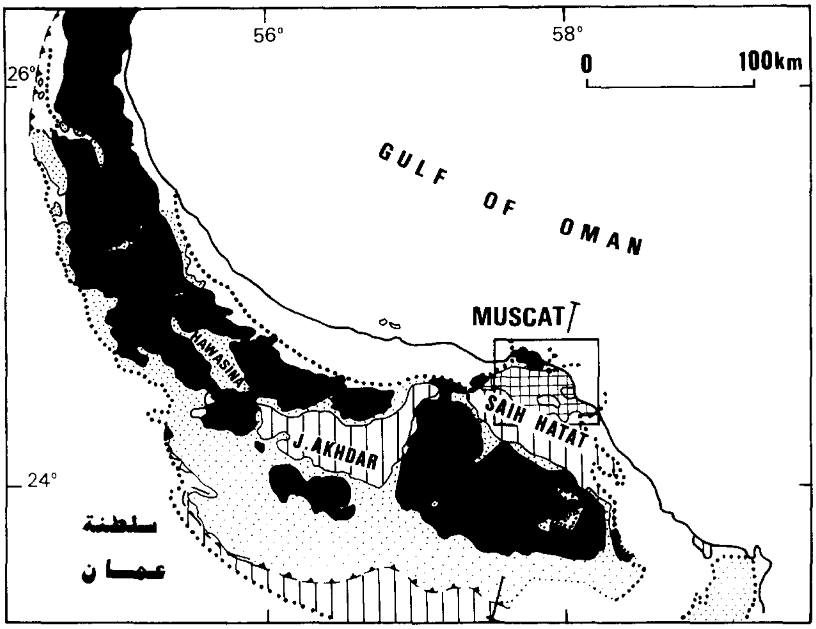


Fig. 1. Simplified tectonic map of the Oman Mountains, after Glennie et al. (1974) except for the H*P*–L*T* metamorphic zone south of Muscat. Black—Sumail ophiolite; stippling—Hawasina nappes and mélange; vertical shading—Arabian continental material, autochthonous and parautochthonous; square hatching—H*P*–L*T* metamorphics; dotted line—erosional limit of the Upper Maastrichtian–Tertiary onlap (white). The studied area around Muscat is boxed (Figs. 3 and 4) and the location of the profile in Fig. 2 is also shown.

pillow basalts), followed by about 300 m of Lower Cambrian stromatolitic dolostones and, following a hiatus, by a thick (more than 3000 m) Lower–Middle Ordovician clastic formation (quartz-rich sandstones and siltstones with a shelly fauna (Glennie et al., 1974; Hopson et al., 1981; Lovelock et al., 1981; Le Métour, 1987). Prior to the deposition of the Upper Permian carbonates and conglomerates (Lovelock et al., 1981; Michard, 1982; Lippard, 1983), the rocks were folded into both open and closed folds associated with an axial plane slaty cleavage.

The Permo-Mesozoic cover consists mainly of up to 3000 m of thick massive shallow-water carbonates ranging from Upper (locally Middle) Permian to Upper Cretaceous. Some other lithologies are intercalated in the carbonate succession. The pelitic and/or volcanic formations are of particular importance, the mineralogy of which is sensitive to the metamorphic conditions: (1) Argil-

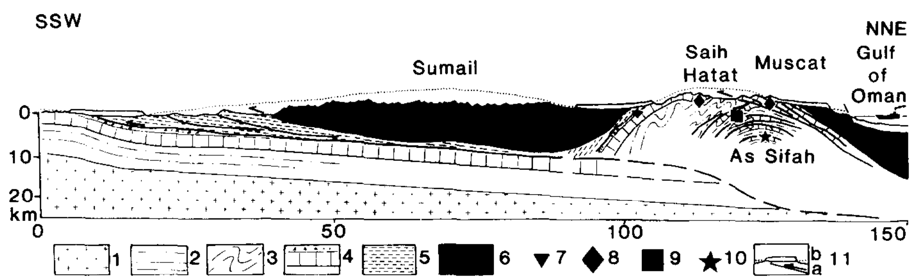


Fig. 2. Schematic cross section along the NNE–SSW traverse shown in Fig. 1. *1* = continental basement rocks; *2* = Upper Proterozoic–Paleozoic sedimentary rocks; *3* = as *2*, folded in the Hercynian orogeny; *4* = Permian–Upper Cretaceous cover (lower part—carbonates, upper part—Turonian–Campanian marls and breccias); *5* = Hawasina nappes and mélange; *6* = Sumail ophiolite; *7* = very low-grade assemblages (kaolinite–chlorite); *8* = low-grade, carpholite-bearing rocks; *9* = medium-grade, chloritoid–chlorite assemblages; *10* = L*T* eclogite; *11* = main thrust faults: *a* = unconformably overlain by the Upper Maastrichtian–Tertiary onlap (*b*).

laceous paleosoils and polymict conglomerates are recognized at the very base of the Permian carbonates (Michard, 1983; Le Métour, 1987; Rabu, 1987) and (2) sandy quartz conglomerates with widespread volcanic flows are found intercalated within the lowest Permian carbonates. The volcanics consist of alkaline to subalkaline basalts associated with intermediate rhyolite and other igneous rocks (Michard, 1983; Le Métour, 1987). Additionally (3), reddish argillaceous sandstones interbedded within the Lower Jurassic limestones and (4) deep-water marls with interbedded turbiditic calcarenites and olistostromal conglomerates, dated from the Early Turonian to Campanian (Glennie et al., 1974; Lippard et al., 1986; Rabu, 1987; Le Métour, 1987) are also recognized. The latter lithological group is typically syntectonic and the most recent deposits are only found in boreholes west of the northern mountains; they contain abundant clasts of oceanic material and are interpreted as having been deposited ahead of the advancing nappes (Glennie et al., 1974). The oldest deposits of the fourth sequence (Muti Formation, Turonian–Santonian) are found all around the continental tectonic windows, just below the oceanic nappes, and only contain autochthonous pebbles and olistolites. The Muti Formation also includes pillow basalt flows (Rabu, 1987) and was deposited on the foundering margin of the continent some time before the emplacement of the nappes.

The allochthonous units of Oman comprise two main rock assemblages (Figs. 1 and 2). The lowest one is mainly composed of deep-water Mesozoic sediments (lime turbidites and radiolarites), ranging in age from Late Triassic to Late Cretaceous (at least Cenomanian) and associated with Triassic alkaline basalts and reef limestones (Oman exotics, Haybi complex). It is usually known as the Hawasina nappe assemblage and ascribed to the Arabo-African Mesozoic passive margin (Glennie et al., 1974; Lippard et al., 1986; Béchennec, 1987; Bernouilli and Weissert, 1987; Cooper, 1987). The thickness of the Hawasina assemblage greatly varies from a few thousand meters in the external parts of the Oman Mountains down to a few tens of meters of highly sheared mélange in the inner parts (Muscat area); indeed the Hawasina nappe sequence may even be locally absent in the inner zone. This lower, proximal allochthon is overlain by another truly oceanic assemblage, namely the Sumail ophiolite, with a basal sheet of *HT* metamorphic slices. The age of oceanic accretion is controversial, either mostly or exclusively Cretaceous (McCulloch et al., 1981; Tilton et al., 1981). Dating of radiolarians gives Albian–Cenomanian ages (about 98 Ma) for the latest tholeiitic pillow basalts, and Turonian–Santonian (about 90–85 Ma) for the latest, differentiated volcanics (Tippit et al., 1981; Beurrier et al., 1987). The more recent deep-sea sediments deposited upon the Sumail volcanics contain Campanian radiolarian faunas (about 81 Ma) (Schaaf and Thomas, 1986). The preserved maximum thickness of the ophiolite reaches 10 km (Shelton, 1984), but the whole section through the Cretaceous oceanic lithosphere is believed to be up to 20 km thick (Boudier et al., 1985; Lippard et al., 1986). This huge crystalline wedge remained practically underformed during its thrusting, except at its base where it was mylonitized, even though it underwent approximately 400 km of transportation, first onto part of the Cretaceous marginal ocean, then onto the Hawasina basin and finally, onto the Arabian continent (see previously mentioned references and Michard, 1987).

Most authors accept that thrusting of the Hawasina nappes was triggered by the southwestward progression of the ophiolitic slab. Thus, their movement began some time before Late Cenomanian which is in good agreement with isotopic dating of the amphibolites of the basal metamorphic sheet (about 95 Ma) (Lanphere, 1981; Boudier et al., 1985). As for the age of emplacement of the Hawasina and Sumail nappes onto the Arabian continental crust, it is estimated between the age of the Muti Formation (about 90–85 Ma) and that of the neoautochthonous, post-thrust shallow-water sediments, the deposition of which began as early as Maastrichtian (about 70 Ma), but still after a period of aerial exposure and lateritic weathering of the ophiolites (Coleman, 1981; Lippard et al., 1986). The large anticlines of the Jebel Akhdar and Saih Hatat tectonic windows rose partly during this late thrusting erosional period, and partly afterwards, during post-Eocene smooth folding and faulting events (Glennie et al., 1974; Searle, 1985).

Metamorphic map

Structural organization of the metamorphics

The HP–LT metamorphic rocks of Oman are visible in the northeastern half of the Saih Hatat tectonic window (Figs. 1 and 2). Except for the occurrence of large dikes and sills of dolerite in the lowest units (Hulw and As Sifah area), the age and significance of which will be discussed later, the protoliths of these metamorphics correspond to the stratigraphic material described above for the southwestern part of the window. Hence, these metamorphic rocks are clearly of local, parautochthonous origin. However, they are severely deformed and divided into several individual tectonic units (Figs. 2 and 3). Along the northern margin of the area, these units consist of large, initially flat-lying duplexes, named the "Muscat nappes" by Michard (1983). They are either overlain by the Sumail peridotites (Matrah massif,

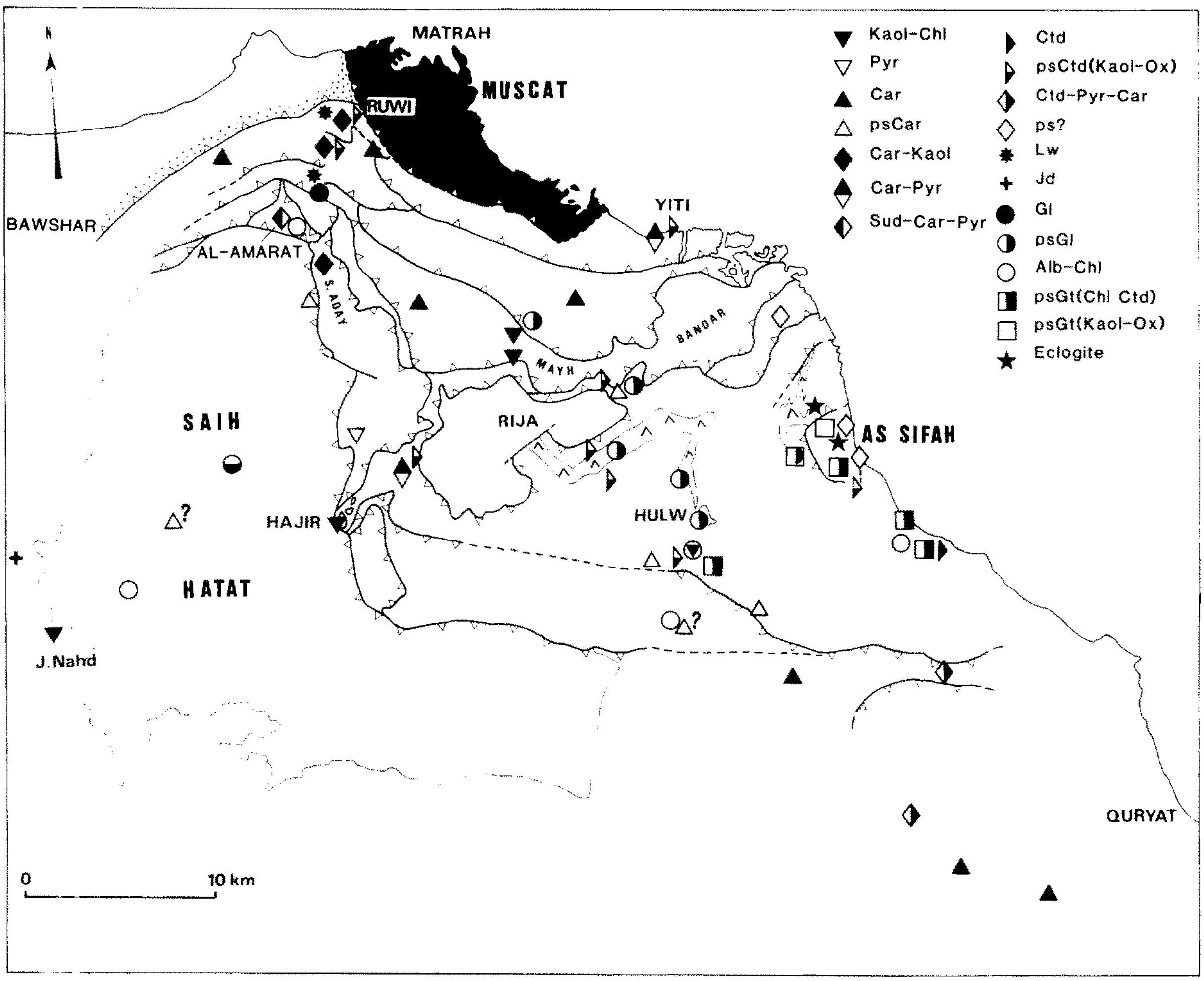


Fig. 3. Metamorphic map of the infraophiolitic, HP–LT metamorphics of Oman (boxed area in Fig. 1). Black—ophiolites; stippling—Hawasina mélange; V-shaped symbols—doleritic sills; white—undifferentiated continental material (Tertiary onlap not represented); bold lines—thrust faults (triangles indicate the down-dip direction); fine line—Permian unconformity in southwest Saih Hatat. Such mineralogical symbol corresponds to several samples from the same area (total number of thin sections = 300; number of microprobe analyses = 650). Abbreviations of mineral names (ps indicates pseudomorphs): Alb = albite; Car = Fe/Mg carpholite; Chl = chlorite; Ctd = chloritoid; Gl = glaucophane and crossite; Gt = garnet; Jd = jadeite (after Le Métour, 1987); Kaol = kaolinite; Lw = lawsonite; Ox = Fe oxides; Pyr = pyrophyllite; Sud = sudoite.

