Lithospheric petrology of the eastern Arabian Plate: Constraints from Al-Ashkhara (Oman) xenoliths

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Mafic granulate and spinel lherzolite xenoliths from Cenozoic alkaline basalts near Al-Ashkhara, eastern Oman, have been selected for a systematic mineralogical, geochemical and Sr-Nd-Pb isotopic study. This is the only place in E Arabia where samples of both lower crust and upper mantle can be examined. Lower crustal xenoliths consist of two mineralogically and chemically distinct groups: gabbronorite (subequal abundances of ortho- and clino-pyroxene and plagioclase) and plagioclase pyroxenite (dominant pyroxene and subordinate plagioclase). Temperature estimates for lower crustal xenoliths using the two pyroxene geothermometer (T-Wells) yield 810–865 °C. The mineral assemblage (spinel–pyroxene–plagioclase) and Al content in pyroxene indicate that plagioclase-bearing xenoliths equilibrated at 5–8 kbar (13 and 30 km depth) in the lower crust. εNd and 87Sr/86Sr calculated at 700 Ma for Al-Ashkhara lower crustal xenoliths (+6.4 to +6.6; 87Sr/86Sr = 0.7028 to 0.7039) are consistent with the interpretation that juvenile, mafic melts were added to the lower crust during Neoproterozoic time and that there was no discernible contribution from pre-Neoproterozoic crust. Upper mantle xenoliths consist of both dry and hydrous (phlogopite-bearing) lherzolites. These peridotites are more Fe-rich than expected for primitive mantle or melt residues and probably formed by pervasive circulation of melts that have refertilized pre-existing mantle peridotites. Mineral equilibration temperatures range from 990 to 1070 °C. Isotopic compositions calculated at 700 Ma are εNd = +6.8 to +7.8 and 87Sr/86Sr = 0.7016 to 0.7025, indicating depleted upper mantle. Pb isotopic compositions indicate that the metasomatism was relatively recent, perhaps related to Paleogene tectonics and basanite igneous activity. Nd model ages for the spinel peridotite xenoliths range between 0.59 and 0.65 Ga. The xenolith data suggest that eastern Arabian lower crust is of hotspot origin, in contrast to western Arabian lower crust, which mostly formed at a convergent plate margin. Geochemical and isotopic differences between lower crust and upper mantle indicate that these are unrelated, possibly because delamination replaced the E Arabian mantle root in Neoproterozoic time.

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1. Introduction

We understand the principal aspects of how Earth’s upper continental crust formed from studying rock exposures, but these samples represent a trivial fraction of Earth’s mass. Deeper understanding of the silicate Earth depends on also understanding the lower continental crust and upper mantle. Because these last two silicate layers are mostly inaccessible, geoscientists must integrate information from large scale geophysical probing with petrologic investigations of lower crust and upper mantle samples. Petrologic investigations focus on samples obtained via two major approaches: exposures resulting from tectonic processes, such as ophiolitic peridotites or peridotite exposures at slow-spreading ridges and study of exhumed ancient granulate terrains (Kempton et al., 1995; Liu et al., 2001) and xenoliths brought to the surface by volcanic eruptions (Griffin and O’Reilly, 1987; Rudnick et al., 1986; Zhou et al., 2002).

Intrusive xenoliths provide key constraints about mineral, chemical and isotopic composition and thermal state of the upper mantle (e.g. Eggers et al., 1998; Griffin and O’Reilly, 1987; Griffin et al., 1999; Kempton et al., 1997; Rudnick, 1992). Similar information about the lower continental crust has also been obtained from xenoliths (Dostal et al., 1980; Downes et al., 1990; Griffin and O’Reilly, 1987; Griffin et al., 1999; Kempton et al., 1997; Rudnick, 1992). Such xenoliths which commonly occur in young volcanic areas that lack nearby outcrops of high-grade lower crustal rocks, permit comparisons to be made between petrological and chemical results of xenolith studies and models based on seismic data (Griffin and O’Reilly, 1987; Loock et al., 1990). Such xenoliths also preserve the fabric and mineral assemblages produced at the ambient pressure and temperature conditions at the site of their formation.

