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Accretion of Oman and United Arab Emirates ophiolite – Discussion of a new structural map

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Abstract

This study is altogether an extended legend for the folded maps incorporated in this volume, a review of the current knowledge on the Oman-United Arab Emirates ophiolite belt, and a new synthesis at the scale of the entire belt. Following a brief description of the petrological and structural units composing the ophiolite, the content of the three structural maps (planar structures, linear structures and dikes) is presented. Next, we discuss the various constraints introduced by these data in view of a synthesis of the ophiolite belt in terms of an ocean floor spreading system. Because they are the link between the factual results summarised in the maps, and the ridge models, these constraints are critical (and are central to the structural analysis of ophiolites). They introduce severe limits to possible ridge models such as those proposed in the conclusion. After being reassembled in a best geometrical and structural fit, the belt is parted into three domains (Figure 1). The south-eastern and central domains (from Wadi Tayin to Haylayn and possibly Sarami massifs) incorporate a 40–50 km-wide and possibly over 200 km long new ridge segment, oriented NW–SE, which is opening into a 1–2 My older lithosphere oriented NE–SW. The northern domain (from Khawr Fakkan to Hilti massifs) is well explained by a model of propagating (Aswad) and failing (Fizh) ridge segments of nearly parallel NNW-SSE orientation which are separated by a 10–20 km-wide transform zone covering the north of Fizh massif. This new synthesis integrates and updates the local syntheses published so far. It illustrates again the contrast between locally simple ridge segments organised around mantle diapirs and the tectonic complexity of the two larger domains, with, as an example, sheared mantle, vertical Moho and dismembered lower crust with hydrous contamination, near the tip of ridge propagators. The relation between the northern and central-southern domains is obscure because the paleomagnetic results suggest that, with respect to the central-southern domain, the northern domain should have rotated 130° clockwise. Such a large rotation of a 200 km long domain is difficult to explain, inasmuch as the age constraints seem to restrict the possible duration of the rotation to a couple of Myr. The preferred model consists in a progressive clockwise rotation during tectonic accretion of increasingly larger blocks. This is initiated in the northern massifs, progressing over 1–2 Myr, southward to finally integrate the entire ophiolite.

Introduction

Modern geological mapping of Oman and its ophiolite was initiated by K.W. Glennie's group of petroleum geologists. In 1974, they produced a remarkable 1:500 000 map of northern Oman and the United Arab Emirates (Glennie et al., 1974). This achievement was followed in the 1980s by a program of the Ministry of Petroleum and Minerals of Oman, involving mainly the Bureau de Recherche Géologique et Minière and the Bishimetal Exploration Co., Ltd. This program published in 1986 and 1987 a set of 1:100 000 and, locally, 1:50 000 maps, covering the entire ophiolite in northern Oman. Regarding specific areas of the ophiolite, a US Geological Survey group produced a detailed, 1:100 000 geological and structural map of the Wadi Tayin massif (Coleman and Hopson, 1981) and the Open University group, a series of 1:100 000 maps in the northern part of Oman ophiolite (Open University, 1982).

Much academic work in the 80's regarding the ophiolite was concerned with its geodynamic environment of origin with discussions whether it was an ocean (Boudier and Coleman, 1981; Boudier et al., 1988) or a back-arc (Lippard, 1986) in relation with ophiolite oceanic detachment and obduction onto the Arabian margin. Since 1980, not much has been added to this debate which is not specially pursued here. The main object of this study is the functioning of the ridge system of origin of the ophiolite, whatever the geodynamic environment.

Since 1981, the beginning of our group's continuous activity in the Oman ophiolite, we have been convinced that a detailed structural and kinematic mapping was necessary before the ophiolite could be correctly interpreted. This long-term project ends with the present paper. It involved, on average, 3–5 geologists mapping during 1–2 months each winter. It has successively resulted first in a special volume (Boudier and Nicolas, 1988) which presented a general structural map of the entire ophiolite nappe, and in a number of publications in the proceedings volume of the International ophiolite conference organised in 1990 in Oman (Peters et al., 1991). The most significant result of the early mapping, which was mainly devoted to mantle structures, has been the discovery of several mantle diapirs. They are characterised in map by the occurrence of locally steep foliation planes and lineations in high-temperature peridotite which point to locally steep mantle flow. Agravity map of Oman (Ravant et al., 1997) shows that the ophiotite nappe is still locally as thick as 10 km.

This intensive study was undertaken because the Oman–United Arab Emirates (O–UAE) ophiolite (initially called by Glennie and co-workers the Semail ophiolite, with a possible confusion with the Sumail massif) is probably the largest and best preserved ophiolite in the world with, among other benefits, excellent outcropping conditions and easy access to the field. After the first phase of mapping, it also became clear that this ophiolite presented other points of major interest. As the internal structure and composition are, on the first order, homogeneous or coherent over large areas, the tectonic "noise" related to obduction and post-obduction events can be considered to be weak. Indeed, recent studies suggest that most tectonic deformations are ridge-related, thus minimising the role of later events (Nicolas and Boudier, 1995). All authors have concluded that the Oman–UAE ophiolite is derived from a fast-spreading paleo-ridge system (Pallister and Hopson, 1981; Tilton et al., 1981; Nicolas, 1989; MacLeod and Rothery, 1992; Nicolas et al., 1994a; Boudier et al., 1997). Spreading rates estimated in independent ways yield results of 5–10 cm yr^{-1} (Hacker et al., 1996), 12 cm yr^{-1} (Nicolas et al., in press), and 6 cm yr^{-1} (see below). This means that the accretion process was dominantly steady-state, a conclusion also explaining the internal homogeneity of the ophiolite over large distances. The assumption of an essentially time-independent functioning helps to understand the fundamental accretion mechanisms.

During the course of early mapping we also realised that further studies of the ophiolite should be conducted at different scales. Most field-oriented studies in Oman have been conducted at the scale of an individual structure, which can be conveniently mapped. As examples, one can cite studies on the root zone of the sheeted dike complex into the upper gabbros (Rothery, 1983; Nicolas and Boudier, 1991), magma chambers (Pallister and Hopson, 1981; Smewing, 1981; Nicolas et al., 1988a; Boudier et al., 1996; Kelemen et al., 1997), wehrlites (Benn et al., 1988; Ernewein et al., 1988; Juteau et al., 1988a), the Moho Transition Zone (MTZ) (Boudier and Nicolas, 1995), or mantle diapirs (Ceuleneer et al., 1988; Reuber et al., 1991; Jousselin et al., 1998). Because observations can be repeated, often from one wadi to the next, in generally excellent outcrop conditions, a high degree of confidence can be attained in the studies of minor structures.

Data at this scale can then be integrated at the larger scale of a ridge segment. This is why a second phase of more detailed mapping, incorporating the dynamic structures observed in the crustal units, was started. This phase was illustrated by the integrated study of the southeastern Nakhl-Rustaq, Sumail and Wadi Tayin massifs (Nicolas and Boudier, 1995). In these massifs, it has been possible to map the evolution of the structures from diapirs, presumably located in the centre of a segment, all the way to the end of such a segment where the magma chamber progressively vanishes, because in the Sumail and Wadi Tayin massifs, the field structures related to the ridge activity are continuous, and because they are not disturbed by subsequent tectonic events. Therefore, the results obtained at this scale are considered to be robust ; subsequent mapping has not significantly changed the structures presented in our earlier publication (Nicolas and Boudier, 1995). Similarly, building on a large number of preceding studies and detailed mapping projects, our recent synthesis of the UAE part of the belt (Nicolas et al., in press) results in models of a propagating ridge system, which are firmly based on field data, their differences reflecting the limits of our present knowledge.

Finally, a synthesis can be envisaged at the scale of the entire Oman–UAE ophiolite, that is, the scale of a 500 km long ridge system. This is what is attempted in this paper. At this large scale, we face considerable complexity because of discontinuities between massifs, because of the intense ridge tectonics, mainly recorded in the central and northern part of the ophiolite, and because of paleomagnetic results implying very large differential rotations (Thomas et al., 1988; Perrin et al., 1994; Weiler, this issue). Consequently, in contrast with the previous, more regional studies, a synthesis at the scale of the whole ophiolite is largely speculative. After a general presentation of the structural maps, enclosed as separate folded plates in the back of this issue, we present the different data sets, which are mostly derived from the structural analysis, and we summarise the current state of understanding of this ophiolite.

Table 1. Area of the different geological formations in the enclosed folded maps, given as a fraction (%) of the total ophiolite surface. The second group of values gives the 5 main formations only.

Geological formation	Area (%)
Plagiogranites	0.2
V2 lavas	2
V1 lavas	3.3
Sheeted dike complex	7.4
Foliated (upper) gabbros	7.4
Laminated and layered (lower gabbros)	23.2
Dunites/wehrlites (HT)	5
Dunites/wehrlites (LT)	0.1
Harzburgites (HT)	34.6
Harzburgites (MT)	1
Harzburgites (LT)	14.1
Metamorphic sole	1.6
Lavas	5.5
Sheeted dike complex	7.6
Gabbros	31.1
Dunites/wehrlites	5.2
Harzburgites	50.6

Presentation of the maps

A new geological map of the entire ophiolite belt is the basis for three distinct structural maps, presenting, respectively, planar features (folded map 1), lineations (folded map 2) and dikes (folded map 3). Simplified contours of the Bougner gravity anomaly map of Oman (Ravant et al., 1997) are drawn on the back of the folded maps. The geological map is largely inspired by the published maps mentioned above, except for some new internal structures, mainly in the peridotites, a partly new drawing of the Moho and a new division of the gabbro section into lower and upper gabbros. A by-product of this digitised map is that it is straightforward to obtain, for the entire ophiolite, the fractional area of each component of the ophiolite (Table 1). The structural maps are based on 6000 field stations (Figure 1), where the data described below have been collected. This mapping applies the concepts and techniques of structural analysis which have been already described (Nicolas, 1989; Nicolas, 1992) and which are recalled in the recent field study of the southern massifs (Nicolas and Boudier, 1995). They are summarised below for the sake of convenience.

Peridotite unit

In all peridotites, the foliation (plane of mineral flattening) and lineation (direction of mineral elongation within this plane) are systematically measured in the field and/or on oriented, bleached, and sliced specimens in the lab. Oriented thin sections are used to deduce the kinematics of plastic flow and the physical conditions of deformation (Figures 2a and 2b). High-Temperature (HT) deformation relates to hypersolidus or solidus temperatures (∼1250 °C), and thus to large-scale plastic flow below the ridge of origin. Low-Temperature (LT, 1000–900 °C) deformations, and locally Medium-Temperature (MT, 1100–1000 °C), relate to localised motion along shear zones and detachment thrusts. LT/MT shear zones inside massifs may relate either to an off-axis activity (see Nicolas and Boudier, 1995), or to the detachment thrusts of the ophiolite, a circumstance best documented in the northern part of the ophiolite belt (Boudier et al., 1988). The LT and MT zones are represented in lighter tunes on maps.

Other mapped structures in peridotites include the dunitic or pyroxenitic banding, which is generally parallel to the foliation as a result of tectonic transposition; when the banding is folded, the foliation

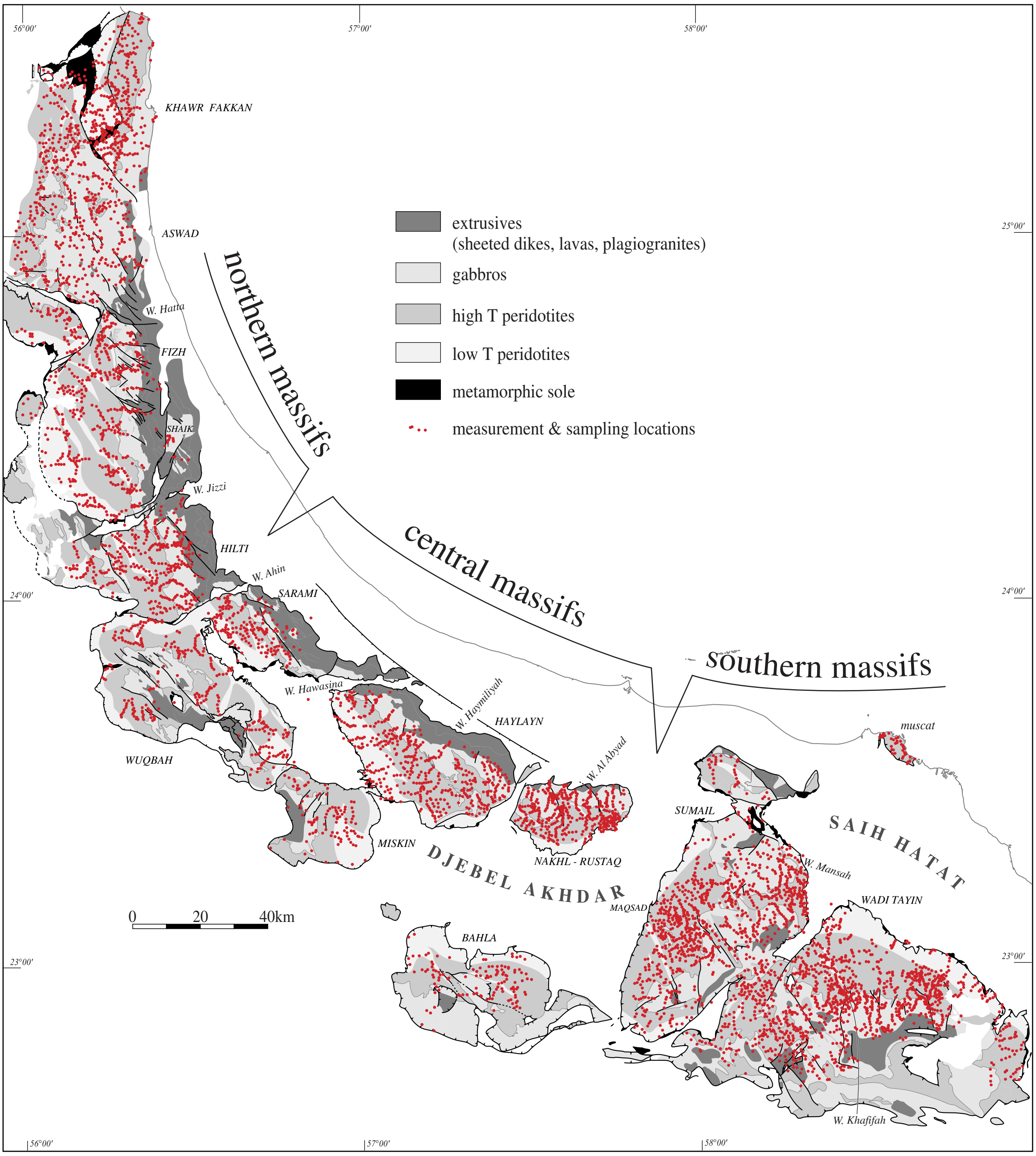

Figure 1. Locations of the 6000 field stations on which are based the structural maps (folded maps 1, 2, 3).

coincides with the axial plane of the fold and the mineral lineation, with the fold axis direction.

MTZ-Moho-Wehrlites

We distinguish a Moho Transition Zone (MTZ), dominantly made of dunites, on top of the harzburgites and below the layered gabbros (Boudier and Nicolas, 1995). Its thickness varies from a few meters to a few hundred meters but it is present everywhere. It is mapped only when it is thicker than 100 m. A thin MTZ marks a straight and locally spectacular limit, whereas in a thick MTZ, large gabbro sills within strained dunites or harzburgites, and large wehrlite sills and intrusions in the lower gabbros render the limit difficult to locate. The boundary between the MTZ and the layered gabbros is considered to be the Moho. Large wehrlite intrusions in the crustal units have been represented in the maps with the same decoration as the dunites and impregnated dunites of the MTZ because of their petrological similarity, and sometimes their physical continuity. Both the dunites and wehrlites are concentrated in areas where melt has actively interacted with the uppermost peridotites (Boudier and Nicolas, 1995; Jousselin and Nicolas, this issue).

