Thick sections of layered ultramafic cumulates in the Oman ophiolite revealed by an airborne hyperspectral survey: Petrogenesis and relationship to mantle diapirism

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A R T I C L E   I N F O

Article history:
Received 5 February 2009
Accepted 9 September 2009
Available online 24 September 2009

Keywords:
Oman ophiolite
Clinopyroxenite
Wehrlite
Ocean crust
Spectroscopy
Remote sensing

A B S T R A C T

Using the HyMap instrument, we have acquired visible and near infrared hyperspectral data over the Maqsad area of the Oman ophiolite (~15×60 km). This survey allowed us to identify and map the distribution of clinopyroxene-rich cumulates (inter-layered clinopyroxenites and wehrlites) whose occurrence was previously undocumented in this area. The cumulates reach several hundred meters in thickness and crop out at distances exceeding 15 km on both sides of the Maqsad former spreading centre. They occur either in mantle harzburgites, as km-sized layered intrusions surrounded by fields of pegmatitic dykes consisting of orthopyroxene-rich pyroxenite and gabbronorites, or at the base of the crustal section where they are conformably overlain by cumulate gabbros. These ultramafic cumulates crystallized from silica- and Mg-rich melts derived from a refractory mantle source (e.g. high Cr#, low [Al2O3], low [TiO2]). These melts are close to high-Ca boninites, although, strictly speaking, not perfect equivalents of present-day, supra-subduction zone, boninites. Chemical stratigraphy reveals cycles of replenishment, mixing and fractional crystallization from primitive (high Mg#) melts, typical of open magma chambers and migration of inter-cumulus melts. The TiO2 content of clinopyroxene is always low (<0.2 wt.%) but quite variable compared to the associated pegmatites that are all derived from a source ultra-depleted in high-Ca boninite. As the diapir continued to rise, the amount of MORB reaching shallow depths increased, together with the surrounding temperature, leading to the formation of magma chambers where the crystallization of layered cumulates became possible. These cumulates remained rich in pyroxene and devoid of plagioclase as long as the contribution of MORB-derived melts was moderate relative to the lithospheric-derived melts. As the contribution of MORB to the refilling of the magma chamber increased, gabbroic cumulates started to crystallize.

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1. Introduction

The Oman ophiolite is one of the best natural laboratories for the detailed study of magma chamber processes related to oceanic spreading (e.g. Pallister and Hopson, 1981; Lippard et al., 1986; Juteau et al., 1988; Nicolas et al., 1988; Browning, 1990; Reuber et al., 1991; Kelemen et al., 1997; MacLeod and Youançq, 2000; Adachi and Miyashita, 2003). Schematic models, widely reviewed in textbooks, consider the deep crustal section of Oman as essentially made of layered gabbroic cumulates crystallized from olivine tholeiite similar to mid-ocean ridge basalts (MORB). Occurrences of ultramafic lithologies associated with these gabbros have been recognized throughout the ophiolite (Pallister and Hopson, 1981; Ernewein and Whitechurch, 1986; Hutin et al., 1986; Rabu et al., 1986; Ernewein et al., 1988; Juteau et al., 1988; Benn et al., 1988; Reuber, 1988; Amri, 1995; Koga et al., 2001). They are collectively referred to as “wehrlites” (strictly speaking, assemblages of olivine and clinopyroxene) by field geologists, although they may contain high amounts of plagioclase, pargasitic amphibole and, less frequently, orthopyroxene. They show large variations in their modal composition, especially in the olivine clinopyroxene ratio. The origin of these ultramafic lithologies remains a topic of discussion. Their occurrence in an ophiolitic massif where most other lithologies are...
In the present study, we reconsider the origin of Oman crustal ultrama.

A popular view is that the ultramafic crustal rocks of Oman occur as “late intrusions” cross-cutting previously crystallized gabbros, a situation actually documented in many outcrops. Wehrlites may also be inter-layered with gabbros and in some places, it can be demonstrated that they are rooted in a nearby intrusion (Reuber, 1988). Where no intrusions are observed, tectonic transposition is commonly invoked to explain the interlayering (e.g. Nicolas et al., 1988).

Different models have been proposed for the genesis of Oman wehrlites. Koga et al. (2001) have deduced from geochemical data that ultramafic cumulates and layered gabbros share a common, MORB-like, parental melt but that the crystallization order of clinopyroxene and plagioclase was inverted due to the introduction of water into the system, in the case of wehrlites. Some authors, following Benn et al. (1988), consider that wehrlitic rocks were actually mixtures between residual mantle olivine and interstitial melts squeezed out from the mantle/crust transition zone in an off-axis setting due to a ‘rolling mill effect’ (Rabinowicz et al., 1984, 1987) and intruded in the overlying layered gabbros in a mushy state. This mechanism was also invoked to account for the formation of olivine-rich “cumulate looking” rocks drilled along present-day mid-ocean ridges (Suhr et al., 2008).

In contrast to these views, wehrlitic intrusions have been related to boninitic magmas derived from hydrated melting of a depleted mantle (Ernewein and Whitechurch, 1986; Lippard et al., 1986; Ernewein et al., 1988). The recent discovery of boninites among the Oman volcanics (Ishikawa et al., 2002) indicates that this hypothesis is plausible, even if the link between the upper crustal boninites and the deep crustal wehrlites is not demonstrated. Early compressive tectonics (intra-oceanic thrusting) have been invoked to account for water introduction and for the intrusion of Oman wehrlites (Ernewein et al., 1988) while other authors argue that a supra-subduction zone setting is a simpler way to explain their field and geochemical characteristics (Lippard et al., 1986).

