INTRODUCTION

Since the first written definition of ophiolites by Brongniart (1821), three distinct episodes of discovery have marked the long history of ophiolites’ relation with oceanic lithosphere. An early historical period was marked by a few premonitory views developed at the turn of the nineteenth century by Italian geologists working in the western Alps and relating the “pietre verdi” to ocean seafloor (see in Nicolas, 1989, p. 3, a review on this period). This period ended in the 1960s, during the early days of plate tectonics, when the new interest in oceanic lithosphere prompted new studies in ophiolites. A milestone was the “ophiolite manifesto” (Anonymous, 1972), in which American and European geologists, starting from very distinct positions, converged to adopt a common definition of ophiolites. The idea that ophiolites represented fossil oceanic lithosphere was not explicitly mentioned, but was largely accepted. Subsequent studies were predominantly geochemical and were rapidly dominated by the question of the oceanic setting of origin. The mid-oceanic ridge (MOR) setting was favored by geologists and supported by the isotopic signature of some ophiolites (Alabaster et al., 1980; Jacobsen and Wasserburg, 1984), whereas Miyashiro (1973), on the basis of major and trace elements of basalts, claimed that the Troodos ophiolite had been generated in an island arc environment. This is conveniently referred to as a supra-subduction zone (SSZ) environment, either an island arc environment or a backarc basin. At the same time, Pearce and Cann (1973) defined the discriminant diagrams that, on the basis of minor and trace elements, were supposed to allow us to distinguish the MOR or SSZ origin of any ophiolite. These diagrams were immediately applied to many ophiolites around the world, driving to the still dominant conclusion that most of these ophiolites were generated in SSZ environments (e.g., Ishiwatari, 1994).

The last period coincides with the discovery by marine geophysicists (Morton and Sleep, 1985; Detrick et al., 1987) of a seismic reflector at 2 km below the seafloor of fast-spreading ridges that was interpreted as a melt lens at the top of a magma chamber. This intrinsically very important discovery had a side effect that is also important in our discussion. Until this point, geophysicists had been reluctant to consider ophiolites as potential fossil remnants of oceanic ridges that were very similar to...
those they were studying. Among their reasons was the difficulty of accepting the existence of a magma chamber at shallow depth below an active ridge. Their reluctance was increased by the fact that most geologists were claiming that this chamber had a very wide top, a view that admittedly is untenable from both a mechanical and a thermal point of view. Ultimately, the melt lens discovery helped to bridge the gap between these two communities. Another development helped to establish a more confident and constructive relation between the ophiolite community and the marine geology and geophysics community—the collective results of the Ocean Drilling Program (ODP) Leg 147 (to the Hess Deep) in 1993 (Gillis et al., 1993). The foliated-layered gabbros of the Hess Deep originated at the East Pacific Rise, and displayed an astonishing petrological and structural analogy with the Oman ophiolite gabbros observed at the same depth. For many geophysicists, this analogy has been a strong incentive to integrate the ophiolite data into their ridge studies.

Where do we stand now regarding the origin of ophiolites? What are the present questions about their genesis and the questions we foresee in the near future? How do we address them? We need first to clarify our motivation in studying ophiolites. Widespread at Earth’s surface, ophiolites can be studied for their record of regional paleogeographic and geodynamic evolution. Here, we are more interested in their contribution to the origin and evolution of the oceanic lithosphere that they represent and in using them to better understand the dynamics of oceanic spreading centers. Slow-spreading ridges are characterized by major topographic gradients, along which the internal structure of in situ oceanic lithosphere can be directly studied and sampled by submersibles and by dredging. In fast-spreading ridges, however, (and with the exception of a few deeps), the internal structure of oceanic lithosphere is masked by a uniform blanket of basalts. The contribution of ophiolites, where the deep structure of fossil oceanic lithosphere is directly exposed, thus is mainly interesting in the latter situation. Unlike slow-spreading activity, fast-spreading activity also tends to be time-independent on the scale of at least a few million years and to generate ridges that are homogeneous over lengths of 50–100 km. To better understand the dynamics of seafloor spreading, it is thus reasonable to consider primarily fast-spreading ridges, mainly the East Pacific Rise, which is the most studied of all. In this respect, structural and petrological studies in ophiolites that originated at fast-spreading centers complement remarkably well the geophysical studies conducted on active fast-spreading ridges.

Our main goal here is to illustrate by a few examples this combined approach of structural studies in selected ophiolites and marine geophysical studies conducted mainly at the East Pacific Rise. Karson (1998) presents an important contribution on this topic that is based on a detailed review of the tectonic windows in ocean floor through which some insight on the structure of the oceanic lithosphere is obtained. Karson emphasizes the contrast between the “stratiform architecture” of the ophiolites and these tectonic windows in the oceans. On the basis of such premises, this contrast is not a surprise. Our mapping over 500 km in the Oman–United Arab Emirates ophiolite (Nicolas et al., 2000) confirms both the general stratiform architecture of this huge ophiolite and the existence of a locally intense ridge tectonic activity. For instance, steepening of the Moho to the vertical at the propagator tips is described below. The associated attenuated ophiolite crust suggests that this situation compares with that of the East Pacific Rise deeps.

Before returning to this topic, we wish to address two controversial issues, and to present viewpoints that are based on detailed structural mapping of 15 ophiolite massifs and on more limited studies in a similar number of other massifs throughout the world. Finally, we call attention to similarities and differences in ophiolites that may help to better situate them with respect to oceanic lithosphere and to enlighten their origin.

**TWO CONTROVERSIAL ISSUES**

We discuss here two subjects that so far have been too loosely accepted as ground truth by most geoscientists interested in ophiolites. In so doing, we update but also substantially repeat earlier discussions (Nicolas, 1989, p. 199–201 and 258).