Michard et al., 1984) or by the Hawasina mélange which includes ophiolitic lenses (west and southeast of the Matrah massif) or, alternatively, by the transgressive Tertiary molasses. These nappes were thrust either onto the Saih Hatat unit or onto the deeper Rija and As Sifah units. The Hulw unit extends between the two latter antiformal units and along their southern side. The Rija–As Sifah–Hulw pile is overlain, both to the east and to the south by the Saih Hatat margin. This complex pattern results from a polyphase deformation (Michard, 1983) with changing vergence directed to the south as well as to the north or northeast (Hopson et al., 1981; Le Métour, 1987). Nevertheless, this structural evolution can be considered as the result of a progressive deformation in the framework of a single major geodynamic event (obduction), since all the units are affected by a pervasive, NNE–SSW trending stretching lineation (Michard et al., 1984; Le Métour, 1987) and they all display a similar synkinematic HP–LT metamorphic evolution, which is probably related to a single, major tectonic event. We feel that this event is the obduction process (see discussion section).

Occurrences of metamorphic minerals

In order to draw the metamorphic map (Fig. 3), we plotted the index minerals and mineral assemblages formed in both metapelites and metabasites on a schematic structural map. The two rock types are classically used for estimating metamorphic conditions but, since 1981, several petrological and experimental studies demonstrated that metapelites are particularly convenient for the reconstruction of HP metamorphic histories (Goffé, 1982; Chopin and Schreyer, 1983; Goffé and Velde, 1984; Schreyer, 1985; Chopin, 1985; Goffé and Chopin, 1986; Schreyer, 1987). Actually, most of the minerals of the metapelites in the present area are characteristic of HP–LT metamorphic conditions. The Muscat nappes offer the first known occurrence of Fe/Mg carpholite associated with kaolinite. Nowhere has so large a range of Fe/Mg substitution in carpholites been found, with Mg/Mg + Fe ratios varying from 0.14 to 0.97. They also offer the only known occurrence of the Mg carpholite–sudoite–quartz assemblage (in the case of Crete sudoite was observed associated with either Mg carpholite or quartz; Seidel, 1983; Theye and Seidel, pers. commun., 1987).

In the metabasites, the typical HP–LT minerals are lawsonite (also present with carpholite in some metapelites), blue amphiboles, and the eclogite assemblage omphacite–garnet. Jadeite was reported in a quartz-free metabasalt (Tegyey and Le Métour, 1987). Two additional metamorphic minerals insensitive to pressure (chloritoid and garnet) were also plotted in Fig. 3 in order to be considered as temperature indicators. Low temperature index assemblages in metapelites are chlorite + kaolinite $\pm$ smectite and in metabasites they are green amphibole $\pm$ albite or chlorite + albite. Pseudomorphs of index minerals were also plotted on this map because they provide valuable information on the earlier evolution of P–T conditions: (1) chlorite + quartz $\pm$ chloritoid as pseudomorphs after garnet; (2) kaolinite + Fe oxides as pseudomorphs after chloritoid and garnet; (3) quartz + kaolinite + Fe oxides $\pm$ smectites and calcite as pseudomorphs after an unknown prismatic mineral, possibly amphibole; (4) chlorite + white phyllite (phengite, paragonite) as pseudomorphs after carpholite; (5) green amphibole + chlorite + albite $\pm$ epidote $\pm$ green biotite as retrograde assemblages in blue amphibole-bearing rocks, and (6) kaolinite + albite as breakdown products of paragonite + quartz.

Metamorphic zones

From the plotted index minerals and mineral assemblages (Fig. 3), and with a reasonable degree of generalization, the following type of metamorphic zoning can be deduced (Fig. 4):

(1) The highest grade of metamorphism (eclogite assemblages) characterizes the deepest and easternmost tectonic units (those of As Sifah). The structurally equivalent, but more westerly located Rija culmination does not show such high-grade metamorphic minerals.

(2) The low-grade, HP–LT assemblages (carpholite-bearing metapelites and lawsonite- and/or crossite–glaucophane-bearing metaba-

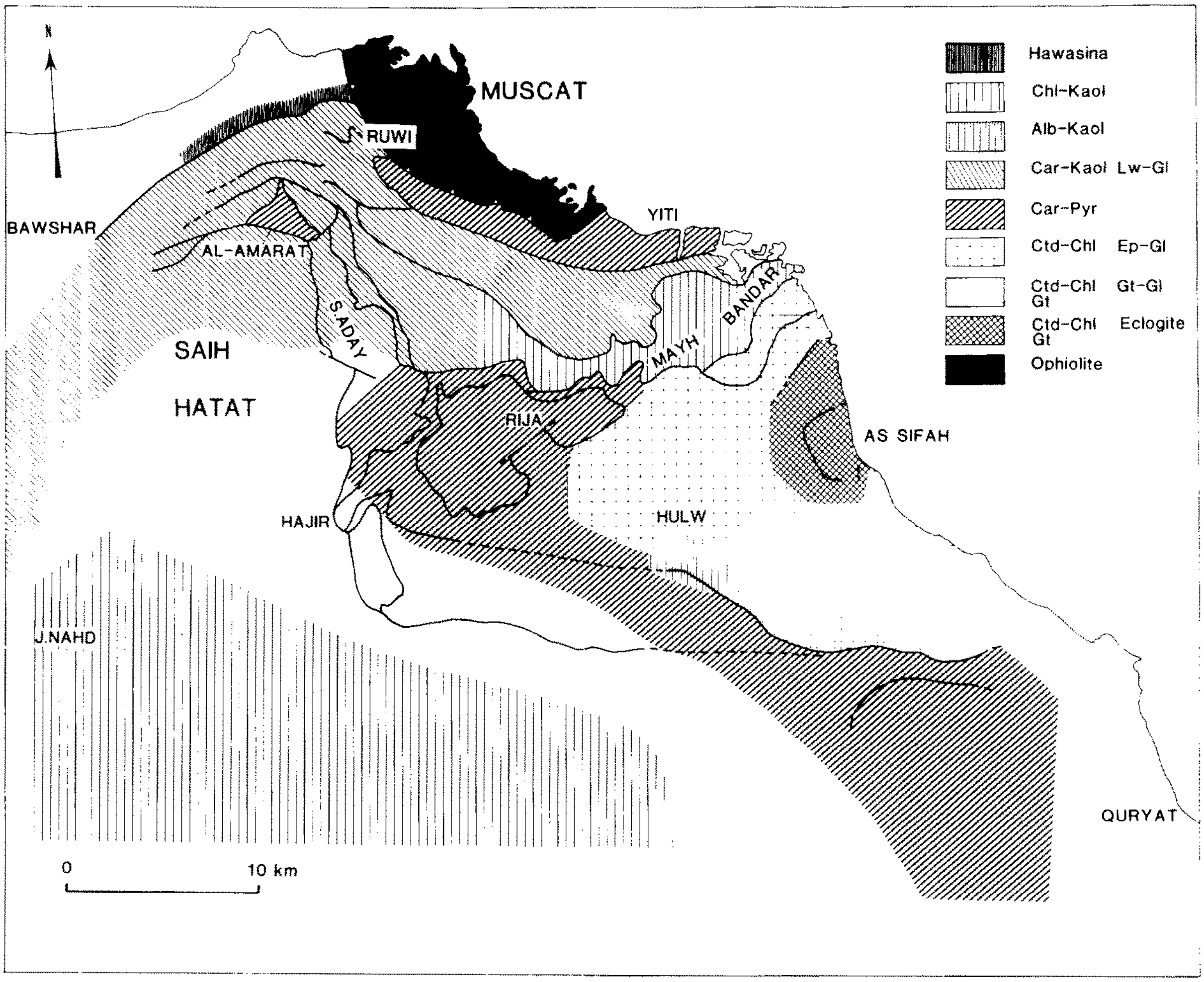


Fig. 4. A generalized metamorphic map inferred from Fig. 3. The Hawasina zone is a very low-grade metamorphic obducted mélange. The ophiolite zone corresponds to the Matrah peridotite. All the other units correspond to continental material (abbreviations as in Fig. 3, except Ep = epidote; metapelitic assemblages to the left and metabasite assemblages to the right. White area represents undetermined metamorphic facies.

sites) are observed within the highest units of detached continental material all around the As Sifah–Rija culmination area, from Yiti to Quryat through the Ruwi–Wadi Aday–Al Amarat (Al Bagriat) section. This is the typical metamorphic grade for the Muscat nappes (Michard, 1983). It also characterizes the Permo-Cretaceous cover (Quryat and Bawshar) and the pre-Permian basement of the northeasternmost part of the Saih Hatat.

(3) The tectonic units located between the two preceding groups (Hulw units and equivalents) show medium-grade assemblages with chloritoid, garnet and glaucophane but without carpholite or lawsonite.

(4) Very low-grade metamorphism (chlorite–kaolinite) is observed in the most external regions of the continental tectonic pile, i.e., in the Permo-Cretaceous cover of southwest Saih Hatat, in the Hawasina mélange to the west of Ruwi (the "anchimetamorphic zone" of Michard et al., 1984), and also within one of the Muscat nappes (the Mayh Bandar unit).

(5) Inverted metamorphic sequences are locally observed in the Muscat nappes: carpholite–pyrophyllite-bearing units may be superimposed onto

(Yiti unit), or intercalated between (Al Amarat unit), carpholite–kaolinite-bearing units (Ruwi unit and southern Aday unit to the southeast of Al Amarat), that are themselves thrust upon a Fe/Mg chlorite–kaolinite unit (Mayh Bandar unit).

(6) The intensity of retrograde alteration of climax minerals varies from one unit to another. It is particularly strong in the intermediate units (Hulw and equivalents), and weaker in upper and deeper units (Muscat and As Sifah). Moreover, contrasted retrograde evolution is shown by juxtaposed units: southern Aday (Mg carpholite preserved), Al Amarat (Mg carpholite replaced by sudoite), and Saih Hatat (Mg carpholite replaced by Mg chlorite).

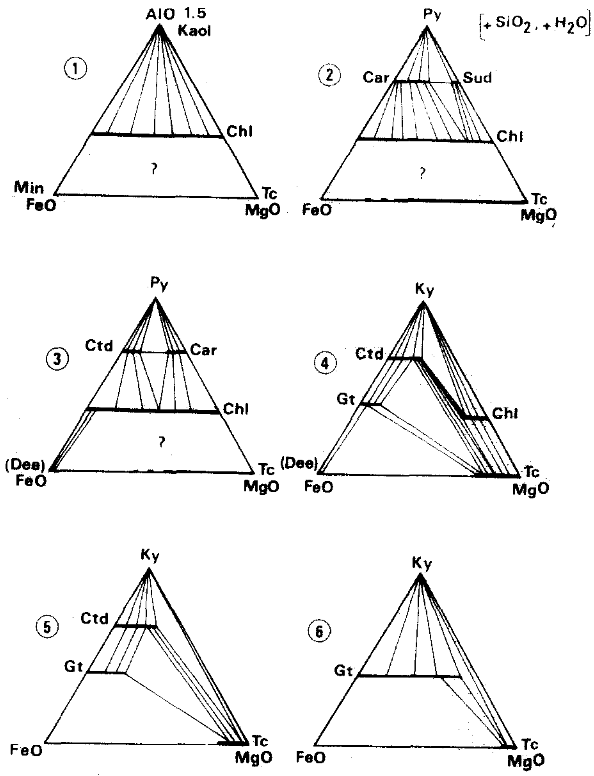


Fig. 6. Six theoretical stages of HP–LT metamorphism of pelites. The sequence of increasing grade (1–6) is taken from Goffé and Chopin (1986). Additional solid phases are quartz and phengite. H_2O is in excess. Stages 4–6 are not known in Oman. Ky—kyanite; Tc—talc; Dee—deerite; Min—minnesotaite; other abbreviations as in Fig. 3.