There is thus clearly a wide diversity of models for the genesis of Oman crustal ultramafic rocks collectively referred to as “wehrlites”. In the present study, we reconsider the origin of Oman crustal ultramafic rocks in light of a field and petrological study of new occurrences discovered during an airborne hyperspectral mapping project over the Sumail massif, in the southeastern part of the ophiolite (Fig. 1). The evolution of this massif is well constrained by detailed structural and petrological studies (Ceuleneer et al., 1988; Amri, 1995; Ceuleneer et al., 1996; Benoit et al., 1996; Jousselin et al., 1998; Benoit et al., 1999; Python and Ceuleneer, 2003). In particular, a former mantle diapir has been mapped in the central part of the survey area, which is crucial to understand the sequence of events that led to the formation of ultramafic cumulates in this area.

2. Geological setting

2.1. The Oman ophiolite

The Oman ophiolite is a fragment of Cretaceous oceanic lithosphere accreted in the Tethys ocean about 100 Ma ago (Tilton et al., 1981; Tippit et al., 1981). The early detachment event (intra-oceanic thrusting) occurred soon after the accretion (35–58 Ma) (Boudier et al., 1985, 1988; Montigny et al., 1988). Its final obduction onto the Arabian margin was completed during the Maastrichtian (circa 70 Ma) (Glennie et al., 1973; Coleman, 1981; Searle and Cox, 1999).

The uplift of the Oman Mountains started during the Miocene (Glennie et al., 1973) and resulted in the present-day exposure of the ophiolite.

The precise context of genesis and emplacement of the Oman ophiolite is still debated: hypotheses range from a mid-ocean ridge setting (e.g. Coleman, 1981; Nicolas, 1989) to a subduction zone setting (e.g. Pearce et al., 1981; Ishikawa et al., 2002). This ambiguity is due to the presence of many lithologies within the ophiolite having an “arc-like” geochemistry, whereas there is a lack of geological evidence for the development of a volcanic arc in the southern Tethyan domain during the Mesozoic era. In order to explain this, it has been proposed that some petrogenetic processes at ocean ridges may mimic those classically attributed to the subduction factory (Amri et al., 1996; Benoit et al., 1999; Python and Ceuleneer, 2003; Godard et al., 2006) or that the Oman ophiolite recorded a polygenetic history (Umino et al., 1990; Arai et al., 2006; Python et al., 2008).

2.2. The Sumail massif

The Sumail massif is located in the southeastern part of the Oman ophiolite (Fig. 1), where a former axial mantle diapir (the Maqsad diapir) can be mapped through the study of high-temperature plastic deformation structures in mantle harzburgites (Ceuleneer et al., 1988; Ceuleneer, 1991) (Fig. 2). Mantle flow planes and lines perpendicular to the mantle–crust transition (whose regional attitude is assumed to mark the paleo-horizontal plane) define a paleo-upwelling zone about 8 km in cross-section. The massif is surrounded by zones of horizontal flow diverging from the upwelling to the NNE on the northern side of the upwelling and to the SSW on its southern side. These directions are perpendicular to the azimuth of the sheeted dyke complex, i.e. to the strike of the paleo-spreading ridge. Diverging mantle flow is observed for distances of about 15 km away from the upwelling. To the North, diverging flow reaches mantle peridotites bearing E–W trending lineations, i.e. accreted from a N–S ridge. To the
Fig. 2. Global RGB composition over the Sumail massif (central part of the picture). Red, Green and Blue correspond respectively to 0.98 µm, 0.87 µm and to 1.03 µm channels. All the values are normalized to the 0.74 µm channel which essentially shows albedo variations. This colored composite enhances the clinopyroxene absorption; clinopyroxene-rich rocks appear in green. On the right of the picture, the two highlighted areas correspond respectively to north cumulates (at the top) and south cumulates (at the bottom). Our detailed cross-section is localized in this second area. The geological sketch (left side of the picture), based on I. Amri (1995) field work, allows a direct comparison between the main geological lithologies and hyperspectral variations.
south, diverging flow abuts a mylonitic shear zone (the Muqbariah shear zone). A clear zoning in the nature of the melt migration structures mirrors this structural map; in the upwelling itself, former porous flow channels made of troctolites are observed, while in the zone of diverging flow, the dominant melt migration structures are olivine gabbro dykes. Further away from the upwelling, dykes are made of olivine gabbro-norrites. This zoning reflects the thermal structure of the area at the time of cooling (Ceuleneer et al., 1996): troctolites crystallize on the high temperature cotectic of MORB (>1190 °C), while pyroxene bearing lithologies are lower temperature crystallization products from MORB.

On a regional scale, the Maqsad diapir constitutes the centre of a much larger structure: a former spreading segment, about 80 km in length parallel to the paleo-spreading axis (N130°E) (Python and Ceuleneer, 2003). This segment is characterized by the abundance, in the mantle peridotites, of former melt channels and dykes made of cumulates of MORB petrological and geochemical affinity (Benoit et al., 1996). This contrasts markedly with the more common situation observed in Oman where mantle dykes essentially consist of crystallization products from magma depleted in Al and the high field strength elements (HFSE) but richer in both MgO and SiO₂ than MORB, being therefore closer to boninites or magnesian anodeses than to MORB (Benoit et al., 1999; Python and Ceuleneer, 2003). At the southern and northern boundaries between the spreading segment centred on the Maqsad diapir and the surrounding peridotites, dykes and intrusions made of such depleted pyroxenitic and gabbronoritic pegmatites are particularly abundant (Benoit et al., 1999). Deformation in the Muqbariah mylonitic shear zone is contemporaneous with the injection of these pegmatites.