**Ophiolites Originate Dominantly in SSZ Settings**

From a very general point of view, the inferred SSZ origin is difficult to sustain for all ophiolites, simply because the ~70 ophiolites for which sufficient information of chemical affinity is available exhibit too much diversity to be so narrowly constrained. For example, a few ophiolites display a felsic extrusive section, such as the Canyon Mountain in Oregon (Misseli and Boudier, 1985) or the Guana Gato in Mexico (Lapierre et al., 1992), fitting well with an island arc environment origin, but most do not. A few ophiolites, such as Karmoy in Norway (Pedersen and Hertogen, 1990), are covered by island-arc related sediments, which points to an island arc environment or backarc basin origin, but many have a deep ocean sedimentary cover that is more compatible with an MOR origin. Incidentally, a fine analysis of the associated marine sediments may turn out to be the best criterion for unraveling the environment of origin of an ophiolite (Pesagno et al., 2000).

The generalized interpretation of SSZ origin of ophiolites has been largely spread out in literature since Pearce and collaborators (Pearce and Cann, 1973; Pearce et al., 1981) pointed out that the Troodos and Oman ophiolite lavas exhibit geochemical characteristics that are transitional between mid-ocean ridge basalt (MORB) and island arc tholeiite (IAT). In Troodos, lower pillow basalts are similar to ocean floor basalts, although upper pillow lavas and subsequent Arakapas ultramafic volcanics have high-Mg, low-Ti, and other high-strength field element (HSFE) concentrations that point to affinities with IAT (Pearce, 1975), besides a large variability of these characteristics (Cameron, 1985). In Oman, the range from MORB to IAT signatures suggests a succession of volcanic episodes (Ala-
baster et al., 1982): the “Geotimes” ridge-axis volcanics (V1 for Ernewein et al., 1988) are very close to MORB, and subsequent seamount-type lavas (V2), “Lasail, Alley, Cpx-phyric” trace an evolution toward an “arc signature.” The chronology of this evolution is well constrained to 1–2 Ma on the basis of biostratigraphy (Tippit and Pessagno, 1981), and is consistent with the time difference between V1 and the initial detachment of the ophiolite from its igneous setting (e.g., Hacker, 1994).

Actually, the arc signature covers a large spectrum of geochemical characters, variably represented in present forearc and arc volcanic compositions, and whose end member, boninite, is well defined (e.g., Juteau and Maury, 1997, p. 33–35) by (1) an extremely depleted character, i.e., high silica and magnesium content and high depletion in rare earth elements (REE) and HFSE (Zr, Ti, Nb, Ta), interpreted as resulting from a high degree of partial melting in the mantle wedge, and (2) an enrichment in highly incompatible large ion lithospheric elements (LILE), interpreted as a subsequent contamination by hydrous fluids. It is the occurrence of one or the other of these characters in volcnicas overlying many ophiolites that supported the common thinking that ophiolites originated in a SSZ environment.

A major problem concerning this model is how to explain the rapid shift from an MOR to an SSZ signature, either in time, as in Oman (~1–2 Ma; Hacker, 1994), or in space, as in Albania (20 km, Shallo et al., 1995). Keeping in mind that the SSZ signature reflects fluid contamination during melting of a depleted mantle source, can we suggest possibilities to produce a SSZ signature other than the arc-related fluid flux in the mantle wedge overlying a subduction zone? Early in the study of the Oman ophiolite, we related the V2 lavas to the initial stage of oceanic thrusting of the ophiolite (Boudier et al., 1988) and explained its occurrence by a water contamination of the melt source due to the dehydration of the underthrust oceanic crust, a process akin to a subduction-related melting. Other hypotheses may be envisaged as well, such as contamination by seawater of the still hot mantle at a distance of a few tens of kilometers from the ridge, in keeping with the recent discovery of deep and hot hydrothermalism near ridge axis (Manning et al., 2000; Nicolas et al., 2003). This second hypothesis has been further developed by Godard et al. (2003) to explain the transitional trace elements and isotopic strontium signatures of V1 Geotimes and V2 Lasail in Oman volcanics. In compilations that include recent data on seamounts from the East Pacific Rise (Niu and Batiza, 1997) and North Fiji Basin (Fleutetolot, 1996), the authors point to the fact that the V2 Lasail lavas, which had been assigned to SSZ origin on the basis of incompatible elements ratios, were lying in the extended field of seamounts. We wish to defend an open-minded attitude on the basis of the fact that the great variability of “arc signature” documented in volcanics from ophiolites suggests that different processes (not necessarily understood yet) may rule this diversity. When the geochemical approach does not meet with data from other origin, an SSZ setting should not be regarded as decisive on the basis only of geochemical signature of the off-axis volcanics.

**Abyssal and Ophiolitic Peridotites**

Many studies have been devoted to the specimens of mantle peridotites collected in the oceans, the most extensive study being that of Dick et al. (1984). Their data are still largely used, remaining the most extensive source for petrological and geochemical modeling of oceanic lithosphere. Comparing their data on abyssal peridotites to those available on peridotites from ophiolites, Dick and Bullen (1984) conclude that there are significant discrepancies. They observe that, on average, the ophiolite mantle is more depleted than the oceanic one and conclude that ophiolites should result from a higher degree of melting, and thus should be good candidates for the hydrous mantle from SSZ environments. We wish to reiterate our doubts on the relevance of the data on abyssal peridotites, recalling that they represent only slow-spreading oceanic lithosphere and associated transform faults, where mantle specimens are readily available on the seafloor. With no basalts attached to them, they may not represent constructional oceanic lithosphere. Mantle specimens from fast-spreading environments are exceptional, being collected along a few transform faults and deeps, which probably are not representative of the normal fast-spreading lithosphere. Since the Dick et al. study, more analyses have been published on the East Pacific Rise mantle. As an example, in Figure 1, we have plotted the characteristic chromium number index measured in peridotites from the fields of abyssal domains, East Pacific Rise, island arcs, and three selected ophiolites. The biased sampling of “abyssal peridotites” in favor of slow-spreading lithosphere is clearly reflected in this figure. The Bay of Islands ophiolite, thought to represent moderate spreading rate conditions (5 cm/yr; Suhr, 1992) plots in this field, and the Oman ophiolite attributed to fast-spreading conditions plots in the East Pacific Rise field. The Josephine ophiolite plots in the SSZ domain, in agreement with the geodynamic evolution of this Jurassic ophiolite in the Mesozoic California. Altogether, abyssal peridotites have a more fertile signature than most ophiolites. This is not surprising, knowing that the majority of ophiolites were derived from moderate- to fast-spreading environments (see below). From this discussion, we draw two conclusions: (1) that the high chromium number of many ophiolites does not necessarily mean that they originated in SSZ environments, and (2) that petrological and geochemical modeling of fast-spreading ridges should rather turn toward ophiolites and stay away from the abyssal peridotites reference.