Stability fields of the mineral assemblages

Assemblages in metapelites

The HP–LT mineral evolution may be considered in the (Fe, Mg)O–SiO_2–Al_2O_3–H_2O system, the relevant minerals of which are plotted in Fig. 5. Many of them are found in Oman (underlined). They correspond to the lower to medium grades of HP–LT metamorphism as shown in Fig. 6, of which assemblages 2 and 3 are the most common. The Fe/Mg carpholite–quartz assemblage occurs frequently, either alone or associated with kaolinite (Ruwi) or pyrophyllite (Yiti). Associations with pyrophyllite and sudoite (Al Amarat), or with pyrophyllite and chloritoid (Quryat) also occur. In contrast, the higher grade assemblages (4, 5 and 6 in Fig. 6) are not present in Oman: even the deepest units contain the chlorite–Fe-rich chloritoid and garnet–chlorite assemblages instead of a chloritoid–talc assemblage.

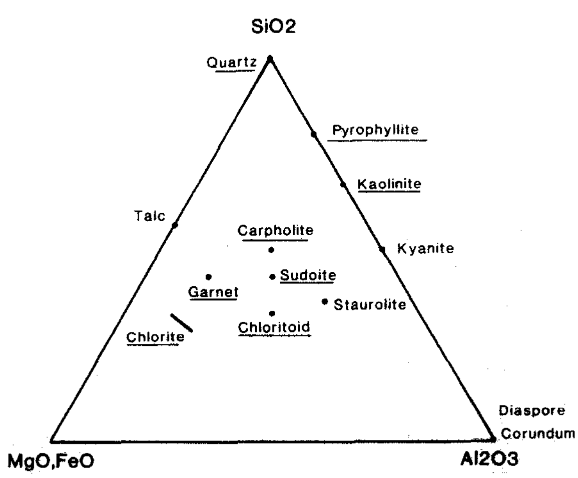


Fig. 5. Phases of interest in HP metapelites considered in the system (Fe, Mg)O–Al_2O_3–SiO_2–H_2O (after Goffé and Chopin, 1986). Underlined minerals are present in Oman metapelites.

On the P–T plane showing the most general stability fields for HP–LT metapelites (Fig. 7), the Oman assemblages occupy the lower P–T area, corresponding to pressures of less than 12 kbar and to temperatures of less than 600 °C. By comparison with a subduction–collision belt such as the Alps (Goffé and Chopin, 1986), the lack of talc–chloritoid assemblages or talc–phengite assemblages (Chopin, 1981) is quite remarkable in Oman.

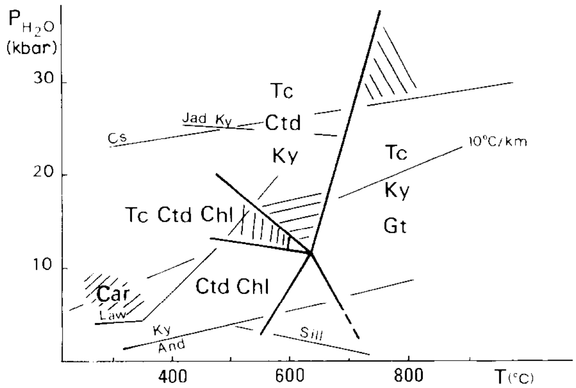


Fig. 7. Stability fields of some critical mineral assemblages with excess quartz or coesite and H_2O (after Chopin, 1985). The ruled areas correspond to the natural assemblages encountered in a typical subduction–collision zone such as the Alps. The Oman metamorphics are concentrated in the lower to medium grades where talc assemblages are not stable. Abbreviations as for Figs. 3 and 6. Additionally, Cs—coesite; Jad—jadeite; Law—lawsonite; And—andalusite; Sill—sillimanite.

Let us now focus on the $P–T$ plane area that applies to the Oman metapelites (Figs. 8 and 10). The sensitivity of pelitic compositions to small P or T changes under HP–LT conditions, i.e., the appearance or disappearance of numerous index minerals over a small P or T range, as well as the large compositional variations of the carpholite solid solution in response to P variations, allows the $P–T$ conditions of the different units to be compared very precisely, even if the absolute $P–T$ location is uncertain in most cases. The low-grade, carpholite-bearing assemblages fall into different areas of the plane which are limited by the kaolinite + quartz = pyrophyllite reaction, the lawsonite stability limit and, moreover, the stability curves of Fe/Mg carpholites themselves according to their Mg substitution ratios (Goffé, 1982; Goffé and Velde, 1984), assuming that water activity is equal to one (see next section). The corresponding temperature varies from 250° to 350°C, but pressure varies from over 8 kbar with $T < 280$°C for the Ruwi-type trivariant assemblages (2, Fig. 10), to only 4–5 kbar for the more external Quryat type ($T = 280$°–350°C, divariant assemblage (3b, Fig. 10). It must be emphasized that the quoted 4–5 kbar value ($T > 280$°C, 3b, Fig. 10) for the Yiti nappe is only a minimum value (trivariant assemblage). As for the 6 kbar pressure indicated for the Al Amarat divariant assemblage ($T = 280$°–300°C, 3c, Fig. 10), this corresponds already to a retrograde stage of the evolution of the Al Amarat assemblage (see next section).

The $P–T$ metamorphic conditions for the higher grade assemblages (Hulw and As Sifah units) lie outside the stability fields of Mg carpholite and lawsonite. Neither these minerals nor their pseu-

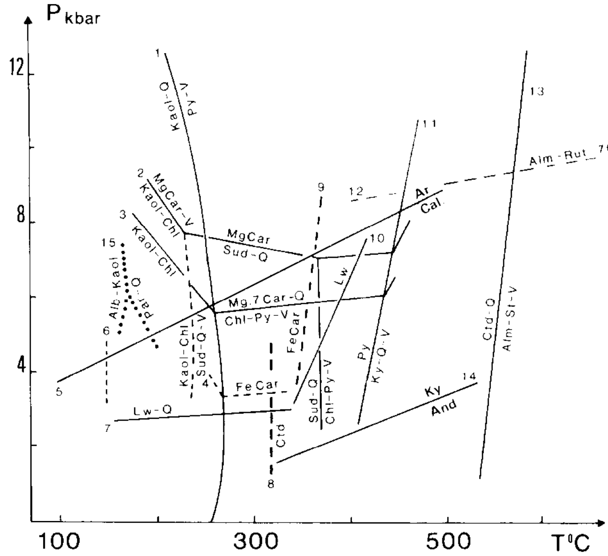


Fig. 8. Stability field of selected minerals of metapelite assemblages. Abbreviations as in Figs. 3, 6 and 7. Additionally, Alm—almandine; Ar—aragonite; Calc—calcite; Par—paragonite; Py—pyrophyllite; Q—quartz; Rut—rutile; St—staurolite; V—H_2O. FeCar, Mg0.7Car, MgCar—0, 70 and 100% respectively of the pure Mg end member of the Fe/Mg carpholite solid solution. *1*—Hemley et al. (1980), Frey (1987). *2*—Fransolet and Schreyer (1984). *3*—Goffé (1982), Goffé and Velde (1984). *4*—Approximate lower limits of the Fe carpholite stability field with univariant curves Chl + Kaol → FeCar + H_2O and Chl + Py + H_2O → FeCar + Q inferred from petrographic evidence (Goffé and Velde, 1984; Goffé, 1984). *5*—Johannes and Puhan (1971). *6*—Temperature kinetic limit of the aragonite–calcite transformation (Carlson and Rosenfeld, 1981). *7*—Lw + Q → laumontite (Nitsch, 1972). *8*—Approximate lower stability limit of the Fe chloritoid inferred from petrographic evidence (Goffé, 1982; Chopin and Schreyer, 1983; Paradis et al., 1983). *9*—FeCar → Ctd + Q + H_2O, approximate location given by Chopin and Schreyer (1983). *10*—Lw → zoisite + margarite + Q + H_2O (Nitsch, 1974). *11*—Chatterjee et al. (1984). *12*—Alm + Rut → Ky + Q + ilmenite geobarometer for a 76% almandine garnet (Bohlen et al., 1983). *13*—Rao and Johannes (1979). *14*—Holdaway (1971). *15*—Possible location of Alb + Kaol → Par + Q + H_2O curve (Chatterjee, 1973).

domorphs have been encountered in these units (see also Lippard, 1983). Hence, the deepest units (As Sifah), where rutile-bearing garnets (80% almandine) are associated with phengite (atomic Si content from 3.35 to 3.55 PFU(per formula unit)), should have experienced temperatures around 450°–500 °C and pressures around or over 8–10 kbar (5, Fig. 10). In the case of the intermediate units (Hulw), where metapelites are characterized by phengite (atomic Si content from 3.3 to 3.4 PFU)–chlorite–chloritoid assemblages locally developed as garnet pseudomorphs, the temperature conditions can be estimated to be in the same range (450 °–500 °C), but at lower pressures (7–9 kbar).

Metabasite assemblages

According to the equilibrium curves plotted in Fig. 9, the glaucophane–epidote assemblages without lawsonite that are found in the Hulw units confirm the $P-T$ estimates (450 °–500 °C and 7–9 kbar) deduced from the associated metapelites (4b, Fig. 10). In contrast, in the case of the deepest units (As Sifah), the eclogitic assemblages (acmit–augite–omphacite, glaucophane, garnet, epidote, phengite, rutile and quartz) give a mean temperature estimate (6, Fig. 10) lower than that derived from metapelites (5). This is 380 °–400 °C from the Na pyroxene–garnet equilibrium, at a pressure of about 11 kbar (intersection with the phengite–garnet equilibrium curve, Fig. 9). The difference between the metapelite and metabasite $P-T$ estimates may be explained by a further equilibration of the metabasites at lower temperature, a process that should not be recognized in metapelites owing to the lack of critical reaction. In the eclogites, the widely varying composition of the Na pyroxenes may represent exsolution features related to a cooling evolution from about 450 °–500 °C to about 350 °–400 °C. This possibility is supported by the fact that glaucophane is often found including earlier omphacite or surrounding garnet. In the metapelites, the chlorite–quartz ± chloritoid assemblages that replace the garnet in some parts of the As Sifah area such as in the Hulw unit could be the result of the same cooling event. The absence of

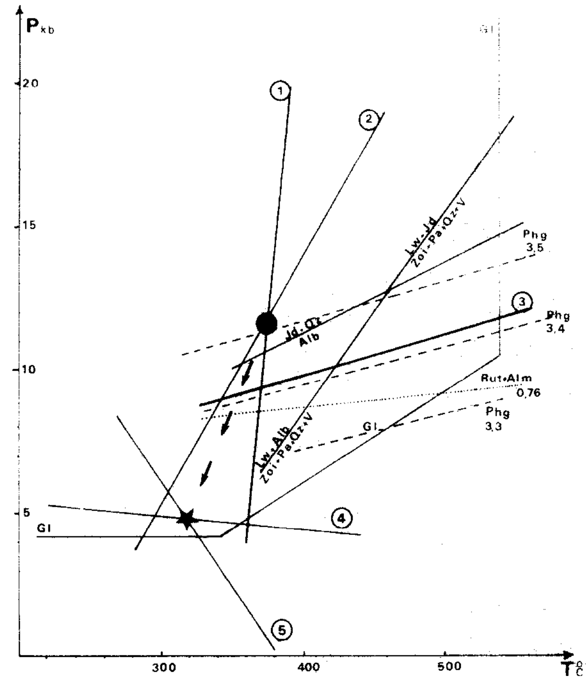


Fig. 9. Stability field of selected minerals of metabasites and geothermobarometric estimates of equilibria in the eclogites (three samples, means of ten calculations on each; dot indicates $P-T$ conditions of the eclogite formation; star indicates $P-T$ conditions of the retrograde alteration of eclogitic assemblages). Abbreviations as in foregoing Figures. Additionally Zoi—zoisite; Qz—quartz; Pa—paragonite. Phg 3.3, Phg 3.4 and Phg 3.5 indicate $P-T$ variations of the Si content of phengites (Massonne and Schreyer, 1987); Gl indicates the largest possible stability field of a natural glaucophane (Maresh, 1977); Jd + Q → Alb (Holland, 1980); Lw + Alb → Zoi + Pa + Q + H_2O (Heinrich and Althaus, 1980); Rut + Alm as in Fig. 8; *1*—garnet–pyroxene KD (Fe–Mg) temperature (Ellis and Green, 1979) (the Fe^{2+} content is calculated with the Vieten and Hamm (1978) method; *2*—phengite–garnet KD (Krogh and Raheim, 1978); *3*—isopleth of jadeite in pyroxene (30% jadeite, mean value) (Holland, 1980); *4* and *5*—pressure and temperature equilibria estimate from the tremolite–edenite content of the later amphiboles associated with chlorite, epidote, albite and quartz (*4*) or albite and quartz (*5*) (Holland and Richardson, 1979).

lawsonite in the metabasites (although they were equilibrated in the stability field of this mineral as shown on Fig. 9) can also reflect an earlier stage of crystallization at higher temperature where zoisite is stable, and the increased stability of the latter mineral at low temperature due to substitution of ferric iron, i.e., the formation of epidote.