3. Contribution of hyperspectral remote sensing to lithological discrimination and mapping of the Sumail massif

3.1. Previous maps

Several geological maps of the Sumail massif have been published during the last thirty years. The map by Glennie et al. (1974) covers all of the Oman Mountains at a 1:500000 scale. Following this pioneering work, the BRGM (French geological survey) has produced more detailed maps, at a 1:100000 scale, of the central part of the ophiolite, including the Sumail massif (Villey et al., 1986; Rabu et al., 1986). These authors accurately mapped the mantle/crust transition zone (the “Moho”), which is peculiarly contoured in the Sumail massif. More recently, we produced a map at a 1:40000 scale (Amri, 1995) that represented a significant improvement in the definition and mapping of the different, frequently complex, lithological units making up the crustal section of the Sumail massif. This work was essentially field-based but used the Landsat and JERS-1 SWIR multispectral data processed by Denniss et al. (1994). The poor spectral resolution of JERS-1 and Landsat did not allow us at the time to spectrally discriminate all the lithologies exposed in the field. In the present study, we show how an airborne hyperspectral survey (combining both high spectral and spatial resolutions (Pinet et al., 2006a,b)) improves the detection of the various lithologies making up the Oman ophiolite in the Sumail massif. In addition, this study includes the southernmost part of the Sumail massif which was not covered by our earlier work (Fig. 2).

3.2. Method

3.2.1. Principle of spectroscopy

Field geology has limitations, particularly when the covered area is quite large. Geologists need to extrapolate detailed observations conducted in different outcrops and this can potentially lead to errors or omissions. However, the interaction between the sunlight and Earth surface materials means that spectroscopy in visible and infrared wavelengths can be a powerful tool (Williams, 1995), particularly when used in support of field work. It is particularly useful for the Oman ophiolite because rock outcrops are almost devoid of vegetation due to the semi-arid climatic conditions.

In the case of mafic and ultramafic minerals, spectrometry permits the detection of mineralogical variations due to electronic transition processes (Hunt, 1977; Burns, 1993). For the detection of olivine and pyroxenes, the absorptions due to cation Fe²⁺ in visible and near-infrared wavelengths are used (Hunt and Salisbury, 1970; Hunt et al., 1973; Adams, 1974; Adams, 1975). Both pyroxenes have two distinct absorption bands: the first one in the 1 μm domain (around 0.9 μm for the low-Ca pyroxene [orthopyroxene] and 1.0 μm for the high-Ca pyroxene [clinopyroxene]) and the second one in the 2 μm domain (around 1.9 μm for orthopyroxene and 2.2 μm for clinopyroxene). In the case of olivine, there is only one feature in the 1 μm domain but this feature is composed of three characteristic absorption bands (at 0.85, 1.05 and 1.25 μm). As these absorptions are specific to each mineral, it is possible to extract information from spectra of natural rocks, to detect the different minerals and to estimate their relative proportions.

Imaging spectrometers are able to detect and record reflectance spectra with both sufficient spatial and spectral resolutions to be used as tools for geological mapping (Goetz et al., 1985; Pinet et al., 2006a,b). Different studies using hyperspectral remote sensing have been carried out on the Oman ophiolite (Pinet et al., 2003, 2006a,b; Jean-Philippe Combe et al., 2006; Clenet et al., 2008; Roy et al., 2008) and on other peridotitic massifs (Chabrilat et al., 2000; Launaeu et al., 2002, 2004). These studies are generally based on the analysis of the spectral shape (i.e. band position, depth and width).

3.2.2. Microprobe analytical method

Major and minor element concentrations in individual mineral phases were determined with a CAMECA SX50 electron microprobe with SAMX automation at the Paul Sabatier University, Toulouse, France, using wavelength-dispersive spectrometry (WDS). Analyses were performed with an accelerating voltage of 15 kV, beam current of 20 nA and a spot size of 4 µm². The following synthetic and natural minerals standards were used: albite (Na), periclase (Mg), corundum (Al), sanidine (K), wollastonite (Si, Ca), pyrophane (Mn, Ti), hematite (Fe), chromium oxides (Cr), and nickel oxides (Ni). Nominal concentrations were corrected by the PAP data reduction method (Pouchou and Pichoir, 1985). Counting time was 10s on the peak and 5s on the background. The detection limits are 0.07% for Al₂O₃, 0.02% for TiO₂ and 0.09% for Cr₂O₃ and Na₂O, for all mineral phases analysed.

Our full data set is available as supplementary electronic files.

3.3. Data set

We used the airborne HyMap instrument to acquire hyperspectral data over the Sumail Massif. The detector has 128 channels between 0.45 and 2.48 μm. The total data set covers a ~15 × 60 km area and has a spatial resolution of 6 m/pixel. It covers the mantle section (essentially harzburgitic) and the crustal section (dunitic to gabbroic) of the massif. The HyMap data were atmospherically corrected by the Mass Spectrometry (ATREM, Gao et al., 1993, CIRES, 1999) and the Empirical Flat Field Transformation (EFFORT, Boardman, 1998).