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formation in the lithosphere (Griffin and O’Reilly, 1987; Satsukawa et al., 2010).

The continental lithosphere of Arabia is a good place to reconstruct the processes that generated and/or modified continental crust and upper mantle during Neoproterozoic time. The nature of the upper mantle and lower continental crust of western Arabia is reasonably well known, because of some seismic refraction studies and an abundance of xenolith-bearing basaltic flows. Cenozoic alkali basaltic lavas in the northern, southern and western parts of the Arabian Plate contain a wide variety of crustal and mantle xenoliths (Abu-Aljarayesh et al., 1993; Coleman, 1993; Ghent et al., 1980; Grégoire et al., 2009; Henjes-Kunst et al., 1990; Kuo and Essene, 1986; McGuire, 1988a,b; McGuire and Stern, 1993; Nasir, 1990, 1994, 1995, 1996; Nasir and Al-Fuqha, 1988; Nasir and Safarjalani, 2000; Nasir et al., 1992, 1993; Nasir et al., 2006; Stern and Johnson, 2010; Stein and Katz, 1989; Stein et al., 1993). In contrast, there is little Cenozoic basalt in eastern Arabia and only few localities contain such xenoliths. However, peridotitic upper mantle xenoliths have been recently discovered in Cenozoic alkaline dykes and flows emplaced along the Oman passive margin facing the Owen Basin (Fig. 1A,B). These basalts and associate upper mantle xenoliths have been studied for their bulk rock and mineral trace element compositions (Gnos and Peters, 2003; Grégoire et al., 2009; Nasir et al., 2006). A recently-discovered Eocene basanite plug near Al-Ashkhara (Fig. 1A,B) contains abundant upper mantle and lower crustal xenoliths, which provide a unique opportunity to explore the composition and nature of eastern Arabian lithosphere to compare with the better-documented lithosphere beneath western and northern Arabia. The question addressed here concerns the nature and evolution of the lithosphere beneath eastern Arabia. We present new petrologic, geochemical, and isotopic data for xenoliths from the Al-Ashkhara plug and compare our results to those for mantle peridotite and lower crustal xenoliths from western Arabia.

2. Geologic setting

The continental Arabian Plate is dominated by Neoproterozoic (1000–543 Ma) crust, although small tracts of older crust are preserved.

Fig. 1. (A) Locality map showing location of the studied Al-Ashkhara (Oman) xenoliths. (B) Xenolith location of studies by Gnos and Peters (2003), Grégoire et al. (2009) and Nasir et al. (2006).
in SW Arabia and Yemen (Stern and Johnson, 2010). Neoproterozoic crust is exposed in the Arabian Shield, which dominates western Arabia, but is deeply buried beneath Phanerozoic sedimentary cover in the east. Eastern and western Arabia both are dominated by Neoproterozoic rocks as old as ~880 Ma but E Arabia has no late Cryogenian and Ediacaran igneous rocks (700–543 Ma) whereas W Arabia has these in abundance, testifying to the stabilization of E Arabian lithosphere in Late Neoproterozoic time. The lithosphere of E Arabia is thicker than that beneath W Arabia, as demonstrated by geophysical studies (Chang et al., 2011; Hansen et al., 2007; Park et al., 2008), and by the fact that this region acted as a shallow marine platform throughout Phanerozoic time (Ziegler, 2001). In contrast, W Arabia has mostly been above sealevel through the Phanerozoic. The Arabian Shield has been relatively stable since late Neoproterozoic (Ediacaran) time, although it was subsequently rejuvenated more than once, the last time during opening of the Red Sea in the last 30 Ma. The crust of Arabia mostly formed as juvenile arcs that were fused together prior to continental collision between E and W Arabia ~600 Ma (Stern and Johnson,
that the host basanites were emplaced 35.3 Ma (Fournier et al., 2008). Neoproterozoic processes ultimately produced a 40-km thick continental crust (Gettings et al., 1986).

