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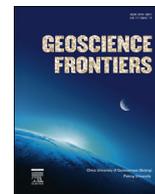


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Research paper

# Masirah – The other Oman ophiolite: A better analogue for mid-ocean ridge processes?

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## ABSTRACT

Oman has two ophiolites – the better known late Cretaceous northern Oman (or Semail) ophiolite and the lesser known and smaller, Jurassic Masirah ophiolite located on the eastern coast of the country adjacent to the Indian Ocean. A number of geological, geochronological and geochemical lines of evidence strongly suggest that the northern Oman ophiolite did not form at a mid-ocean ridge but rather in a supra-subduction zone setting by fast spreading during subduction initiation. In contrast the Masirah ophiolite is structurally part of a series of ophiolite nappes which are rooted in the Indian Ocean floor. There are significant geochemical differences between the Masirah and northern Oman ophiolites and none of the supra-subduction features typical of the northern Oman ophiolite are found at Masirah. Geochemically Masirah is MORB, although in detail it contains both enriched and depleted MORB reflecting a complex source for the lavas and dykes. The enrichment of this source predates the formation of the ophiolite. The condensed crustal section on Masirah (ca. 2 km) contains a very thin gabbro sequence and is thought to reflect its genesis from a cool mantle source associated with the early stages of sea-floor spreading during the early separation of eastern and western Gondwana. These data suggest that the Masirah ophiolite is a suitable analogue for an ophiolite created at a mid-ocean ridge, whereas the northern Oman ophiolite is not. The stratigraphic history of the Masirah ophiolite shows that it remained a part of the oceanic crust for ca. 80 Ma. The chemical variability and enrichment of the Masirah lavas is similar to that found elsewhere in Indian Ocean basalts and may simply reflect a similar provenance rather than a feature fundamental to the formation of the ophiolite.

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## 1. Introduction

For over 40 years the ophiolite model has been used to explain processes operating at mid-ocean ridges. Since the early days of plate tectonics ophiolites have been used as a powerful analogue for the lithologies created at a mid-ocean ridge spreading centre and have been used to provide important insights into the means whereby new ocean crust is created on Earth. In particular they have been used to augment the wealth of information obtained from basalts collected from the ocean floor and provide insights into the deep structure of the ocean crust. A formal definition of the term ophiolite was agreed at a Geological Society of America Penrose field conference in 1972 (Anonymous, 1972; Dilek, 2003) in which an ophiolite was defined as a sequence of lithological units

which from bottom to top comprised an ultramafic complex of harzburgites, lherzolites and dunites, a gabbroic complex, a sheeted mafic dyke complex and at the top a mafic volcanic complex, frequently pillowed. This lithological association quickly became a useful analogue for the seismic velocity structure of the ocean floor and models for how this layered structure was formed have provided a basis for understanding mid-ocean ridge processes (Fig. 1a). More recently, a comprehensive programme of ocean drilling has begun to confirm the 'Penrose' ophiolite model, although to date the ocean drilling programme has only penetrated the upper oceanic crust and the dyke-gabbro transition zone (Larsen et al., 2009). Thus we are still reliant upon ophiolite models of the ocean crust to understand processes in the gabbroic lower oceanic crust and melt migration processes in the underlying mantle.

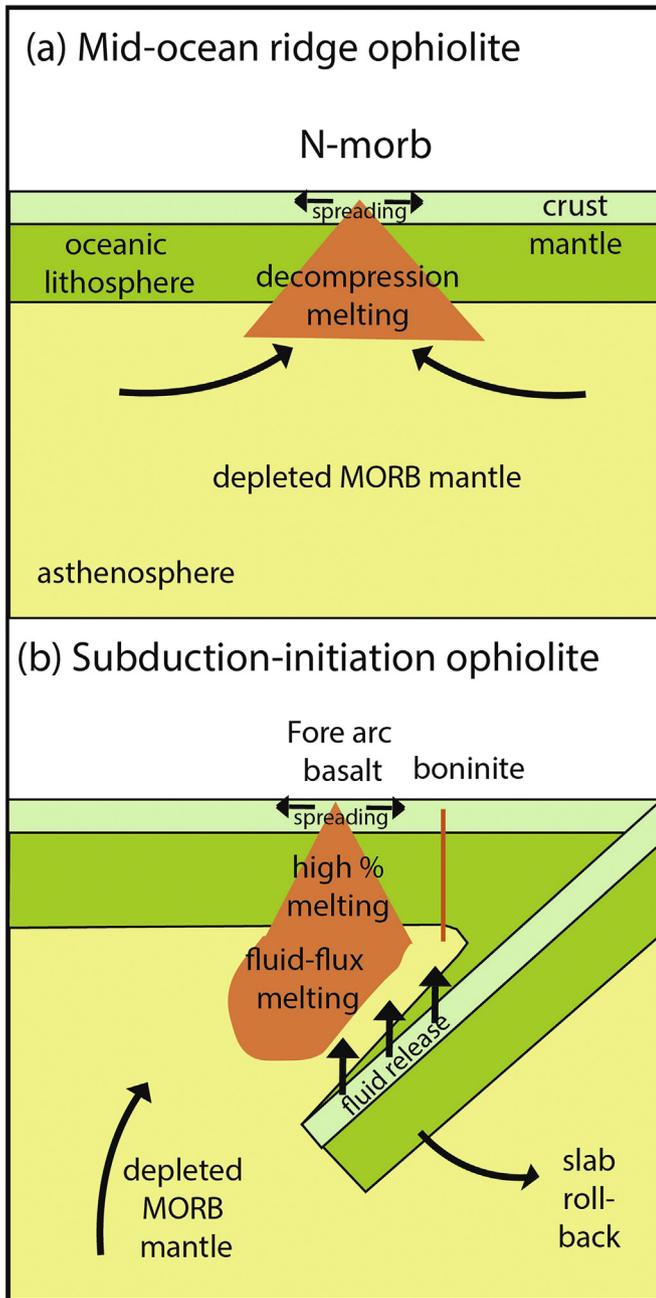
This paper reviews the way in which the ophiolite model has been used to understand mid-ocean ridge processes, in particular in the light of a new and broader understanding of ophiolites popularised by Dilek and Furness (2011, 2014). Specifically, it re-

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**Figure 1.** Competing models for the formation of ophiolites (modified after [Pearce, 2014](#)). (a) The conventional mid-ocean ridge model; (b) the subduction initiation model.

evaluates the evidence that the northern Oman ophiolite was formed at an ocean ridge spreading centre and argues that this ophiolite is better understood in terms of a subduction model. The paper then draws attention to the lesser known Masirah ophiolite one of very few probable ocean-ridge ophiolites. Masirah is located on the eastern coast of Oman and, using previously published data, this paper shows on geological and geochemical grounds that this ophiolite is a fragment of the Indian Ocean floor and is a true mid-ocean ridge ophiolite. However, the Masirah ophiolite is anomalous with respect to its thickness and composition. The significance of these anomalies is explored given the importance of Masirah in unravelling ocean ridge processes in the Indian Ocean.