A first order approach is to use spectral ratios to estimate relative variation in a spectrum. Wavelengths which can highlight the different minerals are chosen. For olivine and pyroxene mixtures, we used 0.98 μm, 0.87 μm and 1.03 μm channels to enhance clinopyroxene absorption features. To minimize the contribution of non-mineralogical effects (e.g. arising from photometric effects induced by the topography and/or sun direction), we normalized those three channels to that of 0.74 μm, which essentially shows variations in albedo (e.g. Chevrel and Pinet, 1992; Cord et al., 2005).
The three ratios are then represented respectively in Red, Green and Blue to obtain a false color composite to show the relative contribution of the clinopyroxene component. The resulting colors of the map in Fig. 2 are interpreted in the following way: where the contributions of olivine and orthopyroxene are weak, and that of clinopyroxene is dominant, the absorption feature is strong in the 0.98 µm and 1.03 µm channels resulting in high 0.87/0.74 µm ratio values, low 0.98/0.74 and 1.03/0.74 µm ratio values. Consequently, high green values correspond to lithologies with the highest clinopyroxene content.

A limitation of our airborne hyperspectral data is, however, the difficulty in mapping the distribution of plagioclase with 0.4–2.5 µm spectrometers. Despite the fact that plagioclase is abundant and fresh, it is spectrally featureless in the visible-near infrared (VNIR) and in the short wavelength infrared (SWIR) spectral range (Hunt and Salisbury, 1970). This issue has been addressed in a recent paper by Roy et al. (2008). A dedicated survey in the thermal infrared (8–12 µm) would be the most appropriate to help resolve this issue but is logistically beyond the scope of this study. Nevertheless, the plagioclase contributes to the overall albedo and this can be used as an indirect indication of its presence, keeping in mind however that albedo can be increased in a number of different ways. Accordingly, the hyperspectral survey was useful in mapping areas where olivine, orthopyroxene or clinopyroxene (but also alteration products like carbonates and serpentines) are the dominant minerals but did not allow us to discriminate between clinopyroxene-rich gabbros (clinopyroxene + plagioclase) and clinopyroxenites. The present work is thus based both on the airborne survey and on field checks of numerous outcrops.

Application of the colored composition techniques, such as described above, on the overall data set allowed us to distinguish areas showing a very strong signature of clinopyroxene. One of these areas is located in the crustal section in the southern part of the massif, outside the area mapped by Amri (1995) (Figs. 2 and 3). The other area is located in the northern part of the massif, in the mantle section (Figs. 2 and 4).

4. Description of the studied outcrops

Given the focus of the present study, we have highlighted two areas characterized by an enhancement of the spectral signature of clinopyroxene with a red–green–blue (RGB) ternary image (Fig. 2).

Fig. 3. Field view of the southern area. Picture is taken from the southern side of the outcrop. Clinopyroxenites and wehrlites crop out below layered gabbros (highlighted in blue [or white] on the picture) which are the most abundant in this massif. Samples have been taken on sections A–B, which has a stratigraphic thickness of about 200 m. One can see alternations of two contrasting rock types: grayish outcrops are principally composed of clinopyroxenites (highlighted in green [or light gray] on the picture) while reddish ones are mostly olivine-rich rocks (wehrlites; highlighted in red [or dark gray]).

Fig. 4. Field view of layered pyroxenites in the northern field area. Layer is typically tens of meters thick. The upper left picture shows that these rocks have a medium-grained (grain size ranging from a few mm to 1 cm) adcumulate texture.
Field checks allowed us to demonstrate that both areas are exceptionally rich in clinopyroxene and are composed of interlayered clinopyroxenitic and wehrlitic ultramafic cumulates (Figs. 3 and 4). The first area lies at the base of the crustal section in the southern part of the massif, the second area is found in the mantle section in the north. Both sites are located in an off-axis position relative to the Maqṣad paleo-ridge (Fig. 2).

The northern outcrop is the most spectacular in the RGB image due to the contrast between the spectral signature of the host mantle harzburgites and that of clinopyroxene-rich lithologies. The hyperspectral survey reveals alternating greenish and purplish bands a few hundred meters wide. In the field, this spectral variation corresponds to the alternation of outcrops with layers of olivine clinopyroxenites, clinopyroxene-rich wehrlites, olivine websterites in the greenish bands and with clinopyroxene-poor cumulates (dunites and clinopyroxene-poor wehrlites) in the purplish bands. All these lithologies contain minor amounts of Cr-spinel. Close examination of the RGB composition reveals a thinner layering composed of bands a few tens of meters in thickness. Our field survey has shown that even smaller-scale (typically meters) alternations of clinopyroxene-rich and clinopyroxene-poor layers also exist but are not resolved by our airborne data. It appears that the relationships between the lithologic succession and the topographic surface are quite complicated in this northern outcrop. Accordingly the construction of an accurate stratigraphic log of this area is not possible.

Thin section examination reveals rocks with adcumulate textures, devoid of any deformation features, contrasting with the host harzburgites that have deformation textures inherited from solid-state mantle flow (Ceuleneer et al., 1988). This outcrop of layered ultramafic cumulates has a lateral extent of about 4 km in a direction parallel to the strike of the layering. It grades progressively into mantle harzburgites that are intruded by abundant pegmatite dykes which are up to several meters thick and are composed of pyroxenites and pyroxene rich gabbronorites equivalent to the ones studied by Benoit et al. (1999). These pegmatites were the only features identified during our previous mapping work (dark purple-band on the map of Amri (1995), Fig. 2). However, their abundance relative to the host harzburgites is too low (<10% of the exposed surfaces), to be identified in our spectral map due, in part, to the pixel size.