W Arabia hosts several Neogene basalt fields locally known as harrats. Most harrats of the Arabian Plate are located on or north of the Arabian Shield. Arabian harrats form a voluminous and widespread volcanic province that extends from Yemen to Turkey, covering about 80,000 km² of western of Saudi Arabia, Jordan and Syria (Coleman, 1993). Many of these harrats contain abundant spinel peridotite (mantle) and gabbroic (lower crustal) xenoliths (Stern and Johnson, 2010).

In contrast to the abundant xenolith-bearing lavas of Western Arabia, few such lavas are found in eastern Arabia; however Cenozoic magmas intruded and erupted through the Batain Melange of eastern Oman. The Batain Melange resulted from the obduction of the Semail ophiolite during late Maastrichtian—Early Tertiary times and from regional uplift that started during the Miocene (Peters and Mercolli, 1998). Regional geological studies have shown that long periods of transtension alternated with shorter transpressional episodes along the E. Oman margin (e.g. Platel and Roger, 1989). The best documented transpressional event occurred during upper Maastrichtian—lower Paleocene times. Since Eocene times, southern Oman has been a passive margin bordering the Gulf of Aden (Platel and Roger, 1989). The opening of the Oman Basin may have complicated the evolution of this margin, although the age of this oceanic crust is still debated (Fournier et al., 2008). Alkaline magmatism affecting the East Arabian margin belonged to two main episodes: a first one contemporaneous with Cretaceous—Tertiary transpression, and more recent dyke injections during Eocene extension.

Clastic and mantle xenoliths studied here were sampled from a large basanite plug near Al-Ashkhara (Fig. 1A,B). 40Ar/39Ar ages of 36 to 40 Ma are reported for Al Ashkhara area Cenozoic magmatic rocks (Worthing and Wilde, 2002). K–Ar whole rock dating indicates that the host basanites were emplaced 35.3 ± 1.5 Ma, at the end of the Eocene time (Electronic appendix 1).

The basanite is a large elliptical plug (250×150 m) with the long axis oriented NE–SW; this was intruded by a smaller plug (20×10 m) (Fig. 2). Both plugs intrude thrust sheets dominated by sandy carbonate rocks of the Triassic Sal Formation. A chilled margin (~20 cm thick) is observed along the contact with the limestone. The basanite contains up to 16% normative nepheline and up to 24% normative olivine. Mg# (＝100Mg/Mg + Fe) varies between 56 and 61 (Electronic appendix 1). The two plugs consist of breccia and diabase which contain abundant mantle and crustal xenoliths. Basanite in the large plug is brecciated, fine-grained, vesicular, and porphyritic – sometimes piloaxitic and glomero-porphyrctic – with ~30% olivine, plagioclase, and clinopyroxene phenocrysts set in a groundmass of plagioclase, clinopyroxene, olivine, phlogopite, analcite and opaque phases. Olivine and clinopyroxene phenocrysts are weakly zoned, with resorbed rims. Xenocrysts of olivine and clinopyroxene derived from mantle peridotite are common. The small plug is diabase displaying porphyritic, holocrystalline, and intergranular textures. Clinopyroxysts (5% vol.) of olivine, plagioclase and clinopyroxene are set in a seriate, fine-grained groundmass of olivine, clinopyroxene, plagioclase, Ti-magnetite, and trace apatite. Red-brown biotite is associated with rims of many olivine and clinopyroxene phenocrysts as well as with grains of Ti-magnetite. The lithologies of the xenoliths in both plugs are the same, comprising mainly peridotite and gabbro. Granitic, metamorphic and sedimentary xenoliths are more common in the large plug. Xenoliths in the small plug are usually less than 10 cm in diameter, whereas the large plug contains abundant brecciated peridotite xenoliths up to 50 cm across (Fig. 3).