## 2. Challenges to the standard interpretation of ophiolite formation

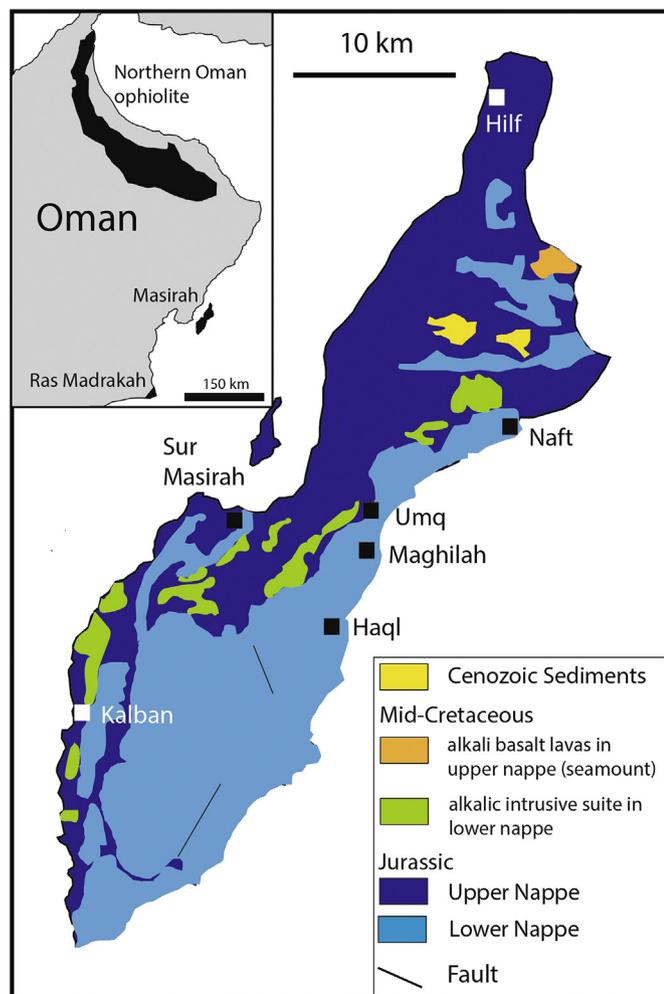
As our knowledge of ophiolites has increased it has become clear that they form in a number of different structural settings and not simply at a mid-ocean ridge. In particular it has been shown on geochemical grounds that many ophiolitic basaltic lavas are different from modern basalts dredged from mid-ocean ridges. Recently these ideas have been synthesised and reviewed by [Dilek and Furness \(2011, 2014\)](#) and [Pearce \(2014\)](#) who show, on the basis of geochemistry and structural setting, that there are five possible structural environments in which ophiolites might form. They argue that ophiolites may originate as part of a volcanic arc, in a supra-subduction zone setting, in a plume environment, at a continental margin or as originally proposed at a mid-ocean ridge. Whilst examples of each of these ophiolite types are known, the supra-subduction type is probably the most abundant and is widespread both geographically and through geological time ([Dilek and Furness, 2011](#); [Whattam and Stern, 2011](#)). In contrast it appears that there are only a very small number ophiolites which were created in an ocean ridge setting, far fewer preserved than has been originally supposed ([Dilek and Furness, 2014](#)).

### 2.1. The Oman example

Perhaps the most serious challenge to the standard ophiolite model comes from recent work on the Oman ophiolite, or the Semail ophiolite as it is sometimes known, one of the most complete, best exposed and best studied ophiolites, located in the north Oman Mountains ([Fig. 2](#)). The northern Oman ophiolite is widely cited as a classic example of an ophiolite which formed at a mid-ocean ridge ([Boudier et al., 1997](#)). Frequently this ophiolite sequence has been compared with a fast spreading ridge such as the East Pacific Rise, particularly because both appear to show evidence of an axial melt lens (see for example [Nicolas and Boudier, 2015](#); [Muller et al., 2017](#)). However, geochemical studies as far back as 1981 suggested that this may not be the full story for some of the lavas in the upper part of the succession have a supra-subduction zone trace element signature indicative of their origin in an arc environment and not at a mid-ocean ridge ([Pearce et al., 1981](#)).

Over the subsequent decades the competing ocean ridge versus subduction models ([Fig. 1](#)) have been strongly debated. However, a consensus is now emerging such that the Oman ophiolite is subduction related. The early work of [Pearce et al. \(1981\)](#) was based upon the trace element geochemistry of pillow basalts in the upper part of the succession. Here four further arguments are presented which support this point of view:

- (1) Recent work on mantle chromitites has shown that some mantle chromitites in Oman formed from boninitic magmas, melts which are uniquely of arc parentage ([Rollinson, 2008](#)). This is consistent with the discovery of boninitic lavas in the northern part of the ophiolite ([Ishikawa et al., 2002](#); [Kusano et al., 2014](#)). Others chromitites of basaltic parentage, originally thought to have formed at a mid-ocean ridge actually represent hydrous melts with high oxygen fugacities, and are quite atypical of basalts formed at a mid-ocean ridge setting; their hydrous nature also suggests a subduction origin ([Rollinson and Adetunji, 2013, 2016](#)). Further, it was noted that large chromitite pods, common in many ophiolites, had not been reported from modern mid-ocean ridge samples obtained either through dredging or drilling (see [Rollinson and Adetunji, 2013, 2016](#)). On this basis it has been proposed that podiform chromitites in the mantle harzburgite section of ophiolites only form in a subduction zone



**Figure 2.** Simplified geological map of Masirah Island, Oman (after Peters et al., 1995; Peters and Mercolli, 1998) showing the Jurassic upper and lower ophiolite nappes and the mid-Cretaceous alkaline basalts and intrusives. Inset map shows the location of Masirah Island and Ras Madrakah in Oman relative to the north Oman Mountains ophiolite.

setting and do not form during mid-ocean ridge magmatism at all. Thus they have the potential to act as a powerful discriminant between the two different tectonic environments (Rollinson and Adetunji, 2013, 2016), although this proposal has been disputed (Arai and Miura, 2015).

- (2) Confirmation of the subduction parentage of the Oman lavas was provided by MacLeod et al. (2013), who showed that the Oman Geotimes lavas (the V1 lavas of Fig. 3) were geochemically distinct from typical mid-ocean ridge (MORB) lavas. Notably on  $TiO_2$  vs  $MgO$  and  $Al_2O_3$  vs  $MgO$  geochemical plots the Oman lavas had consistently lower  $TiO_2$  and higher  $Al_2O_3$  for a given  $MgO$  value than do MORB lavas, but were similar to lavas associated with subduction. MacLeod et al. (2013) interpreted as signifying a greater role for clinopyroxene relative to plagioclase fractionation during the evolution of the magmas. In turn the greater importance of clinopyroxene relative to plagioclase was interpreted as a function of the water content of the melt, for experimental studies show that melts with a higher water content show suppressed plagioclase crystallisation allowing clinopyroxene to precede plagioclase in the crystallisation of the melt (MacLeod et al., 2013). These observations were confirmed through the modelling of possible water contents of the magmas leading to the conclusion that

the water content of the Oman melts was between 0.2 and 1.0 wt.%, higher than was modelled for MORB lavas. The presence of a high magmatic water content in the Oman lavas was interpreted as indicating a subduction zone origin for this water (MacLeod et al., 2013).

- (3) Prior to the work of MacLeod et al. (2013) on Oman lavas, Goodenough et al. (2010) had established that in the northern part of the Oman ophiolite in the United Arab Emirates there was suite of ‘younger’ gabbros which cross-cut earlier layered gabbros and which have the petrological character of hydrous magmas. These gabbros and associated dykes have low Ti and Nb and high Th, trace element attributes which are indicative of a supra-subduction zone setting (Goodenough et al., 2010).
- (4) Further, a recent study of granitic rocks in the mantle section of the northern part of the ophiolite shows that these are not typical oceanic plagiogranites but are true potassic granites and highly aluminous in composition. Geochemical modelling has shown that they represent melted sedimentary material which has been introduced into the mantle from below. The only viable mechanism whereby this could have taken place is through sediment melting during subduction beneath a mantle wedge – further confirming a subduction setting for the origin for this ophiolite (Cox et al., 1999; Rollinson, 2014, 2015).