The southern outcrop is a 150 m wide band extending for more than 1 km in a SW direction. This clinopyroxene-rich band is part of the base of the crustal section (Fig. 3). Higher up, this crustal section is essentially composed of layered gabbros extending to the summits of the massif. The Moho is faulted in this area, resulting in local tilting of the stratigraphic sequence (down to the Moho) cannot be measured due to complex tectonics and so a 200 m thickness is the minimum value. The transition between clinopyroxene-rich and clinopyroxene-poor cumulates is not always clear and does not necessarily correspond to a planar structure. This is illustrated in Fig. 3 where, in the central part of the section, clinopyroxenites and wehrlites seem to be mutually intrusive. In some places, especially in the lower part of the section, the modal variations appear rather chaotic and the layering can be difficult to follow.

Our petrographic study has shown that olivine–clinopyroxenites can contain up to 14% olivine, but that the average value for the whole section is close to 6.5%. These rocks have medium-grained adcumulate textures (millimeter to centimeter grains). Olivine and minor spinels are subhedral while clinopyroxene is generally anhedral. Clinopyroxenites are particularly fresh, contrasting with the more altered olivine. The other dominant rock type is wehrlite with an average clinopyroxene content of 17% but with more variable modal proportions than in clinopyroxenites. The clinopyroxene content of wehrlites can be very low (actually a continuum exists with dunites) and reach a maximum of 43%. The rocks have a medium-grained (up to 10 mm) adcumulate texture with subhedral olivine crystals surrounded by anhedral poikilitic clinopyroxene. In a few samples, clinopyroxene can be subhedral and reach more than 60 mm. Small euhedral grains of Cr-spinel are relatively abundant (2.5%). Orthopyroxene is quite uncommon in this cumulate sequence: a single interstitial orthopyroxene crystal has been observed in a clinopyroxenite (sample Pyr 14A). Gabbros from the top of the section are composed of very fresh plagioclase and clinopyroxene in approximately equal proportions. The two gabbros we have sampled are devoid of olivine or orthopyroxene. They have a medium grained adcumulate texture, without any deformation overprinting.

5. Phase chemistry and chemical stratigraphy

The vertical evolution of the composition of clinopyroxene, olivine and Cr-spinel, and its relationship with the lithostratigraphy is shown in Figs. 6, 7 and 8. The Mg# of clinopyroxenites and olivines are always high and vary within a narrow interval, ranging from 0.93 to 0.89 and from 0.90 to 0.87 respectively (Fig. 6). Although the lowest Mg# values tend to be associated with the clinopyroxenites, the variation ranges of clinopyroxenites and wehrlites largely overlap. In other words, there is no dichotomy in phase chemistry mirroring the bimodal distribution in modal composition (dichotomy between clinopyroxene-rich and clinopyroxene-poor cumulates). On the contrary, variations in mineral chemistry define smooth vertical undulating patterns that are essentially not affected by the rather abrupt lithological transitions. It is worth noting (1) that the “half-wavelength” of the chemical pattern is of the same order of magnitude as the thickness of the mega-layering, i.e. ranging from several meters to a few tens of meters but (2) that the minimum and maximum values of the chemical parameters do not correspond to the major lithological transitions.

The same undulating pattern can be observed for the elements plotted in Figs. 7 and 8. The concentration of the compatible elements is correlated with the Mg#, the one of the incompatible elements is anti-correlated with the Mg#. The concentration of TiO2 in spinel is quite scattered.

There is no general evolution of the phase chemistry along the 200 m sequence of ultramafic cumulates superimposed on the undulating variation pattern (i.e. no overall increase or decrease in the average values of any chemical parameters down- or up-section). A major shift in all these parameters is, however, observed in the
gabbros which contain clinopyroxenes with low Mg# (0.80 to 0.83), a Cr$_2$O$_3$ content below detection limit and with significantly higher TiO$_2$ contents ranging from 0.15 to 0.40 wt.% (Figs. 10 and 11).

Clinopyroxenes are poor in aluminum, with Al$_2$O$_3$ contents ranging from 0.5 to 2.5 wt.%. In an Al$_2$O$_3$ vs Mg# diagram, clinopyroxenes from our ultramafic cumulates define a field almost identical to that of Trinity ophiolite crustal cumulates. This field is included in the much wider variation domain of Oman pyroxenitic mantle dykes (Fig. 9). The Cr$_2$O$_3$ content of clinopyroxenes is consistently high, reaching 1.2 wt.% (Fig. 10). Clinopyroxenes in wehrlites and in dunitic lenses tend to be richer in Cr than clinopyroxenes from clinopyroxenites, but the variation fields of these lithologies largely overlap. In terms of Cr$_2$O$_3$ content, the clinopyroxenites from our ultramafic cumulates is identical to clinopyroxenes from primitive pyroxenitic mantle dykes of Oman (Fig. 10).