3. Xenolith petrography

Xenoliths entrained in Al-Ashkhara basanites can be divided into three groups: (I) ultramafic (upper mantle) xenoliths; (II) mafic (lower crust) xenoliths; and (III) granitic and metasedimentary (upper crust) xenoliths. Only mantle and lower crust xenoliths were studied here.

3.1. Ultramafic xenoliths

Ultramafic xenoliths are the most abundant variety and range between 4 and 50 cm in size. Large xenoliths (~20×50 cm) are usually brecciated and affected by secondary calcite veins (Fig. 3). Smaller xenoliths (~10 cm) are usually fresh. Textural features, modal mineralogy (calculated by a least-squares method using bulk xenolith chemical analyses and constituent mineral analyses) and classification of the xenoliths are listed in Table 1. Ultramafic xenoliths are predominantly dry and hydrous spinel lherzolites, although spinel harzburgite, dunite and wehrlite are also reported from nearby Cenozoic basalts in the Al-Ashkhara area (Gnos and Peters, 2003; Grégoire et al., 2009; Nasir et al., 2006) and all are Cr-diopside group. The mineralogical composition closely resembles the average calculated by McDonough (1990) for world-wide spinel lherzolites. However, most lherzolites have relatively low olivine contents (Table 1), between 40 and 57%, and tend toward websterites (e.g. sample S55, Table 1). In general, textural variations are restricted to granuloblastic and porphyroclastic types. Xenoliths with porphyroelastic textures contain phlogopite, whereas large xenoliths with granuloblastic texture are anhydrous. All minerals in the granuloblastic peridotites belong to the same generation and are recrystallized (Fig. 4A). They display equilibrated texture with ~120° triple junctions and curved...
Table 1
Modal mineralogy of xenoliths from eastern Oman.

A—Mantle xenoliths

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</table>

Cpx: clinopyroxene, Opx: orthopyroxene; DL: dry lherzolite; HL: hydrous lherzolite; Gn: gabbronorite; Ppyxt: plagioclase pyroxenite.

Fig. 4. Photomicrographs of Al-Ashkhara xenoliths, all using cross-polarized light. (A) Granoblastic peridotite (Sample S26); (B) Porphyroclastic peridotite (Sample S33); (C) Porphyroclastic peridotite (Sample S55); (D) Anorthositic (Group A) lower crust granulite (Sample KG1); (E) Anorthositic (Group A) lower crust granulite (Sample KG43); (F) Pyroxene-rich (Group B) lower crust granulite (sample SG20). Ol: olivine; Cpx: clinopyroxene, Opx: orthopyroxene, sp: spinel; plag: plagioclase, serp: serpentine.
grain boundaries. Grains are relatively small (0.2 to 0.5 mm). Some deformed grains show undulatory extinction and kink bands. Porphyroclastic peridotites contain porphyroclasts that are generally 2–4 mm across, with 1–2 mm neoblasts (Fig. 4B, C). Orthopyroxene and olivine crystals are typically altered and contain opaque inclusions. The alteration products are mainly chlorite, iddingsite and serpentine, suggesting that they are mostly derived from olivine and orthopyroxene (Electronic appendix 2). Porphyroclastic texture is observed in the hydrous peridotite samples as manifested by kink bands in olivine and rare aggregates of olivine crystals that were likely produced by recrystallization of larger grains. Phlogopite makes up 1–2% of fresh small porphyroplastic xenoliths. It occurs as small irregular grains (<2 mm in length) in close association with olivine and spinel. Few phlogopite grains show undulatory extinction and kink banding. Interstitial lobate brown spinel is common in all peridotite xenoliths.