## 2.2. A new tectonic understanding of ophiolites

The new views on ophiolites synthesised by Dilek and Furness (2014) are consistent with and supported by a new theoretical understanding of ophiolite formation and its link to ocean floor creation in the work of Whattam and Stern (2011) on subduction initiation. These authors showed that localised sea-floor spreading can take place in a subduction environment. They argue that as the process of subduction begins new asthenospheric mantle rises in front of the sinking slab initiating localised spreading in the forearc. Thus forearc basalts (see Fig. 1b) are created. New studies in the Izu-Bonin-Mariana forearc have revealed a stratigraphy which is analogous to that documented from Oman and other supra-subduction zone ophiolites (Ishizuka et al., 2014). This work is supported by recent geochemical studies which show that basalts from this young forearc are very similar to those found in supra-subduction ophiolites. The similarities in stratigraphy and geochemistry between forearc basalts and those found in supra-subduction zone ophiolites strongly support a subduction initiation model for the formation of supra-subduction ophiolites (Ishizuka et al., 2014; Arculus et al., 2015).

This new model illustrates how the process of sea-floor spreading can take place in multiple environments and builds on recent observations that crustal extension also takes place in a subduction environment. Thus there is the potential for sea-floor spreading to take place within a subduction system so reconciling the apparent paradox of the tectonics of spreading, clearly observed in the stratigraphy of the Oman ophiolite, with the geochemical signatures of subduction.

For these reasons the Oman ophiolite can no-longer be regarded as having formed at a mid-ocean ridge spreading centre, but rather in a localised spreading centre associated with subduction. This of course poses a dilemma. For if one of the type-examples of ocean crust creation at a mid-ocean ridge is incorrect, then where can we study an example of an ophiolite that faithfully records mid-ocean ridge spreading processes? Here it is argued that the Masirah ophiolite is structurally and geochemically different from the northern Oman ophiolite and is a more suitable candidate for former ocean crust created at a mid-ocean ridge spreading centre.

### 3. The Masirah ophiolite

A number of authors have previously described the lithologies of the Masirah ophiolite (Moseley and Abbotts, 1979, 1984; Moseley, 1990; Peters and Mercolli, 1997, 1998; Peters, 2000) and these data have principally been used to better understand the breakup of Gondwana and the formation of the Indian Ocean. The differences between Masirah and the northern Oman ophiolite have not previously been discussed in the context of a possible diversity of ophiolite 'environments'. Hence the significance of the Masirah ophiolite as possible example of an ocean ridge ophiolite has not been noticed apart from a brief discussion of the geochemistry by Dilek and Furness (2014). In part this may also be because the Masirah ophiolite is more hydrothermally altered and tectonically disrupted than its well preserved northern neighbour. The purpose of this study therefore, is to re-examine the geological relationships and geochemistry of the Masirah ophiolite in the light of what we now know about the multiplicity of environments in which ophiolitic crust can be created. In part this will be done through a comparison with the geology and geochemistry of the better known supra-subduction zone ophiolite in northern Oman.

The Masirah ophiolite is located on the eastern coast of Oman, adjacent to the Indian Ocean and is thought to be part of the Batain Group, a mélangé comprising Permian to Cretaceous sediments and volcanic rocks obducted onto the Oman passive continental margin (Shackleton and Reis, 1990; Schreurs and Immenhauser, 1999). These rocks also contain a suite of Cretaceous alkaline volcanic rocks including the  $137 \pm 1$  Ma Asseelah ultramafic lamprophyre-carbonatite complex (Nasir et al., 2011). The Masirah ophiolite has been dated as ca. 150 Ma on the basis of U-Pb in zircon ages in gabbro dykes ( $156 \pm 5$  Ma and  $150 \pm 14$  Ma, Peters et al., 1995), Ar-Ar ages in plagioclase from gabbro dykes (Smewing et al., 1991) and latest Jurassic (Tithonian) radiolaria in the immediately overlying sediments. However, the obduction of the Batain nappes and the emplacement of the Masirah ophiolite are much later and thought on stratigraphic grounds to have taken place during the late Maastrichtian/early Palaeocene at ca. 65 Ma. The transport direction is thought to have been from the ESE to the WNW, from a former 'Batain Basin'.

The ophiolite is exposed over 700 km<sup>2</sup> on the island of Masirah (Moseley and Abbotts, 1979) and in the nearby headland of Ras Madrakah on the Arabian mainland (Shackleton and Reis, 1990) (Fig. 2). In detail there are two ophiolitic sheets exposed, thrust one above the other, both preserving in places a near-complete ophiolitic sequence (Peters and Mercolli, 1997). The overlying sediments preserve a long history and indicate that the ophiolite was part of the ocean floor for ca. 80 Ma (Peters and Mercolli, 1998) – very different from suprasubduction zone ophiolites such as the one in northern Oman where the creation to obduction history may be as short as 1 Ma (Rioux et al., 2012, 2016). The sedimentary history of the two nappes indicates that the thrusting was in the upper Maastrichtian to lower Palaeocene shortly before the obduction of the ophiolite onto the Arabian continental margin. The lower nappe preserves a complete ophiolite sequence comprising a peridotitic mantle to pillow lava sequence overlain by a thick sequence of sediments, whereas the upper nappe has a mantle to pillow lava sequence but no overlying sediment (Peters and Mercolli, 1998).

#### 3.1. The Masirah ophiolite – stratigraphy

The total thickness of the Masirah ophiolite varies, but it is thin, for the total crustal section never exceeds more than 2–2.5 km (see Fig. 3). Peters and Mercolli (1997, 1998) described several sections through the ophiolite and showed that the mantle sequence has a

maximum thickness of about 100 m and is primarily composed of harzburgite with rare dunitic bands. For the most part the mantle rocks are heavily serpentinised. Podiform chromitites are not reported although there are occasional aggregates of chromite-rich dunite (ca. 5% chromite) in serpentinised harzburgite. Harzburgites and dunites are also described from the Ras Madrakah headland on the Oman mainland, also thought to be part of the Masirah ophiolite (Shackleton and Reis, 1990).

Mantle harzburgites transition upwards into the lower crustal section through a layered gabbro-peridotite complex comprising a sequence of plagioclase dunite, plagioclase wehrlite and melanoctroctolite intrusions interlayered with the gabbro. The gabbro sequence is up to 500 m thick and shows significant internal variations although there are no discontinuities and no evidence that any part of the section is missing. These gabbros are layered in the lower part of the crustal section and foliated in the upper part. The foliation defined by the plagioclase and clinopyroxene is sub-parallel to the gabbroic layering lower in the succession. Clinopyroxene is the dominant pyroxene phase.

The overlying dyke sequence is up to 1000 m thick and again there is a gradual transition from the gabbro sequence into the dykes. Most dykes are  $1 \pm 2$  m thick and three different types of mafic dyke were recorded by Peters and Mercolli (1998) – microgabbroic dolerite, porphyritic dolerite and aphyric black dolerite. The microgabbroic dolerite and porphyritic dolerite dykes contain abundant euhedral plagioclase phenocrysts, pseudomorphs after olivine phenocrysts and a smaller volume of clinopyroxene phenocrysts.

The lavas in the upper part of the ophiolite vary in thickness from 250 to 500 m and comprise sheet flows, volcanic breccias and pelagic sediments in addition to pillow lavas. There are large, metre-sized pillows without interpillow material at the base of the lava sequence and these grade upwards into smaller pillows and lava tubes. These basaltic rocks are microcrystalline to porphyritic and do not preserve a glassy matrix. Plagioclase, olivine and clinopyroxene or Cr-spinel phenocrysts are present. They have been metamorphosed to prehnite-pumpellyite and greenschist facies although there is little evidence of intense hydrothermal alteration and no significant massive sulphide deposits are recorded. A summary of the stratigraphy is given in Fig. 3.