Gabbro unit

We follow our earlier structural distinction (Nicolas et al., 1988a) between lower, layered gabbros and upper foliated gabbros, with an intermediate facies of laminated gabbros in-between. Gabbros are ubiquitously deformed by magmatic flow which induces a magmatic foliation, and a mineral lineation (Figure 2c) ; both are carefully mapped in the field. Magmatic foliation is parallel to the compositional layering, and mainly developed in the lower gabbros. Upsection, and typically in the foliated gabbros, the layering tends to disappear. On average, the foliated gabbros represent one third of the gabbro sequence (Table 1) ; they coincide with a progressive steepening of the layering-foliation (see below). Kinematic analysis relies on field (tiling, magmatic shear bands) and thin section (mainly tiling) observations ; thin sections also allow an estimate of the amount, generally negligible, of HT plastic deformation following the magmatic flow (Figure 2d). Amphibolite facies shear bands, occurring as steeply dipping mylonites with horizontal lineations, are altogether uncommon and generally only meter-thick. They are exceptional in the southern massifs, but comparatively more abundant and wider in the northern Fizh massif. It should be pointed out that normal listric faults and shear zones associated with flaser gabbros are unknown in Oman, in contrast with ophiolites related to slow-spreading environments (Boudier and Nicolas, 1985; Boudier et al., 1989; Cannat and Lécuyer, 1991; Alexander and Harper, 1992; Nicolas et al., 1999), and with oceanic gabbros from slow-spreading ridges (Cannat et al., 1991).

Upper crustal unit

For structural investigations on the paleo-ridge of origin, the attitude of the sheeted dike complex is particularly important, as it gives the local trend of the ridge (Nicolas, 1989, p. 9). With exceptions such as the northern Fizh massif, where this complex has been rotated to moderate dips before being cut by a new steep dike complex (MacLeod and Rothery, 1992), and locally disturbed zones close to the front of the ophiolite nappe, the sheeted dike complex is generally steep and at large angle to both the overlying lavas and to the underlying Moho (Nicolas et al., 1996). As pointed out by MacLeod and Rothery (1992), this is another difference with slow-spreading ridges and with ophiolites related to such environments, where diabase dikes are variably rotated. Because the sheeted dikes indicate the presumed ridge trend and point to the paleo-vertical, this unit has been measured systematically (about 10 to 100 individual measurements for each field station). In the northern massifs, for both the sheeted dikes and the extrusives, we often rely on measurements made by the other groups and represented in the 1:50 000 and 1:100 000 Oman geological maps.

Dikes and veins

Several generations of dikes and veins of various composition are measured in the different ophiolite units. They are described in more detailed in Nicolas et al. (this issue; see also pictures on folded map 3). In peridotites, the earliest ones are emplaced in melting peridotites, as olivine gabbros, troctolites, websterites, and more exceptionally chromite dikes bounded by a dunite reaction rim. They are called 'reactive dikes'. Dunite veins are included in the same group since they are thought to represent the walls of a melt conduit which, during closure, totally expelled its residual melt. Other mafic dikes are emplaced in cooling peridotites at progressively decreasing temperatures. These dikes comprise websterites, coarse grained to fine grained norites or olivine gabbros, and, eventually, diabase dikes; diabase dikes emplaced early have no

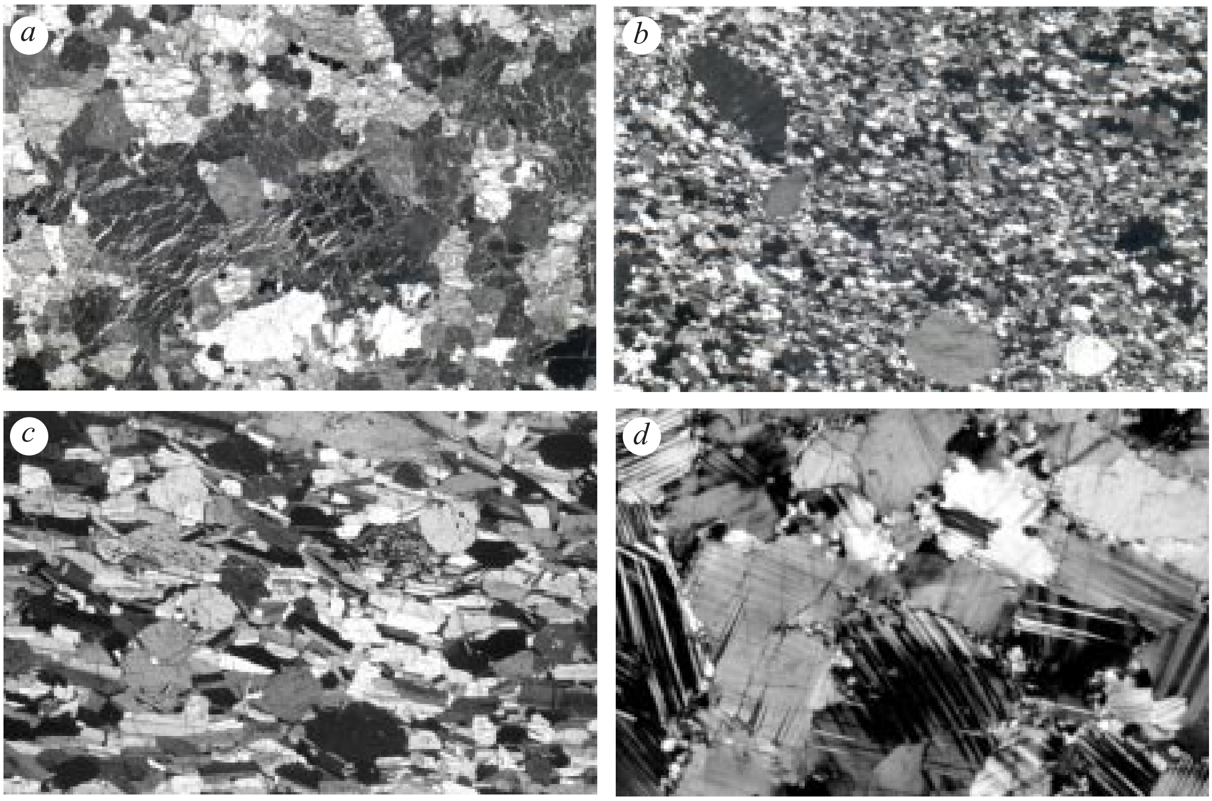

Figure 2. Microstructures in peridotites (a and b) and gabbros (c and d), cross-polarisers, plane normal to foliation and parallel to mineral lineation, both traces horizontal. (a) Hypersolidus, HT deformation ($\approx$ 1250 °C) in harzburgites. (b) Subsolidus, LT deformation ($\approx$ 1000 °C) in harzburgites. (c) Magmatic flow deformation ($\approx$ 1200 °C) in gabbros. (d) Plastic flow deformation ($\leq$ 1100 °C) in gabbros. Field of view : (a) 15 mm; (b) 8 mm; (c) 12 mm; (d) 5 mm.

chilled margins and those emplaced later, at host-rock temperatures $\leq$450 °C, do have chilled margins.

Mafic dikes in gabbros are mostly diabase dikes. Noritic dikes are less common than olivine-gabbro dikes but appear locally in abundance, suggesting water contamination in the source zone (Benoit et al., 1996). On the other hand, swarms of hydrothermal veins with a whitish alteration in greenschist facies conditions are very common ; one set is generally parallel to the overlying sheeted dike complex. Dikes and irregular bodies of hydrous plagiogranite are principally developed at the top of the upper gabbros. They commonly constitute magmatic breccias with a basaltic component. Dikes of amphibole gabbro or diorite are present in all massifs, at all levels, including the lowermost gabbros. They are related to the hydrothermal veining, and indicate that there is a HT hydrothermal circulation down to the bottom of the crust near the ridge (Nicolas et al., this issue; see also below).

Occurrences of diabase dikes and sills in the uppermost mantle and lower crustal formations have been identified on maps, when they may reflect a renewed magmatic activity in the crust of a cooling lithosphere (tips of propagators, off-axis seamounts, see below).

Faults

Peridotites are moderately (about 30 to 50% on average), but ubiquitously, serpentinized with a mesh structure pointing to water penetration and reaction in static conditions. Bands of schistose serpentinites, a few meters to several tens of meters wide, are related to local shearing and faulting. Only the major faults have been represented on the maps and we recognise that the role of brittle LT deformation in the mantle section has been underplayed here. The coherence of foliation and lineation trajectories in mantle formations, even where they are affected by bands of schistosed serpentinites, suggest that these bands induce mainly moderate translation displacements without significant rotations. For this reason, the serpentinite bands in the mantle section unlikely affect the large-scale structures. This subject, however, deserves further study. On the other hand, in the lower crustal units where shear displacements can be better evaluated because of the contrasted lithology, most faults represented in some former geological maps happen to be hydrothermal fractures with limited, or no, displacement (see below). Typically these fractures in aerial photographs appear as straight lineaments which are parallel to the overlying sheeted dike complex.

Environment of the ophiolite

We have chosen to represent in maps only the metamorphic sole below the ophiolite (Ghent and Stout, 1981; Lippard et al., 1986; Boudier et al., 1988) and the underlying Hawasina formations, whose imbricated nappes represent the sedimentary and volcanic content of the marine and epicontinental domains located between the paleo-ridge of origin and the Arabian present coast line (Bechennec et al., 1990; Pillevuit, 1993). All these formations have been thrust onto the Arabian margin and its proximal formations which are not mapped here ; neither have we mapped all the sedimentary formations deposited onto the obducted ophiolite.

Design of the maps

The structural maps (folded maps in this issue) have been made following the same principles used in Nicolas et al. (1988b) and Nicolas and Boudier (1995). In the mantle section, an average density of one station for every square kilometre is sufficient to keep track of the structural continuity. In the more heterogeneous gabbros, this density is more irregular.

The location of our main field stations (Figure 1), shows a contrast in the density of measurements. The higher density in the Aswad (Nicolas et al., in press) and in the southern and central massifs (Nicolas and Boudier, 1995; Jousselin et al., 1998) is justified by the exemplary character of these massifs in understanding the functioning of ridge segments. In some areas, the density of measurements is lower, mainly along the western front of the nappe where the ophiolite is thinner, and intensely deformed during obduction, and where we considered that the overall disturbed structure would have required too much involvement for limited benefit. For the maps presenting a vast number of individual measurements, it becomes necessary to draw trajectories in order to visualise the trends of the corresponding structures. We realise that this exercise is partly subjective. The maps contain an enlarged inset of the southern massifs with both the measurements and the trajectories, to allow the reader to estimate the pertinence of the drawn trajectories. Field measurement data are available upon request, or in the laboratory web page (http://www.isteem.univ-montp2.fr/TECTONOPHY). Eventually, in keeping with the principles of kinematic analysis in rocks affected by large, solid-state flow (Nicolas and Poirier, 1976), or magmatic flow (Nicolas, 1992), trajectory maps of foliations and lineations are considered to be good proxies for flow planes and flow lines, respectively.

Map of planar structures

The folded map 1 presents the field orientations of planes of plastic foliation in the peridotite section, of magmatic layering and foliation in the lower and upper gabbros, and of bedding planes in basaltic flows and overlying sediments. In peridotites, we did not distinguish high-temperature (HT), low-temperature (LT) and locally medium-temperature (MT) foliation trajectories by different colours or symbol, as these different temperature conditions are already identified in the geological background. In the trajectory map of foliations (folded map 1), mainly foliations steeper than 40–45° have been represented because the trajectories drawn for flat-lying planes are too sensitive to minor orientation changes and, consequently, not meaningful. For flat foliations, only general orientations are indicated on the map. Steep foliation trajectories in the HT peridotites, where they are contorted and associated with steep lineations and thick MTZ, have been ascribed to uprising flow inside mantle diapirs (Nicolas and Violette, 1982; Nicolas and Boudier, 1995; Jousselin et al., 1998). In such areas, the 45° and 70° isodip curves delineate the contours of the diapir. Otherwise, steep foliations are generally seen in transcurrent shear zones associated with flat lineations. Within these shear zones, LT peridotites are highly strained with abundant syntectonic recrystallization. There is, however, one noteworthy exception, which is a steep SW–NE band, 20 km wide, in the middle of the Wadi Tayin massif with HT mantle peridotites and moderate deformation. This structure has been interpreted previously as a transform fault (Ceuleneer et al., 1988). Eventually, other areas, mainly in the central and northern massifs, present contorted steep foliations with variably oriented lineations. They correspond to large-scale folding of the mantle section in MT/LT conditions.

In gabbros, LT trajectories are distinctly represented by dashed lines of different colour. Steep HT layering is common and, in areas where the Moho is regionally flat lying like in the southern massifs, the steepening is due either to large-scale magmatic folds such as those in the Khafifah area of Wadi Tayin massif and in Al-Abiad oasis of Nakhl-Rustaq massif (Nicolas and Boudier, 1995), or to the intrusion of wehrlites in still hot and hardly deformable lower crust. Moreover, in upper, foliated gabbros, the foliation steepens progressively and tends to become parallel to the sheeted dikes, near their rooting zone (Smewing, 1981; Nicolas et al., 1988a; MacLeod and Rothery, 1992; Nicolas et al., 1996). We have conducted a specific study to check this foliation steepening and to trace the magmatic flow direction as deduced from the mineral lineations. In areas throughout the ophiolite where foliations and lineations could be related to the trend of the overlying sheeted dike complex and the distance to the base of this complex determined, the foliations and lineations have been projected in the sheeted dike reference frame (local average), itself rotated to the N–S and vertical orientation. Three categories are reported in Figure 3, depending on the depth, less than 100 m, 100–500 m, and over 500 m, with regard to the base of the sheeted dike complex. At all depths, the gabbro foliation display a dominant orientation close to that of the sheeted dikes with, surprisingly, no significant flattening with depth in the investigated depth range (Figure 3a). Thus, the flattening of foliation occurs below the level of foliated gabbros, and at depths significantly greater than 500 m. The lineations are rather steep in all three diagrams (Figure 3b); at shallow depth they tend to fan in the direction of the sheeted dikes, and at greater depth, in the direction perpendicular to the sheeted dike strike. As discussed below, the sense of rotation of the foliation in upper gabbros has been used to infer the side of the paleo- ridge from which the considered section issued. Eventually, for all these reasons, areas of steep foliation in gabbros have not been contoured on the map.

Map of linear structures

Folded map 2 presents the field orientations of mineral lineations, induced by plastic or magmatic deformation, in peridotites and in gabbros. By contrast with foliations, trajectories of lineations are more consistent when they are horizontal, or moderately plunging, than when they are steep. Lineations steeper than 30°, 45° and 60° have been contoured. As previously discussed (Nicolas and Boudier, 1995; Jousselin and Nicolas, this issue), the steep lineations are present mainly in diapiric areas.

The trajectory map of lineations shows no orientation discontinuity between solid-state flow in mantle formations and magmatic flow in lower gabbros. This observation has been made repeatedly in Oman and has led to the concept of a strong mechanical coupling between flows in the two domains (Nicolas et al., 1994a; Chenevez et al., 1998). This way, the direction of mantle flow near the Moho can be inferred, even in areas where the mantle formations are concealed below the lower gabbros.

Finally, the sense of shear has been systematically investigated. In oriented thin sections of plastically deformed peridotites, this information is deduced from the obliquity between lattice and shape fabrics (Nicolas and Poirier, 1976). In the magmatic gabbros, the sense of shear is deduced from heterogeneous deformation, mainly tiling and shear bands (Blumenfeld and Bouchez, 1988). The large number of shear sense determinations made in mantle formations precludes presentation of individual measurements on the map. After regional grouping of these measurements, the arrows on the map point to the dominant number of top/bottom motion measurements over the area surrounding the arrow in cases of shear flow related to flat and moderately dipping planes. A double-head arrow indicates cases of transcurrent shear senses along steep planes and dominantly flat, or weakly plunging, shear lineation. Comparing dominant shear measurements on the map, it can be seen that the shear sense is generally consistent in each area. A shear sense inversion, represented in the map by a red dotted line is common at shallow depth below the Moho (Ildefonse et al., 1995; Michibayashi et al., this issue) (Figure 4). These kinematic indicators are of paramount importance in interpreting the paleo-ridge system, and their use is discussed below. In the gabbros, a given outcrop often yields opposite shear directions in nearly comparable proportions. This suggests that shear is accompanied by a significant compaction component.