3.2. Mafic lower crustal xenoliths

Mafic xenoliths are subangular to round and 2 to 10 cm across. These xenoliths consist mainly of variable amounts of clinopyroxene, orthopyroxene, plagioclase and spinel and display cumulate igneous textures. The xenoliths are subdivided based on their mineralogy into two groups: plagioclase-rich gabbronite (Group A; Fig. 4D,E) and plagioclase-poor pyroxenite (Group B; Fig. 4F). Mineral assemblage and estimated modal mineral proportions of these mafic xenoliths are listed in Table 1. In thin section the xenoliths are medium to coarse-grained (1–3 mm) and reasonably fresh, with granoblastic fabrics or near-equigranular textures. Most crystals show strain shadows. Pyroxene is usually smaller than plagioclase. Both pyroxenes (Opx, Cpx) have exsolution lamellae of the complementary pyroxene. Green spinel occurs in all xenoliths, mainly as individual grains associated with pyroxene. Accessory minerals are titanomagnetite, apatite, and titanite. Secondary minerals include Mg-rich chlorite, calcite and calcinite.

Gabbronite xenoliths are mainly composed of cumulus plagioclase and pyroxene (both Cpx and Opx), and are similar in texture and mineral composition to plagioclase-rich xenoliths from western Arabia. Cumulate pyroxenes form dark concentrations and bands in plagioclase peridotite xenoliths. Cumulate textures without any evidence of recrystallization indicate the magmatic nature of both kinds of xenoliths. Plagioclase pyroxenite xenoliths are more enriched in pyroxene (both Opx and Cpx) compared with lower crustal xenoliths from other Arabian Plate localities. In all xenoliths, Opx is more or less altered to chlorite and serpentine.

4. Analytical methods

4.1. Mineral analyses

Mineral analyses were carried out utilizing a CAMECA SX100 microprobe with five wavelength-dispersive spectrometers at the Institute of Mineralogy, Universität Stuttgart, Germany. The analytical conditions were an acceleration voltage of 15 kV, a beam current of 20 nA, and a peak integration time of 20 s. Data were corrected using the PAP procedure (Pouchou and Pichoir, 1985). Precision for major and minor elements is estimated to be 1–2% and 5%, respectively. At least three determinations were averaged for each analysis quoted. Fe**3+** was calculated from charge balance, assuming ideal site occupancy.

4.2. Bulk-rock major, trace element and isotopic compositions

Sample whole rock powders were prepared in an agate planetary ball mill. The samples were analyzed for major and trace element concentrations, loss on ignition (LOI) and Pb, Sr, and Nd isotopic compositions at the Activation Laboratories Ltd., Ancaster, Canada. Major element compositions were determined using fused samples analyzed with ICP/OES. Selected trace and rare earth elements (REE) were analyzed by Perkin Elmer Sciex Elan 6100 ICP-MS after lithium metaborate/tetraborate fusion-digestion. Several standards were included in each batch to check the accuracy of analyses (Electronic appendix 3 shows the quality control). Three blanks and five controls (three before sample group and two after) were analyzed per group of samples. Duplicates were fused and analyzed every 15 samples. An estimate of the precision for the ICP analyses 1-sigma errors for major elements is <1% relative and the minor elements ±2% or better (Electronic appendix 3). Powdered samples were weighed to obtain approximately 100 to 200 ng of Sr, Nd and Pb. A leaching step with 6 N HCl during 30 min at 85 °C was done before acid digestion. Samples were dissolved during 36–48 h on a hot plate with a mixture of HF and HNO₃. After evaporation to dryness, 1 ml of HNO₃ was added to the residue and kept at about 90 °C for 12–24 h.

Strontium isotopes were separated using Sr Eichrom resin. Analyses were performed on a Triton-MC mass-spectrometer (MC-TIMS). Total blanks are 0.3–0.7 ng for Sr. Sr isotopic ratios were fractionation corrected to ⁸⁷Sr/⁸⁶Sr = 0.7285. During the period of work the weighted average of 15 SRM-987 Sr-standard runs was 0.71024 ± 2(2σ) for ⁸⁷Sr/⁸⁶Sr.