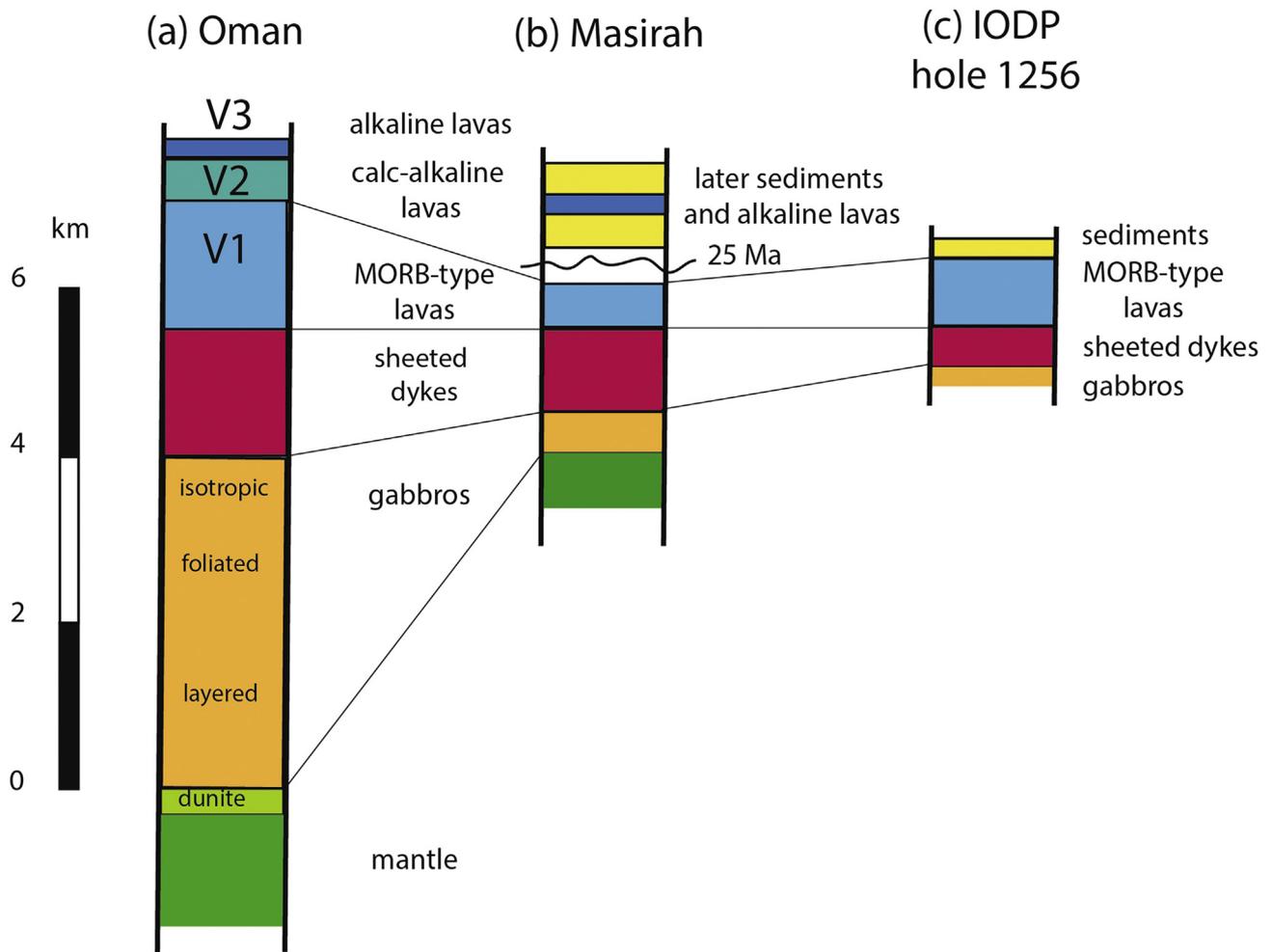
Structurally overlying the pillow lavas of the lower nappe are alkali volcanic rocks of mid-Cretaceous age. These are thought to represent former seamount structures developed on the Jurassic ocean floor and have the geochemical character of ocean island basalts (Meyer et al., 1996). They are red-brown in colour, distinguishing them from the green basalts of the underlying ophiolite sequence. Intercalated sediments indicate that they formed at ca. 130–125 Ma. In the upper nappe there is an alkali gabbro-diorite-granite suite of the same age (Peters, 2000), representing a plutonic substructure to the seamounts.

#### 3.2. The Masirah ophiolite – geochemical character

Peters and Mercolli (1998) and Mahoney et al. (1998) provided the only modern geochemical data for the Masirah ophiolite. Here their data are used and compared to data from the northern Oman ophiolite from (Godard et al., 2006; MacLeod et al., 2013) and lavas from the Atlantic Bank region of the southern Indian Ocean floor reported by Coogan et al. (2004).

##### 3.2.1. Major element chemistry

Peters and Mercolli (1998) showed that basaltic lavas and dykes from Masirah have a relatively narrow range of MgO and SiO<sub>2</sub> values (5.5–12.1 and 44.2–54 wt.% dry, respectively). A positive correlation on a Ni-MgO plot suggests olivine fractionation from a



**Figure 3.** Stratigraphic sections for (a) the subduction related Oman ophiolite from northern Oman (Lippard et al., 1986; Nicolas et al., 1996); (b) the Masirah ophiolite (Peters and Mercolli, 1998) and (c) the ocean crust at IODP hole 1256 in the Guatemala Basin, in the eastern Pacific Ocean (after Larsen et al., 2009).

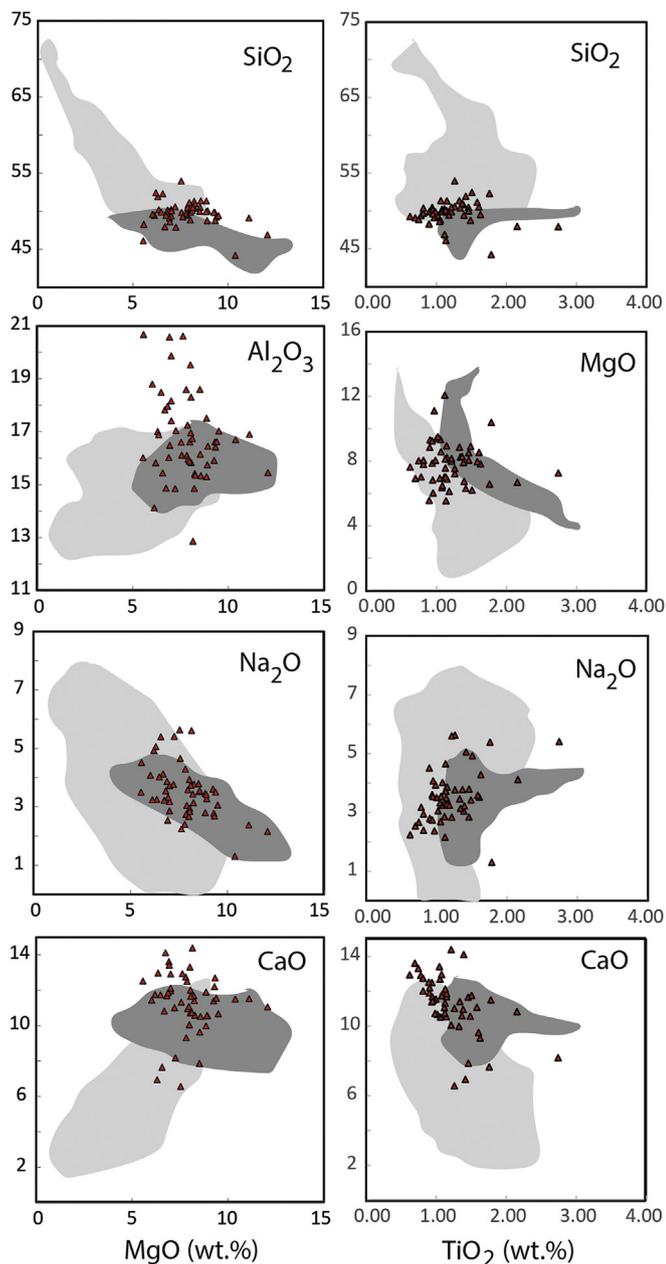
parental melt composition of 10–12 wt.% MgO. These compositions are similar to lavas from the southern Indian Ocean floor (Coogan et al., 2004) and when compared to compositions from the northern Oman ophiolite the Masirah samples are relatively unevolved (see Fig. 4). Scatter plots of major elements vs MgO and TiO<sub>2</sub> show that the Masirah lavas and dykes have a range of CaO and Al<sub>2</sub>O<sub>3</sub> values which are similar to the range found in Indian Ocean floor but which have a very different distribution from the range found in the northern Oman ophiolite (Fig. 4). This suggests that these mafic magmas experienced some plagioclase fractionation in contrast to the northern Oman lavas where the geochemical evidence suggests that plagioclase fractionation was suppressed in favour of clinopyroxene fractionation (MacLeod et al., 2013). Whilst some of the scatter in these diagrams may reflect some major element mobility post-dating the magmatic history of these lavas (see for example the K-Zr plot, Fig. 5b) the large differences between the composition of the lavas from the northern Oman ophiolite and those from Masirah strongly suggest that their fractionation history was different. In contrast the similarity between major element compositions in the Indian Ocean lavas and those at Masirah suggest a possible affinity. Basaltic lavas and dykes from Masirah also have a similar range of Na<sub>2</sub>O values to that of the Indian Ocean lavas, see Fig. 4 (although some of the higher values at Masirah may reflect the spilitisation of some pillow basalts), but less extensive than the range in the Oman suite (Fig. 4). Peters and