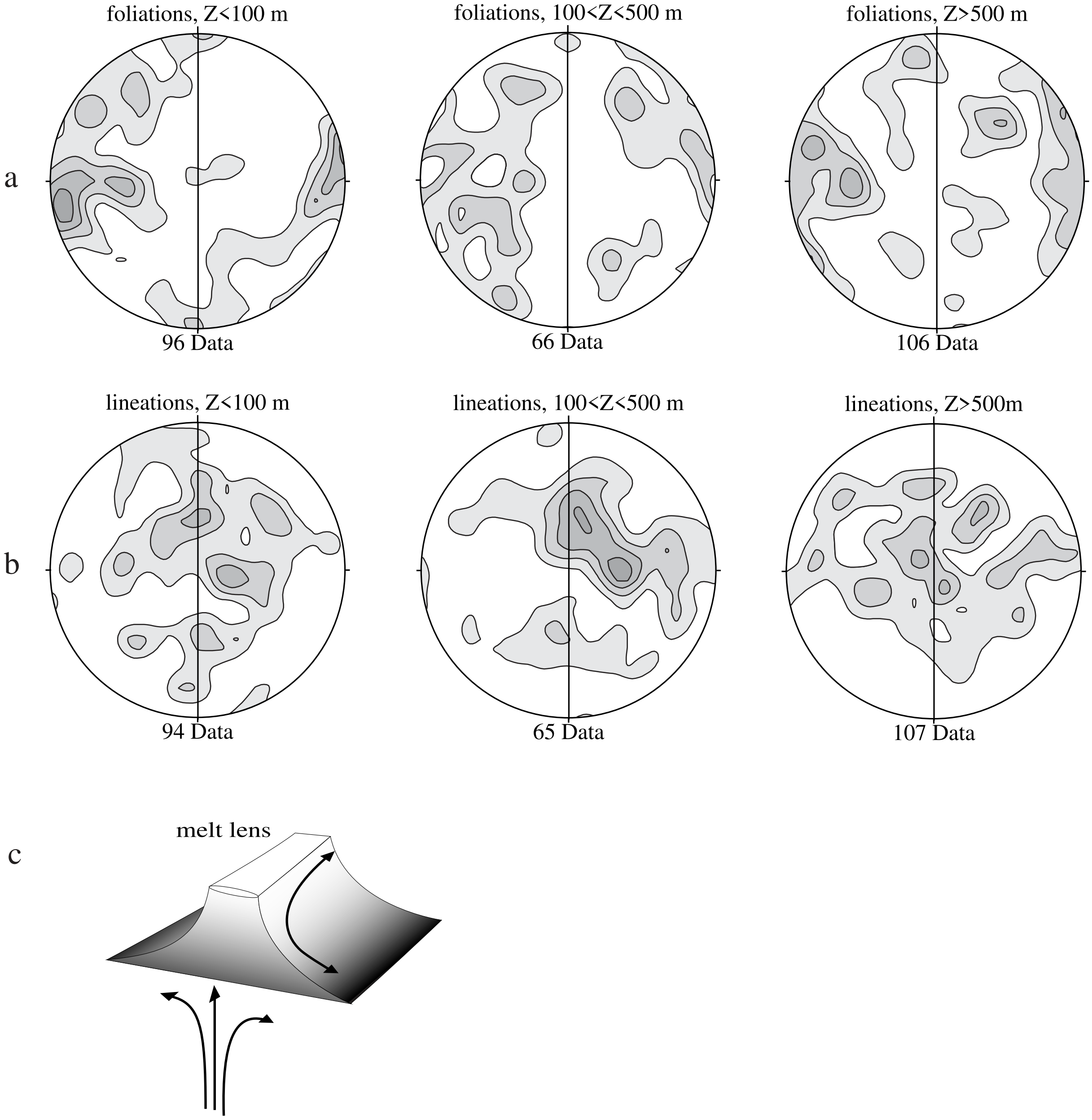

Figure 3. Stereoplots of magmatic foliation (a) and lineation (b) for the upper foliated gabbros after rotation of the close-by diabase sheeted dike complex to vertical and N–S (NS line on stereonet). Each series of stereonets corresponds to measurements at three increasing depths below the sheeted dike root zone. Contours are at 1, 2, × times uniform. (c) Sketch showing upsection steepening of foliations and rotation of lineations from steep-transverse to steep- longitudinal, as deduced from stereoplots a and b.

Considering the time involved in these measurements and the uncertain result about a dominant shear sense , we have given up systematic estimations of shear flow in gabbros.

Map of dikes and sills

Folded map 3 presents the trajectories of dikes. Detailed comments on the maps and stereonets of individual dikes within each massif of the ophiolite are shown in Nicolas et al. (this issue). Drawing a map of dike trajectories rather than representing individ-

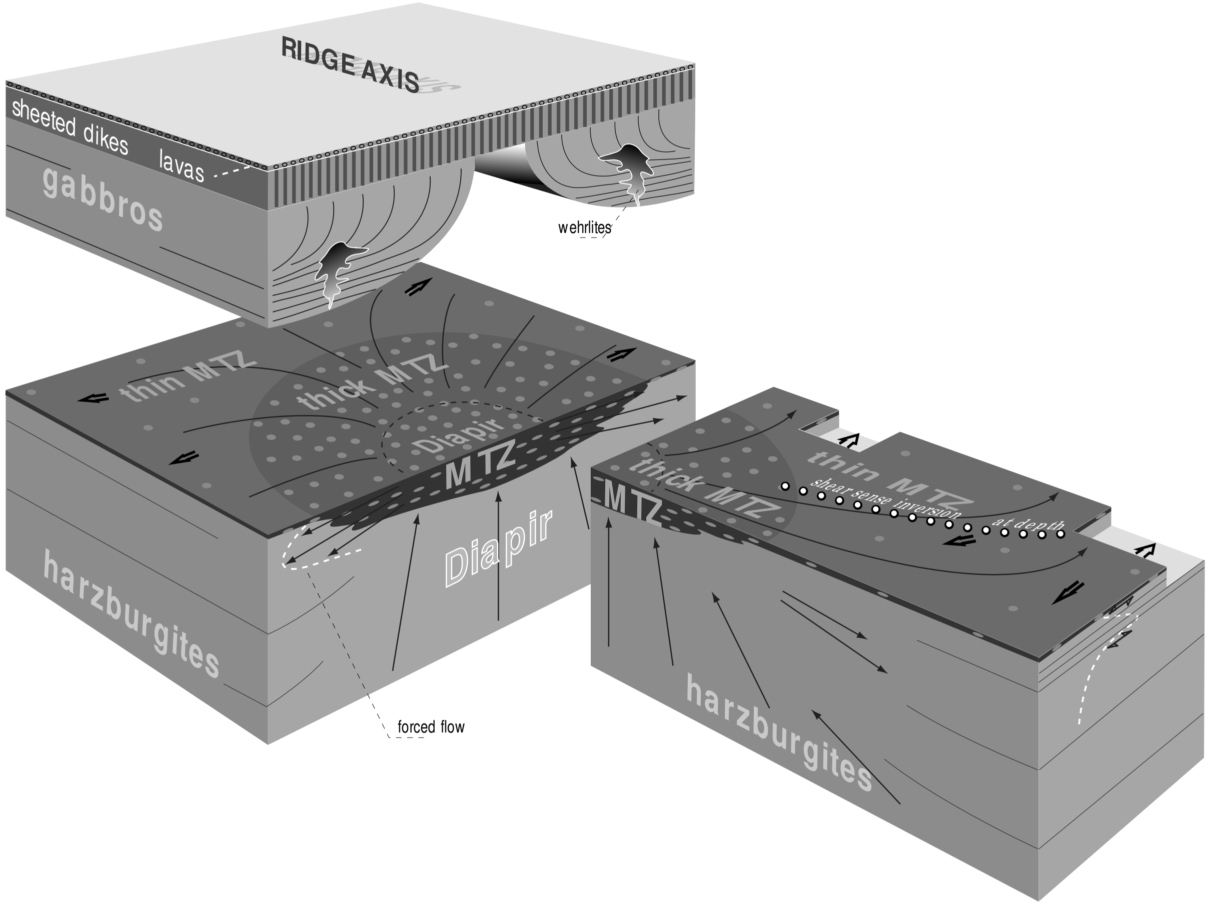

Figure 4. Sketch of a ridge deduced from the study of the O–UAE ophiolite. Below a fairly cylindrical crustal section, the mantle illustrates the contrast between a diapiric head covered by a thick MTZ, from which HT flow diverges with a top-away-to-the-ridge shear, and the surrounding areas associated with thin MTZ and a shear sense inversion at $\approx$ 1 km below the Moho.

ual measurements, which was technically unfeasible, has the merit of pondering individual or local orientations of dikes and sills. We have also been encouraged by the observation that in spite of all uncertainties exposed above, fairly clear local patterns emerge; such patterns are important in view of ascribing the considered area to a spreading direction.

All categories of dikes are oriented dominantly parallel to the overlying sheeted dike complex, whatever their nature and their distance or timing of emplacement with respect to the active paleo-ridge (as deduced from estimation of temperature of country rocks at the time of dike emplacement). Only a second-order analysis can detect local departures from this general conclusion (Ildefonse et al., 1993; Nicolas et al., 1994b; Nicolas et al., this issue).

Principal units within the ophiolite nappe

Ophiolite nappe

The ophiolite nappe is divided into a dozen massifs which are partly or entirely separated by Hawasina stringers. The northern massifs extend from the United Arab Emirates (UAE) to the Wadi Jizzi at the south end of Fizh massif, comprising Khawr Fakkan, Aswad, Fizh and Hilti massifs; the central massifs extend SE, from there to the Jebel Akhdar, comprising Sarami, Wuqbah, Haylayn, Miskin, Nakhl-Rustaq and Bahla; on the eastern side of Jebel Akhdar, the southern massifs comprise Sumail and Wadi Tayin. The small Muscat massif is isolated, but the gravity map (Ravaut et al., 1997, folded maps) suggests that it is continuous with the northern tip of the Sumail massif beneath a cover of post-emplacement sediments. The northern and central massifs, with the exception of

Bahla, Wuqbah and Miskin have in common a moderate eastward dip which probably directly reflects their obduction onto the Arabian margin. In these massifs, an average 20° eastward dip of the nappe has been estimated from the similar dip of the LT harzburgites overlying the metamorphic aureole of the base of the ophiolite, and of the volcanics and associated cherts at its top, as well as from the steep westward dip in most sheeted dikes. The southern massifs have a synform structure with a dominant 10–20° SE dip in Sumail and 20–30° S dip in Wadi Tayin.

Former gravimetric modelling of the structure of the ophiolite (Manghnani and Coleman, 1981; Shelton, 1984) and the new gravity map of Oman (Ravaut et al., 1997; this issue) confirm both eastward dips of 20–30° of the northern and central massifs and the synform structure of the southern massifs. It also indicates that the ophiolite thickness is in the range of 5–7 km. A high gravity spine, which follows the coast SE of Muscat beyond Wadi Tayin, suggests that the true extent of the Oman nappe is about 800 km rather than 500 km. Observation of peridotite outcrops deep within wadis, below thick Tertiary deposits in wadi Tiwi, near Sur (150 km SE in straight line from Muscat), supports the idea of such a SE extension of the ophiolite nappe below Tertiary sediments.

Hawasina formations

The Hawasina formations, covering large areas west of the ophiolite nappe, are considered as having been bulldozed in front of, and below, the initially 10–15 km thick ophiolite nappe, progressing through its marine basin of origin. Accordingly, the Hawasina formations extend beneath the ophiolitic nappe and its metamorphic aureole, and appear as tectonic windows between the ophiolite massifs. It could be envisaged that these Hawasina windows within the ophiolite nappe were opened by locally deeper erosion, provided that such windows are entirely bordered by their naturally overlying formations (namely the metamorphic aureoles and the LT peridotites of the base of the ophiolite nappe). This is partly the situation for the large Hawasina Window separating the Sarami, Haylayn, Miskin and Wuqbah massifs. However, detailed structural studies (Graham, 1980; Bechennec et al., 1990) point to an intrusive contact at least along the NE part of this window (Figure 5). Out-of-sequence contacts between Hawasina and ophiolite units in the Semail massif have been explained by Gregory et al. (1998) as resulting from the ophiolite being a thin–skinned thrust above a truncated surface. This in contradiction with our mapping and with the gravity data mentioned above. In most localities the Hawasina formations are intruding the ophiolite nappe, as shown by mutual contacts occurring at any level of the nappe. This is supported by detailed field observations inside and around the Wadi Tayin massif : deformation cleavages and lineations at the periphery of Hawasina inclusions are steep, and kinematic analysis, however uncertain in this type of formation, suggests normal rather than reverse motion, thus confirming the extrusion interpretation. In Wadi Tayin, Coleman (1981) also considered the Hawasina observed in the Ibra synform as an extrusion, which he ascribed to underlying Triassic salt diapirism. A puzzling situation is met in the northern and the central parts of the ophiolite nappe where the Hawasina formations crop out east of the ophiolite, and are in tectonic contact with the upper ophiolite lavas instead of being buried below the ophiolite. Their interpretation as being extruded through the ophiolite is compatible with their contact with the ophiolite nappe, being a normal fault as documented by Nicolas (1989, p. 42). The Hawasina extrusion through a supposedly uniform ophiolite nappe has been related by Chemenda et al. (1996) to the occurrence of high-pressure metamorphism along the coast of Oman in the Saih Hatat region. Buoyancy-driven and rapid uplift of the deeply subducted crustal units would have locally exhumed them as high-pressure metamorphic rocks. It would also cause the doming of the Arabian margin formation which is responsible for the Jebel Akhdar and Saih Hatat surrections. Eventually, it would have forced the overlying Hawasina formation through zones of weakness in the ophiolite nappe. This is an alternative to the obduction- related ramp anticline tectonics proposed earlier by Graham (1980) for the Hawasina Window, and by Bernoulli and Weissert (1987) and Goffé et al. (1988) for the Djebel Akdar and the Saih Hatat. Ramp tectonics and extrusion are not mutually exclusive. We favour the extrusion process for the Hawasina windows on the basis of a few observed intrusive contacts and because of their apparently random occurrence inside the ophiolite nappe.