Neodymium isotopes were separated using AG50WX12 cation exchange resin followed by two steps of Nd purification using the HDEHP technique. Total blanks for Nd are 0.1–0.5 ng. Analyses were performed on a Triton multicollector mass-spectrometer (MC-TIMS). Nd isotopic compositions are normalized to ¹⁴⁶Nd/¹⁴⁴Nd = 0.7219 and are relative to ¹⁴⁶Nd/¹⁴⁴Nd = 0.511860 for the La Jolla standard. During the period of work the weighted average of 10 La Jolla Nd-standard runs yielded 0.511872 ± 15(2σ) for ¹⁴²Nd/¹⁴⁴Nd. Measurement precisions for individual runs commonly are better than 0.00001.

For Pb separation, after complete evaporation, 0.5 ml of 8 N HBr was added to the sample and kept at 70 °C for 2–3 h before evaporation. Sample powders were leached twice (6 N tridistilled HCl) during 30 min at 85 °C before acid digestion (36–48 h in ultrapure HF and HNO₃). The chemical separation of Pb was done using 50 μl of anion exchange resin (Biorad AG1X8, 200–400 mesh) with samples being loaded and washed in 0.5 N HBr. Lead was then eluted in 6 N HCl. Pb-isotope analyses were performed on a Thermo Finnigan-High Resolution ICP/MS. Total procedural blanks are less than 30 pg Pb and are considered negligible. External reproducibility of Pb isotope analysis of BCR-2 performed during the time of this work are: 206Pb/204Pb = 0.2%, 207Pb/204Pb = 0.2%, 208Pb/204Pb = 0.2% — at the 2σ level.

5. Results

5.1. Mineral chemistry

Chemical compositions of the major silicate minerals are listed in (Electronic appendix 5–7). Primary mineral compositions in individual xenoliths are nearly homogeneous, except for minor zoning of Al, Cr, Ca and Na in pyroxenes.

5.1.1. Olivine

Olivine is only found in the peridotites. This has relatively low Fo contents (85–86%), compared to olivine from peridotite xenoliths elsewhere in the Al Ashkhara area (Fo 90–93; Gnos and Peters, 2003; Grégoire et al., 2009; Nasir et al., 2006; Electronic appendix 4) as well as from western Arabia (Fo 90–93%; McGuire, 1988a, 1988b; Nasir, 1990; Nasir and Saffarjani, 2000). This is also lower than expected for mantle olivines, which are typically Fo 89–93. The CaO content in Al-Ashkhara peridotite olivines is low: 0.06 to 0.09%. NiO content reaches 0.39% (Electronic appendix 5).
5.1.3. Clinopyroxene

Clinopyroxenes from both dry and hydrous spinel lherzolites are aluminous diopside (Morimoto, 1988) with Mg# = 88–91 (Electronic appendix 6) that correlates with Mg# of co-existing Opx. Cpx from the hydrous lherzolites is distinguished from those in dry lherzolites by low Cr2O3 contents (0.07 to 0.48%) than Opx in hydrous lherzolite (4.0–4.2%). CaO varies from 0.79 to 0.9%. No significant compositional variation between large porphyroclasts and fine-grained Opx is observed.

Gabbroic xenolith Opx is enstatite with Mg# = 78 in plagioclase pyroxenite and 79–80 in gabbroinite. Gabbroic Opx has high Al2O3 (4.3–5.0%) and low CaO contents (0.29–0.45%).