**MARINE GEOPHYSICAL AND OPHIOLITIC STRUCTURAL DATA: THE NECESSARY CONFRONTATION**

As mentioned above and stressed by Karson (1998), the flat relief of fast-spreading ridges masks all internal structures. The geophysical, mainly seismological, tools provide only indirect information. This is where structural data from ophiolites with a clear fast-spreading center origin can provide the missing information; however, matching the results from both sources
Diapir Controversy

From the mapping of steep and concentric high temperature mantle flow structures in the Zambales (Philippines) and the Cyprus ophiolites, Nicolas and Violette (1982) have proposed that the asthenospheric mantle flow beneath oceanic ridges was diapiric (three-dimensional). The East Pacific Rise segmentation was discovered at the same time (MacDonald and Fox, 1983; Lonsdale, 1983; Whitehead et al., 1984) and its scale of 50–100 km, comparable to the spacing of transform faults along the Mid Atlantic Ridge, prompted the idea that the uprising mantle could be partitioned accordingly. This resulted in a number of three-dimensional models of mantle upwelling (Whitehead et al., 1984; Rabinovich et al., 1984, 1987; Scott and Stevenson, 1989). The increasing confinement for lower spreading rates predicted by the last model has been confirmed by subsequent analog (Magde et al., 1996) and numerical models (Lin and Morgan, 1992; Sparks and Parmentier, 1994; Magde et al., 1997) and by the discovery of a strong mantle Bouguer negative anomaly in the middle of the Mid Atlantic Ridge segments (Detrick et al., 1995), suggesting that a hot and melting asthenosphere is rising from below. In contrast, the East Pacific Rise has no clear mantle Bouguer anomaly, suggesting a two-dimensional mantle uprising. The remarkable MELT experiment was run in the fastest and most linear part of the East Pacific Rise to test the mantle flow and melting below through a combination of seismology, gravity, and electromagnetism. With the resolution of the experiment limited to a depth of 100 km, the mantle uprising is clearly two-dimensional (Melt Seismic Team, 1998). Finer scale geophysical investigations conducted at the 9°N East Pacific Rise natural laboratory do not rule out that, on the scale of 10 km, mantle upwelling may have a three-dimensional structure (Barth and Mutter, 1996; Dunn et al., 2000; Canales et al., 2003) (Fig. 2).

In the meantime, increasingly detailed structural studies in the Oman ophiolite (Ceuleneer et al., 1988; Nicolas and Boudier, 1995; Jousselin et al., 1998) (Fig. 3) confirmed the occurrence of several frozen diapirs. Because the Oman ophiolite is regarded by all authors as being issued from a fast, possibly a super fast (Nicolas et al., 2000) spreading center, a comparison is possible with the East Pacific Rise. The most apparent explanation regarding the discrepancy between two- and three-dimensional mantle uprising is that the resolution of most East Pacific Rise geophysical data is not yet sufficient to image small diapirs, the size of those mapped in Oman. Another explanation is that the small mantle uprising diapirs in Oman, breaking just below Moho where they are mapped, may be local heads of larger three-dimensional mantle upwellings whose flow would rotate at a greater depth, below the 5–10 km depth sampled by the ophiolite. Following this idea, Nicolas and Boudier (2000) have mapped in this ophiolite the domains of intense melt circulation that should be related to mantle uprising. Such domains are characterized by the development of thick Moho transition zones.
Where ophiolites come from and what they tell us

(MTZ) where the mantle harzburgites are replaced by dunites and by a similarly large development of olivine gabbros and wehrlitic intrusions into the lower crust. This mapping has revealed the existence of a few areas, some 50 km across, incorporating the small mapped “structural” diapirs of Figure 3. These larger domains, in which the MTZ is thicker on average and the gabbro unit is richer in olivine, could be the petrological signature of larger mantle diapirs whose size seems more compatible with the scale of segmentation expected below oceanic ridges.

Seismic Anisotropy And Mantle Tectonics

In Figure 2, P-wave anisotropy, deduced from crossed seismic lines, shows a fast-velocity axis remarkably parallel to the spreading direction at a large distance from the ridge, which is comparable to the findings of the MELT experiment. This fast seismic velocity is now interpreted as being parallel to the plastic flow direction in the mantle. At the finer scale of the ellipse of attenuated seismic velocities seen in Figure 2, which is that of the diapirs mapped in Oman, the expected mantle flow should not be so regularly oriented, but rather should be diverging. The measurements necessary to check this point seem now in view and this question may serve as an entry into a new and promising research domain.

The interest for new seismic waves, such as the SKS shear waves, which rise vertically from the core mantle boundary and whose splitting can be used to know the local seismic anisotropy in the horizontal plane (Silver, 1996), has paved the way for new studies on what has been called mantle tectonics. Because it is possible to obtain a large number of such data on a regional scale, it becomes possible to relate the mantle flow pattern deduced from them to the regional tectonics deduced from surface studies. This has been illustrated by beautiful examples incorporating the upper mantle into continental crust tectonics (Barruol and Souriau, 1995). It should not be long before such techniques are commonly applied to the oceanic lithosphere, revealing the patterns of upper mantle flow. This will make a direct comparison possible, at the same 10 km scale, with similar flow patterns deduced from structural studies in ophiolites. Spectacular progress in better understanding of mantle flow below oceanic ridges is in view.

What is a Magma Chamber?