If the Hawasina formations are intrusive, the division of the ophiolite nappe in massifs may be fortuitous, at least for those massifs separated by narrow Hawasina stringers, thus suggesting a continuity from one massif to the next. This is obviously the case of Sumail and Wadi Tayin which are still attached in the central part. Internal structures are also clearly contin-

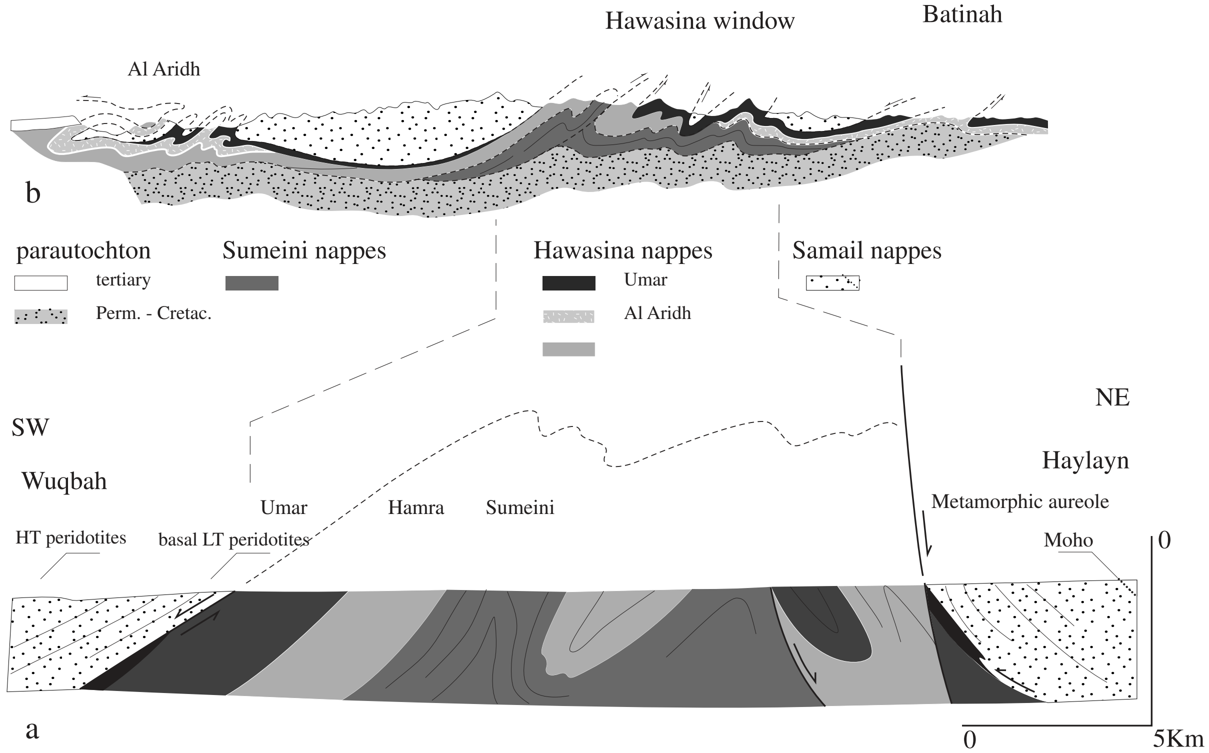

Figure 5. (a) Cross-section through the Hawasina window between the Haylayn and the Wuqbah massifs, illustrating the extrusion of Hawasina formations along the Haylayn massif (modified from Rabu, 1993). (b) Cross-section through the belt along the same profile in Rabu (1993).

uous (Nicolas and Boudier, 1995). The Nakhl-Rustaq and Haylayn massifs are separated by a stringer of Hawasina, but once their 2–6 km gap is closed and the Haylayn massif is displaced 10 km NNE along the gap, a perfect coincidence is obtained for both the tip of a rift propagator and the extension of the Rustaq mantle diapir (Nicolas and Boudier, 1995). Haylayn massif is again separated from Sarami by a 2–4 km-wide stringer of Hawasina. Both massifs are in apparent continuity. The Sarami massif is separated from the northern massifs by a major structural discontinuity which has been identified in several ways (thickness of gabbro unit, general trend of sheeted dikes, internal structures).

In the northern massifs, the nappe fragmentation of massifs by Hawasina extrusions is only clear in the mantle and lower crust units. At the upper crustal level, the Hilti, Fizh, Aswad and Khawr Fakkan massifs are continuous in exposure. The internal structures in Hilti and Fizh are in continuity and the Fizh-Aswad and Aswad-Khawr Fakkan massifs are only separated by accretion-related shear zones and thrusts (Nicolas et al., in press). This suggests that the Hawasina formations in these northern massifs would not have intruded the ophiolite throughout and would not have destroyed its internal structure.

At the scale of the entire ophiolite, the structural continuity is less obvious, as the central and northern internal massifs are separated from the three western and external massifs (Wuqbah, Miskin and Bahla) by the large Hawasina Window, and the two southern Sumail and Wadi Tayin massifs are separated from the central and northern parts by the large anticlinal dome of the Djebel Akhdar. Whatever the origin of the doming, the large separation between massifs may be due either to their gravity sliding or to their progressive erosion during the doming. If the massifs are now separated by gravity sliding as proposed by Hanna (1990), it is justified to close the gap created by the domes and to look for a total structural continuity throughout the ophiolite nappe. In the case of erosion of the ophiolite above the domes, parts of the ophiolite nappe are now missing and there is no continuity between the massifs.

Best fit

In conclusion, we structurally divide the ophiolite into two domains, with some uncertainty about the place of the western Wuqbah-Miskin and Bahla massifs. Bearing the stigmas of a intense ridge-related internal tectonic activity (see below), the northern domain ranges from Kawhr-Fakkan to Hilti. The massifs are still continuous and this vast domain has not been torn into pieces by obduction/related tectonics. A second domain comprises the central massifs (Sarami-Haylayn and Nakhl-Rustaq) on the one hand, and the southern massifs (Sumail and Wadi Tayin) on the other. By moving together the central and southern massifs, the structural boundaries and trajectories in the Sumail-Wadi-Tayin and Nakhl-Rustaq massifs become aligned, and seem to reveal the presence of parallel ridge segments belonging to a propagating system (Nicolas and Boudier, 1995; see also below). In Figure 6, we have tentatively displaced all massifs to obtain the best possible fit. This fit may represent the internal structures of the ophiolite nappe before it was dispersed by obduction-related processes. We will use it to unravel the ridge-related accretion activity.

Discussion. Data constraints on models

Thermal constraints on deformational events

From the study of microstructures and associated mineral parageneses, it is possible to bracket the temperature related to a given deformation event (Nicolas, 1986; 1989, p. 29; Nicolas and Boudier, 1995). This is of paramount importance because it allows us to constrain the geological and deformational history of the ophiolite and, eventually, to conclude that it records mainly the ridge and near-ridge activities.

Ridge activity

It is particularly easy to identify HT structures resulting from ridge activity, because they occur at temperatures above the solidus in the upper mantle (1200–1250 °C) and lower crust (1150–1200 °C). Plastic flow in the mantle took place in the presence of some melt resulting now in the occurrence of plagioclase and clinopyroxene impregnation textures, dikelets and lenses (Figure 2a). Magmatic flow in gabbros is equally easy to distinguish from plastic flow (Figures 2c and 2d). Surprisingly, only little plastic deformation is documented in the gabbros, mainly in the lowermost gabbros just above Moho; shear zones with flaser gabbros are quite exceptional. It seems that, as soon as they cross the isotherm 1150 °C, which is close to the rigidus temperature (Kelemen and Aharonov, 1998), gabbros become especially rigid, probably more rigid than the underlying peridotites (Seront, 1993). Temperatures associated with subsequent deformations occurring below the solidus are more difficult to bracket.

Near-ridge activity

In the folded maps, we have locally (Sarami, Haylayn and Fizh massifs) distinguished mantle peridotites displaying a MT deformation, estimated to correspond to temperatures of 1000–1100°C. Otherwise, LT deformation is ascribed to temperatures of 900–1000 °C, 900 °C typically corresponding to the very fine- grained peridotite mylonites in contact with granulitic metamorphic aureoles equilibrated at or below 850 °C (Ghent and Stout, 1981; Gnos and Kurz, 1994; Hacker and Mosenfelder, 1996). Below these temperatures, deformation in the mantle sequence is blocked, unless water ingresses the system and weakens the peridotite in connection with the development of hydrated phases, first in the amphibolite facies. Talc and chlorite develop pervasively in the Khawr Fakkan massif (Gnos and Nicolas, 1996) and in a more scattered way in other massifs (Ceuleneer, 1986). Further alteration occurs in greenschist facies and at lower temperature conditions with the development of serpentine minerals.

The weak and local overprint of plastic deformation in the crustal gabbros being ascribed to the immediate vicinity of the magma chamber, lower temperature deformation, locally observed in the gabbros, corresponds to amphibolite facies shear bands which in most situations are steep and narrow (meter-sized or less), and bear a horizontal mineral-stretching lineation. These bands are totally recrystallised, pointing to ingression of water in the gabbros and to temperatures of the order of 600°C. Relays to greenschist facies ($\approx$ 400 °C) are locally observed. Commonly, such shear zones are in the continuation of LT shear zones in the mantle section. Otherwise, hydrothermal fractures are present everywhere in the gabbro section down to the Moho (Nehlig and Juteau, 1988; Nehlig, 1994). Ingression of water at temperatures around 900°C at the base of the crust, below the ridge, is indicated by the occurrence, at all depths, of gabbro and amphibole-diorite dikes (Nicolas et al., this issue). These dikes grade into hydrothermal fractures.

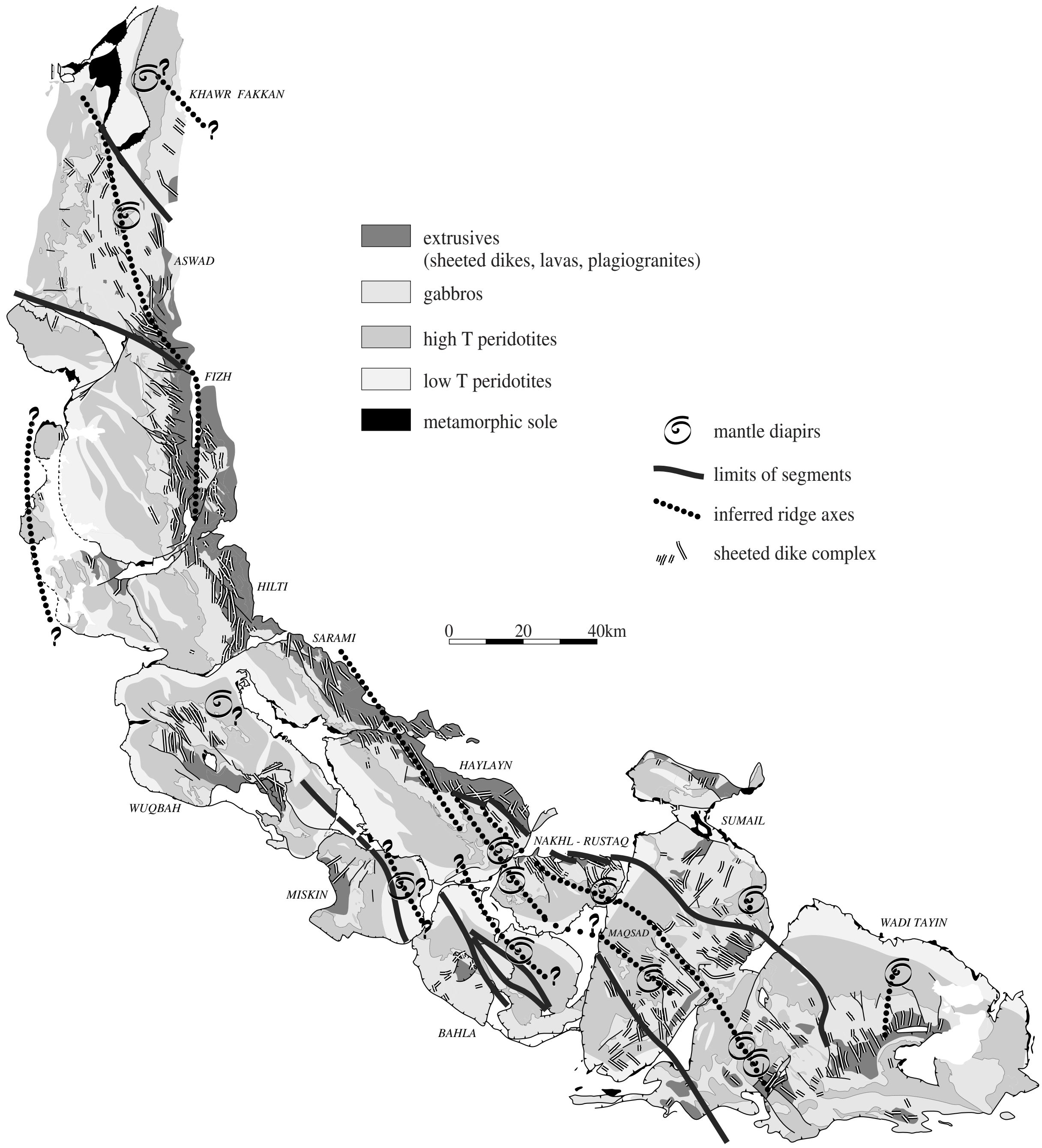

Figure 6. Main ridge-related structures in the Oman–UAE ophiolite reassembled into a best fit of the central and southern massifs. This fit is obtained by minimising voids and overlaps and ensuring the continuity of internal structures. It is assumed to represent the ophiolite nappe assemblage before its dismembering by local Hawasina extrusion and gravity sliding aside the Djebel Akhdar dome.

The highest temperature fractures are very narrow and tightly spaced, marked by stringers of green actinolitic hornblende, corresponding to temperatures estimated at 450–500 °C (Juteau et al., this issue). These are followed, in greenschist facies, by zoisite-epidote-quartz-prehnite assemblages in which fluid inclusion studies yield temperatures ranging from 390° to 110° (Nehlig and Juteau, 1988). Referring to the thermal model of Cormier et al. (1995), temperatures of 600 °C in lower to middle crust point to ages between 0.5 and 2 Myr after accretion, while temperatures of 400 °C correspond to ages larger than 2.5 Myr, but probably younger than 15 Myr (Denlinger, 1992). From hornblende $^{40}Ar/^{39}Ar$ ages, Hacker et al. (1996) also conclude that the crust in Oman was cooled to 550 °C in 1–2 Myr.

Thermal structure and spreading rate

When lithosphere drifts away from the ridge of origin, the thermal gradients in the upper mantle and lower crust are small compared with those in the upper crust. This allows us to put time constraints on successive deformations, and because Oman was a fast- spreading ridge system, to grossly estimate the distance from the ridge where they occurred. We assume an average half-spreading rate of 5 cm/yr and we adopt the thermal profile of Cormier et al. (1995). With these parameters, MT deformation in the mantle, a few kilometres below the Moho, would occur in a lithosphere which is around 1 Myr old; that is, around 50 km away from the axis. LT deformation observed at similar depth could occur no farther than 100 km away from the ridge of origin. We can test these estimates in the Khafifah area of the Wadi Tayin massif where a shear zone defining the northeastern limit of the Maqsad segment has been mapped. This shear zone is located 25 km from the axis of the ridge of origin (folded maps). At 5 km below the Moho, its structures locally reflect a MT deformation which, being in the range of 1100 °C, corresponds to an age of 0.8 Myr. Considering the distance to the ridge of origin, the half-spreading rate would be $25/0.8 \approx 30$ km Myr^{-1} $= 3$ cm yr^{-1}, to be compared with our a priori estimate of 5 cm yr^{-1}. We conclude that however crude are the estimates deduced from the thermal analysis, they help to constrain models for the Oman–UAE ridge system.

Dealing with the LT deformation ($\sim$1000 °C) related to basal thrusts ($\sim$6 km below the Moho), the same type of analysis shows that the detachment thrusting occurred in a lithosphere not older than 2 Myr, which is consistent with radiometric ages (Hacker et al., 1996).

Kinematic constraints

HT deformation

As discussed above, kinematic analysis is restricted to the mantle units because shear indicators deduced from solid-state flow analysis generally point, with no ambiguity, to a given shear direction (folded map 2), in contrast with the confused situation in magmatically deformed gabbros. In mantle formations, each massif presents a dominant shear direction at depth. However, at 1 km or less beneath the Moho, an inversion in the shear direction is observed in Haylayn, Sarami, Hilti, Fizh and Aswad massifs that is over a cumulative distance of 200 km along the Moho in a SE-NW direction. This inversion, also documented in other ophiolites (Prinzhofer et al., 1980; Violette, 1980; Girardeau and Nicolas, 1981; Suhr, 1992; Hoxha and Boullier, 1995; Nicolas et al., 1999), is attributed to the forced flow which is expected at the outskirts of mantle diapirs (Figure 4). Possibly because many more structural data are available in Oman than in other massifs, it is now suspected that the limit of shear inversion does not correspond to a simple planar surface, but that it draws fingers into the underlying mantle, suggesting that the forced flow could localise and dig channels into the underlying asthenosphere (Ildefonse et al., 1995; Mishibayachi et al., this issue). In the northern part of the Fizh massif and in the Khawr Fakkan massifs, where the inversion is not visible, the Moho is steep and continuing deformation has sheared the uppermost mantle, erasing any possible trace of inversion. In Nakhl-Rustaq, Sumail, Wadi Tayin massifs (Nicolas and Boudier, 1995) and possibly in Wuqbah massif, where ridge segments have been trapped and a paleo-ridge axis has been traced, there is no clear shear inversion, and the shear direction in flat-lying mantle flow is directed top-away from the inferred ridge axis. Possibly, the inversion in these massifs occurred within either the locally thick MTZ or the deforming gabbros of the lower crust, as suggested by a local study of shear indicators in the gabbros of Wadi Tayin (folded map 2). This may be a consequence of these areas being rheologically weaker, because of the local melt impregnation, compared with areas located farther along axis from the diapir head.