In the western (external) part of the Saih Hatat area, jadeite is present in a quartz-free metabasite

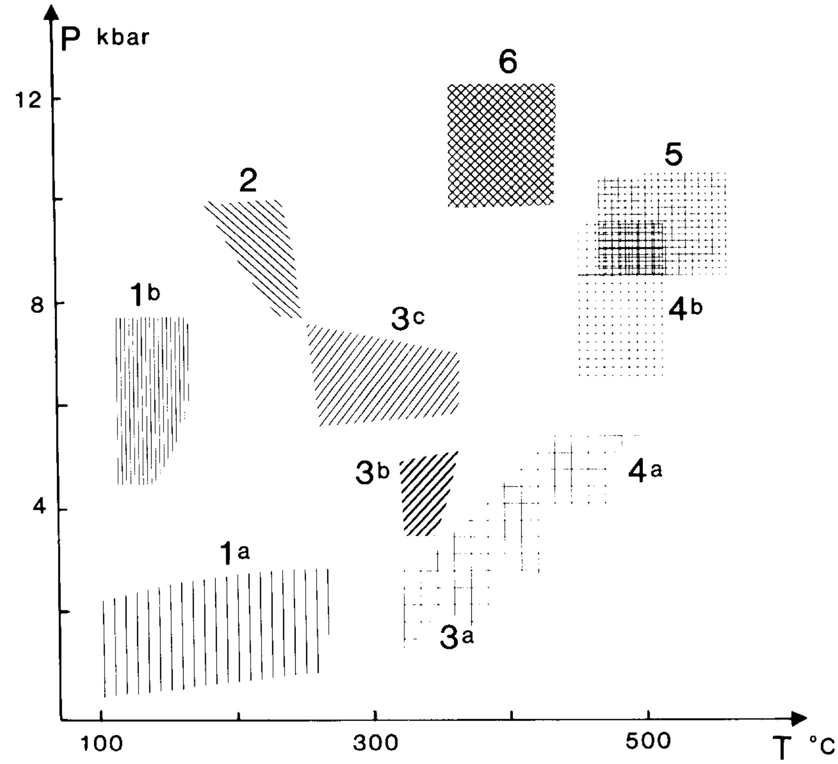


Fig. 10. Estimated fields of formation of the Oman H*P* metamorphic assemblages as deduced from the petrogenetic grids in Figs. 8 and 9. The fields are plotted with the same ornamentation as that used in Fig. 4, with additional subdivisions: *1a*—kaolinite–chlorite assemblages (external zone and Mayh Bandar area in the Muscat nappes; *1b*—albite–kaolinite retromorphic area (top of the Hulw unit); *2*—carpholite–kaolinite assemblages (most of the Muscat nappes, northern Saih Hatat); *3a*—chlorite–chloritoid early assemblages; *3b*—Fe carpholite–chloritoid–pyrophyllite assemblages (Quryat); *3c*—Fe/Mg carpholite–pyrophyllite assemblages (Yiti); *4a*—chlorite–chloritoid without garnet, of the north Hulw area; *4b*—phengite (Si = 3.3–3.4)–chlorite–chloritoid–rutile–(garnet) and glaucophane–epidote assemblages (Hulw area); *5*—phengite (Si = 3.4–3.5)–garnet–rutile assemblages (As Sifah area); *6*—eclogite assemblage of the As Sifah area.

(Tegyey and Le Métour, 1987). The occurrence of jadeite in this area, close to the limit of the kaolinite–chlorite zone (Figs. 3 and 4), requires the existence of relatively H*P*–L*T* conditions, e.g., 3 kbar at 200 °C (jadeite → nepheline + albite univariant curve; Robertson et al., 1957). The conditions consistent with the analcime destabilization hypothesis (Tegyey and Le Métour, 1987) would imply pressure higher than 20 kbar at 200 °C (Griggs and Kennedy, 1956), which is unrealistic.

Evidence for the prograde to retrograde evolution events

The Oman H*P*–L*T* metamorphics are exposed within a restricted geographical area but they are remarkably widely scattered on the *P*–*T* plane (Fig. 10). This can be explained by post- or late-metamorphic displacement of tectonic units that first equilibrated at different levels and/or positions, and by a complex evolution of the metamorphic conditions (*P*, *T* and fH_2O) with time. Evidence for the first phenomenon were deduced above from the metamorphic map (Figs. 3 and 4). Let us now examine the evidence for the second phenomenon that can be deduced from mineralogical observations in thin sections.

Prograde evolution: From chloritoid to carpholite growth

Chloritoid prisms included in carpholite phenoblasts are observed within all the carpholite-bearing units. In the southern part of the Quryat

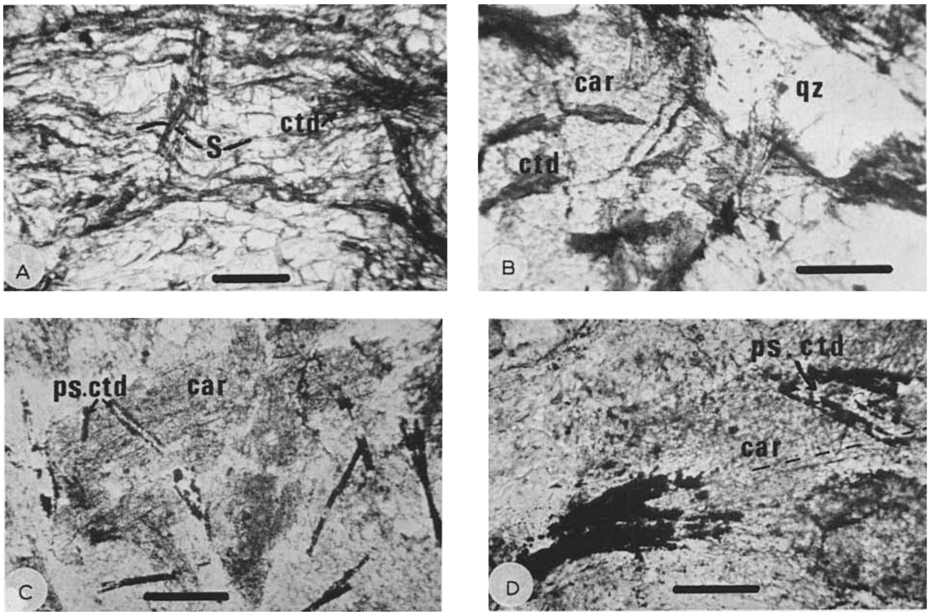


Fig. 11. Petrographic evidence for the earliest stage of metamorphism in the carpholite-bearing units. A. Fanned chloritoid aggregates cutting across an earlier foliation (S). Permian metapelites, Quryat unit (Wadi Mijlas). Scale bar: 100 μm. B. Inclusion of similar chloritoid prisms within ferrocarpholite crystals. Same location. Scale bar: 25 μm. C. Inclusion of oxidized chloritoid prisms within large Fe/Mg carpholite crystals. Jurassic metapelitic layers, Ruwi Hills. Scale bar: 100 μm. D. Detail of Fig. 11C; the chloritoid pseudomorphs consist of oxides (black) and kaolinite (white) associated with carpholite. Scale bar: 25 μm.

unit, chloritoid prisms are not altered (Figs. 11A and B). In the northern part of the same units as well as in the Hajir area, the chloritoid prisms are partly replaced by a kaolinite–Fe oxides assemblage. Still further to the north, in the Muscat nappes, chloritoid no longer exists but its kaolinite–Fe oxide pseudomorphs can be easily identified (Figs. 11C and D). Moreover, this chloritoid alteration was achieved before the Fe/Mg carpholite crystallization (Figs. 11B–D). These chloritoid–carpholite relationships are remarkably different from their relationships in the Alps where carpholite crystallized prior to chloritoid (thus following the H*P*–L*T* prograde evolution shown in stages 1–3 in Fig. 6, Goffé and Velde, 1984; Goffé, 1984). The early growth of chloritoid associated with pyrophyllite before the carpholite crystallization indicates that metamorphism began in a L*P*–H*T* gradient ($P < 3$ kbar and $T >$ 300°–320°C, Field 3a, Fig. 10) followed by an increase of pressure at roughly constant or weakly decreasing temperatures (Fields 3b and 3c or 2, Fig. 10). As suggested by Seidel and Theye (pers. commun. 1987) and Theye (1988), at low water activity, chloritoid may be formed instead of carpholite during progressive high *P*–*T* metamorphism. With rising fH_2O a post-chloritoid formation of carpholite may result even at constant pressure and temperature. In our case, however, we consider that fH_2O was always equal to one because the quartz crystals associated with chloritoid and carpholite contain numerous water inclusions without visible CO_2 or CH_4, and because the widespread occurrence of kaolinite-bearing rocks suggests that fH_2O was not significantly reduced, otherwise kaolinite should be always replaced by pyrophyllite in the acceptable range of *T* conditions (Frey, 1987).

The oxidized minerals

Oxidized pseudomorphs are usually found in the studied area, except in the southernmost part (Fig. 3). Some of them are not unambiguously identified, e.g., the large pseudomorphs (up to 2–3 cm) of prismatic phenoblasts (replaced by quartz–Fe oxide kaolinite–calcite ± smectite) in the Hulw and As Sifah units. (These may represent ancient Ca amphiboles). Others are clearly identified as former garnets (replaced by

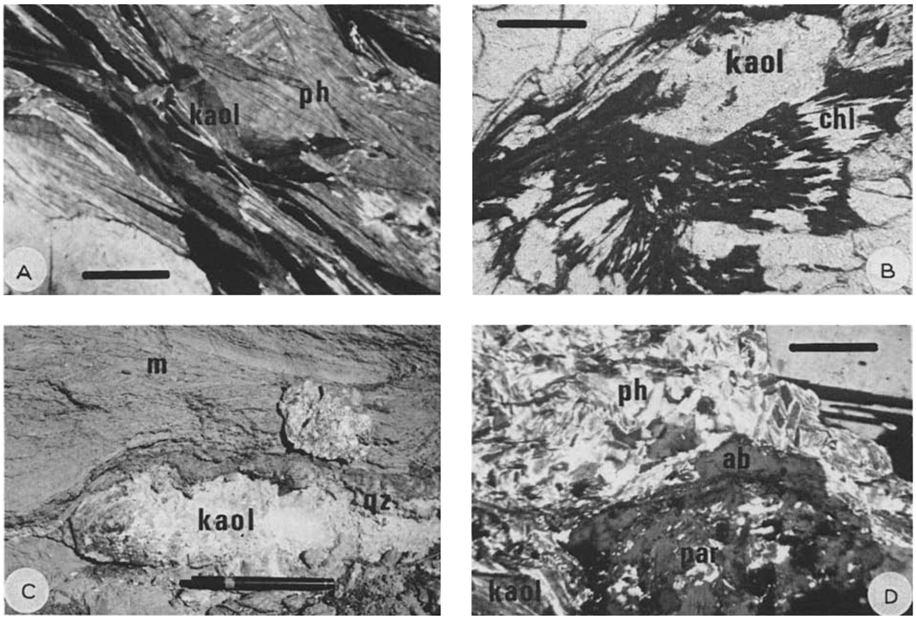


Fig. 12. Petrographic features of the metapelites in the intermediate and deep units. A. Syntectonic kaolinite–oxide association, probably after chloritoid, in a phengite schist (Hulw unit). Scale bar: 25 μm. B. Magnesiochlorite–Fe oxide–kaolinite association after ferrochlorite (As Sifah metapelites). Scale bar: 25 μm. C. Quartz–kaolinite lens in the phengite micaschists (m) (Hajir thrust plane). Pen: 13 cm. D. Late-tectonic kaolinite–albite assemblage after paragonite–quartz in a phengite schist (top of the Hulw unit). Paragonite relicts are shown in the poikilitic albite. Scale bar: 25 μm.

kaolinite–Fe oxides) in the As Sifah unit. This is also the case for total or partial chloritoid pseudomorphs (see previous section), and for Fe/Mg chlorites which are replaced by kaolinite–Fe oxide–Mg chlorite assemblages (Fig. 12B). These features can be observed in the more internal areas (As Sifah) as well as in the more external regions (Jebel Nahd) of the studied area. The intensity of the oxidation of these ferrous minerals progressively increases from south to north. This oxidation process occurred not only before carpholite growth (see preceding section) but also during this growth, as locally shown by the replacement of Fe carpholite by an Mg carpholite–kaolinite–Fe oxide assemblage (Fig. 13).

Such a retromorphic oxidation can be ascribed to the cooling of the metamorphic rocks. Cooling actually favors reactions such as (1) chloritoid + quartz + H_2O + O_2 → kaolinite + Fe oxide, (2) Fe chlorite + O_2 → kaolinite + Fe oxide + quartz + H_2O and (3) Fe carpholite + O_2 + H_2O → kaolinite + Fe oxide (Ganguly, 1969; Bryndzia and Scott, 1987). In the P–T conditions where the Mg term of these minerals can be stable, the Mg term can precipitate if the initial mineral is substituted toward the Mg pole (which is the case of chlorite and carpholite; Figs. 12B and 13).