Mercolli (1997) following Klein and Langmuir (1987) used the mean Na<sub>2</sub>O at MgO = 8 wt.% of 3.45 wt.% Na<sub>2</sub>O, to infer a crustal thickness of ca. 2 km. If this value is adjusted downwards to allow for some Na<sub>2</sub>O enrichment during alteration the estimated crustal thickness is still small.

In summary, the major element chemistry of the Masirah lavas and dykes is similar to that of lavas from the southern Indian Ocean but different from the compositions of lavas from the northern Oman ophiolite. The only exception to this are for the oxides TiO<sub>2</sub>, Fe<sub>2</sub>O<sub>3</sub><sup>T</sup> and P<sub>2</sub>O<sub>5</sub> where the concentrations are lower in the Masirah lavas than in those from the Indian Ocean, although scatter plots of Fe<sub>2</sub>O<sub>3</sub> vs TiO<sub>2</sub> and P<sub>2</sub>O<sub>5</sub> vs TiO<sub>2</sub> (not shown) demonstrate that the two lava suites lie on a common trend and have a similar petrogenetic affinity.

### 3.2.2. Rare earth element chemistry

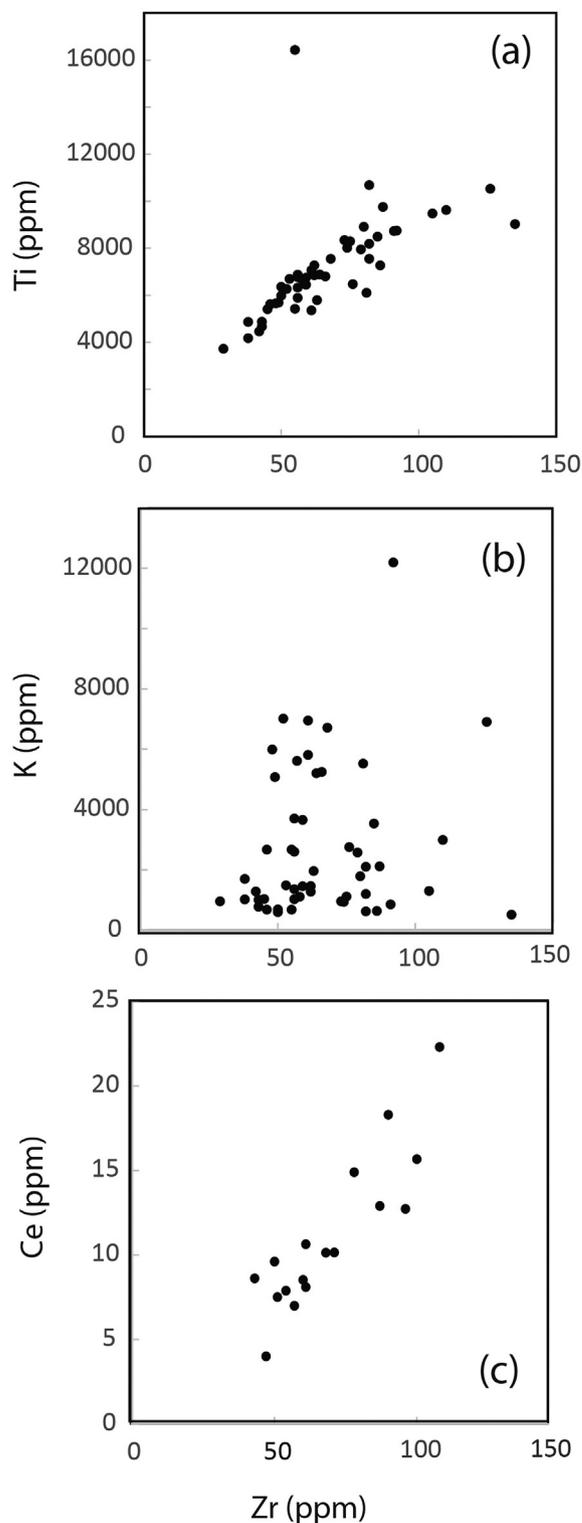
Rare earth element plots for the Masirah lavas are flat, with concentrations between 10 and 30 × chondrite (Mahoney et al., 1998; Peters and Mercolli, 1998). In most samples the light REE are slightly enriched with normalised La/Sm ratios between 1.02 and 1.32, with one sample strongly enriched with (La/Sm)<sub>n</sub> = 1.86 indicating that the parental melts were derived from an enriched mantle source (Fig. 6a). Other samples are not enriched in the light REE and are more typical N-MORB, with concentrations between 16 and 25 × chondrite, and have normalised La/Sm ratios between 0.95 and



**Figure 4.** MgO plots and TiO<sub>2</sub> plots for lavas and dykes from the Masirah ophiolite (red triangles), compared with samples from the Indian Ocean (dark grey) and the Oman ophiolite (light grey) (Data from [Peters and Mercolli, 1998](#); [Coogan et al., 2004](#); [Godard et al., 2006](#); [MacLeod et al., 2013](#)).