In conclusion, the shear sense of HT flow in the mantle is expected to be top- away from the spread-

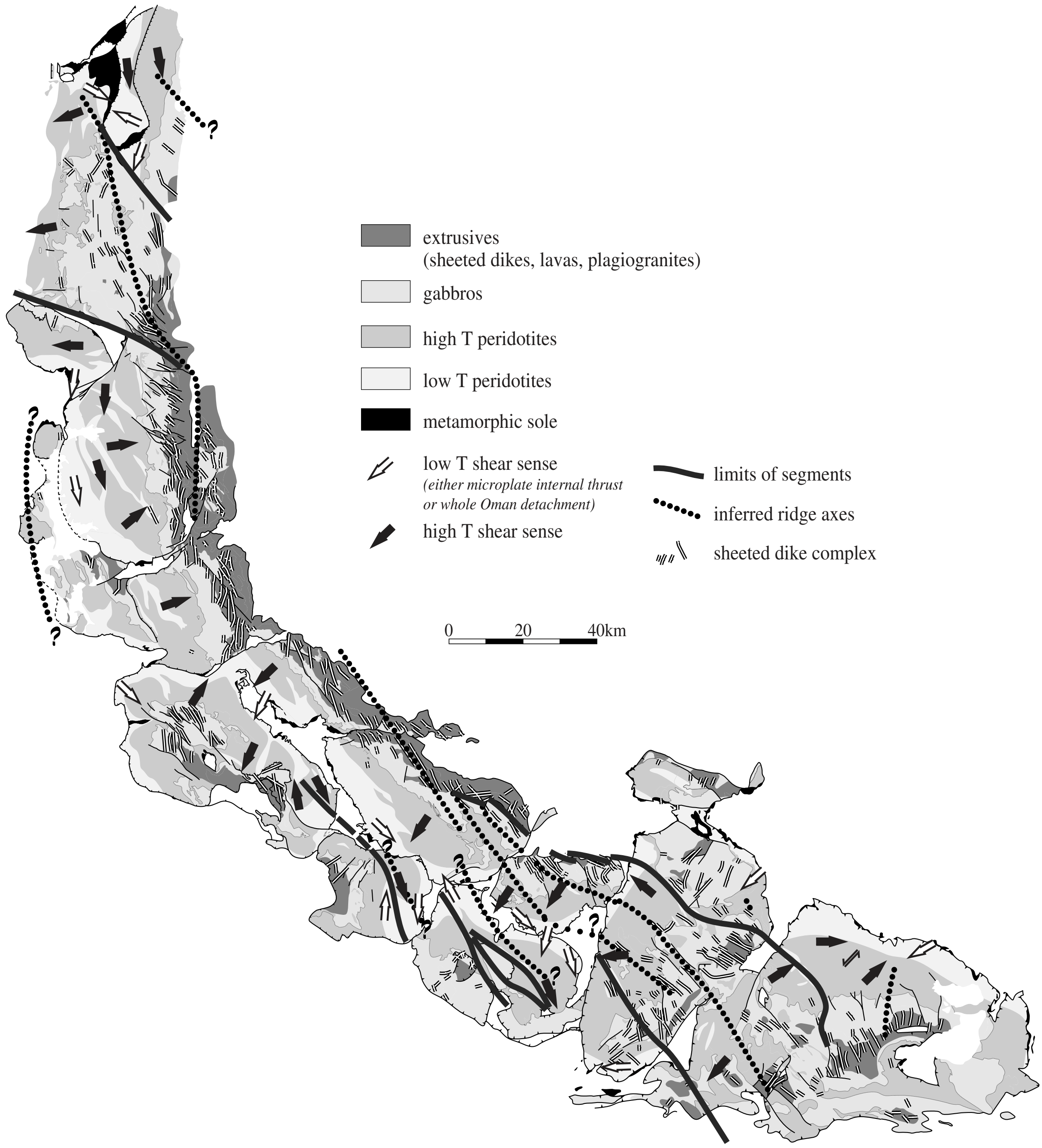

Figure 7. Simplified map of HT and LT shear directions in the mantle units of the various massifs. This map is the fit of Figure 6 with representation of the major structural features in the belt.

ing axis direction, except for the first kilometre below the Moho in areas issued from a ridge segment located away from a diapir (Figure 4). This provides a criterion to locate the ridge axis of origin for a given massif. Figure 7 summarises the results of this analysis at the scale of the Oman belt. It shows that, in the southern massifs, where ridge segments have been frozen, opposite senses are recorded on each side of the presumed ridge axis. These segments end in Nakhl-Rustaq and the southern part of Haylayn. In these massifs, mantle exposures are mostly restricted to the western side of the paleo-ridge axis, thus explaining the dominant top-to-the-SW shear direction. North of these massifs, and up to the Aswad massif in the UAE, there is no more ridge segment documented by the occurrence of diapirs within the ophiolite, with possible exceptions in Wuqbah and Miskin. As a consequence, these parts of the ophiolite may derive from one flank or the other of a paleo-ridge which was oriented, in the present reference frame, NW–SE to N–S, as deduced from the trend of sheeted dikes, and was located eastward or westward with respect to the ophiolite belt. In the main part of Haylayn and in Sarami, the dominant SW-directed shear sense indicates that the ridge of origin was located to the northeast. Hilti and Fizh record a dominant E-directed shear sense and thus would derive from a ridge located to the west. In the western part of the Fizh massif, and south of the large Wadi Hatta shear zone, a dominant S-directed shear sense in HT and MT peridotites corresponds to a complex system, analysed below. North of Wadi Hatta, the Aswad massif in the UAE, seems to have incorporated a ridge segment and its western mantle formation displays, accordingly a dominant W-directed shear sense (Nicolas et al., in press).

LT-MT deformation

LT-MT deformations are mainly observed within 1–2 km above the basal sole and the underlying metamorphic aureole. They are mostly related to the detachment and early thrusting of the ophiolite. In the Khawr Fakkan (Gnos and Nicolas, 1996; Nicolas et al., in press) and Fizh (Boudier et al., 1988) massifs, the continuity between the thrust motion and vertical shear zones has been demonstrated in the field. There are other steep shear zones which are not directly related to the detachment, but presumably to ridge segmentation and propagation. We describe them independently, although we realise that they may be ultimately related through processes which are not yet well understood, such as microplate rotations and thrusts (Boudier et al., 1997).

Dealing with detachment-related thrusts, the best kinematic indicators for LT deformation are again found in peridotites, in contrast with metamorphic aureoles where it is difficult to obtain reliable shear-sense indications, except in quartzites. Since the metamorphic aureoles outcrops are often rotated during late tectonic thrusting, lineations and shear directions are also less reliable than in the overlying peridotites. For these reasons, nothing is added here to the results from earlier studies in the metamorphic aureoles (Lippard et al., 1986; Boudier et al., 1988). Just above the metamorphic aureole, the first peridotites are mylonitic over a few tens of meters. Above this, they are porphyroclastic with a LT structure which grades upsection into MT and HT structures.

A continuum from ridge-related to detachment-related flow is suggested for the following reasons: (1) In most massifs there is a continuity in foliation and lineation orientations from the lower LT-MT to the overlying HT peridotites. A good example of continuity is found in the Haylayn massif, with a progressive rotation occurring through MT facies, from flat and N–S foliations in LT facies, to steep and E-W foliations in HT facies. In other massifs such as Khawr Fakkan, Wuqbah, Miskin, Nakhl-Rustaq, Sumail and Wadi Tayin, the HT and LT lineations are in the same general direction. (2) An important magmatic activity still occurs during the detachment as the associated shear zones inside the peridotite massifs are invaded by gabbro and norite dikes which were injected during deformation (Boudier et al., 1988; Boudier et al., this issue). It is also tempting to explain the local occurrence of clinopyroxene in mylonites as resulting from tectonic scattering of mafic dikes into the enclosing peridotite.

In the northern and central massifs there is a consistent SSE-directed thrusting direction which is recorded in all massifs. The SSE shear sense is more consistent in the northern massifs, including Hilti, with a total balance of 105 SSE shears against 18 NNW shears, than in the central massifs with 123 SSE against 71 NNW shears. The most striking exception is the western part of the Bahla massif where the basal thrust is entirely NW-directed (Figure 7). Dextral and steep NNW-SSE shear zones in Fizh, Hilti, Sarami and Haylayn are in physical continuity with the SSE-directed thrusts. The SE-directed thrust is present at the base of all massifs except Khawr Fakkan, Sumail and Wadi Tayin where W-directed thrusting is recorded. For this reason, the SE-directed thrust is tentatively considered as corresponding to the main detachment for the entire ophiolite nappe. In the last section of the paper, we discuss why only the W-directed LT deformation is recorded in the southern massifs, both in peridotites and in the underlying metamorphic aureole (Boudier and Coleman, 1981; Boudier et al., 1988).

Shear zones related to ridge segmentation

Hundreds of meters- to kilometre-thick steep shear zones with flat-lying lineations in the mantle section have been documented in the southern and central massifs where they separate a younger NW–SE segment from an older NE–SW system (Nicolas and Boudier, 1995; Figure 7 and below). Like the detachment-related thrusts and shears, these segmentation-related shears correspond to LT-MT deformation conditions in the mantle. They are commonly traced in the lower crust as amphibole to greenschist facies mylonites. The shears which cross-cut the massifs, such as the NW–SE sinistral shears in Fizh are km-wide and MT in deeper mantle, 100 m-wide and LT in shallower mantle, and 10 m-wide and amphibole facies in gabbro, recording decreasing T-conditions from deeper to shallower horizons below the ridge of origin.

We explain these mantle and lower crust shear zones as the result of the local cooling of a new uprising asthenosphere at the contact with a colder lithospheric boundary. As the HT asthenosphere flowing along this contact is cooled to MT and LT conditions, melt issued from the opening segment is frozen in such contacts resulting in a profusion of mafic dikes (Ceuleneer, 1991). As discussed earlier, the consideration of deformation structures in the Wadi Tayin shear zone allows to introduce age constraints: the new segment opened into a lithosphere which is estimated to be no more than 1 Myr older. This conclusion can be extended to similar situations in other massifs.

In the northern massifs, we also ascribe to ridge segmentation the NW–SE sinistral shear zones inside the Fizh massif and along the Wadi Hatta which separates it from the Aswad massif. There, MacLeod and Rothery (1992) describe, as in the southern massifs, mafic intrusions associated with crustal shears. A coherent explanation can be offered for the N–S sinistral system internal to Aswad (Nicolas et al., in press), but not for the NW–SE sinistral shear inside Hilti and Bahla, or the smaller NW–SE dextral shear inside Nakhl- Rustaq. There are also, within many massifs, isolated shears which may relate to such systems, or record local instabilities in the ridge activity.

Distinguishing detachment-related from segmentation-related shears is possible when they are in continuity with basal thrusts; we have also pointed out that the former seem to display flow trajectories in continuity with HT flow. The segmentation-related shears, in contrast, may cut the HT flow trajectories as in Fizh, Semail, and Wadi Tayin massifs.

Structural constraints

Lithospheric deformations

Most commonly, the Moho is planar, and it was horizontal in the ridge reference frame; upsection, the gabbro unit is only locally deformed in relation with wehrlite intrusions, and the lower gabbros are parallel to the Moho. The sheeted dike complex is, or was, uniformly steep, being inclined 80–70° away from the ridge axis (Nicolas et al., this issue). Contrasting with this situation, the Moho is locally steep or vertical in a few massifs, being underlined in most places by shear deformation in the uppermost mantle and in the lower crust. In such areas, the gabbro section is locally plastically deformed and the layering is tilted; it is invaded by gabbro-norite intrusions, locally brecciated (Juteau et al., 1988b), transected by amphibolite facies shear zones and finally heavily altered by hydrothermal circulation in greenschist facies conditions; intrusions and breccias of plagiogranites often associated with diabases are also common. As seen above, this situation was created close to the ridge of origin because of the 900–1000 °C required for mantle shearing, and because such zones are commonly invaded by intrusion of mafic dikes and are hydrothermally altered. Moreover, the steep structures below and above the Moho fan out at depth into the flat-lying basal aureole, and toward the surface into the similarly flat-lying upper volcanics and associated sediments (Nicolas, 1989, p. 62). The sheeted dike complex is generally not tilted except in Fizh massif, where it is cut by new vertical dikes (MacLeod and Rothery, 1992). In northern parts of the belt, areas with steep Moho comprise the southern part of Khawr Fakkan, the northern part of Fizh where the lower crust is particularly affected by shears and plagiogranite intrusions, Sarami and Wuqbah where the crust is tilted and heavily faulted. In the central and southern parts of the belt, planar Moho and overall simple lithospheric structures predominate. In Haylayn massif, the gabbro unit displays a synform, the axis of which steepens to the SE, associated with a steepening of the Moho; in the SE Haymiliyah area, both the Moho and the axis of the large crustal fold are very steep.

With the exception of the western massifs, where steep Mohos could be due to emplacement tectonics, and of the Mansah area of the Sumail massif, where local tilting and alteration of the crust are ascribed to the intrusion of an off-axis mantle diapir (Jousselin and Nicolas, this issue-b), it seems that steep Mohos are associated with the tips of propagating rifts. This is now documented for the northern part of Fizh (Nicolas et al., in press), for the Haymiliyah area in Haylayn (Reuber et al., 1991; Nicolas and Boudier, 1995), and for the western part of the Nakhl-Rustaq massif. Steepening of the Moho along a propagating rift may be explained in two ways, which may be combined. The cracking lithosphere, though it is very young, may have steep walls at the Moho level along which the intruding asthenosphere will be plated, resulting in steep MT to LT foliations in the uppermost new mantle. It can also be envisaged that this new asthenosphere has a dragging effect on its lithospheric walls and tilts them, in a way similar to that proposed by MacLeod et al. (1996) for the Hess Deep rift valley (equatorial East Pacific Rise). In our reconstruction of the paleo-ridge system, we use the possible association of steep Mohos with propagators to locate a propagator close to the steep Moho mapped in northern Haylayn and in Sarami.

From this summary of lithospheric deformation in the ophiolite nappe, we draw the following conclusions :

– These deformation events occurred within lithospheres which were probably not older than 1 Myr.

– They may have a local tectonic origin such as the disruption of a very young lithosphere by a propagating rift (Reuber et al., 1991; Nicolas and Boudier, 1995; Nicolas et al., in press) or by the intrusion of a mantle diapir below a seamount (Jousselin and Nicolas, this issue, b). They may also have a more general cause related to compressive forces exerted on a large piece of young lithosphere. This could explain the deformation at the scale of massifs in the central and northern parts of the belt.

– By contrast, the absence of lithospheric deformation in other massifs may be explained either by the absence of tectonic forces or, more probably, by their rigidity, itself related to their age being in excess of 1–2 Myr. A similar situation is described in present-day microplates. The domain located between propagating segments deforms by internal shear until it is 1–2 Myr old. Beyond, it becomes rigid and starts to rotate as a microplate (Kleinrock et al., 1991; Korenaga and Hey, 1996; Bird et al., 1998).

– Eventually, the major deformation events are thus related to the earlier history of the ophiolite : only little and very LT internal deformation can be ascribed to subsequent events, e.g., obduction and deformations subsequent to emplacement onto the Arabian margin.

Foliations and lineations in upper gabbros

Foliations and lineations in upper gabbros progressively steepen 1 km below the base of the sheeted dike, in relation to the presumed shape of the magma chamber (Figure 3). The way the gabbros steepen allows one to predict the flank of the ridge from which the considered area is issued, and has been successfully tested in the southern massifs and in the Nakhl-Rustaq massif where the ridge axis was identified (Figure 6) (Nicolas and Boudier, 1995). This confirms our earlier tent-shaped magma chamber model associated with active mantle flow (Nicolas et al., 1988a), which has been tested further by a 2D numerical model (Chenevez et al., 1998).

In Haylayn and Sarami massifs, this geometry is blurred by ridge tectonics. The deformation is tentatively explained by the weakness of this area in relation with the paleo-ridge being located underneath or nearby to the northeast (Figure 6; see also below). In the Hilti massif, the dominant eastern dips in upper gabbros point to a ridge located to the west, also in keeping with the results of the kinematic analysis in the mantle (Ildefonse et al., 1995; Michibayashi et al., this issue). This analysis cannot be pursued in the other northern massifs because of the increased importance of deformations in the gabbro unit which masks the direction of the foliation steepening. In contrast with the mantle unit of the Aswad massif in the northern part of the belt which is exceptionally undeformed, the gabbros are contorted, and no consistent steepening of foliation in upper gabbros has been noted.