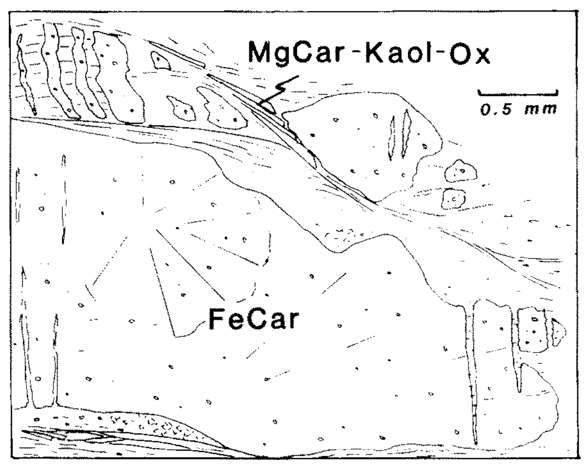


Fig. 13. Petrographic features of the Jurassic metapelites, Ruwi Hills. The early, fan-shaped Fe carpholite crystals include a weakly marked foliation and numerous rounded quartz grains. They are stretched and sheared during the progression of strain. Mg carpholite, kaolinite and oxides crystallize within the corresponding shear planes.

Mg carpholite–sudoite evolution

In the Al Amarat unit, Fe/Mg carpholite (X_{Mg} * = 0.69–0.74) is associated with sudoite (X_{Mg} = 0.81–0.93), pyrophyllite and quartz. The P–T stability conditions of this divariant assemblage can be estimated to lie at 280°–350°C and 6–7 kbar (Fig. 8). In some thin sections, the replacement of preexisting carpholite fibers by sudoite can be observed (Fig. 14B). The Mg content of relict carpholite in these pseudomorphs is always high (X_{Mg} = 0.74–0.75), but lower than that of the newly formed sudoite. These carpholite relicts may correspond to the less Mg-substituted parts of preexisting, longitudinally zoned prisms of Mg carpholite. Similar features have already been described in the Ligurian Alps (replacement of the ferrous parts of longitudinally zoned carpholite prisms by chlorite during a prograde metamorphic evolution (Goffé, 1984)). In the Al Amarat unit, the replacement of Mg carpholite by sudoite can be interpreted as a retrograde evolution from higher to lower pressure (Fig. 8). The temperature conditions of the HP stage in the Al Amarat unit are not precisely constrained. We can assume that they were either the same as in the other Ruwi nappes (field 2, Fig. 10), or as in the Yiti nappes (T > 280°C, P > 8 kbar). In the first case, the retrograde evolution should occur at increasing temperature and in the second case at constant temperature (toward field 3c).

* X_{Mg} = Mg/Mg + Fe + Mn.

Retrograde evolution of the eclogites

The eclogitic assemblage is partly replaced by a blue-green amphibole–quartz–albite–chlorite–epidote assemblage. Using the method of Holland and Richardson (1979), we can estimate the P–T conditions of this alteration stage to be around 5 kbar and 320°C (Fig. 9). With respect to the eclogitic conditions (400°C at 10–11 kbar), these P–T values indicate a decrease in temperature and pressure.

Aragonite–calcite transition

Aragonite should have been stable during the peak metamorphism of the Muscat nappes but was not found in our sampling. We suspect that it was entirely transformed into calcite during the latest stages of metamorphism. This hypothesis is supported by the occurrence of prismatic, polycristalline calcite pseudomorphs of syntectonic phenoblasts in the Cretaceous Ruwi metapelites (Fig. 14A). These pseudomorphs can be interpreted as ancient aragonite prisms. Following this argument, the retrograde path for the Ruwi units must enter the calcite stability field at temperatures higher than 150°C between 4 and 5 kbar according to kinetic models of calcite–aragonite transformation (Carlson and Rosenfeld, 1981; Madon and Gillet, 1984).

Late LT reactions

Late-metamorphic thrusting was already demonstrated from the map in Fig. 3 and from the

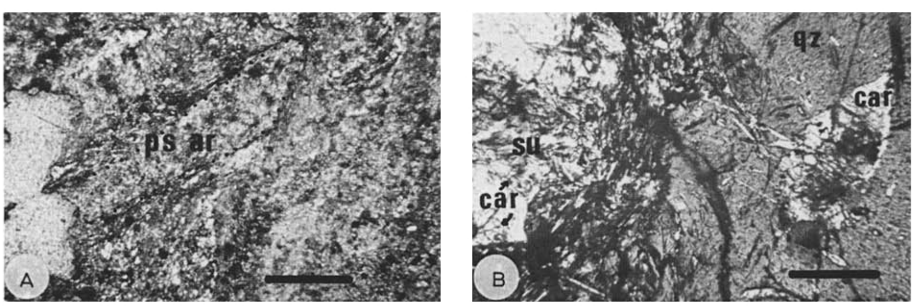


Fig. 14. Petrographic features of some Muscat nappes. A. Calcitic pseudomorph, probably after an earlier aragonite prism (Upper Cretaceous calc metapelites, central Ruwi Hills. Scale bar: 250 μm. B. Replacement of Mg carpholite fibers by sudoite and quartz (left) associated with large Fe carpholite prisms (right) (Permian metapelites, Al Amarat). Scale bar: 25 μm.

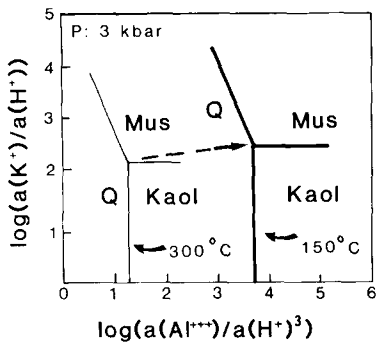


Fig. 15. Theoretical phase relations in the system K_2O–SiO_2–Al_2O_3–H_2O at 3 kbar and 150° and 300°C. Thermodynamic data from Bowers et al. (1984).

widespread boudinage features of climax phenoblasts such as carpholite in the Muscat nappes (Michard et al., 1984). Two mineral reactions related to cooling evolution in thrust fault zones complete this demonstration of the late-metamorphic thrusting.

(1) Close to Hajir, nearly vertical strata of Cambrian carbonates (southern Aday unit) overlie micaschists and Permian carbonates owing to a southwestward, gently dipping thrust fault. Many quartz–kaolinite–oxide lenses are scattered within the fault plane (Fig. 12C). This precipitation of quartz and kaolinite in voids in the rocks can be ascribed to a hydrothermal process related to the metamorphic cooling history (Goffé et al., 1987). In a hydrated system such as the present one (note the abundant water inclusions in the quartz crystals), a decrease of temperature results in an increase in the K:H and Al:H ion activity ratios in the aqueous solution in equilibrium with muscovite (phengite), kaolinite and quartz (Fig. 15). Hence, in order to maintain this equilibrium during a period of cooling, kaolinite and quartz must precipitate while muscovite must dissolve.

(2) At the southern top of the Hulw units, close to the basal thrust of the Hatat–Quryat units, silvery phengite, kaolinite-bearing schists are rich in poikiloblastic albite (Fig. 12D). The albite phenoblasts crystallized in syn- or post-tectonic conditions and contain inclusions of quartz and rare paragonite crystals. The paragonite is remarkably absent in the surrounding matrix in which syn- or post-kinematic kaolinite is found in contact with albite. These assemblages of kaolinite–albite–quartz–relict paragonite may reflect the breakdown of paragonite through the reaction envisaged by Chatterjee (1973); i.e., paragonite + quartz + H_2O → kaolinite + albite. This rare albite–kaolinite–quartz assemblage was, up to now, only observed in rocks which underwent deep diagenesis (Chatterjee, 1973). The Oman occurrence is the first example of this assemblage related to metamorphism. Its stability field is not clearly established but must be located at low temperature (in the kaolinite stability field) and probably at high pressure (up to 7 kbar; Chatterjee, 1973). The breakdown of paragonite in the Hulw unit corresponds to a late drastic cooling event occurring after the rocks underwent higher grade metamorphic conditions.

It is important to note that most of the mineral equilibria which record the cooling evolution of this area imply an excess of water. Water was actually abundant in most of the observed rocks (particularly in the upper and intermediate units), as shown by the numerous fluid inclusions trapped in the quartz crystals of the metapelites.

The metamorphic *P–T* paths

The preceding observations allow us to reconstruct the metamorphic *P–T* paths specific to the

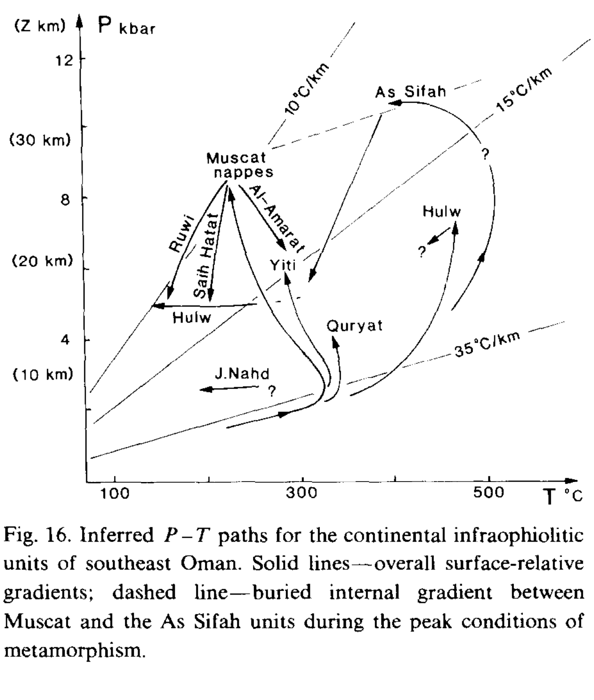


Fig. 16. Inferred *P–T* paths for the continental infraophiolitic units of southeast Oman. Solid lines—overall surface-relative gradients; dashed line—buried internal gradient between Muscat and the As Sifah units during the peak conditions of metamorphism.

various tectonic units (or groups of units) in the area (Fig. 16).

Prograde metamorphic paths

The $P-T$ prograde evolution is fairly constrained for the upper units (Muscat nappes, Quryat area). Metamorphism began there at moderate pressure conditions (less than 3 kbar), up to temperatures of about 300°–350°C (chloritoid crystallization, Field 3a, Fig. 10) and thereafter pressure increased, allowing Fe carpholite growth (Fields 3c or 2, Fig. 10). In the Muscat nappes, except for the Yiti unit, this pressure increase was accompanied by cooling to about 250°C as shown by the kaolinite + Fe oxide pseudomorphs after chloritoid and the kaolinite + Fe oxide + Mg carpholite crystallization after Fe carpholite. The Yiti unit underwent a more moderate cooling (oxidation of chloritoid with pyrophyllite remaining stable). The more external Quryat unit endured only a weak increase in pressure (up to around 4 kbar) within the pyrophyllite and chloritoid stability field ($T = 320°-350°C$, Field 3b, Fig. 10).

The prograde histories of the Hulw and As Sifah units are not so well documented. It seems that the peak conditions (epidote–glaucophane and eclogite conditions, fields 4b and 6 in Fig. 10, respectively) were reached without crossing the lawsonite and Fe/Mg carpholite stability fields. The lack of these minerals and their pseudomorphs is significant, in contrast with the similar metamorphic zones in the Alps and Corsica where they are abundant and easily recognizable (e.g., Caron et al., 1981; Goffé and Chopin, 1986; Pognante and Kienast, 1987). Hence the prograde paths of these deep southeasterly units should be approximately of the same type as those of the upper units. They all initially exhibit a rather high thermal gradient (about 35°–30°C/km) and then a mean gradient of about 10° (Muscat)–15°C/km (As Sifah), quite typical of $HP-LT$ metamorphism.

Retrograde metamorphic paths

Three types of retrograde paths can be distinguished for the upper westerly units (Muscat nappes and northern Hatat, Fig. 16). Most of these units remained in the kaolinite stability field while pressure decreased. However, in the northern Saih Hatat metamorphics, the alteration of Mg carpholite to Mg chlorite shows that pressure fell at nearly constant temperature (outside the sudoite stability field). This case is similar to that of the Al Amarat unit, where pressure decrease was accompanied by constant temperature or weak reheating, as shown by partial replacement of Mg carpholite through sudoite + Fe carpholite.