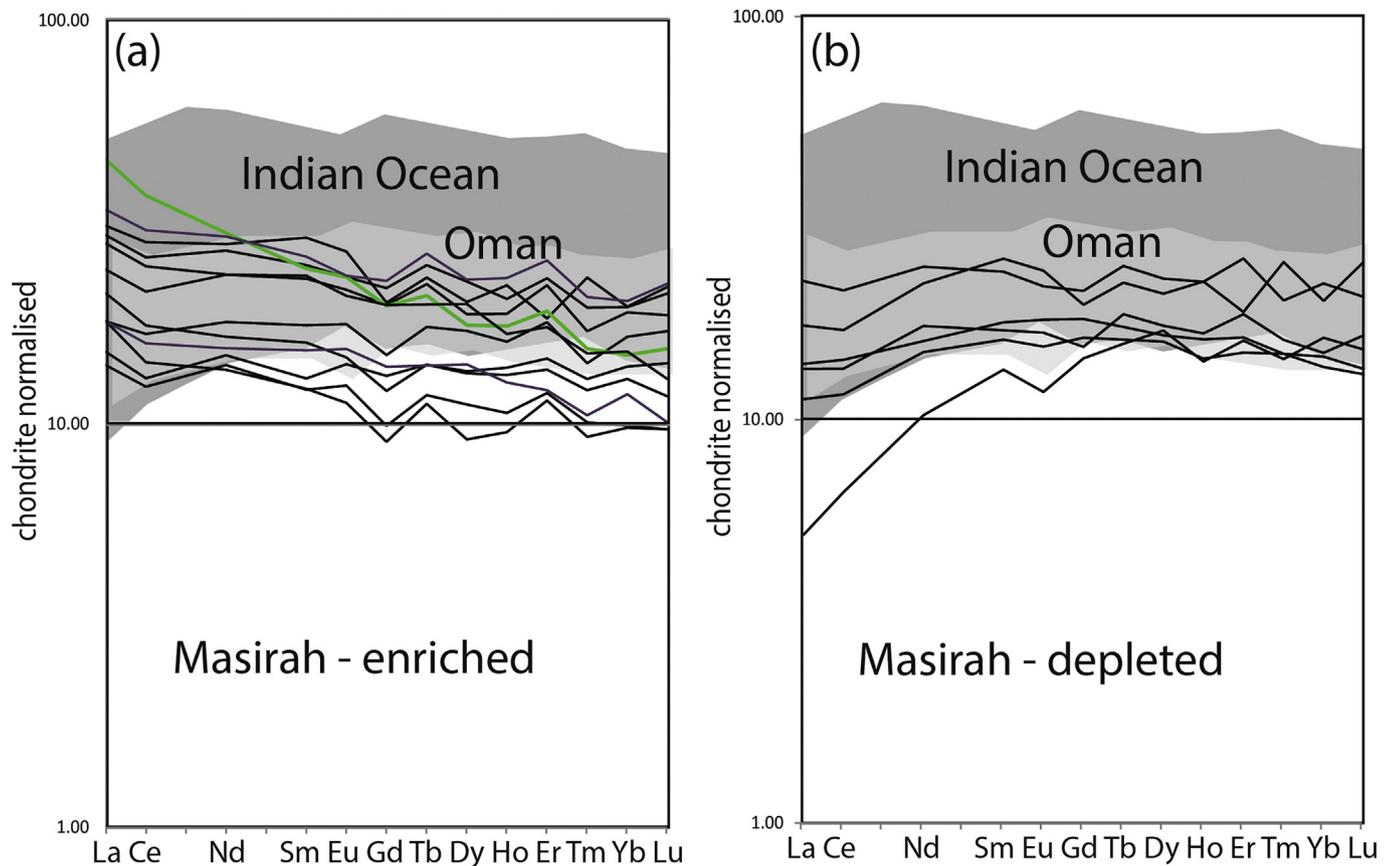
0.39 (Fig. 6b). The positive correlation between Ce and the immobile element Zr (Fig. 5c) indicates that there is no significant element mobility in the light REE, despite the observations of [Peters and Mercolli \(1998\)](#). However, there is no simple relationship between the MgO content of the mafic rocks and their REE concentrations, indicating that the lavas sampled do not represent a simple cogenetic magmatic suite, but are from a number of different fractionation series and therefore different parental magmas.

Relative to basaltic lavas from the southern Indian Ocean (data from [Coogan et al., 2004](#)) the majority of the Masirah lavas have lower concentrations of middle and heavy REE, but are relatively enriched in light REE (Fig. 6). In the western Indian Ocean, however, drillhole lavas are more variable in composition and include samples of enriched and depleted MORB ( $(\text{La}/\text{Sm})_n = 0.35\text{--}3.8$ ,



**Figure 5.** Plots of the trace elements Ti, K and Ce against the immobile element Zr in the Masirah lavas and dykes. (a) Ti vs Zr indicates immobility of the two elements; (b) the scatter between K and Zr suggests that K has been mobilised; (c) the positive correlation between Ce and Zr indicates minimal light REE mobility (Data from [Peters and Mercolli, 1998](#)).

[Mahoney et al., 1998](#)). A smaller number of Masirah lavas conform to the more typical light REE depleted N-type MORB patterns of the Indian Ocean samples and the northern Oman lavas ([Coogan et al., 2004](#); [Godard et al., 2006](#)), see Fig. 6.



**Figure 6.** Rare earth element (REE) plots for lavas and dykes from the Masirah ophiolite relative to data for lavas from the northern Oman ophiolite (light grey field) after Godard et al., 2006) and lavas from the southern Indian Ocean (dark grey field) after Coogan et al. (2004). (a) Lavas showing light REE enrichment typical of E-MORB; the sample plotted in green shows the most extreme REE fractionation; (b) lavas showing light REE depletion typical of N-MORB.

### 3.2.3. Other immobile elements

Peters and Mercolli (1998) and Mahoney et al. (1998) reported immobile incompatible element data for the Masirah mafic rocks. Zr/Nb ratios vary between 8.4 and 51.6 and Y/Nb ratios between 2.3 and 23.4. In both cases the higher values are associated with mafic rocks with depleted LREE patterns. Zr/Y ratios are also variable between 2.4 and 4.0. This variability cannot easily be explained by fractional crystallisation of plagioclase, clinopyroxene or olivine and reflects some variability in either the amount of melting or in the source.  $\epsilon_{\text{Nd}}$  values calculated for 150 Ma are highly variable and range from +6.4 to +10.5 in Masirah basalts and from +6.7 to +11.0 in the associated gabbros (Nagler and Frei, 1997; Mahoney et al., 1998) indicating a variety of magma batches from a heterogeneous mantle source.

Using the discriminants for the six different ophiolite types proposed by Pearce (2014) the Masirah lavas and dykes plot in the MORB field on the V vs Ti (ppm)/1000 diagram (Fig. 7a). On the Nb/Yb vs Th/Yb plot (Pearce, 2014) the Masirah samples plot in the main MORB-OIB oceanic array, between the positions of average N-MORB and E-MORB, whereas the northern Oman samples plot in the subduction zone field (Fig. 7b). N-MORB normalised trace element patterns, using the data of Peters and Mercolli (1998) and Mahoney et al. (1998) and Rb in the place of Th, show that most Masirah samples display enrichment typical of E-type and P-type MORB (Fig. 8), however, there is obvious Rb mobility. However, it is of some importance to note that there is no indication of an arc signature in these rocks, for there is no negative Nb-anomaly and these samples have relatively high Ti at a given V content (Fig. 7a).

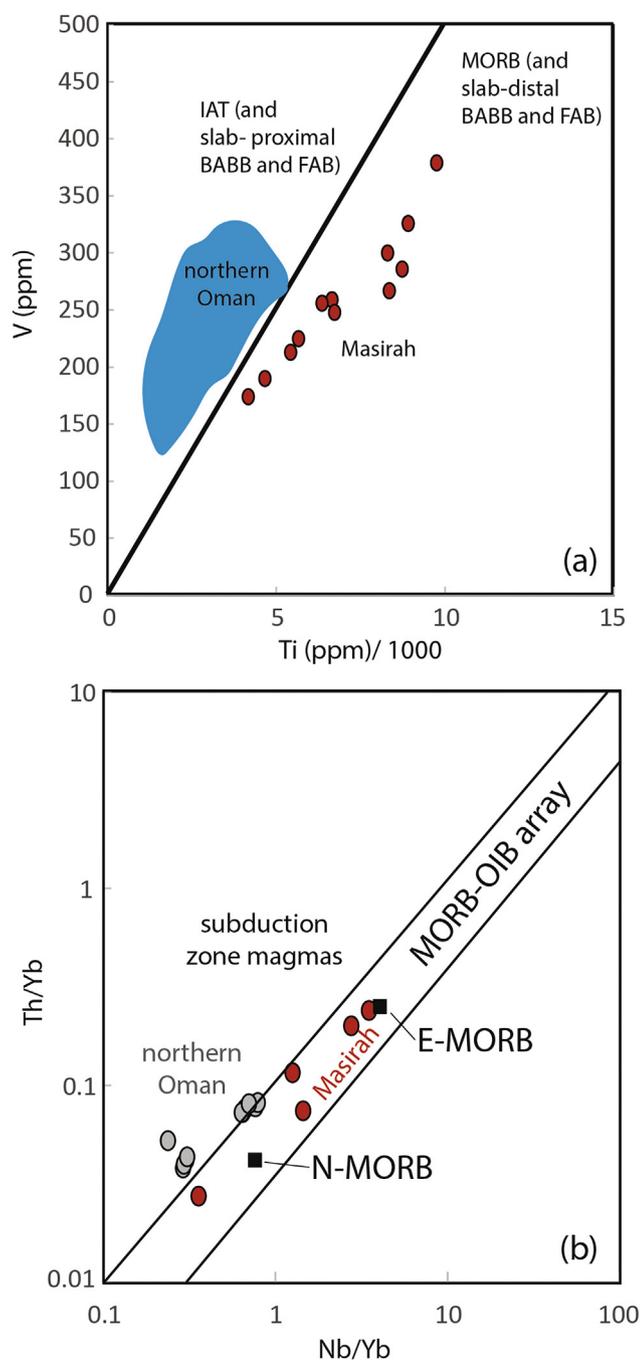
Thus the immobile trace element data indicate that the Masirah lavas and dykes have a close affinity to MORB lavas and do not carry