The magmatic lineations in the upper gabbros of Figure 3 point to a dominant down-the-dip flow direction, which should be ascribed to the subsidence of the floor of the melt lens in relation to the continuing settling of crystallising gabbros (Phipps Morgan and Chen, 1993; Quick and Denlinger, 1993; Chenevez et al., 1998). The dominant along-axis girdle at shallower depth, near the presumed position of the paleo-melt lens, and the transverse girdle at depths greater than 500 m below the lens, suggest that there is a component of longitudinal flow in the melt lens which fades downward as a result of the subsidence-related flow.

Petrological constraints

In terms of mineralogy and petrology, the Oman ophiolite is remarkably homogeneous. Only two points are noteworthy and contribute to our general synthesis. The first one deals with the nature and thickness of the dunite transition zone below the Moho and with the abundance of wehrlite intrusions into the overlying crust (Nicolas and Boudier, in press). The second one deals with the nature and occurrence of secondary volcanics on top of the ophiolite-related MORB-type volcanics, and their possible association with a hydrous contamination of mantle (Lippard et al., 1986; Boudier et al., 1988; Ernewein et al., 1988).

Moho transition zones varies in thickness from a few meters to several hundred meters. Concomitantly, thickest MTZ tend to display more gabbro and pyroxenite dikes, sills, and diffuse impregnation, pointing to extensive melt circulation below the ridge of origin (Boudier and Nicolas, 1995). MTZ thicker than a few hundred meters are associated with a thinner gabbro unit and with an increased number of wehrlite sills and intrusions into the crust (Nicolas et al., 1996). This effect has been ascribed to compaction of the MTZ at the outskirts of diapirs, with injection into the crust of an olivine-rich mush (Nicolas et al., 1988a; Boudier and Nicolas, 1995). This has been quantified in the Maqsad diapir area (Jousselin and Nicolas, this issue). In order to evaluate the importance of this wehrlite magmatism into the gabbroic unit, we have represented in the map of Figure 8 the 'cumulate interlayered peridotite gabbro' (CPG) unit of the 1:100 000 map of Oman. This unit appears on this map, instead of the 'cumulate layered gabbros' (CLG), in areas which are specially rich in wehrlite sills and intrusions.

In the southern and central massifs, as well as in the Aswad massif, where diapirs have been mapped, it is also observed that thick MTZ and wehrlite-rich gabbro units overlie diapirs, an observation easily explained by the fact that melt extracted from the mantle should preferentially migrate through the diapirs. However, these olivine-rich zones extend along the ophiolite belt over distances on the order of 50 km, and thus, in the south, cover two or three individual diapirs. This suggests that these zones represent the true dimensions of diapiric areas, within which smaller mantle fingers may be protruding, each one being represented by the 10 km-across mapped diapirs (Nicolas and Boudier, in press).

Secondary volcanism

The early geochemical data on effusive components of the Oman ophiolite (Pearce et al., 1981; Alabaster et al., 1982) show specific geochemical signatures, which, combined with textural and petrological features (rate of hydrothermal alteration, vesicular textures, presence and types of phenocrysts), and relative chronology, lead to the identification of 5 units, then condensed into 3 units on the 1:100 000 geological map.

The Geotimes Unit, named later V1 (Ernewein et al., 1988), or SE1 on the 1:100 000 geological map, is tholeiitic and displays N-MORB REE patterns, assigned to ridge accretion processes. This unit encompasses the sheeted dike complex, and is represented along the whole Oman ophiolite, although poorly exposed in the southern part, outside the limits of the Sumail ridge segment.

The Lasail Unit, more intensely altered, is represented only north of the Haylayn massif. The Alley Unit formed by vesicular pillows overlying Geotimes or Lasail Units, and the Cpx-phyric Unit are grouped together with the Lasail unit as V2 or SE2 on the 1:100 000 geological map. The lava flows of these units are separated from the Geotimes-V1 lavas by a veneer of umbers, not thicker than a few meters, and sulfide mineralised gossans. Biostratigraphic dates show that V1 and V2 were erupted during a narrow interval ranging from early Cenomanian to early Turonian (96-91 my) (Tippit et al., 1981; Schaaf and Thomas, 1986). V2 volcanics were interpreted as representing seamounts (Pearce et al., 1981), based on their discontinuous distribution and geochemical variability. Geochemically, they display LREE depleted patterns and slight to strong HFSE (Zr, Hf, Nb, Ta) negative anomalies with respect to neighbouring trace elements or N-MORB normalised patterns. These signatures are transitional between MORB and IAT. Subduction-related hydrous melting of already depleted mantle has been invoked in order to explain the IAT signature, leading to the arc-environment model proposed for the genesis of the Oman–UAE ophiolite (Lippard et al., 1986). Another possible shallow source for hydrous remelting of the mantle has been proposed: it results from the early thrusting at the ridge axis, or close to it, along the flat lithosphere-asthenosphere boundary, of a hot mantle wedge over-

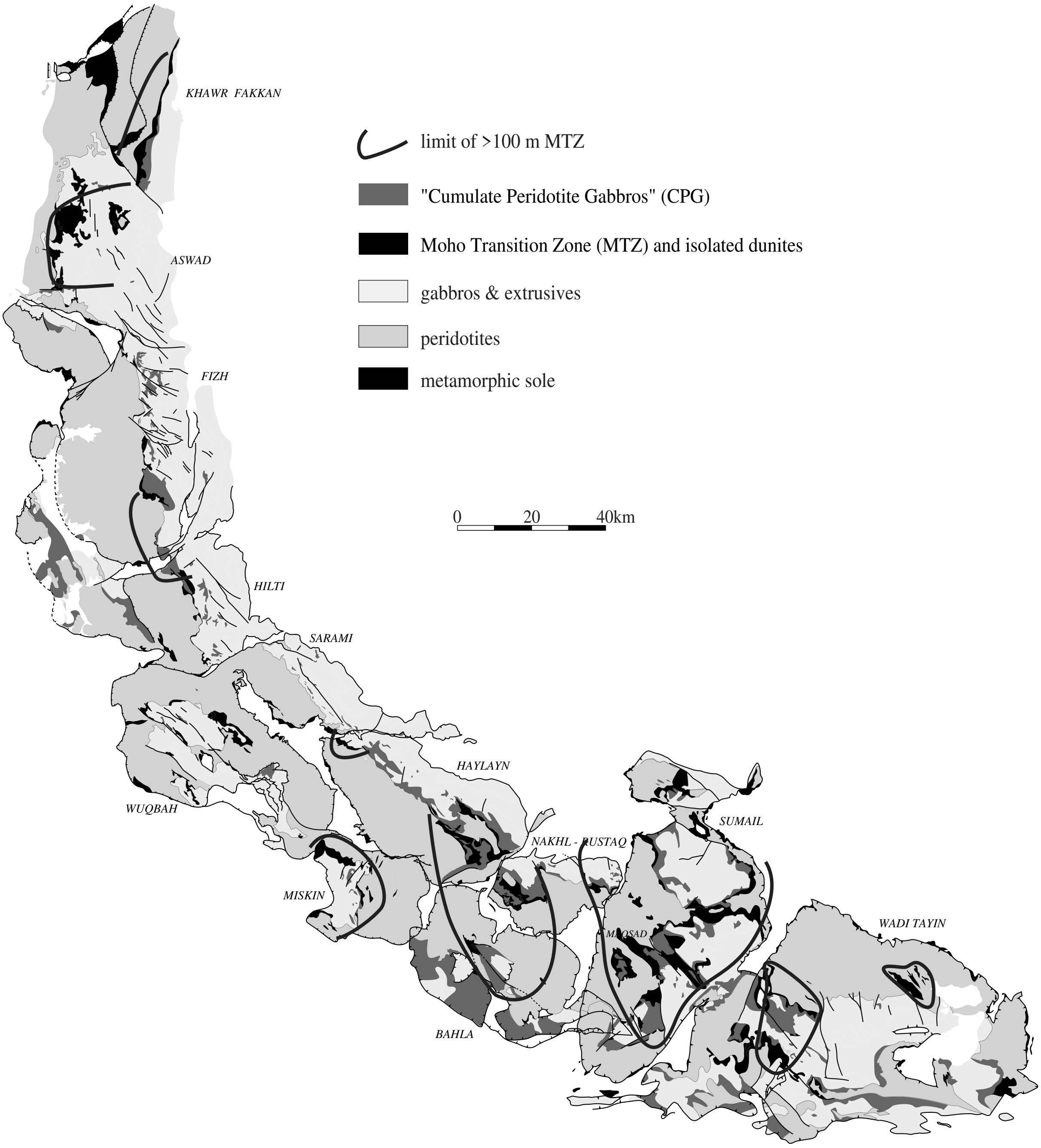

Figure 8. Map illustrating the contrasted contents of olivine in the MTZ and lower gabbros along the belt. MTZ thicker than 100 m are associated with olivine-rich lower gabbros. These gabbros are enriched in olivine, mainly by injection of wehrlites, either as sills or discordant intrusions; they are represented in the Geological map of the Sultanate of Oman (1987) as 'Cumulate Peridotite Gabbro'. This figure, derived from Nicolas and Boudier (in press), suggests the presence of a few upwelling mantle areas in the ophiolite belt.

riding a zero age crust (Boudier et al., 1988; Hacker et al., 1996). We conclude that the significance of the IAT signature of V2 effusives is not straightforward, and that it would probably gain from further investigation of their possible genetic link with the different members of the lower crust and MTZ. Incidentally, the absence of arc- related sediments above the ophiolite and in the Hawasina formation (Béchennec et al., 1990) does not militate in favour of a back-arc for the environment of origin of the Oman–UAE ophiolite.

The last Salahi or V3 Unit, restricted to a unique site in Wadi Salahi, shows a distinctive enrichment in LREE associated with a mildly alkaline signature. Interestingly, the rare, and a few meter-sized, biotite-granite intrusions cross-cutting the mantle section have similar geochemical trends (Lippard et al., 1986; Briqueu et al., 1991). These intrusions are abundant in the Khawr Fakkan massif where they are dated 85 to 90 my (Gnos and Peters, 1993), like the V3 volcanics (Tippit et al., 1981), and thus are inferred to represent an obduction-related magmatism. They are also present along the Moho of the Haylayn massif, and along the southern shear zone bounding the Sumail NW–SE segment.

Age constraints

Age constraints on the accretion of the Oman paleoridge system, on the oceanic detachment-related metamorphic sole and on the volcanic events subsequent to the accretion have been recently compiled and complemented by Hacker et al. (1996). We summarise here the salient results of this new study (Figure 9).

– Accretion, dated at 94–95 Myr, is followed within 1–2 Myr by the intraoceanic detachment and thrusting. This is in excellent agreement with the time lapse deduced from the cooling history (see above). It is recalled that emplacement onto the Arabian margin occurs at $\sim$78 Myr.

– It is also during such a short time that the deep crust is cooled to 550°, in agreement with our thermal constraints, and with the assumption of an early stage of magnetisation (Weiler, this issue).

– Scenarios based on short-time lapses and sharp age gradients predict either a thrusting of the ophiolite initiated away from a more easterly-located ridge, or a thrusting parallel to the ridge axis, as deduced from structural analysis in LT basal peridotites and metamorphic soles by Boudier et al. (1988). Eventually, the second scenario is preferred as it fits better with the rapid cooling time (1–2 Myr) in the metamorphic sole. To achieve this, and in keeping with the structural analysis of Boudier et al. (1988), we propose that, after 1–2 Myr of ridge parallel thrust, the motion becomes normal to the ridge. The preferred scenario also implies that the spreading rate is in the range of 50 to 100 mm yr^{-1}, and the thrusting rate in the range of 100 to 200 mm yr^{-1}.

Paleomagnetic constraints

New paleomagnetic studies (Feinberg et al., 1999; Weiler, this issue; Perrin et al., this issue) confirm previous data (Luyendyk and Day, 1982; Luyendyk et al., 1982; Shelton, 1984; Thomas et al., 1988; Perrin et al., 1994) regarding the various components of the ophiolite: radiolarites and umbers, volcanics, sheeted dikes, gabbros and peridotites. The most recent data applying to peridotites, gabbros and volcanics state that low-titanium magnetite is the dominant magnetic marker corresponding to very high Curie temperatures in the range of 550 °C, and that the northern and southern domains of the Oman–UAE ophiolite carry a different paleomagnetic record. In the northern massifs, Perrin et al. (this issue) confirm that the mean direction of magnetisation recorded by the V2 lavas consistently lie in a SE sector while slightly different directions are recorded, in the different massifs, in the V1 lavas. Interestingly, the V2 paleomagnetic directions coincide with those of the gabbros (Weiler, this issue). Since it takes $\sim$1 Myr for the gabbros to cool to 550°C, they passed the Curie temperature after emplacement of the V2 volcanism ($\sim$1 Myr after the accretion-related V1). In the southern domain, the most recent paleomagnetic data obtained in gabbros and peridotites (Feinberg et al., 1999; Weiler, this issue) confirm previous data showing that orientations of the mean direction of magnetisation lie in a NNW to NNE sector.

Based on these data, two models are proposed.

(1) Relying mainly on his paleomagnetic results obtained in gabbros, and ascribing the age of magnetisation to an hydrothermal alteration occurring prior to detachment, Weiler (this issue) concludes that the southern massifs have not been rotated with respect to the Arabian plate, in contrast with the northern massifs which have been rotated 120–130° clockwise. This results in a model (Figure 10a) which maximises rotations.

(2) Relying on their paleomagnetic results obtained in the volcanic sequence, mainly in the northern domain, Perrin et al. (this issue) question the unity of

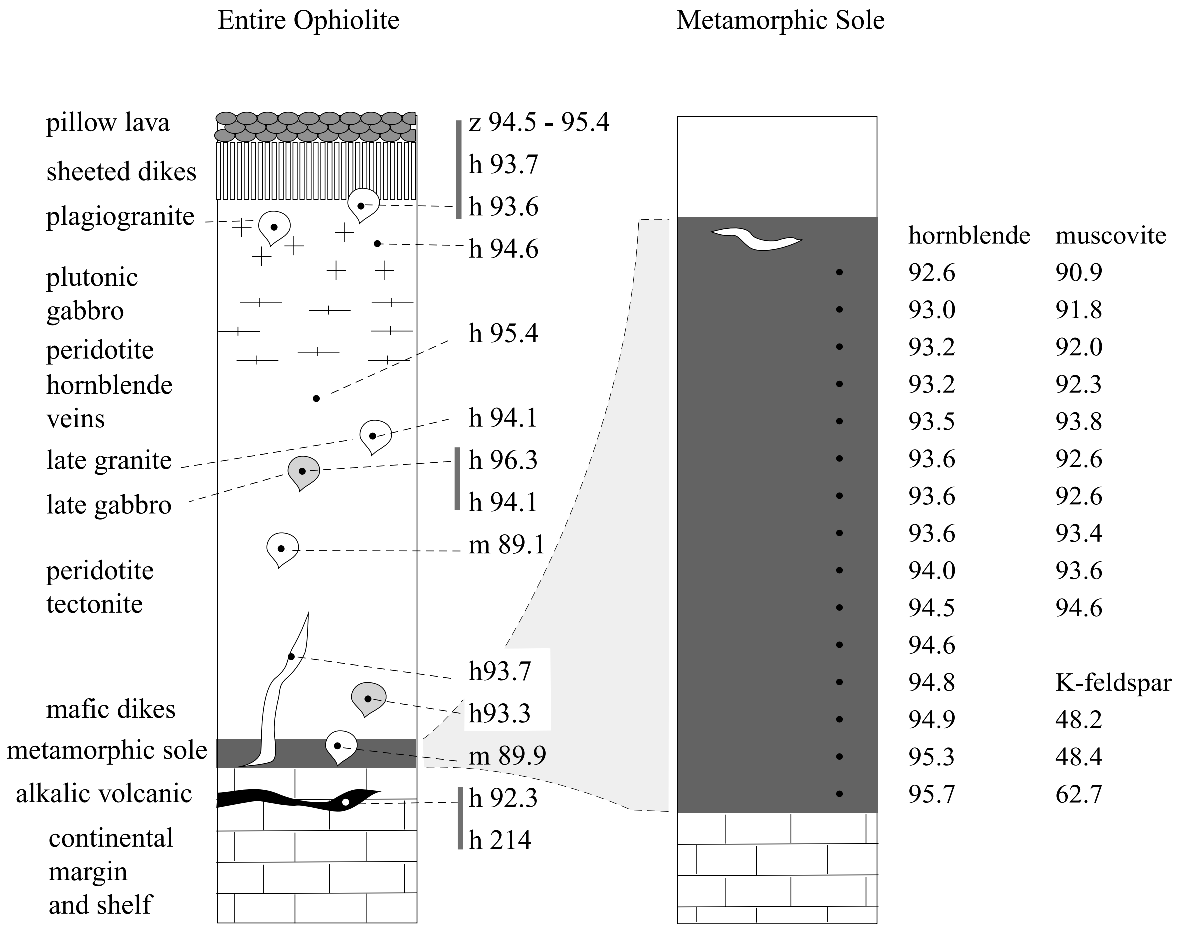

Figure 9. Summary of radiometric ages for the accretion of the ophiolite (upper left column) and its detachment registered in the metamorphic aureole (lower left, detailed in right column). Pb/U zircon (z) ages and ^{40}A/^{39}A hornblende (h), muscovite (m), and K-feldspar ages (simplified from Hacker et al., 1996).

this northern domain at the time of the emplacement of V1 lavas. They suggest that these massifs could rotate one with respect to the other until V2 emplacement (20° counterclockwise for Aswad, 40° clockwise for Hilti, unknown for Fizh), and that, afterwards, a possible rotation on the order of 90° could have occurred between the northern and southern domains. The resulting model emphasises progressive rotations (Figure 10b).