The magma chamber in which the layered gabbros of ophiolites were generated traditionally has been regarded as a large melt pouch in which crystals would be sorted by gravity after nucleation at the colder ceiling and the heavier crystal mush settled on the floor. This view was shattered by the discovery of the perched melt lens at fast-spreading ridges <2 km wide and only a few tens of meters thick (Singh et al., 1998). Large-angle seismic reflection studies have shown that between the melt lens and the Moho there is a wide triangular domain of strong
seismic attenuation (Fig. 2). This attenuation has been ascribed to the presence of hot gabbros with trapped melt pockets. For marine geologists and geophysicists, the magma chamber thus is reduced to a melt lens. The vast attenuated domain underneath would have been filled with the gabbros that started crystallizing within the melt lens before subsiding as a gabbro mush and eventually evolving into a deforming solid (Quick and Dellinger, 1993; Phipps Morgan and Chen, 1993). This view is incompatible, however, with the ophiolite record showing that the gabbro pile still displays magmatic textures and not solid-state deformation textures. Structural studies conducted in the gabbro unit of the Oman ophiolite show that these gabbros were intensely deformed during their crystallization by a process akin to suspension flow, with no trace of plastic deformation, except locally at the Moho level (Nicolas, 1992). The critical melt fraction above which suspension flow is possible is ~40%, far above the fraction initially accepted for the attenuated domain below ridges, which was <10%.

By combining different approaches, a solution to this dilemma seems foreseeable. Moreover, the studies have resulted in significant advances in the fields of both petrophysical modeling of seismic data and physical properties, and rheology of gabbro mush. Taking into account the large structural anisotropy induced in the gabbros by their large deformation and by the observed flat shape of melt pockets in Oman gabbros, it is possible to calculate the seismic attenuation for any melt fraction. Melt fractions have been estimated by comparing the attitudes of the gabbros known from field studies to the seismic paths below the ridge. They are currently found to be >10% and possibly as high as 24% (Lamoureux et al., 1999) (Fig.4).
In the meantime, analog experiments have shown that the critical liquid fraction could be reduced to ~25%–30% when flat particles were submitted to a large shear flow, because of their strong alignment compared to their random initial assemblage (Nicolas et al., 1993). In these experiments, it was also observed that the flat particles would slip one on top of the other until blocked by disoriented laths. The discovery in the gabbros of microtextures of impingement of one plagioclase lath into another, indicative of pressure solution, has been interpreted as nature’s response to the problem of blockage. A model has been proposed in which crystals slip by suspension flow until blockage and activation of pressure solution (Nicolas and Ildefonse, 1996). With this model, the melt fraction can be reduced drastically, the absolute limit being the few percent of melt where pressure solution would operate alone. The mush viscosity can be estimated from the model, and can be as high as $10^{15}$ Pa s$^{-1}$, a value indepen-

Figure 4. A: Estimations of melt fractions in the crustal low-velocity zone (LVZ) below East Pacific Rise on the basis of the East Pacific Rise seismological data and structural measurements in the Oman gabbros. B: These data are integrated with petrological modeling of the seismic anisotropy induced by both the crystallographic fabrics and the geometry of melt lenses in the gabbros. Because the lower gabbros are dominantly horizontal, the seismic velocity is close to the modeled $V_p X$; because the upper gabbros are dominantly vertical, the seismic velocity is close to the modeled $V_p Z$. The modeled $V_p X$ and $V_p Z$ correspond to melt fractions of 12%–24% (light gray area) and 20%–25% (dark gray area), respectively (in Mainprice, 1997).
dently derived from a physical numerical modeling of the Oman magma chamber (Chenevez et al., 1998).

We conclude that the attenuated domain below fast-spread- ing ridges should be considered to be the magma chamber as the continuation of the melt lens above; however, it is filled with a mush nearly as viscous as a plastically deforming body. Comparing the structures in the Oman gabbros with those of the Skaergaard Complex in Greenland, the best-documented crustal magma chamber, we observed no significant differences between them and suggested that the crystallizing gabbros deposited below the free surface in Skaergaard were rapidly forming a thick mush during their gravity-driven flow, much like the Oman gabbros (McBirney and Nicolas, 1997).

Tectonic Activity Below the Basalt Blanket of Fast-Spreading Ridges

Except for special areas such as fractures zones or micro- plates, the seafloor of fast-spreading ridges is remarkably flat with only few tens of meters of throw associated with normal faults parallel to the ridge, the small nodal basins between overlapping spreading centers and, of course, seamounts. The gravity signal also is very weak. These features suggest that the deep crustal and uppermost mantle structures are also simple and presumably are dominantly horizontal. Turning to Oman again, this is exactly what has been mapped there, particularly within the Aswad massif in the northern United Arab Emirates. In the western part of this massif, the Moho is absolutely horizontal with the lower gabbros defining cuestas and buttes over distances of tens of kilometers. In other parts of the belt, however, the Moho can be vertical more than 10–20 km along-strike, where it is commonly underlain by shear zones that are a few tens of meters thick, reflecting conditions of ~1000 °C in the mantle. Above, the lower crust has suffered an intense hydrous recrystallization; it is also strongly deformed and cut by amphibolite facies shear zones that are branching on the steep Moho shear zones. Such peculiar areas are also invaded by mafic dikes that are mostly noritic and that evolve to websterites in the mantle section. All these features point to an activity occurring very close to the ridge axis. In contrast to these vertical structures observed in the middle of the ophiolite section, both the basal contact and the basalt flows and sediments on top are moderately inclined and nearly parallel to each other (Fig. 5). These areas characterize the limits of segments and are most developed at the tips of propagators, as typical in the western part of the Nakhl massif, the eastern part of the Haylayn massif, and the center of the Fizh massif. The Mansah area in the Semail massif where an off-axis diapir has been mapped (Jousselin and Nicolas, 2000a) displays similar features. They are similarly interpreted to be a result of the uprising mantle punching a soft lithosphere, not older than 1 Ma. The asthenospheric pressure would tilt the Moho, fracture the crust, allow penetration of seawater to this depth, and thus favor the development of hydrous magmatism, as proposed by Boudier et al. (2000). To conclude, we suggest that the deep crust and uppermost upper mantle of fast-spreading ridges might be locally intensely tectonized, as observed in the Oman ophiolite, but invisible from the seafloor because of the very efficient ductile relaxation in this hot lithosphere.