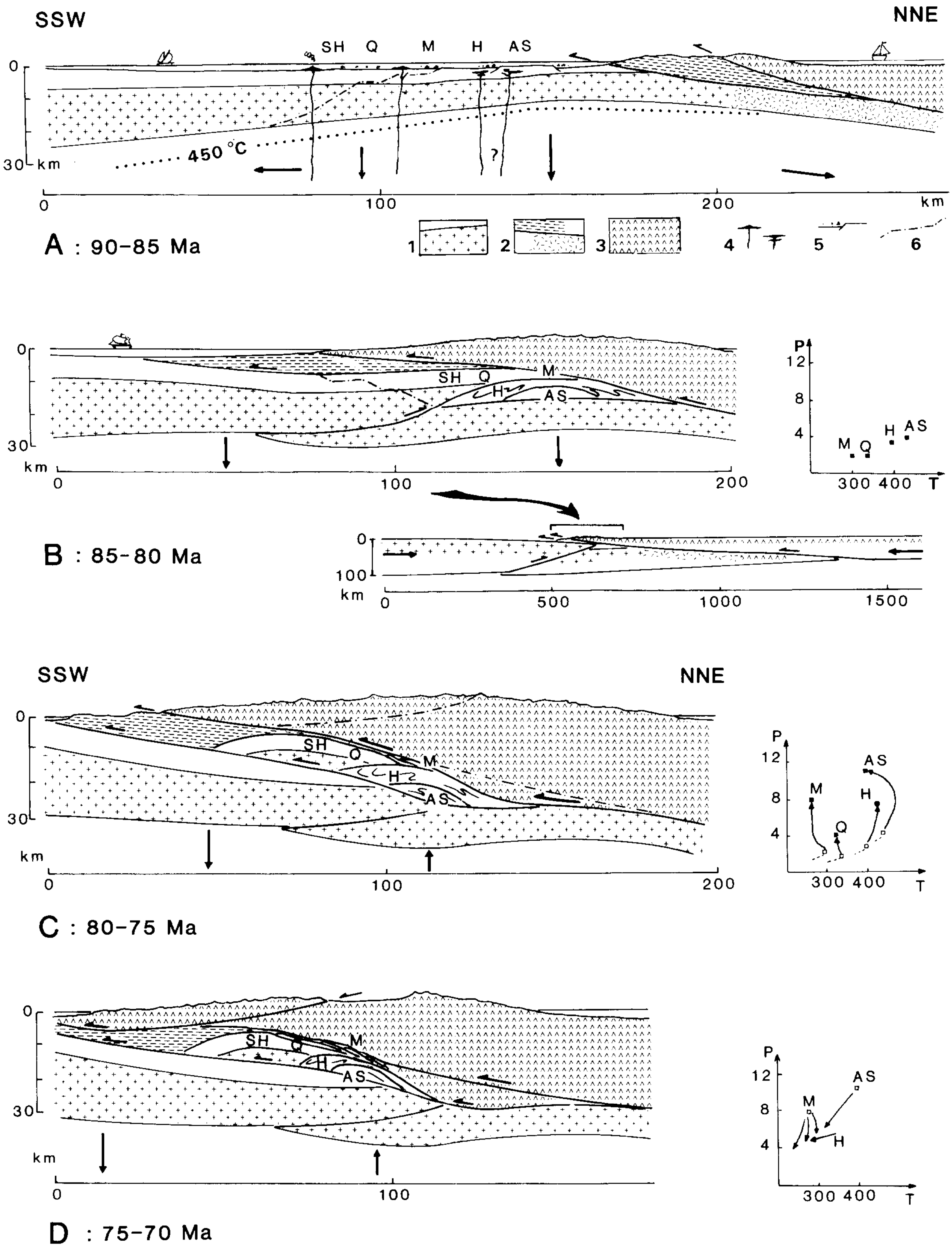
The retrograde path of the As Sifah unit is represented according to the data obtained from the eclogites (Fig. 9). The Hulw path is poorly constrained, except for its latest part (field 1b, Fig. 10).

Interpretation and discussion

A "hot and cold" evolution during pressure increase

In the $P-T$ plane, the metamorphic evolution of the subophiolitic continental slices of Oman is represented by a complex pattern of curves (Fig. 16). Prograde history, which is rarely documented in the literature, is unusual in having two distinct stages: an intermediate pressure stage characterized by a 30°–35°C/km thermal gradient, followed by a typical $HP-LT$ stage with a

Fig. 17. Tectonic interpretation of the $HP-LT$ metamorphism of the southeastern Oman Mountains. Metamorphism affected the Saih Hatat (SH), the Quryat (Q), the Muscat (M), the Hulw (H) and the As Sifah (AS) units. *1* = continental crust (crosses) and sedimentary cover (white); *2* = Hawasina pre-oceanic basin (intermediate crust and detached sedimentary cover); *3* = ophiolitic wedge; *4* = tholeiitic basalts and dolerites within the continental margin; *5* = chaotic breccias in the Turonian–Santonian Muti marls; *6* = incipient failure of the crust, activated in the next stage. A. Turonian–Santonian. B. Santonian–Campanian. At this point, accommodation at depth of the observed shortening of the metasedimentary units is schematically represented as the result of a single SSW-dipping shear zone throughout the continental crust, but it was probably more widely distributed in the deepest, ductile part of the crust. C. Campanian. D. Late Campanian–Maastrichtian. See text for further explanation.


SSW
NNE
SH Q M H AS
0
30 km
450 °C
?
0 100 200 km
A : 90-85 Ma
1
2
3
4
5
6
0
SH Q M
H
AS
30
km
0 100 200
P
12
8
4
M Q H AS
300 400 T
0
100
km
0 500 1000 1500
B : 85-80 Ma
SSW
NNE
0
SH
Q
M
H
AS
30
km
0 100 200
P
12
8
4
M
Q
AS
H
300 400 T
C : 80-75 Ma
0
SH
Q
M
H
AS
30
km
0 100
P
12
8
4
AS
M
H
300 400 T
D : 75-70 Ma

10°–15°C/km gradient. The transition between these stages is marked by cooling of some of the units (Muscat nappes), while in the others (Hulw and As Sifah) temperature remained broadly constant during pressure increase. However, the peak conditions for both types of units are still plotted on a 30°–35°C/km line in the P–T plane (dashed line in Fig. 16). If we accept that these peak conditions are broadly coeval (see discussion section) this line represents an internal thermal gradient (i.e., between the buried upper and lower continental units). This internal gradient is nearly the same as the overall surface-relative gradient which held sway during the beginning of the metamorphism.

It is possible to explain this two-stage prograde metamorphism in the framework of the progressive advance of the Tethyan obduction upon the Arabian continental margin (Figs. 17A–C).

Initial thermal setting

The studied metamorphic units correspond to the inner and upper part of the Arabian continental margin which is a moderately and recently thinned area which underwent Tithonian but especially Turonian subsidence (Glennie et al., 1974; Rabu, 1987) (Fig. 17A). It may have been characterized by a high geothermal gradient of about 40°C/km (e.g., Alvarez, 1984; Gaulier et al., 1988). Such a structural and thermal reconstruction is consistent with the occurrence of Turonian–Santonian basaltic flows intercalated in the deep-water marls of the Saih Hatat–Jabal Akhdar area (the Muti Formation, see the section on the geological setting). This marginal Late Cretaceous magmatism may even be responsible for the large doleritic dikes and sills of the As Sifah–Hulw area which are considered as being of Permian age by Le Métour (1987) but which are actually undated (note that Permian volcanics do occur in the Muscat area, but they are of alkaline to transitional chemistry unlike the intrusions in this case which are tholeiitic.

The earliest metamorphic stage (85–80 Ma)

Since the highest levels of the continental sedimentary cover (the Muti Formation, 90–85 Ma) are metamorphosed even in the uppermost tectonic units (Quryat and Muscat), metamorphism must have been triggered under an allochthonous overburden. We believe that this overburden was simply the advancing ophiolite and Hawasina nappes reaching the continental margin at the end of the Muti sedimentation (Fig. 17B). Initial shearing of the oceanic lithosphere had taken place earlier, at about 95 Ma (Lanphere, 1981; Boudier et al., 1985; Lippard et al., 1986; Montigny et al., 1987a,b) and, in the Hawasina basin, sedimentation had ceased before the Turonian at approximately 90 Ma (Lippard et al., 1986; Béchennec, 1987; Cooper, 1987). At the end of the Muti sedimentation (approximately 85–80 Ma), the upper continental units (Quryat and Muscat) were overburdened by obducted material, the thickness of which may be estimated at about 10 km from the petrological data. At the same time (see additional commects section), the deepest units (Hulw and As Sifah) were experiencing a still heavier overburden of approximately 15–20 km thick. We assume that this total overburden resulted from both the obducted ophiolite and the piling up of continental units beneath the oceanic lithospheric wedge. According to the available structural data (Michard et al., 1984; Le Métour, 1987; Le Mer, unpubl. data), the synmetamorphic recumbent folds and thrusts display a complex arcuate pattern and verge either to the south or south-southwest (northern units), to the north (Hulw area) or to the northeast (Hajir and Quryat). We suggest that this pattern would result from the shortening of the overburdened and horizontally stressed continental lithosphere (i.e., the lower slab of the obduction zone) by means of a SSW-dipping shear zone (Fig. 17B). This shear zone should be broadly coeval with the ophiolitic basal thrust and should represent a conjugate feature created by the plate convergence. Similar patterns of broadly coeval thrusts and backthrusts are observed in other belts (Platt, 1986) as well as in experimental analogs (Balé and Brun, 1987; Ballard et al., 1987). Our interpretation also evokes the deep seismic profiles across the Western Europe Hercynian front and across the Western Alps which show externally dipping reflectors beneath the internally dipping main basal thrusts (Bayer et al., 1987; Hirn and Matte, 1987).

However, two alternative interpretations may be suggested for the accommodation at depth of the observed shortening of the Saih Hatat–As Sifah traverse. One can imagine either that all the metasedimentary units were detached from the underlying continental basement which was progressively underthrust beneath the oceanic lithosphere, or that the continental basement underwent a ductile shortening distributed throughout the whole metamorphic zone. Nevertheless, we favor the model represented in Fig. 17B for the reasons of structural geology referred to above and because it seems to more appropriately match the thermal constraints (the HT–LP gradient), deduced from the petrographical data, according to isotherm distribution in thrust belts (England and Thompson, 1984; Davy and Gillet, 1986).

The HP stage (80–75 Ma)

The second prograde stage, leading to the peak of metamorphism, was characterized by a strong pressure increase in the innermost units (Muscat and As Sifah), a weaker pressure increase in the intermediate and external units (Hulw and Quryat), and almost no temperature change or even a slight cooling (Muscat nappes). Our tectonic interpretation (Fig. 17C) follows the thermal model already discussed for the Alps by Goffé and Velde (1984) and Gillet and Goffé (1988). It is consistent with the general discussion of the P–T–time paths in orogenic belts by England and Thompson (1984), Davy and Gillet (1986), Chamberlain and Karabinos (1987) and Thompson and Ridley (1987).

We ascribe the pressure increase to the thrusting of an ophiolitic slab which becomes progressively thicker and the cooling effect to the simultaneous thrusting of the entire detached tectonic pile toward more external and cooler areas of the continental crust. Any break down in the progression of the thrusts would have caused a temperature increase within the overburdened units (England and Thompson, 1984; Davy and Gillet, 1986). The thrust plane migration down to the base of the continental sedimentary cover should have caused part of the Saih Hatat anticlinal bending by the mechanism of ramp folding advocated by Hanna (1984) and Searle (1985). The maximum thickness of the ophiolite, at least over the Muscat nappes, should have been approximately 25 km. The eclogite-bearing As Sifah units experienced a still heavier overburden, around 30–35 km thick, due to the piling up of continental slices at the hanging wall of the As Sifah units, under the ophiolite. The difference in the metamorphic grade of the As Sifah and Rija culminations may reflect some eastward thickening of the ophiolite and of the continental units beneath.

Retrograde evolution (75–70 Ma)

The retrograde evolution of most of the units is typified by simultaneous unloading and cooling. This evolution can be explained by (1) thrusting of the whole pile upon more external and cooler areas and (2) coeval erosional and tectonic gravity-driven unloading (low-angle extensional faulting). The necessity of rapid, tectonically controlled exhumation of HP–LT terranes for the minerals in these terranes to be preserved has been repeatedly stressed (Dal Piaz, 1974; Draper and Bone, 1981; Caron, 1984; Rubie, 1984; Gillet et al., 1984; Goffé and Velde, 1984; Davy and Gillet, 1986; Platt, 1986; Thompson and Ridley, 1987; Gillet and Goffé, 1988).

Late-metamorphic displacements within the Oman tectonic pile are demonstrated by the occurrence of erratic inverted P–T gradients and pervasive late-crystalline deformation of the HP–LT minerals (Ouazzani-Touhami, 1986). We suggest that these internal displacements results from the ongoing thrust movement of the ophiolite. Thrusting may occur partly along a new, intraperidotitic shear zone (Fig. 17D), resulting in no further increase of the ophiolitic load upon the continental units. The persistence of pressure at decreasing temperature, demonstrated in the southern part of the Hulw unit, can be ascribed to the reactivation of NE-verging thrust faults. The occurrence of a SW-dipping slaty cleavage in the uppermost, external "autochthon" (Muti marls; Le Métour, 1987; Rabu, 1987) can possibly be linked to this late NE-verging tectonism, but could alternatively be ascribed to earlier backthrusting (Fig. 17B).

Pre-Late Maastrichtian tectonic evolution is characterized by further NNE–SSW shortening of the infraophiolitic units, bringing about more bending of the major anticlines (Saih Hatat–As Sifah and Jebel Akhdar windows). The emplacement of the very low-grade Hawasina mélange upon the northwestern Muscat nappes (Michard, 1983; Le Métour, 1987) probably occurred shortly before. In order to explain the contrasting metamorphic grades of these superposed units, it was suggested that this Hawasina mélange was emplaced (together with the Matrah peridotites) from an area to the northwest of Muscat where metamorphism was weaker due to the reduced thickness of the oceanic slab (Michard et al., 1985).

Additional comments

In the preceding reconstruction, the tectonic and metamorphic histories of the upper and deeper subophiolitic units (Saih Hatat–Muscat and Hulw–As Sifah, respectively) are coeval and their entire prograde and retrograde evolution lasted approximately 15 Ma from about 85 to about 70 Ma according to the stratigraphic data. However, it was recently suggested that the HP–LT metamorphism of the As Sifah–Hulw units may be as old as Early Cretaceous, hence older than the ophiolite thrusting onto the Campanian marls of the external units (Montigny et al., 1987a,b). This view was based on isotopic ages (K/Ar and $^{39}Ar/^{40}Ar$ methods) which are scattered between 239 ± 8 and 57.5 ± 11 Ma, but which cluster between 110 and 80 Ma (phengite ages) or 80 and 67 Ma (glaucophane ages) (Le Mer et al., 1986; Montigny et al., 1987a,b).