Since Weiler's and Perrin et al.'s models require large rotations of the northern/southern domains which are difficult to explain (see below), it is worth to consider the option of different geomagnetic polarities between the northern and southern domains. In this interpretation, the northern massifs would have a reversed polarity and the southern massifs, a normal polarity. This hypothesis has the merit of minimising to 45° or less the rotation of the northern domain with respect to the southern one (Figure 10c).

It seems however difficult to envisage that the inversion occurred during or immediately after accretion because its age is well constrained at 94–95 Myr (Figure 9), that is during the long period of Late Cretaceous normal geomagnetic field. Magnetisation of the northern massifs during a reverse period would imply the existence at 94–95 Myr of such a reverse period lasting about 1 Myr (time lapse necessary for

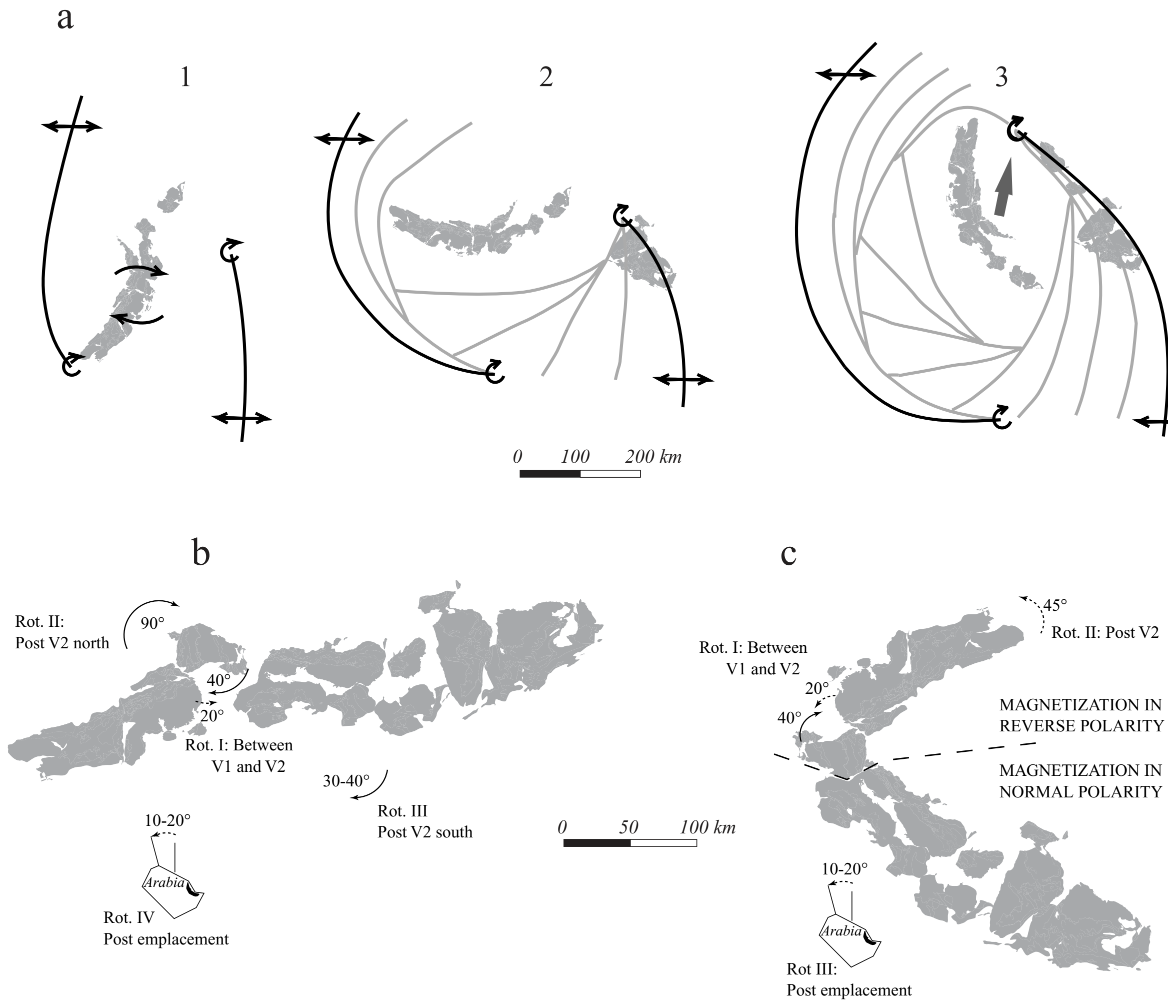

Figure 10. Paleomagnetic models for the evolution from the Oman paleo-ridge system to the ophiolite. (a) An EPR microplate model (after Weiler, this issue). Black lines are active ridge segments; grey lines are crust isochrons. The northern massifs are located in the microplate and rotate 130°, from stage (1) to stage (3), with respect to the southern massifs (south-east propagator). This is succeeded by a translation of the northern massifs relative to the southern ones (grey arrow), and by their obduction onto the Arabian margin. See Weiler (this issue) for further discussion. (b) A model of progressive clockwise rotation (after Perrin et al., this issue). The first rotation occurs in the northern massifs; once assembled, the northern domain rotates 90° clockwise with respect to the central-southern domain. Then, both domains rotate 40° during or immediately after the last stage of accretion, and before emplacement of the V3 lavas and ophiolite obduction. (c) Model minimising the rotation between the northern and central-southern domains, by assuming a reversed polarity of the magnetic field in the northern domain. The assembly of the northern massifs is the same as in model (b); the rotation of the northern domain with respect to the central-southern one is 45° counterclockwise; the central-southern domain is not rotated, as in model (a).

the V1-V2 rotation of 40°), which is not documented in literature.

Another possibility is that the massifs have been remagnetised, the hypothesis proposed for the southern massifs by Feinberg et al. (1999), which is discussed and discarded by Weiler (this issue). These authors, arguing on similar paleomagnetic orientation of the high- pressure metabasites from the continental margin, which are dated 80–78 Myr (El-Shazly and Lamphere, 1992), propose that the southern domain lithologies were remagnetised during obduction due to advective fluids moving up from the metamorphosed continental margin into the faulted ophiolite nappe. Unpublished results obtained by J. Besse (pers. comm.) on a Hawasina Permian to Triassic exotic block from the Baid window (between Sumail and Wadi Tayin massifs) yields an uniform NNE magnetic orientation, in the range of that obtained in the southern ophiolitic domain. This clear remagnetisation is ascribed to the ophiolite being thrust over this Hawasina block during obduction (~85 Myr). If we assume that the ophiolite was also remagnetised in relation with obduction, that the southern domain was remagnetised at ~85 Myr at the end of the long normal period, and that the northern domain was remagnetised soon after, during an inverse period, models can be proposed in which the northern massifs would have suffered moderate individual counterclockwise rotations to come in their present alignment with the southern massifs (Figure 10c).

Unfortunately, the hypothesis of a general rem agnetisation during obduction is not favoured because the magnetite, for most part coeval of the development of the lizardite network in olivine (Yaouancq and MacLeod, this issue), is similar in peridotites and gabbros from both the southern and the northern domains. Moreover the same type of alteration is described in peridotites and gabbros from the present-day ocean floor, particularly in the drilled lithosphere from Hess Deep (Gillis et al., 1993; Mevel and Stamoudi, 1996), which is not older than 1 Myr, and where similar high-T magnetite-carried magnetic remanence has been investigated (Pariso et al., 1996; Richter et al., 1996). In addition, a late remagnetisation would neither explain the V1-V2 rotations documented in the northern massifs (Perrin et al., this issue), which suggest that the magnetisation was acquired very early, during the volcanics extrusion, nor the V3 orientation which corresponds to basalt emplacement also predating obduction and which is close to the present-day north. For these reasons, we tend to discard also this scenario.

Conclusions

A structural synthesis

In the preceding discussion of the data on spreading and early thrusting activity in the Oman–UAE ophiolite paleo-accretion system, local interpretations based on the analysis of typical structures have been proposed. We wish here to extend these analyses to the entire ophiolite belt and discuss a general interpretation, from the spreading history to the oceanic detachment and obduction. Before, we need to present a new synthesis for the northern domain, to recall a former synthesis for the central and southern domain (Nicolas and Boudier, 1995), and to possibly extend it to the entire belt.

Aswad-Fizh propagating system

Following a recent and detailed structural study of the Aswad massif (Nicolas et al., in press), a simple model showing how Aswad could represent a propagating ridge segment was introduced. We elaborate here on this model and introduce the structural constraints from the Fizh massif (Figure 11a) to propose integrated models (Figures 11b and 11c).

The starting assumption is that the Fizh lithosphere was not generated by the Aswad ridge segment because the fossil HT mantle flow was E-directed and not W-directed as in the Aswad lithosphere. This points to the Fizh spreading axis as being located to the west of the massif, and to the Wadi Hatta and northern Fizh shear zones as being issued from a sheared basin between competing ridge segments. Previous workers have already noted the abundance, in the northern Fizh crust, of amphibolite-facies sinistral shear zones. The occurrence of an intense hydrothermal alteration in this crust is confirmed by a specially deep and wide zone of $\delta^{18}O$ (Stakes and Taylor, 1992), of orthopyroxene-rich intrusions (Boudier et al., this issue), and of successive injections of diabase dikes, some largely tilted. This area has been successively interpreted as a leaky, sinistral, transform fault (Smewing, 1980), a model close to that of Figure 11b, a large-offset overlapping spreading centre (Reuber, 1988), a model close to that of Figure 11c, and/or a propagating rift structure (MacLeod and Rothery, 1992); assumed propagation is southward as in our earlier model (Nicolas et al., in press). Our analysis complements these studies by tracing the continuity between the crustal shear zones and kilometre-wide, LT shear zones in the mantle section, also showing a dominant sinistral shear sense (Figure 11a). We have also mapped the various mafic dikes present in the mantle section; these dikes and the average diabase sheeted dike direction have been represented in the same structural sketch (Figure 11a). In order to derive propagator models (Figures 11b and 11c) from this sketch, two assumptions are made:

– The Alley rift which separates, along a NS direction, the Fizh massif from the small Shaik massif is a rift, as proposed by Lippard et al. (1986), and it is ascribed to the southward propagating tip of the Aswad segment.

– In order to align the NNW-SSE Aswad ridge axis with the N–S Alley rift, we assume that the Hatta sinistral fracture zone was still active soon after the accretion of the whole system, and that it has displaced the Aswad ridge axis some 10 km westward.

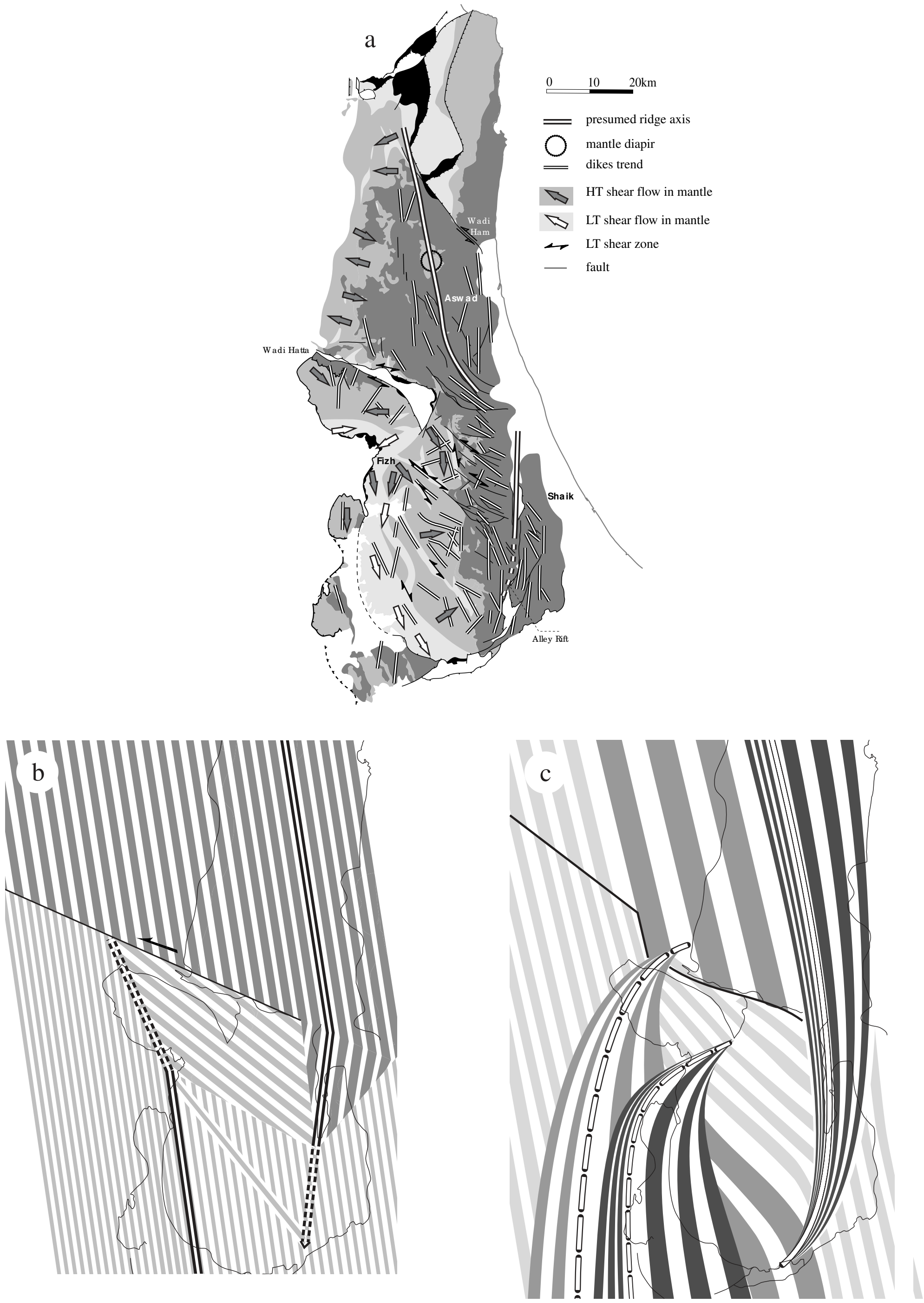

Figure 11. (a) Structural and kinematic sketch of the Aswad-Fizh area and (b and c) models of a system associating a propagating ridge (Aswad-Alley) and a doomed ridge (Fizh). The two ridges are separated by a shear domain (Wadi Hatta-northern Fizh crust). The contour of the structural sketch (a) has been figured in the models (b and c), after a 10 km ESE displacement of the nothern area (Aswad) along the Wadi Hatta fault. Model (b) represents a stable ridge system, with two segments progressively increasing their mutual distance, and with, in-between, the tip of a growing shear domain (Hey et al., 1980). Model (c) represents a mobile ridge system in which the tip of the western ridge jumps and curves inwardly in order to maintain its distance with a stable and propagating eastern rift (modified from Korenaga and Hey, 1996). The model shows two successive ridge jumps. Only the last one (darker decoration) is somewhat constrained by the structural data.