The TiO$_2$ content of Cr-spinel (Fig. 7) and of clinopyroxene (Fig. 11) is also quite low, but rather variable, ranging from 0.1 to 0.4 in Cr-spinel and from below detection limit to 0.2 wt.% in clinopyroxene. Their Cr# ranges from 58 to 74. Cr-spinel from wehrlites tends to have a higher Cr# than Cr-spinel from pyroxenites.
Fig. 6. Stratigraphic variations of Mg# \(100 \times \frac{Mg}{Mg + Fe_{tot}}\) in clinopyroxene (left) and olivine (right) from the southern outcrop. The color scale on the left shows the end-member rock types (green [or light gray]: clinopyroxenite; red [or dark gray]: wehrlite; blue [or white]: gabbro). Circles, triangles and squares correspond to pyroxenites, wehrlites and dunite lens samples respectively.
Fig. 7. Stratigraphic variations of Cr$_2$O$_3$ (wt.%) in clinopyroxene (left) and Cr-spinel from the southern outcrop (symbol signification is the same as Fig. 6).
Fig. 8. Stratigraphic variations of TiO₂ (left) and of Al₂O₃ (right) in Cr-spinel from the southern outcrop (symbol signification is the same as Fig. 6).
The Cr# is broadly correlated with the Fo of olivine, although the scatter in this diagram (not shown) is quite high.

6. Discussion and conclusions

This study has shown that visible and near-infrared reflectance data are of great help in detecting lithologies that may be overlooked during the geological mapping of large areas (~1000 km²) due to their resemblance to other, more common, rock types. An airborne hyperspectral survey of the Maqsad area has allowed us to establish the presence of clinopyroxene-rich ultramafic cumulates — an occurrence which has not been previously documented, nor expected, in this essentially gabbroic environment. These ultramafic cumulates are distributed in off-axis positions on both sides of the Maqsad paleo-ridge axis.

6.1. Crystallization history

The Maqsad ultramafic cumulates have primitive compositions as attested by the high Mg# (i.e. Fo) of their olivines and by the high Mg# and Cr content of their clinopyroxenes. The 0.02–0.03 difference in the Mg# of clinopyroxene and olivine is typical of cumulate phases in equilibrium with the same melt (e.g. Grove et al., 1992). The most primitive among these melts were close to equilibrium with mantle
peridotite, while the most evolved ones have undergone only moderate degrees of fractional crystallization. The degree of fractional crystallization is, on average, slightly higher in the cumulates from the northern outcrop than in those from the southern outcrop (Fig. 12).

The construction of thick, continuous sequences of primitive cumulates implies frequent refilling of a magma chambers with melts extracted from the mantle. The alternation of mega-layers of clinopyroxene-poor and clinopyroxene-rich cumulates is most likely related to complex cycles of refilling, compaction–migration, and/or fractional crystallization (Fig. 13). The maximum Mg# values are not found at the base of the olivine-rich (wehrlitic) mega-layers nor are the minimum values located at the top of clinopyroxenite mega-layers, as might be expected from fractional crystallization models. The most evolved cumulates are preferentially found in the middle of

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**Fig. 11.** Variations in TiO$_2$ versus Mg# \([100 \times Mg/(Mg + Fe_{tot})]\) in clinopyroxene from the southern outcrop compared to various lithologies from the Trinity ophiolite (Ceuleneer and le Sueur, 2008).

**Fig. 12.** Variations of TiO$_2$ versus Mg# \([100 \times Mg/(Mg + Fe_{tot})]\) in clinopyroxene from the southern and northern outcrops. Values are compared to those of the pegmatitic dykes surrounding the northern outcrop (defining the “boninitic fractionation trend”) and to those of layered gabbroic cumulates cropping out in along the axis of the Maqṣād paolo–spreading structure (defining the “MORB fractionation trend”). The fields were constructed with data from Benoit et al. (1999) and with as yet unpublished data.
clinopyroxenite mega-layers. The most primitive cumulates do not systematically occur at uniform distances from the lithological transitions, and can occur at the base, at the top or in the middle of wehrlitic mega-layers. This may be due to progressive mixing between evolved and primitive melts: although the chemical parameters like Mg# and Cr# evolve gradually as the mixing progresses. The introduction of a primitive melt batch does not immediately affect the proportions of crystallizing phases; this change occurs later on, when the melt, evolving on a given cotectic line of the phase diagram, reaches a new cotectic line. In addition to variations in major element parameters Mg#/100 variations in olivine along the cross-section. Right column: interpretation of this trend (together with those presented in Figs. 6–8) in terms of successive stages of replenishment, mixing and fractional crystallization.

Fig. 13. Left column: Mg#/100 variations in olivine along the cross-section. Right column: interpretation of this trend (together with those presented in Figs. 6–8) in terms of successive stages of replenishment, mixing and fractional crystallization.
composition, the water activity in the magma is another parameter that may influence the position and nature of the coticetic lines.

The number of these refilling episodes can be deduced from the undulating patterns observed in the geochemical log (Fig. 6). In Fig. 13, we propose an interpretation of this "stratigraphic" evolution in terms of cycles of replenishment and fractional crystallization. It appears that four cycles contributed to the formation of the 200 m section exposed in this outcrop. As mentioned above, the refilling events do not induce an abrupt kink in Mg# and in other chemical parameters diagnostic of fractional crystallization. The resetting to primitive values occurs progressively in a transition zone of up to 50 m. A simple explanation for this observation would be that the refilling is not a sudden event. The delivery of primitive melts into the magma chamber could be more or less continuous, although its rate would not be constant. Intervals of decreasing Mg# would correspond to low rates of injection, i.e., periods where the low melt supply cannot balance the overall cooling of the magma chamber. Conversely, intervals of increasing Mg# would correspond to periods of higher rates of melt supply with efficient mixing leading to progressive resetting of the magma composition to primitive values. Alternatively, the delivery of primitive melts may be quite sudden but the mixing between these melts and the more evolved melt, stagnating in the magma chamber, may not be instantaneous. Another and more complex way to account for the observed systematic shift between the mineralogical transitions and the discontinuities in the geochemistry is to invoke compaction of the cumulus minerals (olivine + clinopyroxene), extraction of the evolved inter-cumulus melt from the compacting layers and the migration of this melt in the less compacted overlying layers leading to what has been called "inflation metastatism" (cf. Irvine, 1980).