the fingerprint of arc magmatism. However, in detail whilst a small number of samples have REE patterns similar to those of N-MORB, many appear to be derived from a mantle source typical of E-MORB or P-MORB. Together the data from major elements, incompatible elements, REE and Nd-isotopes suggest that the relatively thin crustal section of the Masirah ophiolite contains batches of melt which are the product of multiple melting events and that these melts are derived from a number of different mantle source regions.

### 3.3. Distinctive features of the Masirah ophiolite

The description of the Masirah ophiolite above highlights a number of important differences between it and subduction-related ophiolites such as the northern Oman ophiolite. These are summarised as follows. First, the relatively thick sequence of sediments and lavas which overly the Masirah ophiolite show that it preserves a very long (80 Myr) history of igneous and sedimentary activity prior to emplacement onto the Arabian continental crust. In contrast in the northern Oman ophiolite the sedimentary cover is thin, only ca. 100 m thick, and recent U-Pb zircon geochronology indicates that there was a very short time interval of ca. 1 Ma between the formation of the mafic crust and its obduction (Rioux et al., 2012, 2016).

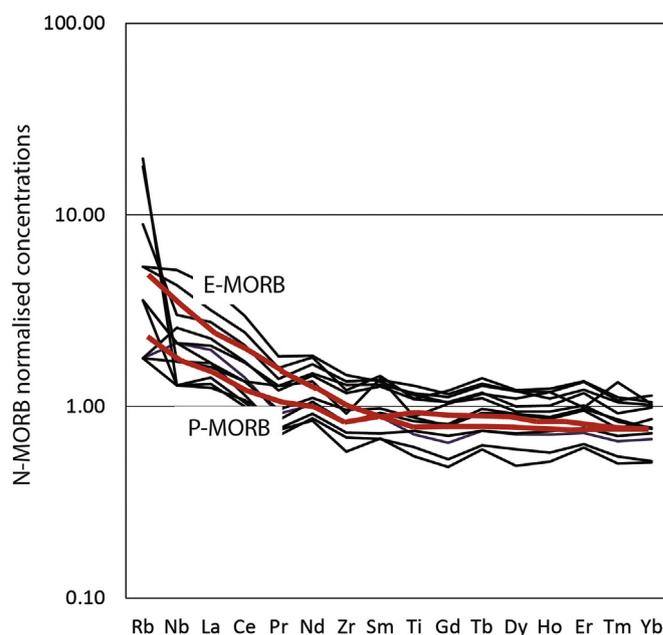
Secondly, the Masirah ophiolite is about 2 km thick – from the petrological Moho to the top of the pillow basalts, and has a thin gabbro sequence (ca. 500 m) and a thin lava sequence (250–500 m), see Fig. 3. In contrast the northern Oman is ca. 6–8 km thick (Lippard et al., 1986; Nicolas et al., 1996), more representative of typical thicknesses quoted for average ocean crust



**Figure 7.** Immobible element discrimination diagrams for lavas from Masirah and the northern Oman ophiolite. (a) Ti (ppm)/1000 vs V. The Masirah data are from [Peters and Mercolli \(1997\)](#). The data for the northern Oman ophiolite are for the upper pillow lavbas and have been extracted from Fig. 8 of [Kusano et al. \(2014\)](#). (b) Nb/Yb vs Th/Yb diagram. The Masirah data are from [Mahoney et al. \(1998\)](#), the northern Oman data are from [Godard et al. \(2006\)](#). The templates are from [Pearce \(2014\)](#).

([White et al., 1992](#)). In particular the northern Oman ophiolite has a gabbro sequence which is on average 3.6 km thick, although a range of thicknesses from 1.5 to 5.4 km have been measured in different parts of the ophiolite, and a lava and sequence which is up to 2 km thick ([Lippard et al., 1986](#); [Nicolas et al., 1996](#)). Relative to the IODP hole 1256 drilled into fast spreading crust from the Guatemala Basin in the Eastern Pacific Ocean, the Masirah lavas are thinner but the dyke sequence thicker than those recorded ([Larsen et al., 2009](#)).

Further, in Masirah the dominant pyroxene phase in the gabbros is clinopyroxene whereas in northern Oman orthopyroxene



**Figure 8.** N-MORB normalised concentrations of trace elements in the Masirah ophiolite lavas and dykes, relative to concentrations in E-MORB and P-MORB (red lines), data after [Pearce \(2014\)](#). The plot emphasises the enriched nature of most of the Masirah magmas and the relative mobility of the element Rb. Normalising values after [Sun and McDonough \(1989\)](#).

gabbros (gabbro-norites) are common. This may be indicative of a slightly more silica-rich melt composition in the northern Oman ophiolite than in Masirah. Also, the lava sequence in the two ophiolites is very different. On Masirah MORB-like basalts are overlain by a suite of alkaline lavas, with a time interval of ca. 20 Ma between the two events. In Oman there is a geochemical progression from MORB-like basalts to evolved arc like basalts, over a short interval of time (ca. 1 Ma).

Geochemically the Masirah ophiolite is distinctive. The major element chemistry, supported by phenocryst evidence, indicates that plagioclase fractionation is common, whereas in the northern Oman ophiolite clinopyroxene fractionation was dominant. [MacLeod et al. \(2013\)](#) showed that this difference in fractionation pattern is indicative melts with different water contents and implies that the Masirah lavas were drier than those in northern Oman. The Masirah lavas are also distinctive in their trace element character. REE patterns show both light REE depletion and enrichment in contrast to the universally depleted character of the northern Oman lower pillow lavas. The Masirah lavas also show variable incompatible element ratios, which together with the variable REE and variable Nd-isotope compositions imply many geochemically different magma batches most probably from different mantle sources. Finally, the discriminants proposed by [Pearce \(2014\)](#) for ophiolites indicate that there is no geochemical evidence that the Masirah dykes and lavas formed in a subduction environment. MORB-normalised trace element patterns show no negative Nb-anomaly, and these samples have relatively high Ti at a given V content.

## 4. Discussion

### 4.1. Masirah is structurally and geochemically a part of the Indian ocean crust

Based on the observations of [Barton et al. \(1990\)](#) and [Beauchamp et al. \(1995\)](#), [Peters \(2000\)](#) proposed that the Masirah

ophiolite nappes are part of a sequence of mafic rocks which extend 200 km eastwards into the Indian Ocean and are present beneath the Masirah Ridge, the Owen Basin (where typical oceanic crust is clearly identified, [Barton et al., 1990](#)) and extend to the Owen Ridge. In detail this crust/nappe complex varies in thickness from 5 km beneath the Masirah Basin to 14 km beneath the Masirah Ridge ([Barton et al., 1990](#)). In other words [Peters \(2000\)](#) proposed that there is some connectivity between the ocean crust of the Indian Ocean floor and the nappes of the Masirah ophiolite. However, the ocean crust east of Masirah is younger than expected from plate reconstructions ([Barton et al., 1990](#)) for the crust beneath the Owen Basin is late Cretaceous in age ([Mountain and Prell, 1990](#)) in contrast to the late Jurassic age of the Masirah ophiolite.