OPHIOLITES AND SPREADING RATES

Ishiwatari (1985a) and Boudier and Nicolas (1985) independently proposed to classify ophiolites according to how the nature of their mantle constituents reflects the degree of partial melting in the mantle of origin. Ishiwatari considered petrological and chemical criteria in gabbros and basalts, whereas Boudier and Nicolas introduced more general criteria, largely structural. Going a step further, these authors related mantle depletion to spreading rate of the ridge of origin, a point that is discussed below. Boudier and Nicolas gave the names LOT (for lherzolite ophiolite type) to ophiolites whose mantle was lherzolitic and less depleted, and HOT (for harzburgite ophiolite type) to ophiolites whose mantle was harzburgitic and more depleted. Ishiwatari proposed three types: his “Liguria” type coincided with LOT and his “Papua” type with HOT, and his “Yakuno” type was in between. Several criteria are attached to the LOT and HOT types, the most prominent being the reduction or even the absence of the gabbro unit in LOT compared to HOT. The reduction affects mainly the layered gabbros that, when present, are usually poorly structured compared to those in HOT (Table 1).

Boudier and Nicolas (1985) related the degree of partial melting to the spreading rate, a position that has been and is still disputed because excess or reduced melting with respect to an average asthe- nospheric mantle uprising below an oceanic ridge can be explained

Figure 5. Cross-section through the Oman ophiolite nappe in the northern Fizh massif (modified from Boudier et al., 1988). The Moho has been sheared and tilted to the vertical, whereas the lower and upper contacts of the thrust nappe are gently east-dipping. The revealed internal tectonic activity has no visible consequence on the seafloor. LT—low temperature; HT—high temperature.
### TABLE 1. OPHIOLITES TYPES

<table>
<thead>
<tr>
<th>Harzburgite ophiolite type (HOT)</th>
<th>Harzburgite lherzolite ophiolite type (LHOT)</th>
<th>Lherzolite ophiolite type (LOT)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>MASSIFS CONSIDERED</strong></td>
<td>Oman-UAE (1), Papua (2), New Caledonia, southern massif (3)</td>
<td>Bay of Islands (4), Yakunoo (5), Mirdita (6), Troodos (7), Antalya (8), Kuzildag (9), Aladag (10), Xigatse (11), Muslim Bagh (12), Josepnie (13), Vourinos (14), Zambales (15)</td>
</tr>
<tr>
<td><strong>CRUSTAL SECTION</strong></td>
<td>Thickness: 4–6 km</td>
<td>Thickness: 2–3 km</td>
</tr>
<tr>
<td><strong>Volcanics overlying sheeted dikes</strong></td>
<td>Low-alumina tholeiite</td>
<td>High-alumina tholeiite</td>
</tr>
<tr>
<td><strong>Sheeted dikes</strong></td>
<td>Well expressed and steeply dipping</td>
<td>Steeply to moderately dipping (4, 7, 9, 12, 13, 14, 15), horizontal (11, 15), absent or poorly organized (5, 6, 8)</td>
</tr>
<tr>
<td><strong>Gabbros</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>exposure</strong></td>
<td>Thick, continuous (1, 2)</td>
<td>Thin, continuous (4, 12, 13, 14, 15), nestled (7, 9), locally absent (5, 6, 9, 11)</td>
</tr>
<tr>
<td><strong>lithology</strong></td>
<td>Lower: Ol-gabbro (1, 3) Upper: Ol-gabbro and gabbro-norite</td>
<td>Dominant gabbro-norite (4, 5, 8, 10, 13, 14, 15), Ol-gabbros (6, 7, 9, 11, 14) and ferrogabbro</td>
</tr>
<tr>
<td><strong>structure</strong></td>
<td>Lower layered gabbro Upper folioted gabbro (1, 2)</td>
<td>Well layered (8, 9, 10, 13, 14, 15), poorly layered-foliated gabbro (4, 6, 7, 11, 12, 13), isotropic gabbro</td>
</tr>
<tr>
<td><strong>penetrative deformation shear bands</strong></td>
<td>Magmatic deformation</td>
<td>Magmatic to plastic (flasergabbro) deformation</td>
</tr>
<tr>
<td><strong>wehrlite intrusions</strong></td>
<td>Abundant (1, 3)</td>
<td>Present (4, 6, 7, 12, 14)</td>
</tr>
<tr>
<td><strong>MANTLE SECTION</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>lithology</strong></td>
<td>Harzburgite and dunite</td>
<td>Cpx-harzburgite, locally Sp-,Plag-lherzolite and dunite (5, 6, 11, 15)</td>
</tr>
<tr>
<td><strong>temperature of deformation</strong></td>
<td>Very high to high temperature</td>
<td>High temperature to moderate temperature</td>
</tr>
<tr>
<td><strong>internal structures</strong></td>
<td>Flat foliation (except diapirs)</td>
<td>Flat (4, 6, 11, 12, 13, 14) to steep (7, 8, 14, 15) foliation</td>
</tr>
<tr>
<td><strong>shear zones</strong></td>
<td>Uncommon, vertical with flat lineation</td>
<td>Common, steep with steep lineation (8, 10, 11, 14)</td>
</tr>
<tr>
<td><strong>diabase occurrence</strong></td>
<td>Uncommon</td>
<td>Common (8, 10, 11, 12, 14, 15)</td>
</tr>
<tr>
<td><strong>moho transition zone</strong></td>
<td>Interlayering of dunite, and gabbro, websterite lenses, variable thickness &lt;0.5km</td>
<td>Locally thick (up to 2 km) dunite, interlayered with wehrlite, websterite (4, 5, 6, 10, 13, 15)</td>
</tr>
<tr>
<td><strong>chromite pods</strong></td>
<td>Uncommon (1, 2), common (3)</td>
<td>Abundant</td>
</tr>
<tr>
<td><strong>serpentinization</strong></td>
<td>Lizardite</td>
<td>Lizardite and antigorite</td>
</tr>
</tbody>
</table>