In our opinion, this view based on isotopic data is not conclusive and the apparent Early Cretaceous ages could also be explained by the occurrence of inherited argon in younger metamorphic minerals. The Late Cretaceous age of metamorphism is clearly established for the Muscat and Saih Hatat–Quryat units (upper units), where the Upper Cretaceous Muti Formation (calcschists with olistostromal layers (see section on geological setting) is clearly identified among the HP–LT metamorphics. As for the deeper, Hulw–As Sifah units, we consider that they were metamorphosed at the same time, based on the following arguments: (1) They underwent a metamorphic history broadly similar to that of the upper units, characterized by the succession of "hot" and "cold" stages, i.e., by an anticlockwise P–T path, (2) they show a similar structural pattern which is characterized by polyphase syn- to late-metamorphic folds and thrusts associated with a generally flat-lying schistosity and a pervasive, NNE–SSW trending stretching lineation (Michard et al., 1984, unpubl. data), and (3) while the Late Cretaceous orogeny is clearly registered in the sedimentary record of the continental margin, the same record does not show any trace of a possible orogenic event during the Early Cretaceous time span (Glennie et al., 1974; Rabu, 1987).

Hence we generally agree with Lippard et al. (1986) who ascribe the HP–LT metamorphism to the obduction process, but some aspects of their model are unacceptable to us. Let us here emphasize that our petrological data preclude the concept of a "hot ophiolitic slab" during the epicontinental stage of the obduction: the peridotites were cooler than 400 °C when they reached the Arabian continental margin. Neither do we feel it necessary to suggest a duplication of the ophiolitic slab above the As Sifah eclogites (Boudier et al., 1985; Lippard et al., 1986). Our mineralogical and structural data support the idea of a wedge-shaped slab, about 25 km thick, above the As Sifah eclogites (Fig. 17).

Finally, our model is consistent with the data presented by Le Métour et al. (1986), Le Métour (1987) and Rabu (1987), who also consider a purely Late Cretaceous metamorphic evolution. However, in their opinion this evolution and the associated deformation result from an early "subduction" process, poorly defined but distinct from the further obduction of the Sumail ophiolite. We feel that our model of progressive deformation and metamorphism of the continental margin in the obduction zone better matches the restricted available time span (around 15 Ma, see above).

Conclusion

The Oman HP–LT metamorphic belt appears relatively simple since it results from a single tectonic process—obduction (in the sense of Coleman, 1971). At the same time, it also appears rather complex, since it evolved over 15 m.y. during the progressive development of the large lithospheric thrust. Most isograds correspond to tectonic contacts and the entire pile was shifted tens of kilometers toward the stable continent (Fig. 17). However, due to the lack of a subsequent collision, the obduction-inherited metamorphic pattern remained virtually unmodified and broadly simple: low-grade assemblages characterize the uppermost and external units, and high-grade assemblages characterize the inner and deepest. Inverted gradients are locally shown in the highest nappes. These are ascribed to late-metamorphic thrust movements which are demonstrated in the structures, and not to any "hot slab" effect (the "hot iron" model).

The P–T evolution is fairly well constrained using the metapelitic assemblages for low-grade units (Fe/Mg carpholite-bearing rocks), and the metabasite assemblages for high-grade units (LT eclogites). The most remarkable point is that prograde evolution began with a rather "hot" gradient (30°–35°C/km). The main pressure increase set in later on at a constant or even decreasing temperature. During the peak conditions (nearly 8 kbar and 270°C for the upper units and 11 kbar and 400°C for the deepest units), the surface-relative gradient was actually low (10°–15°C/km), but the internal gradient at depth (between upper and deeper units) remained rather high. This is interpreted in the framework of the Late Cretaceous obduction process. When the oceanic lithosphere reached the continent, it overrode a relatively "hot" passive margin, recently thinned, with coeval basaltic volcanism. Pressure subsequently increased due to the wedge shape of the advancing ophiolitic slab. The required thickness of the slab above the As Sifah eclogite is about 25 km. Beneath the slab, the overburden was increased by nearly 5 km of folded and sliced continental cover. Detachment of the latter material from the underlying crust occurred close to the 300–400°C isotherm which corresponds to a major mechanical discontinuity between brittle and ductile lithosphere (Le Pichon et al., 1987) and to high pore-fluid pressures due to dehydration reactions (Ayrton, 1980). The temperature of the overlying tectonic pile was kept low by being thrust above more external and cooler areas of the continent. This kinematic framework persisted throughout the retrograde evolution of the metamorphic pile and was accompanied by erosional and tectonic unloading, allowing the preservation of most of the HP minerals.

Acknowledgements

Field work was supported by grants from the Institut National des Sciences de l'Univers, the Université Louis Pasteur, Strasbourg, and the Centre National de la Recherche Scientifique, ER 224, Paris. Facilities from the Oman Ministry of Petroleum and Minerals represented by M.H. Kassim and from the Bureau de Recherches Géologiques et Minières represented by J. Caïa and M. Villey are also acknowledged. We thank J.M. Caron, C. Chopin, R.G. Coleman, W. Schreyer, E. Seidel and Th. Theye for their constructive criticism.

Crystallochemical data on the studied minerals are available at the first author's laboratory.

References

Alleman, F. and Peters, T., 1972. The ophiolite–radiolarite belt of the North Oman Mountains. Eclogae Geol. Helv., 65: 657–697.

Alvarez, F., 1984. Etude de l'évolution thermique des bassins sédimentaires formés par extension. Conséquences thermo-mécaniques du rifting. Thèse Spéc., Univ. Paris, Paris 6, 220 pp.

Ayrton, S., 1980. High fluid pressure, isothermal surfaces, and the initiation of nappe movement. Geology, 8: 172–174.

Bailey, E.H., 1981. Geological map of the Muscat-Ibra area, Sultanate of Oman. J. Geophys. Res., 86 (B4): pocket map.

Balé, P. and Brun, J.P., 1987. Centrifugal versus centripetal collisional arcs: an experimental approach. Tectonophysics, in press.

Ballard, J.F., Brun, J.P., Van den Driessche, J. and Allemand, P., 1987. Propagation des chevauchements au-dessus des zones de décollement: modèles expérimentaux. C.R. Acad. Sci. Paris, Sér 2, 305: 1249–1253.

Bayer, R. et al., 1988. Premiers résultats de la traversée des Alpes occidentales par sismique réflection verticale. C.R. Acad. Sci. Paris, Sér. 2, 305: 1461–1470.

Béchennec, F., 1987. Géologie des nappes Hawasina dans la partie orientale et centrale des montagnes d'Oman. Thèse d'Etat, Univ. Paris, Paris, 412 pp.

Bernouilli, D. and Weissert, H., 1987. The upper Hawasina nappes in the Central Oman Mountains. Stratigraphy, palinspastics and sequence of nappe emplacement. Geodin. Acta, 1: 47–58.

Beurrier, M., Bourdillon de Grissac, Ch., De Wever, P. and Lescuyer, J.-L., 1987. Biostratigraphie des radiolarites associées aux volcanites ophiolitiques de la nappe de Samail (Oman). C.R. Acad. Sci. Paris, Sér 2, 304: 907–910.

Bohlen, S.R., Wall, V.J. and Boettcher, A.L., 1983. Experimental investigations and geological applications of equilibria in the system $FeO-TiO_2-Al_2O_3-SiO_2-H_2O$. Am. Mineral., 68: 1049–1058.

Boudier, F. and Michard, A., 1981. Oman ophiolites: the quiet obduction of oceanic crust. Terra Cognita, 1(2): 109–118.

Boudier, F., Bouchez, J.L., Nicolas, A., Cannat, M., Ceuleneer, G., Misseri, M. and Montigny, R., 1985. Kinematics of oceanic thrusting in the Oman ophiolite; model of plate convergence. Earth. Planet. Sci. Lett., 75: 215–222.

Bowers, T.S., Jackson, K.S. and Helgeson, H.C., 1984. Equilibrium Activity Diagram. Springer, Berlin 397 pp.

Bryndzia, L.T. and Scott, S.D., 1987. The composition of chlorite as a function of sulfur and oxygen fugacity: an experimental study. Am. J. Sci., 287: 50–76.

Carlson, W.D. and Rosenfeld, J.L., 1981. Optical determination of topotactic aragonite–calcite growth kinetics: metamorphic implications. J. Geol., 89: 615–638.

Caron, J.M., 1984. The diversity of geodynamic regimes leading to high pressure–low temperature metamorphism (Western Alps and Corsica). Terra Cognita, 4: 39–43.

Caron, J.M., Kienast, J.R. and Triboulet, C., 1981. High pressure–low temperature metamorphism and polyphase alpine deformation at Sant'Andrea di Cotone (Eastern Corsica, France). Tectonophysics, 78: 419–451.

Chamberlain, C.P. and Karabinos, P., 1987. Influence of deformation and pressure–temperature paths of metamorphism. Geology, 15: 42–44.

Chatterjee, N.D., 1973. Low-temperature compatibility relations of the assemblage quartz–paragonite and the thermodynamic status of the phase rectorite. Contrib. Mineral. Petrol., 42: 259–271.

Chatterjee, N.D., Johannes, W. and Leistner, H., 1984. The system $CaO-Al_2O_3-SiO_2-H_2O$: new phase equilibria data, some calculated phase relations and their petrological applications. Contrib. Mineral. Petrol., 88: 1–13.

Chopin, C., 1981. Talc–phengite: a widespread assemblage in high-grade pelitic blueschists of the Western Alps. J. Petrol., 22: 628–650.

Chopin, C., 1985. Les relations de phases dans les metapélites de haute pression. Thèse d'Etat, Univ. Paris, Paris, 6, 80 pp.

Chopin, C. and Schreyer, W., 1983. Magnesiocarpholite and magnesiochloritoid: two index minerals of pelitic blueschists and their preliminary phase relations in the model system $MgO-Al_2O_3-SiO_2-H_2O$. Am. J. Sci., 283(A): 72–96.

Coleman, R.G., 1971. Plate tectonic emplacement of upper mantle peridotites along continental edges. J. Geophys. Res., 76: 1212–1222.

Coleman, R.G., 1981. Tectonic setting of ophiolite obduction in Oman. J. Geophys. Res., 86 (B4): 2497–2508.

Cooper, D.J.W., 1987. Hamrat Duru group: revised stratigraphy of a Mesozoic deep-water passive margin in the Oman Mountains. Geol. Mag., 124: 157–164.

Dal Piaz, G.V., 1974. Le métamorphisme de haute pression et basse température dans l'évolution structurale du bassin ophiolitique alpino-apenninique. Bull. Suisse Minéral. Pétrogr., 54: 399–424.

Davy, Ph. and Gillet, Ph., 1986. The stacking of thrust slices in collision zones and its thermal consequences. Tectonics, 5: 309–320.

Draper, G. and Bone, R., 1981. Denudation rates, thermal evolution and preservation of blueschist terrains. J. Geol., 89: 601–613.

Ellis, D.J. and Green, D.H., 1979. An experimental study of the effect of Ca upon garnet–clinopyroxene exchange equilibria. Contrib. Mineral. Petrol., 71: 13–27.

England, P.C. and Thompson, A.B., 1984. Pressure–temperature–time paths of regional metamorphism. I. Heat transfer during the evolution of thickened continental crust. J. Petrol., 25: 894–928.

Fransolet, A.M. and Schreyer, W., 1984. Sudoite, di/trioctahedral chlorite: a stable low-temperature phase in the system $MgO-Al_2O_3-SiO_2-H_2O$. Contrib. Mineral. Petrol., 86: 409–417.

Frey, M., 1987. The reaction-isograde kaolinite + quartz = pyrophyllite + H_2O, Helvetic Alps, Switzerland. Schweiz. Mineral. Petrogr. Mitt., 67: 1–11.

Ganguly, J., 1969. Chloritoid stability and related parageneses: theory, experiments and applications. Am. J. Sci., 267: 910–944.

Gaulier, J.M., Le Pichon, X., Lyberis, N., Avedik, F., Geli, L., Moretti, I., Deschamps, A. and Salah Hafez, M., 1988. Seismic study of the crust of the northern Red Sea and Gulf of Suez. In: X. Le Pichon and J.R. Cochran (Editors), The Gulf of Suez and Red Sea Rifting. Tectonophysics, in press.

Gillet, Ph. and Goffé, B., 1988. On the significance of aragonite occurrence in the Western Alps. Contrib. Mineral. Petrol., 99: 70–81.

Gillet, Ph., Ingrin, J. and Chopin, C., 1984. Coesite in subducted continental crust: $P-T$ history deduced from an elastic model. Earth Planet. Sci. Lett., 70: 426–436.

Glennie, K.W., Boeuf, M.G.A., Hughes-Clarke, M.W., Moody-Stuart, M., Pilaar, W.F.H. and Reinhardt, B.M., 1973. Late Cretaceous nappes in the Oman mountains and their geologic evolution. Bull. Am. Assoc. Petrol. Geol., 57: 5–27.