In the models of figure 11b and 11c, it should be kept in mind that the location of the western ridge axis is not fixed. We tentatively locate it near the front of the ophiolite nappe, either inside the nappe, along a major shear domain as in the model (a), or along the front of the nappe as in the model (b). Because the ridge axis is a mechanically weak band, these tectonic lines may have been the locus of detachment for the early oceanic thrusting (Nicolas, 1989, p. 62). The Fizh ridge axis could, however, have been located farther west. In the general setting of competing ridge segments with the eastern segment propagating and the western one receding (Nicolas et al., in press), the models of Figures 11b and 11c both explain the data of Figure 11a. The model in Figure 11b illustrates the geometry of fixed segments with the development of a well fixed shear domain, issued from an instantaneous transform fault migrating south through time, a system described along the EPR by Hey et al. (1980) as a 'non-transform ridge offsets' system. The model in Figure 11c illustrates the geometry of a jumping western ridge axis; thus the western and eastern ridge axes keep their initial distance with, in-between, a rotated and sheared overlapping basin, a model inspired from Korenaga and Hey (1996) for the 29°5 overlappers of the EPR.

In the model of Figure 11b, the tip of the western ridge (the doomed rift tip) is rotated away from the eastern segment; the geometry of the ridge system is such that the transform zone is transpressional, as confirmed by the existence of two systems of sinistral shears and faults: the dominant WNW–ESE one, and a second one trending NNW–ESE to N–S (Smewing, 1980). The latter is best exemplified by the large LT-mantle shear zone following the vertical Moho in the northern Fizh massif. The Moho inside this zone may have been steeply rotated in relation with the propagation of the Aswad rift, as proposed above for similar situations. The deformation and rotation of the ridge structures within the transform zone would operate by sinistral shear and not by dextral bookshelf faulting as documented in oceans by Reiser Wetzel et al. (1993), possibly because of the compressive environment. However, this would not predict the injection of WNW–ESE diabase dikes, cutting and disrupting the earlier, tilted dike complex (MacLeod and Rothery, 1992).

In the model of Figure 11c, the western ridge is seen as having jumped eastward to its final position and its tip has rotated towards the eastern ridge segment which is assumed here to be fixed, for the sake of simplicity and in the absence of any structural evidence. We also assume that the tip of the eastern and propagating rift penetrates farther south than in the model of Figure 11b, and rotates to the SW, a possibility which is compatible as well with the model of Figure 11b. From the divergence of HT mantle flow trajectories, we tentatively locate a diapir where now is the large Hawasina window, at the tip of the western segment. The eastern limit of this segment follows the major N–S to NNW–SSE dextral shear zone, which is separating two domains of distinct HT flow trajectories. The southward shear flow directions are related to the last activity of this western segment and the eastward flow directions, to a former accretion issued from the same ridge system. Overall, the model of Figure 11c seems to fit better the data, but we feel that presenting both models reflects our present limits in the understanding of this ridge system.

The NW–SE new ridge opening in the central and southern massifs

The results of detailed structural mapping in the Wadi Tayin , Sumail and Nakhl-Rustaq massifs lead to the description of a NW–SE ridge system which is opening in a lithosphere with a NE–SW sheeted dike direction (Nicolas and Boudier, 1995). This younger segment has been actively studied (Jousselin et al., 1998) because it is centred on several diapirs in the mantle which have been sampled by the ophiolite and which are now frozen within it. In the mantle and lower crust, the boundary with the older lithosphere, which is the limit of extension of NE–SW diabase dikes, is sheared. From the study of such shear zones, we deduced above that the age difference between the older NE–SW lithosphere and the younger NW–SE one was ≈ 1 Myr.

More recent mapping (Boudier et al., 1997), and reassembly of the massifs as in Figure 6, allows us now to propose a reconstitution of the NW–SE system throughout the southern and central massifs. In the Bahla and Miskin massifs, occurrences of large shear zones parallel to the general NW–SE direction, and constituting a limit between NE–SW and NW–SE diabase dikes suggest that there may be opening branches, southwest of the main segment. Because of their external position close to the front of the ophiolite nappe, Bahla, Miskin and Wuqbah (whose fit is not understood) are affected by many late thrusts making the structural data difficult to interpret.

Structurally, the simplest interpretation of the Haylayn and Sarami massifs is that they have been created by a farther northern extension of the main NW–SE segment, the axis of which is located below the crust of the exposed ophiolite, or closeby to the east (Figure 6). This interpretation is based on the following reasons:

– North of the mapped segments, the sheeted dikes in these two massifs strike steadily NW–SE, in continuity with the mapped segment strike.

– The HT mantle flow in both massifs is W-directed, indicating that they are issued from a ridge axis located east of the mantle exposures.

– In the Haylayn massif, the foliated gabbros dip on average to the east with respect to the sheeted dike complex, suggesting that, by contrast with the preceding conclusion, the axis of the magma chamber is located west of the gabbro exposures.

The location of the ridge axis inside the massifs adopted in Figure 7 accounts for these points, in particular for the last one. If this segment was located far away to the northeast, the HaylayN–Sarami lithosphere would be comparatively cold and rigid by the time of oceanic detachment. However, the mapping has revealed, mainly in Haylayn, that the lower crust was deformed in a large scale syncline, and that the Moho was generally steep and the mantle section, invaded by MT and LT deformation zones. This suggests that the lithosphere was still plastically deforming, and consequently has been newly accreted from an axis located nearby. The steep Moho may alternatively reflect a propagating rift environment (see above in the dicussion chapter).

In conclusion, the young NW–SE segment seems to be composed of smaller individual ridge segments, possibly opening en-échelon. As the segment is 40–50 km wide and could be extending over 200 km, opening rates would be even higher than the assumed fast spreading rate. A dextral shear component accompanying the NW–SE expansion is suggested by the observation of deviated flow trajectories around the Maqsad diapir and of a N–S diking direction, subsidiary with respect to the dominant NW–SE direction (Nicolas et al., this issue).

Structural relations between the two ridge systems

In summary, there is a continuous and fairly consistent ridge pattern in the northern massifs from Aswad to Fizh, and possibly farther south to Hilti, thanks to a reasonable extension of the Fizh ridge segment southwards. Only the Khawr Fakkan massif, thrust over the north of Aswad, takes no clear place in this ridge system. Its ridge axis has been tentatively located outside the massif in a NE direction (Gnos and Nicolas, 1996).

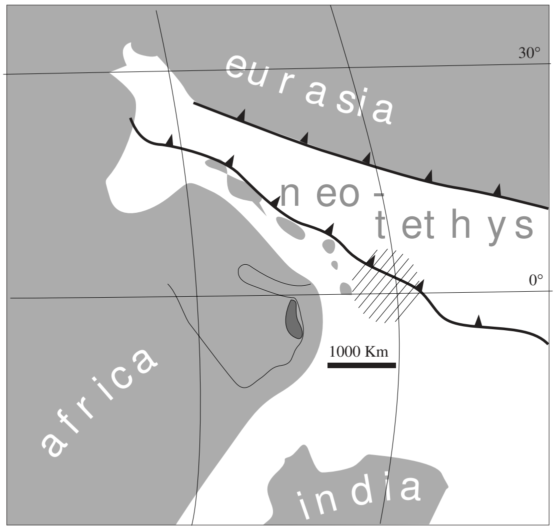

Figure 12. Regional paleo-geographic sketch at 94 Myr, the age of final accretion and oceanic detachment (barbed line) of the Oman–UAE ophiolite. The original site of the ophiolite is presumably located within the hatched circle (redrawn from a synthesis by Van der Voo et al., 1999)

Though they are now largely separated, the remarkable structural continuity of the southern massifs allows them to be fitted together in the scheme of a large (140–200 km $\times$ 40–50 km) NW–SE ridge system opening inside a subcontemporaneous lithosphere accreted in a NE–SW direction; the NW–SE segment is composed of individual ridge segments approximately organised en-échelon. We try below to incorporate the northern and southern domains into a single model for the entire Oman–UAE ophiolite. However, it should pointed out that, if a major structural boundary is required between an independent northern domain and the rest of the ophiolite, as suggested by the paleomagnetic study, structurally it should not be placed between Sarami and Haylayn, but between Hilti and Sarami (Figure 7).

Looking for a model integrating structural and paleomagnetic data

As a frame for the forthcoming discussion, we use available informations from paleo-geographic studies (Figure 12), showing that the ophiolite cradle is poorly located somewhere near the paleo-equator and within $\sim$1000 km from the Arabian margin. After detachment close to the ridge of origin, the ophiolite was thrust in a SW direction, with possibly a dextral shear component, related to the convergence of Africa and Eurasia plates.

Perrin et al. (this issue) and Weiler (this issue) conclude that there has been large (90°–130°) rotations of the northern domain with respect to the central-southern domain. The invoked sharp structural boundary between the two domains is a challenge, in particular because there is no marked structural and petrological discontinuity between the northern and southern ridge systems, neither in sheeted dike orientations, nor in HT and LT flow directions in the mantle, as well as in the common LT transport direction of most massifs toward the SE.

Weiler (this issue) proposes that the northern domain may be issued from a contemporaneous but independent part of a larger ridge system, detached soon after accretion, possibly as a piece of a microplate, rotated by impinging a continental promontory during motion, and finally emplaced against the southern domain. We note however that the rotations and translations envisaged in Figure 10a represent grossly 1000 km in displacement, which would require 10 Myr at a rate of 100 km Myr^{-1}. Such a time span is incompatible with the time span limited to $\sim$2 Myr between accretion, massifs assemblage of the northern and central- southern domains and their common LT emplacement-related thrusting. This reasoning, which is based on the thermal constraints on the successive deformations in peridotites, is confirmed by radiometric ages (Hacker et al., 1996; Figure 9). Again, within 1–2 Myr, the ages are similar for accretion and basal thrust in the northern and central-southern domains. For this reason, we need to discuss further the alternative model developed by Perrin et al. (this issue).

With their new data on the volcanics, and integrating the ages constraints of Hacker et al. (1996), Perrin et al. (this issue) propose a progressive magnetisation, first for the northern lavas, next for the southern lavas (located outside the NW–SE segment), and together with the southern gabbros (located inside the NW–SE segment), during a clockwise rotation of the ophiolite, operating during the accretion episode (Figure 10b); the main rotation of the northern domain with respect to the southern domain is 90° clockwise, following V2 emplacement in this domain. Once the two domains have been assembled, a common 40° clockwise rotation occurs, following V2 emplacement in the southern domain, and coeval with the Wadi Tayin, Sumail, and Nakhl-Rustaq gabbro magnetisation.

A geodynamic situation consistent with our structural and kinematic data and with the progressive clockwise rotation of Perrin et al. (this issue) could be a Manus-type microplate (Martinez and Taylor, 1996), a type of microplate opening in a transtensional regime and close to a continental margin. Although the Manus microplate is opening in a back-arc basin, such an environment does not seem to be, tectonically, a requirement, and it is not favoured for the Oman–UAE ophiolite (see above). The model of Figure 13 would fit with the presumed location of the original ridge system of the Oman–UAE ophiolite, close to the Arabian margin (Figure 12), a location accounting also for the subcontinental nature of marine sediments associated to the ophiolite (Béchennec et al., 1990) and for the potassic signature of some granites intruding the ophiolite (Briqueu et al., 1991); the transtensional regime would account for the dominant regime of dextral shear and clockwise rotation which are recorded in the ophiolite. The progressive clockwise rotation of the northern part of the belt would be followed by the dextral, emplacement-related, SSE-directed thrusts and shears, subparallel to the ridge system. The next stage is a WSW-directed thrust towards the older oceanic lithosphere and the Arabian margin, as documented by Boudier et al. (1988). The W-directed basal deformation in the newly accreted and still hot Wadi Tayin and Sumail massifs records this new thrusting direction whereas in the central and northern massifs which were accreted earlier, this W-directed thrust is only recorded in the greenschist facies metamorphic aureole. This scenario of successive stages of local accretion, shearing, thrusting and related rotation, which progress from north to south of the belt during a time lapse of 1–2 Myr, accounts well for the continuity of HT structures related to accretion and LT structures related to ridge tectonics and early detachment, which is a persistent image resulting from this structural study. The model of Figure 13 satisfies all major structural, kinematic and paleomagnetic constraints, but it is obviously too simplistic, betraying both the paucity of present-day ridge analogues, and the limits of our understanding of this unique ophiolite.

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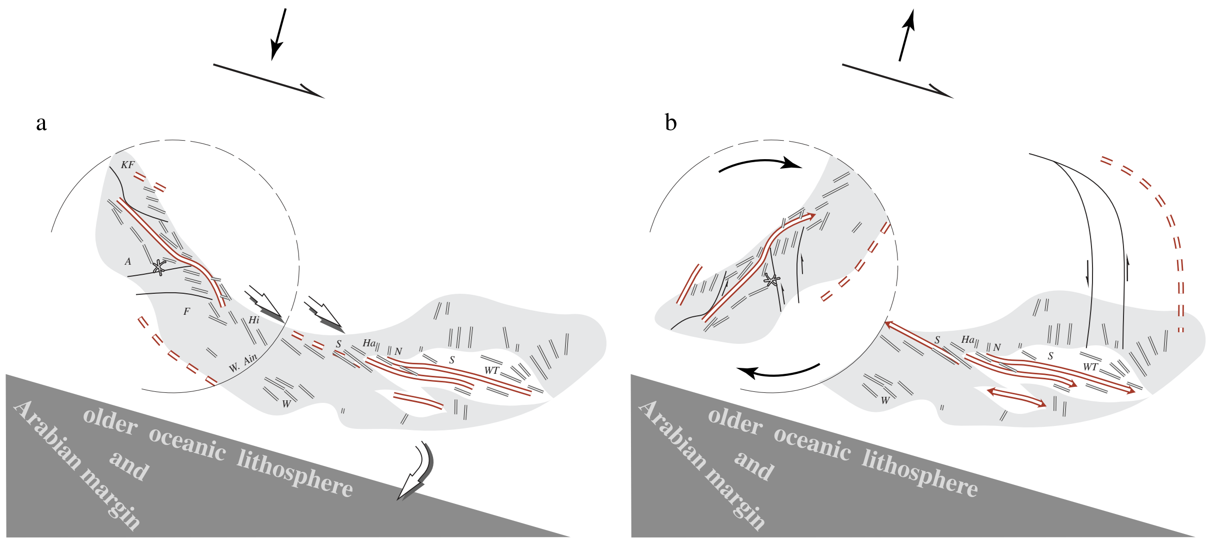

Figure 13. A simplistic model for the Oman–UAE paleo-ridge system. The geodynamic setting is a NW–SE dextral and transtensional oceanic environment along the rigid boundary of the Arabian passive margin and its associated old oceanic lithosphere. Accretion occurs during the transtension phase, followed by a transpression phase responsible for detachment and obduction. The ridge-related structures are simplified from Figure 6. In (a), the system is represented soon after accretion (or possibly still during accretion for the southern domain) during the main stage of SSE-directed thrust, before the late 40° clockwise rotation proposed by Perrin et al. (this issue; rotation III in Figure 10b) and the SW-directed thrust over the Arabian margin. In (b), the system is represented ∼1 Myr earlier, before the 90° clockwise rotation of the northern domain (Perrin et al., this issue; rotation II in Figure 10b). The accretion of the northern segment, and associated internal rotations, occur before the stage represented in (b).

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