**Note:** Revised classification of ophiolites (Boudier and Nicolas, 1985) based on nature of their mantle section. Data from: 1—Nicolas et al. (2000); 2—Davies (1977); 3—Cassard (1980); 4—Suhr (1992); 5—Ishiwatari (1985); 6—Hoek and Foullier (1995); Nicolas et al. (1999); 7—George (1978); Thy and Moores (1988); 8—Juteau (1974); 9—Dilek et al. (1999); Dilek and Thy (1998); 10—Cakir (1978); Dilek et al. (1999); 11—Girardeau and Mercier (1988); 12—Mahmood (1994); 13—Dick (1977); Harper (1982); 14—Rassios et al. (1983); 15—Evans and Hawkins (1989); Abrajano et al. (1989); 16—Boudier (1978); Bodinier (1988); 17—Abbate et al. (1994); Rampon and Piccardo (2000); 18—Quick (1981); Le Sueur et al. (1984); Cannat and Boudier (1985); Boudier et al. (1989); 19—Menzies and Allen (1974); Ferrière (1985); Dijkstra et al. (2001).
equally well by thermal or compositional differences in the rising asthenosphere. For example, because it is under the influence of the Iceland hot spot, the Reykjanes Ridge in the northern Atlantic Ocean has a fast-spreading ridge structure for a slow spreading rate; the hydrous melting of the mantle slab located above a subduction zone illustrates the role of composition, readily generating depleted mantle rocks whatever the spreading rate is. The discussion developed above on SSZ or MOR origin of ophiolites shows that it is not easy to take into account the compositional factor. The situation is more difficult with the thermal factor. We are not aware of any study of ophiolites that discusses the relation of ophiolites with hot spot or cold spot environments. Another difficulty relating ophiolite facies to spreading rate, also emphasized by Karson (1998), appears with the heterogeneous structure of slow-spreading ridges, where there are major differences between the segment centers and the vicinity of fracture zones. As documented by Karson and Rona (1990) and Tucholke et al. (1998), tectonically denudated mantle “core complexes” may occur in fractures zones that contrast with “normal” crustal structure in the middle of ridge segments. This difficulty is overcome in the Mirdita ophiolite in Albania (see below), which is large enough to encompass both a “core complex” and a “normal” lithosphere; but in a smaller ophiolite displaying either of these components, a wrong diagnosis of LOT and HOT is possible. This being acknowledged, differences between oceanic lithosphere accreted at slow- and fast-spreading ridges are so great (Karson, 1998) that Boudier and Nicolas’ (1985) early classification of ophiolites in terms of their inferred spreading rates, referring mainly to their mantle signature, was largely valid. In Table 1, we compare the revisited HOT and LOT classification, and in Table 2, we summarize the new and rapidly expanding database derived from the various oceanic floors.

With more data and experience, however, we need to reappraise the meaning of LOT in terms of spreading rate. The abyssal peridotites, which are issued from slow-spreading ridges, are normally composed of harzburgites, and strictly speaking the corresponding ophiolites should be considered as HOT. A closer examination shows, however, that the abyssal peridotites are less depleted than the harzburgites that correspond to true HOT. They commonly contain a small percentage (<5%) of clinopyroxene, either as unmelted relics from the mantle of origin or as secondary melt impregnations, commonly in association with plagioclase (Cannat, 1996). The true LOTs are characterized by a mantle section of plagioclase lherzolites, still partly fertile and having preserved mantle microstructures, or issued from a secondary melt impregnation of a depleted harzburgite (Nicolas, 1989, p. 21). We are now tempted to restrict the origin of the former fertile lherzolites to very slow-spreading settings. Oceanic rifts seem to be good candidates, as illustrated by the Zabargad Island formations in the Red Sea (Nicolas et al., 1985). Another illustration could be the newly described Gakell-Lena ridge in the Arctic Ocean where similar lherzolites have been dredged extensively (Dick et al., 2002; Snow et al., 2002).

The refertilized lherzolites might be issued from fracture zones in slow-spreading environments, as illustrated by the

massifs du Nord (Tiebaghi) in New Caledonia (Nicolas, 1989, p. 128) or by the Othrys ophiolite in Greece (Dijkstra et al., 2001). We propose to introduce a new category, the lherzolite-harzburgite ophiolite type (LHOT), whose mantle section is made of abyssal peridotites corresponding to slow-spreading situations. The 1985 classification of Boudier and Nicolas also needs to be revisited because a number of detailed studies in new ophiolite massifs (see Table 1) have since confirmed and enriched the criteria then proposed.

The gabbro unit, when fully developed as in the HOT Oman ophiolite, is composed of, from bottom to top, layered gabbros, foliated gabbros, and irregular patches of isotropic gabbros just below the sheeted dike complex. Layered and foliated gabbros display a magmatic fabric, with only traces of penetrative plastic deformation. In LHOT ophiolites, these features are attenuated and disappear, whereas penetrative magmatic deformation features tend to be replaced by partitioned deformation structures (flaser gabbro bands). The gabbro section is plastically deformed in its lower horizons in BOI (LHOT or HOT) and devoid of plastic strain in Oman (HOT). Typically, in the LHOT Mirdita ophiolite, layered gabbros are thin and poorly layered. These gabbros are present in the eastern part of the belt, where the crust is “normal.” They are represented only by highly strained lenses tectonically mixed with lherzolites in the western part, which has been interpreted as a “core complex” formed in the vicinity of a transform zone. In the Xigaze ophiolite (LHOT), the crust is composed mainly of diabase sills and basalts, and the gabbros are either absent or restricted to lenses that are a few hundred meters wide.

Dunites originated from the MTZ, and wehrlites intrusive into the crust also differ. In Oman, the MTZ varies greatly in thickness but does not exceed 400 m, whereas in the LHOT Bay of Islands ophiolite, it can attain a thickness of 2000 m; in contrast, wehrlites are much more abundant in Oman. Relating a large development of dunites and wehrlites to a high degree of melting and, thus, to a fast spreading rate, Jousselin and Nicolas (2000b) have interpreted the greater thickness of dunites in BOI compared to Oman to be a consequence of the lithosphere-asthenosphere limit being steeper in slower spreading conditions, which would result in the mantle flow and the associated, locally more constrained melt delivery (Fig. 6).