Glennie, K.W., Boeuf, M.G.A., Hughes-Clarke, M.W., Moody-Stuart, M., Pilaar, W.F.H. and Reinhardt, B.M., 1974. Geology of the Oman Mountains. Verh. K. Ned. Geol. Mijnbouwkd. Genoot. Geol. Ser., 31: 423 pp.

Goffé, B., 1982. Définition du faciès à Fe-Mg-carpholite–chloritoide, un marqueur du métamorphisme de $HP-BT$ dans les métasédiments alumineux. Thèse d'Etat, Univ. Paris, Paris 6, 232 pp.

Goffé, B., 1984. Le facies à carpholite chloritoïde dans la couverture briançonnaise des Alpes Ligures: un témoin de l'histoire tectono-métamorphique régionale. Mem. Soc. Geol. Ital., 28: 461–479.

Goffé, B. and Chopin, C., 1986. High-pressure metamorphism in the Western Alps: zoneography of metapelites, chronology and consequences. Schweiz. Mineral. Petrogr. Mitt., 66: 41–52.

Goffé, B. and Velde, B., 1984. Contrasted metamorphic evolutions in thrusted cover units of the Briançonnais zone (French Alps): a model for the conservation of HP–LT metamorphic mineral assemblages. Earth Planet. Sci. Lett., 68: 351–360.

Goffé, B., Murphy, W.M. and Lagache, M., 1987. Experimental transport of Si, Al and Mg in hydrothermal solutions: an application to vein mineralization during high-pressure, low-temperature metamorphism in the French Alps. Contrib. Mineral. Petrol., 97: 438–450.

Griggs, D.T. and Kennedy, G.C., 1956. Simple apparatus for high pressures and temperatures. Am. J. Sci., 253: 722–735.

Hanna, S., 1984. The tectonic evolution of the Oman Mountains thrust belt and the rate of the ophiolite emplacement. Colloq. Int. C.N.R.S., 41: (Abstr.).

Heinrich, W. and Althaus, E., 1980. Obere Stabilitätsgrenze von Lawsonit + Albit bzw. Jadeit. Fortschr. Mineral. Beih., 58: 49–50.

Hemley, J.J., Montoya, J.W., Marinenko, J.W. and Luce, R.W., 1980. Equilibria in the system Al_2O_3–SiO_2–H_2O and some general implications for alteration/mineralization processes. Econ. Geol., 75: 210–228.

Hirn, A. and Matte, Ph., 1987. Seismic signature and tectonic cross section of the Variscan crust in Western France. Terra Cognita, 7: 156 (Abstr.).

Holdaway, M.J., 1971. Stability of andalusite and the aluminum silicate phase diagram. Am. J. Sci., 271: 97–61.

Holland, T.J.B., 1980. The reaction albite = jadeite + quartz determined experimentally in the range 600–1200 °C. Am. Mineral., 65: 129–134.

Holland, T.J.B. and Richardson, S.W., 1979. Amphibole zonation in metabasite as a guide to the evolution of metamorphic conditions. Contrib. Mineral. Petrol., 70: 143–148.

Hopson, C.A., Coleman, R.G., Gregory, R.T., Pallister, J.S. and Bailey, E.H., 1981. Geologic section through the Samail ophiolite and associated rocks along a Muscat-Ibra transect, Southeastern Oman Mountains. J. Geophys. Res., 86: 2527–2544.

Johannes, W. and Puhan, D., 1971. The calcite–aragonite transition reinvestigated. Contrib. Mineral. Petrol., 31: 28–38.

Krogh, E.J. and Raheim, A., 1978. Temperature and pressure dependence of Fe–Mg partitioning between garnet and phengite with particular reference to eclogites. Contrib. Mineral. Petrol., 66: 75–80.

Lanphere, M.A., 1981. K–Ar ages of metamorphic rocks at the base of the Samail ophiolite, Oman. J. Geophys. Res., 86 (B4): 2777–2782.

Le Mer, O., Ouazzani-Touhami, M., Montigny, R., Goffé, B. and Michard, W., 1986. Chronologie du métamorphisme à haute pression sous l'obduction omanaise. In: Réuion des Sciences de Terres, 11. Soc. Géol. Fr., Clermont-Ferrand, p. 110.

Le Métour, J., 1987. Géologie de l'autochtone des montagnes d'Oman dans la fenêtre du Saih Hatat. Thèse d'Etat, Univ. Paris, Paris 6, 420 pp.

Le Métour, J., Rabu, D., Tegyey, M., Béchennec, F., Beurrier, M. and Villey, M., 1986. Le métamorphisme régional crétacé de faciès éclogites-schistes bleus sur la bordure omanaise de la plateforme arabe: conséquence d'une tectogenèse précoce anté-obduction. C.R. Acad. Sci. Paris, Sér 2, 302: 905–910.

Le Pichon, X., Bergerat, F. and Roulet, M.J., 1987. Plate kinematics and tectonics leading to the Alpine belt formation. Geol. Soc. Am. Spec. Pap., 218: 111–131.

Lippard, S.J., 1983. Cretaceous high pressure metamorphism in NE Oman and its relationship to subduction and ophiolite nappe emplacement. J. Geol. Soc. London, 140: 97–104.

Lippard, S.J., Shelton, A.W. and Gass, I.G., 1986. The ophiolites of northern Oman. Geol. Soc. London Mem., 11: 178 pp.

Lovelock, P.E.R., Potter, T.L., Walsworth-Bell, E.B. and Wiemer, W.M., 1981. Ordovician rocks in the Oman mountains: The Amdeh formation. Geol. Mijnbouw, 60: 487–495.

Madon, M. and Gillet, Ph., 1984. A theoretical approach to the kinetics of calcite–aragonite transition: application to laboratory experiments. Earth Planet. Sci. Lett., 67: 400–414.

Maresh, W.V., 1977. Experimental studies on glaucophane: an analysis of present knowledge. Tectonophysics, 43: 105–125.

Massonne, H.J. and Schreyer, W., 1987. Phengite geobarometry based on the limiting assemblage with K-feldspar, phlogopite and quartz. Contrib. Mineral. Petrol., 96: 212–224.

McCulloch, M.T., Gregory, R.T., Wasserburg, G.J. and Taylor Jr., H.P., 1981. Sm–Nd, Rb–Sr, and $^{18}O/^{16}O$ isotopic systematics in an oceanic crustal section: evidence from the Samail ophiolite. J. Geophys. Res., 86: 2721–2735.

Michard, A., 1982. Contribution à la connaissance de la marge nord du Gondwana: une chaîne plissée paléozoïque, vraisemblablement hercynienne, en Oman. C.R. Acad. Sci. Paris, Sér. 2, 295: 1031–1036.

Michard, A., 1983. Les nappes de Mascate (Oman), rampe épicontinentale d'obduction à faciès schiste bleu, et la dualité apparente des ophiolites omanaises. Sci. Geol. Bull., 36: 3–16.

Michard, A., 1987. L'obduction. La Recherche, 186: 312–322.

Michard, A., Bouchez, J.L. and Misseri, M., 1981. Les nappes métamorphiques de Mascate, nouvel élément infra-ophiolitique en Oman. Terra Cognita, 1: 19 (Abstr.).

Michard, A., Goffé, B. and Ouazzani, M., 1983. Obduction related high pressure, low temperature metamorphism in upper crustal materials, Oman. Terra Cognita, 3: 187–188 (Abstr.).

Michard, A., Bouchez, J.L. and Ouazzani-Touhami, M., 1984. Obduction-related planar and linear fabrics in Oman. J. Struct. Geol., 6: 39–49.

Michard, A., Juteau, T. and Whitechurch, H., 1985. L'obduction: revue des modèles et confrontation au cas de l'Oman. Bull. Soc. Géol. Fr., 8 (1): 189–198.

Montigny, R., Le Mer, O., Thuizat, R. and Whitechurch, H., 1987a. K–Ar time table for the obduction of the Oman ophiolite. Terra Cognita, 7: 105.

Montigny, R., Le Mer, O., Thuizat, R. and Whitechurch, H., 1987b. Chronologie de l'obduction ophiolitique en Oman. In: Géologie de l'Oman. Séance Spéc. Soc. Géol. Fr. (Abstr.).

Nitsch, K.H., 1972. Das P-T-XCO_2 stabilitätsfeld von Lawsonit. Contrib. Mineral. Petrol., 34: 116–134.

Nitsch, K.H., 1974. Neues Erkenntnisse zur Stabilität von Lawsonit. Fortschr. Mineral., 51: 34–35.

Ouazzani-Touhami, M., 1986. Structures et cristallisations associées dans des zones de cisaillement: nappes de Mascate (Oman) et nappes de Federico (Rif interne, Maroc). Thèse Doct., Univ. Louis Pasteur, Strasbourg, 170 pp.

Paradis, S., Velde, B. and Nicot, E., 1983. Chloritoid–pyrophyllite–rectorite facies rocks from Brittany, France. Contrib. Mineral. Petrol., 83: 342–347.

Pearce, J.A., Alabaster, T., Shelton, A.W. and Searle, M.P., 1981. The Oman ophiolite as a Cretaceous arc–basin complex: evidence and implications. Philos. Trans. R. Soc. London, Ser. A, 300: 299–317.

Platt, J.P., 1986. Dynamics of orogenic wedges and the uplift of high pressure metamorphic rocks. Geol. Soc. Am. Bull., 97: 1037–1053.

Pognante, U. and Kienast, J.R., 1987. Blueschist and eclogite transformation in Fe–Ti gabbros: a case from the Western Alps ophiolites. J. Petrol., 28: 271–292.

Rabu, D., 1987. Géologie de l'autochtone des montagnes d'Oman dans la fenêtre du Jabal Akhdar. Thèse d'Etat, Univ. Paris, Paris, 420 pp.

Rao, B.B. and Johannes, W., 1979. Further data on the stability of staurolite + quartz and related assemblages. Neues. Jahrb. Mineral. Monatsh., 10: 437–447.

Reinhardt, B.M., 1969. On the genesis and emplacement of ophiolites in the Oman mountains geosyncline. Schweiz. Mineral. Petrogr. Mitt., 49: 1–30.

Robertson, E.C., Birch, A.F. and Macdonald, G.J.F., 1957. Experimental determination of jadeite stability relation to 25,000 bars. Am. J. Sci., 255: 115–137.

Rubie, D.C., 1984. A thermal–tectonic model for high-pressure metamorphism and deformation in the Sesia Zone, Western Alps. J. Geol., 92: 21–36.

Schaaf, A. and Thomas, V., 1986. Les radiolaires campaniens du Wadi Ragni (nappe de Semail, Oman): un nouveau repère chronologique de l'obduction omanaise. C.R. Acad. Sci. Paris, Sér. 2, 303: 1593–1598.

Schreyer, W., 1985. Metamorphism of crustal rocks at mantle depths: high-pressure minerals and mineral assemblages in metapelites. Fortschr. Mineral., 63: 227–261.

Schreyer, W., 1987. Experimental studies on metamorphism of crustal rocks under mantle pressures. Hallimond Paper. Mineral. Mag., 364 (52): 1–26.

Searle, M.P., 1985. Sequence of thrusting and origin of culminations in the northern and central Oman mountains. J. Struct. Geol., 7: 129–143.

Searle, M.P., Lippard, S.J., Smewing, J.D. and Rex, D.C., 1980. Volcanic rocks beneath the Semail ophiolite nappe in the northern Oman mountains and their significance in the Mesozoic evolution of the Tethys. J. Geol. Soc. London, 137: 589–604.

Seidel, E., 1983. Fe–Mg-carpholite, indicator of high P/low T metamorphsim. Terra Cognita, 3: 190.

Shelton, A.W., 1984. Geophysical studies on the northern Oman ophiolite. Ph.D. Thesis, Open Univ., Milton Keynes, 323 pp.

Tegyey, M. and Le Métour, J., 1987. Mise en évidence de jadéite dans un basalte à augite primaire préservée, Permien autochtone du Saih Hatat, Oman. Bull. Mineral. Suppl., 110: 27 (Abstr.).

Theye, Th., 1988. Aufstiegende Hochdruck-metamorphose im Sedimerten der Phyllit–Quartzit—Einheit Kretas und der Peloponnes. Dissert., Brawnschweig, 223 pp.

Theye, Th. and Seidel, E., 1987. Indikatorminerale in hochdrukmetamorphen Marmoren Kretas: Sudoit, Lawsonit, Aragonit. Fortschr. Mineral., 65: 181.

Thompson, A.B. and Ridley, J.R., 1987. Pressure–temperature–time histories of orogenic belts. Philos. Trans. R. Soc. London, Ser. A, 321: 27–45.

Tilton, G.R., Wright, J.E. and Hopson, C.A., 1981. Uranium–lead isotopic ages of the Samail ophiolite, Oman, with applications to Tethyan ocean ridge tectonics. J. Geophys. Res., 86: 2763–2775.

Tippit, P.R., Pessagno, E.A. and Smewing, J.D., 1981. The biostratigraphy of sediments in the volcanic unit of the Samail ophiolite. J. Geophys. Res., 86: 2756–2762.

Vieten, K. and Hamm, H.M., 1978. Additional notes on "the calculation of the crystal chemical formula of clinopyroxenes and their content of Fe^{3+} from microprobe analysis". Neues Jahrb. Mineral. Monatsh., 2: 71–83.