Whatever the details of the petrological processes, the first order, robust conclusion remains: several large batches of primitive melt extracted from the mantle were injected in the shallow lithosphere and in the lower crust around the Maqsad diapir. Fractional crystallization of these melts has led to the formation of ultramafic, essentially wehrlitic–clinopyroxenitic cumulates. The degree of fractional crystallization was moderate (less than 10%) as indicated by the slight evolution of the Mg/(Mg + Fe) ratio of the cumulus minerals. Consequently, the total volume of melt that was injected exceeds, by at least one order of magnitude, the volume of the presently exposed rocks. Although an accurate quantitative determination of this volume would be difficult given the number of unknowns and poorly constrained parameters, a few cubic kilometres of melt looks a realistic estimate.

6.2. Nature of the parent melt of the ultramafic cumulates

The abundant crystallization of wehrlites and clinopyroxenites at low pressure requires parental melts rich in MgO, CaO and SiO₂ and poor in Al₂O₃, and possibly water, relative to tholeiitic melts of MORB affinity. Primitive high-Ca boninites is the best present-day analog of these parental melts. These lavas have been sampled along the Tonga trench, i.e. in an intra-oceanic subduction zone environment (Fallon and Crawford, 1991). Their phenocryst assemblage is composed of olivine + clinopyroxene + orthopyroxene + Cr-spinel; high-An plagioclase is present in the most evolved lavas. The analogy is also supported by similarities in the mineral composition of Oman ultramafic cumulates and phenocrysts of high-Ca boninites indicating that they are derived from a refractory mantle source. Oman Cr-spinel that crystallized in equilibrium with olivine and clinopyroxene to form the ultramafic cumulates has Cr# values ranging from 0.6 to 0.7 which are high compared to typical values for Cr-spinel crystallizing from MORB, that have Cr# scattered around an average value of ~0.5 (e.g. Araí et al., 2006; Python et al., 2008). However the compositional fields of Cr-spinel from MORB and boninites largely overlap and the high Cr# values of our ultramafic cumulates are in the intersection between these two fields; accordingly the value of Cr# alone is not sufficient to infer the composition of their source. However, other parameters (e.g. the low concentrations of Ti in both clinopyroxene and Cr-spinel and the low concentration of Al in clinopyroxene) point to parental melts derived from quite depleted peridotites. These depleted compositions are much closer to those encountered in cumulates from supra-subduction zone ophiolites and in phenocrysts of present-day boninites than in MORB cumulates or phenocrysts (e.g. Dick et al., 1991; Falloon and Crawford, 1991; Murton et al., 1992; van der Laan et al., 1992; Ross and Elton, 1997; Natland and Dick, 2001, Dantas et al., 2007; Python et al., 2008; Ceuleeneer and le Sueur, 2008).

A puzzling difference between the Maqsad ultramafic cumulates and supra-subduction zone boninites is the lack of orthopyroxene in the cumulates, in contrast to its ubiquitous occurrence in boninites. Orthopyroxene, frequently associated with high-An plagioclase is, however, abundant in the pegmatites surrounding the ultramafic cumulates. Trace element and isotopic evidence led us to propose that the parent melts of these pegmatites are derived from hydrated melting of a thin, altered lherzolitic lid triggered by the rise of a hot asthenospheric diapir to a shallow depth (Benoit et al., 1999). The absence of orthopyroxene from the wehrlitic–clinopyroxenitic cumulates indicates subtle but significant differences between their parent melts and those of the pegmatites, although the two are certainly co-genetic given their close association in the field. We consider that the pegmatites (websterites and gabbronorites) crystallized from a "pure" boninite-like endmember while the wehrlites and clinopyroxenites are cumulates from hybrid melts formed by mixing between lherzolitic melts and MORB extracted from the Maqsad diapir. MORB are poorer in silica than boninites, which would contribute to the displacement of the crystallization path away from a cotectic involving orthopyroxene. As long as the contribution of MORB remains moderate, it will not significantly affect the generally depleted characteristic of the resulting melt. Trace element and isotopic data are necessary to further constrain the exact nature of the melts and of the mantle source that led to the formation of Maqsad ultramafic cumulates, and to test and quantify the validity of the mixing hypothesis.

The abrupt, but magmatic, transition to the gabbros capping the ultramafic sequence records a drastic and sudden change in the nature of the melt supplying the magma chamber. These gabbros are formed from a much more evolved melt than the parent melts of ultramafic cumulates in terms of Mg# and Cr contents, and crystallized along the plagioclase + clinopyroxene cotectic, typical of evolved MORB (Dick et al., 1991; Ross and Elton, 1997; Natland and Dick, 2001).