Chromite deposits may be excellent indicators. They are rare in Oman, uncommon in BOI, and remarkably abundant in the typically LHOT eastern part of Mirdita. Nicolas and Al Azri (1991) interpret this phenomenon through a combination of two causes: (1) the necessary melting-out of clinopyroxene in order to render chromium available, a condition which is achieved in both LHOT and in HOT; and (2) thermal or redox conditions that favor the crystallization of chromite in the uppermost mantle below Moho, where the chromite can also be trapped (Lago et al., 1982). These conditions would be met in LHOT uppermost mantle due to cooling related to slow-spreading and/or to seawater penetration at such depth near the ridge; in the HOT-fast-spreading setting, the conditions for chromite crystallization are normally met only above, within the magma chamber, where the
### TABLE 2. OCEANIC LITHOSPHERE

<table>
<thead>
<tr>
<th></th>
<th>Fast-spreading</th>
<th>Slow-spreading</th>
<th>Very slow-spreading oceanic rifting</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>OCEANIC RIDGES</strong></td>
<td>East Pacific Rise: Garrett FZ (1),</td>
<td>Mid-Atlantic Ridge: Fracture zones(5)</td>
<td>SW Indian Ridge (6); Gakkel-Lena Ridge (7); Zabargad (8); Galicia bank (9)</td>
</tr>
<tr>
<td></td>
<td>Hess Deep (2), Costa Rica Rift (3)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Easter microplate (4)</td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Spreading rate</strong></td>
<td>10 cm/yr</td>
<td>3 cm/yr</td>
<td>&lt;1 cm/yr</td>
</tr>
<tr>
<td><strong>CRUSTAL SECTION</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Sheeted dikes</strong></td>
<td>~1.5 km thick (3), steep to</td>
<td>Uncommon, thickness: 0–1 km</td>
<td>Diabase sills and dikes (8)</td>
</tr>
<tr>
<td></td>
<td>moderately dipping (2, 3)</td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Gabbros</strong></td>
<td></td>
<td>Thickness: 0–3 km</td>
<td>Thickness 0–1.5km (6)</td>
</tr>
<tr>
<td><strong>lithology</strong></td>
<td>Ol-gabbro (2, 4), ferrogabbro (4),</td>
<td>Ol-gabbro, gabbronorite</td>
<td>Ol-gabbro, gabbronorite, ferrogabbro (6)</td>
</tr>
<tr>
<td></td>
<td>gabbro-norite (2), minor wehrlite (2)</td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>deformation</strong></td>
<td>Magmatic (2, 3), plastic deformation</td>
<td>No deformation or high-</td>
<td>Magmatic and plastic foliation,</td>
</tr>
<tr>
<td></td>
<td>in shear zones (2)</td>
<td>temperature plastic (flasergabbro)</td>
<td>subhorizontal or steep (6)</td>
</tr>
<tr>
<td><strong>shear bands and faults</strong></td>
<td>Cataclastic listric faults</td>
<td></td>
<td>Mylonitic bands, normal faults,</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>granulite facies (6)</td>
</tr>
<tr>
<td><strong>MANTLE SECTION</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>lithology</strong></td>
<td>Harzburgite (1, 2, 4),</td>
<td>Cpx-harzburgite (2–3% Cpx),</td>
<td>Plag-Sp-lherzolite (4–7%Cpx)</td>
</tr>
<tr>
<td></td>
<td>Plag-dunite (1, 2, 4)</td>
<td>dunite, Sp-, Plag-lherzolite</td>
<td>(6,8,9), occurrence plag-rimed Sp</td>
</tr>
<tr>
<td><strong>internal structures</strong></td>
<td>High-temperature plastic foliation (1, 2, 4), gently to steeply dipping (2)</td>
<td>Moderate-temperature plastic foliation gently dipping</td>
<td>Moderate-temperature plastic foliation, high strain</td>
</tr>
<tr>
<td><strong>shear zones and faults</strong></td>
<td>Few cataclastic shear bands (2)</td>
<td>Low-angle detachment faults</td>
<td>High- to low-angle mylonitic shear bands, normal faults(6)</td>
</tr>
<tr>
<td><strong>serpentinitization</strong></td>
<td>Lizardite</td>
<td>Lizardite and antigorite</td>
<td>Antigorite</td>
</tr>
</tbody>
</table>

*Note: Criteria of Table 1 applied to oceanic ridges with different spreading rates. Data sources: 1—Hebert et al. (1983); Cannat et al. (1990); Niu and Hekinian (1997); 2—Hékinian et al. (1993); Mevel et al. (1996); 3—Alt et al. (1993); 4—Constantin et al. (1995); 5—Dick et al. (1984); Michael and Bonatti (1985); Cannat and Casey (1995); Karson (1998); 6—Dick et al. (1984); Meyer et al. (1989); Cannat (1991); Dick et al. (1999); 7—Dick et al. (2002); Snow et al. (2002); 8—Bonatti et al. (1986); Nicolas et al. (1987); 9—Evans and Girardeau (1987); Girardeau and Mercier (1988); Cornen et al. (1996).*
chromite would be dispersed within crystallizing gabbros.