All the geochemical indicators discussed above show that the Masirah ophiolite is MORB-like. There are no geochemical indicators of arc magmatism. In a detailed trace element and isotopic study [Mahoney et al. \(1998\)](#) showed that in many respects Masirah conforms to typical Indian Ocean crust. Trace element signatures show evidence of both enrichment and depletion, as is seen in basalts from the Indian Ocean, and Pb- and Nd-isotope studies place the Masirah lavas firmly in the Indian Ocean field ([Mahoney et al., 1998](#)).

#### 4.2. Masirah is anomalously thin ocean crust

The crustal section of the Masirah ophiolite is thin relative to normal ocean crust. It is ca. 2 km thick and there is no evidence to suggest that the crustal section is incomplete ([Peters and Mercolli, 1998](#)). So, whilst the Masirah ophiolite represents crust created at an ocean ridge it appears to be anomalously thin.

The thickness of the ocean crust is determined by the amount of mantle melting, which was shown by [McKenzie and Bickle \(1988\)](#) is a function of mantle potential temperature. Normal oceanic crustal thicknesses determined from seismic measurements are ca. 7 km ([White et al., 1992](#)) and these thicknesses are recorded for both fast and slow spreading ridges. This 'normal' ocean crust thickness reflects a mantle potential temperature of ca. 1300 °C. Thinner crust occurs where there is less melt either because the mantle is cooler due to conductive heat loss, or where melt has been removed in an earlier melt-extraction event ([Singh et al., 2011](#)). Thin crust associated with cooling is found in normal spreading centres at segment ends, in association with ocean fracture zones, or in the early stages of spreading associated with continental rifting ([White et al., 1992](#)).

There is also a relationship between crustal thickness and spreading rate. For example, [Wilson et al. \(2006\)](#) showed that there is an inverse relationship between the top of the axial melt lens and the spreading rate so that the thickness of the oceanic upper crust decreases with increased spreading rate. In the case of a fast spreading ridge (10 cm/yr – half-rate) such as is sampled in the IODP hole 1256 the upper crust may only be about 1 km thick, although the relative thicknesses of lavas and dykes can, locally, be very variable ([Larsen et al., 2009](#)). The lava section sampled in the IODP hole 1256 fast spreading ridge is dominated by massive sheet flows, fragmented flows and breccias and pillow lavas are comparatively rare ([Tominaga et al., 2009](#)), not dissimilar to Masirah. Slow spreading ridges on the other hand may be magma-starved, such that there is no magma chamber and no upper oceanic crust ([Phipps Morgan and Chen, 1993](#)). Alternatively, melt may be created but may stagnate within the mantle leading to a thin oceanic upper crust and shallow gabbro intrusions in peridotitic mantle. These intrusions do not appear to crystallise as layered gabbros but rather react with the enclosing mantle peridotite for where mantle has been penetrated by drilling in these areas there are examples of diffuse melt-mantle interaction ([Larsen et al.,](#)

[2009](#)). This is in contrast to what is commonly observed in ophiolites where dunitic veins in harzburgitic mantle indicate strong mantle-melt interaction and pyroxene dissolution.

Applying these observations to Masirah it would appear that the Masirah ophiolite formed at a 'cool' spreading centre. The stratigraphy of the ophiolite indicates that Masirah was not the product of slow spreading because there is an upper crust section which includes a modest thickness of gabbro, some of which is layered, indicating the former presence of an axial magma chamber. These features may be best interpreted in the light of the original setting of the Masirah ophiolite. [Peters and Mercolli \(1998\)](#) and [Peters \(2000\)](#) suggested on the basis of its 150 Ma age and its palaeomagnetic position that the Masirah ophiolite was formed off the eastern coast of Africa at a spreading ridge associated with the early stages of separation between eastern and western Gondwana. At 150 Ma it may have been relatively close to the African continent and either Madagascar or Antarctica. It is possible that this location, as a young spreading centre, in an intercontinental rifting environment can explain the reduced the mantle potential temperature, necessary for the reduced melt production. This is consistent with the work of [Pérez-Gussinyé et al. \(2006\)](#) who modelled melt production at the west Iberian continent-ocean transition and who showed that in this type of rifting environment reduced melt thickness is a function of both a reduced spreading rate and a lower mantle potential temperature.

#### 4.3. Masirah is anomalously enriched ocean crust

The Masirah ophiolite preserves a very variable geochemistry in a highly condensed crustal section. In particular trace element ratios, REE and mantle normalised trace element patterns show a great deal of variability from typical N-MORB to E-MORB. Nd isotopes also show greater than expected variability with  $\epsilon_{Nd,150}$  values ranging from +11.0 to +6.4, i.e. from mantle that is depleted to that which is more enriched. This isotopic evidence indicates that this variability cannot simply reflect varying degrees of partial melting but must indicate compositional variability in the mantle source.

Trace element enrichment in the MORB source has been explained in a number of different ways. For example, source enrichment might reflect the proximity of a mantle plume, although none is known in the vicinity of Masirah at 150 Ma. Other models include source enrichment through mixing. For example, [Hémond et al. \(2006\)](#) proposed the mixing of an enriched basaltic component into otherwise depleted mantle. Alternatively [Donnelly et al. \(2004\)](#) have suggested that enrichment is the consequence of the infiltration of small fraction melts into the source. What is not permissible however is the mixing of continental material into the mantle source. This is precluded by the trace element concentrations in the basalts. The very nature of mantle enrichment is that it predates the time of eruption of the associated lavas, for in the case of Nd isotopes there needs to be time for samples with contrasting Sm/Nd ratios to evolve ([Mahoney et al., 1998](#)). This is supported in the Nd-isotopic data from the southeast Indian Ridge which suggest that the enrichment took place at about 300 Ma, long before spreading began ([Donnelly et al., 2004](#)).

Thus on Masirah, the enriched nature of the mantle source is not a feature that can be associated specifically with the opening of the Indian Ocean, rather it predates it. What is not clear is whether the heterogeneous source now reflected in the Masirah lavas, and created at least 150 Ma before the initiation of seafloor spreading ([Donnelly et al., 2004](#)), is fundamentally linked to that spreading process, perhaps as an essential precursor to continental rifting. On the other hand, however, there appears to be sufficient similarity between the isotopic chemistry of the Masirah lavas and the basalts from the northwest Indian Ocean floor to suggest that the Masirah

lavas are not in any way special but simply reflect the isotopic diversity of the Indian Ocean source (Mahoney et al., 1998).

## 5. Conclusions

The conclusion of this review is that the Masirah ophiolite is a suitable analogue for an ophiolite created at a mid-ocean ridge, whereas the northern Oman ophiolite is not. The original setting of the Masirah ophiolite was as a young ridge formed during the rifting and separation of eastern and western Gondwana, whereas the emerging evidence indicates that the Oman ophiolite was formed at a fast-spreading centre associated with subduction initiation in a supra-subduction environment.

The stratigraphic history of the Masirah ophiolite shows that it remained a part of the oceanic crust for ca. 80 Ma prior to obduction. Its condensed stratigraphy is therefore best understood as forming at a young mid-ocean ridge in which the mantle potential temperature was reduced because of the proximity of nearby continental crust and the associated subcontinental mantle. The chemical variability and enrichment of the Masirah lava source predates the formation of the ophiolite and is similar to the chemical variability found elsewhere in Indian Ocean basalts. It may therefore reflect a similar provenance for the Masirah basalts to those in the Indian Ocean.

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