6.3. Tectonic setting for the genesis of the ultramafic cumulates

It may be tempting to relate the occurrence of thick sequences of clinopyroxenite and wehrlites to petrological processes developing in the subduction factory. A simple interpretation would be to consider that their genesis is independent of the formation of the gabbroic Oman crust, and to interpret them as "late intrusive", following classical views concerning the genesis of ultramafic crustal lithologies of Oman (e.g., Lippard et al., 1986; Reuber, 1988; Ernewein et al., 1988). In these models, the Oman ophiolite would have recorded a polygenetic history, form an early accretion stage along an ocean-spreading centre (mid-ocean ridge or back-arc basin), to a late stage where water introduction into the mantle and the production of boninitic melts was a natural consequence of subduction and/or thrusting (e.g. Ernewein et al., 1988; Umino et al., 1990; Ishikawa et al., 2002).

In the case considered in the present study, however, it clearly appears that the ultramafic crustal lithologies are not randomly located relative to a former mantle diapir containing MORB melts. This strongly supports the view that the formation of the cumulate clinopyroxenites and wehrlites is related to the emplacement of this diapir, and that the ultramafic cumulates and the gabbroic cumulates forming the bulk of the crustal section in the Maqsad area are co-genetic.
The ultramafic cumulates cropping out in mantle peridotites (Northern outcrop) lack a high-temperature plastic deformation overprint. This indicates that they crystallized after the accretion of the asthenospheric (i.e. flowing) mantle into the lithosphere. On the other hand, they are presently restricted to an off-axis setting on both sides of the Maqsad paleo-ridge — although these structural relationships are complicated by the Muqbariah shear zone in the case of the southern outcrop. These structural observations, together with the petrological constraints, can be used to propose a possible scenario illustrated in Fig. 14. Following a period of magma-starved spreading, the Maqsad asthenospheric diapir reached shallow depths beneath the ridge and impinged on a formerly accreted and hydrothermally altered lithospheric lid. Evidence for penetration of very high temperature hydrothermal fluids at mantle depths is abundant in this area (Leblanc et al., 1991; Python et al., 2007). The temperature in partially molten asthenospheric diapirs beneath mid-ocean ridges is classically estimated around 1300 °C. This is significantly above the low pressure (~0.2 MPa) wet solidus of residual peridotites that would be refractory in dry conditions (≤1000 °C, e.g. Wyllie, 1988). Hydrated melting of the lithosphere will lead to the formation of melts ultra-depleted in incompatible elements but enriched in silica and water. Low-degree melting will produce trondhjemitic melts (e.g. Koepke et al., 2007) while higher degrees of melting will lead to the formation of melts rich in both silica and MgO, similar to boninites (e.g. Falloon and Crawford, 1991). The trondhjemitic pods with high-An% plagioclase scattered in the mantle section of the Maqsad area have been attributed to such processes and may record the stage of incipient hydrated melting of the altered, depleted peridotites (Amri et al., 1996). As melting proceeded, the lithospheric melts became boninitic-like in composition (Benoit et al., 1999). At this stage, the melt production was high enough to trigger hydraulic fracturing of the surrounding mantle rocks; however, due to the low ambient temperature, these melts were not able to migrate over long distances and crystallized in situ as pegmatitic dykes (cf. O’Driscoll et al., 2007).

The Maqsad diapir was saturated with MORB melts. As it continued to rise, the amount of MORB delivered at shallow depth increased, together with the surrounding temperature, leading to the formation of magma chambers when melt delivery occurred at a higher rate than crystallization. At this stage, the formation of layered cumulates became possible. The cumulates remained rich in pyroxene and devoid of plagioclase as long as the contribution of MORB was moderate relative to the one of the boninitic melts. Mixing between MORB and boninitic-like melts may account for the pronounced scatter in the incompatible elements content of clinopyroxene in the ultramafic cumulates. This is supported by the Ti content of these clinopyroxenes which are intermediate between the ultra-depleted values characterizing the most primitive pegmatites and the higher values typical of primitive MORB cumulates (Fig. 12).

As spreading continued, the contribution of MORB to the refilling of the magma chamber increased and gabbros started to crystallize. The magmatic discordance between the primitive ultramafic cumulates and the quite evolved gabbros observed in the southern outcrop, likely results from complex interactions between the dynamics of melt delivery and petrological evolution in the magma chamber and/or in the underlying mantle plumbing system. The Maqsad asthenospheric diapir eventually reached Moho level and at this stage, the lithospheric lid was fully eroded and the axial magma chamber only contained MORB-type melts. Although they formed in the lithosphere in an axial position, the ultramafic cumulates were passively transposed to off-axis positions, where they can be observed at present. This transposition does not imply plastic deformation since these ultramafic cumulates crystallized in the shallowest part of the mantle and at the base of the crust where the mantle was quickly cooled by hydrothermal circulation and this thin lithospheric lid was passively rafted away from axial position.

The hypothesis described above provides a possible explanation for the puzzling observation that although cumulates derived from andesitic melts are not common features along present-day mid-ocean ridges, they do exist (e.g. Nonnotte et al., 2005). Models of ocean spreading need to take this observation into account.

**Acknowledgments**

We are very indebted to Dr. Hilal Al Azri from the Ministry of Commerce and Industry of Oman for his support for the organization of our airborne survey and to Dr. Sabine Chabrillat for her invaluable contribution to the initiation of the project. The interpretation of the hyperspectral data has benefited from collaboration with Patrick Launee, Serge Chevrel, Christine Rosenberg, Regis Roy and Jean-Philippe Combe. Thin sections were produced by Fabienne de Parseval.