**Metamorphic Aureole, Signature of Ophiolites Origin**

Moores (1982) and Coleman (1984) proposed to divide the ophiolites into two families depending on whether they had been incorporated into the continental crust at a passive margin (their Tethyan ophiolites) or at an active margin (their Cordilleran ophiolites) (Fig. 7). In the former case, ophiolites rest as giant thrust sheets upon a continental substrate; in the latter case, they are incorporated, often dismembered as a mélange, into accretion terranes mainly issued from trench sediments and seamount fragments (Coleman, 2000). From an earlier review (Nicolas, 1989, p. 289–309), we retain that more than 60 ophiolite massifs that comprise most, if not all, of the least dismembered ophiolites belong to the Tethyan group, with which they share a fascinating feature, the presence of a metamorphic aureole at their base. This aureole grades upsection over a few hundred meters from the underlying undeformed basalts and sediments to intensely strained amphibolites and granulites that are equilibrated above 800 °C. These amphibolites and granulites are overlain by concordant mylonitic peridotites deformed at 900–1000 °C that are grading over 1–2 km into normal mantle peridotites. Tectonic reconstructions (Boudier et al., 1988) and models (Hacker et al., 1996) show that this metamorphism was produced by the ironing effect of a ~10-km-thick lithospheric slab that represents the future ophiolite. Following thermal models of oceanic ridges (Phipps Morgan and Chen, 1993; Henstock et al., 1993; Cormier et al., 1995), we conclude that the age of detachment of the lithospheric slab cannot be older than 2 Ma. Accordingly, in Oman, the age of the metamorphic aureole is 1–2 m.y. younger than the ridge of origin. Young ages in metamorphic aureoles are the rule, with the few exceptions of 10–15 m.y. younger ages (review in Wakabayashi and Dilek, 2001). Because both the geological and the thermal modeling information are compelling, we assume that the few datings with 10–15 Ma differences are dubious (Hacker et al., 1996).

Major consequences can be drawn from the most common existence of basal aureoles below ophiolites. These ophiolites were detached at no more than 10–100 km from their ridge of origin, depending on the spreading rate and the age difference between igneous accretion and metamorphism in the basal aureole. At their time of detachment, they were thin and hot pieces of oceanic lithosphere. This extraordinary feature raises as many questions as it helps to solve. Is it not possible that the SSZ secondary signature of many ophiolites would be due to water penetration into the overriding future ophiolite? (Boudier et al., 1988) Are there such active sites in present day oceans, or do they belong to peculiar periods in earth history? Paleomagnetic studies in Oman (Perrin et al., 1993) show 30° rotations within 1 Ma, suggesting that this SSZ secondary signature of many ophiolites would be due to water penetration into the overriding future ophiolite? (Boudier et al., 1988) Are there such active sites present in the Easter or Juan Fernandez microplates along the East Pacific Rise (Boudier et al., 1997). This model has a few appealing features: a microplate is a “pre-cut” piece of oceanic lithosphere, confined by ridge segments and compressive boundaries. With these active boundaries, it might be more easily detached from its site of origin than a normal piece of lithosphere. This future ophiolite also contains internally active ridge segments, a feature that accounts for the presence of frozen diapirs in some ophiolites. In addition, tip of propagating segments, such as Pito or Endeavour Deeps from Easter and Juan Fernandez, respectively, create locally very slow-spreading conditions in the overall fast-spreading environment; such contrasted situation might have its equivalent in the different characteristics pointed out in the Oman ophiolite between active segments and their propagating tip (Nicolas and Boudier, 1995). In an ophiolite as small as Troodos in Cyprus, the slow-spreading situation documented on the base of evidence for a rifted valley (Varga and Moores, 1985) could be reconsidered if the ophiolite were originated from the tip of a propagator. There are, however, many pending questions. Can the compressive zones in microplates generate the large thrusts responsible for the metamorphic aureoles of ophiolites? Why and how is the microplate-ophiolite detached and overthrust?

**CONCLUSION: CALL FOR AN INTEGRATED APPROACH IN OPHIOLITE STUDIES**

As studies of ophiolites turn toward a better understanding of the functioning of oceanic spreading centers, the first problem
is to identify the type of center where the considered ophiolite has formed. We have recalled a number of criteria that can be used to estimate the spreading rate of this center (Tables 1 and 2). Spreading rate is clearly the most important parameter to account for the diversity of the oceanic lithosphere (Karson, 1998), and it is also clearly reflected in the structure and composition of ophiolites. In the study of slow-spreading oceanic ridges, the ophiolite reference is of lesser importance because, thanks to the submarine reliefs that can be >4 km, direct observation of deep crust and mantle is possible. This reference becomes compulsory in fast-spreading situations where the subdued reliefs allow only exceptionally deep structures to be exposed. We have concluded above that many ophiolites were likely originated from spreading centers corresponding to intermediate to fast rates and, by a few examples showing the interplay between structural studies in ophiolites and geophysical studies in oceanic ridges, we have tried to illustrate the potential of this cross-fertilizing approach.

Spreading rate is not the unique parameter, however, and a closer analysis should consider the effects of different temperatures or compositions in the mantle sources. The environment of origin (MOR or SSZ) that is discussed above and that has much—too much—mobilized the ophiolite community, belongs to this category of problems. It has been approached from the standpoint of geochemistry, but with disputable results and more elementary pieces of information, e.g., the associated sedimentary record often having been overlooked.

This drives us to our main conclusion, which is a call for integrated studies. Through a few examples, we have illustrated the fruitful cooperation between structural geologists in ophiolites and marine geophysicists. A similarly rich cooperation is possible between geochemists and petrologists working in ophiolites and marine environment. Dealing with the relative roles of structural and geochemical studies in ophiolites, we insist on the common sense idea that these approaches are complementary, but that geochemistry should be applied within the framework of a thorough understanding of the structures, or at the very least, of the general geology. Indeed, many ophiolites have been dismembered extensively by collision-related tectonic deformation, making it difficult to conduct ridge-related structural studies. One might question the results of geochemical studies that are not supported by structural information in such ophiolites. In contrast, in an ophiolite whose internal structure has been preserved and decrypted, the solution to further problems depends largely on focused geochemical studies. Typically, the answer to the question evoked above regarding a mantle source outside the thermally or chemically average mantle source can be addressed by geochemistry.

As a concluding remark, we would like to encourage young scientists to undertake more structural studies in selected ophiolites, in association with and opening the way for geochemists. Evidently, they should keep an eye on corresponding marine research, but when their results are at odds with marine studies, as has happened in the past, they should remember that careful observations made on relevant outcrops in the field and the complementary laboratory studies provide the most compelling evidence.

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