5.1.3. Clinopyroxene

Clinopyroxenes from both dry and hydrous spinel lherzolites are aluminous diopside (Morimoto, 1988) with Mg# = 88–91 (Electronic appendix 6) that correlates with Mg# of co-existing Opx. Cpx from the hydrous lherzolites is distinguished from those in dry lherzolites in having lower Mg# (88–89 vs. 91) as well as lower Al2O3 (5.15–5.27% vs. 6.1–6.2%) and Cr contents (0.7–0.72% vs. 0.84–0.87%).

Cpx in lower crustal mafic xenoliths is diopside with Wo 48–50, En 42–43, Fs 7–9 (Morimoto, 1988), Mg# = 82–85, 0.48–0.89% TiO2, and 0.85–0.98% Na2O. High Al2O3 content (6.0–7.8%) is consistent with high Mg# in associated Opx. Lower crust Cpx is characterized by low Cr2O3 contents (0.07–0.48%), typical for lower crust xenoliths (Rudnick, 1992). Like Opx, Cpx in pyroxenites has higher Mg# (~85) and Cr# (11–22) than in gabbronorites (Mg# = 82–83, Cr2O3 = 0.07–0.1%). Pyroxenes from gabbroic xenoliths are compositionally similar to lower crustal xenoliths of western Arabia (Stern and Johnson, 2010).

5.1.4. Spinel

Peridotite xenoliths contain brown Cr-spinel with Mg# between 73 and 78, and Cr# (100Cr/Cr + Al) between 11 and 22 (Electronic appendix 5) and 0.12 to 0.27% NiO. Green Al-spinel is common in the gabbroic xenoliths. It has low Cr# (1–5 in gabbroinite and 7–9 in plagioclase pyroxenite xenoliths). Mg# varies between 49 and 66.

5.1.5. Phlogopite

Phlogopites in hydrous peridotite xenoliths have high TiO2 (4.8–7.8%), FeO (17.8–18.7%) and BaO (0.45–2.63%) contents and low Mg# (42–55) (Electronic appendix 6).

5.1.6. Plagioclase

Plagioclase is mostly labradorite (An 57–68) in gabbroinite and labradorite–bytownite (An 66–74) in the pyroxenite (Electronic appendix 7). Grains are generally homogeneous; however, minor normal zoning (2–5 mol% An) was detected in a few samples. Plagioclase is usually altered to analcrite.

6. Geothermobarometry

Temperatures of equilibration have been calculated for all xenoliths (Electronic appendix 8) using the two-pyroxene calibrations of Brey and Koehler (1990a, 1990b) and Wells (1977) and the Ca-in-orthopyroxene thermometer of Brey and Koehler (1990a, 1990b) assuming P = 15 kbar. Temperatures of spinel lherzolites range from 965 to 1025 °C using the two-pyroxene thermometers and from 990 to 1025 °C using the Ca-in-orthopyroxene thermometer. These temperature estimates are mostly higher than those for plagioclase pyroxenites which range between 615–855 °C using the two pyroxene geothermometer and 765–835 °C using the Ca-in-orthopyroxene geothermometer.

Pressure determinations for Al-Ashkhara mafic and ultramafic xenoliths are difficult because these lack garnet. However, approximate pressure for Al-Ashkhara lherzolites may be estimated by comparing observed mineral assemblages to mineral stability fields determined for the spinel- to garnet-peridotite transition (Green and Hibberson, 1970; O’Hara et al., 1971), and from phase diagrams for the gabbro to eclogite transition (Green and Ringwood, 1967; Herzberg, 1978; Irving, 1974; Ito and Kennedy, 1971; Wood, 1987). Such considerations establish lower and upper pressure limits for the depth sampled by the xenoliths. The mineral assemblage of the Al-Ashkhara spinel peridotite xenoliths indicates a pressure range of 12–16 kbar (40–55 km depth) if the FeO contents of coexisting olivine (Fo50) are accounted for (O’Neill, 1981). This corresponds to the uppermost part of the subcontinental lithospheric mantle.

The absence of olivine in the gabbroic xenoliths suggests pressure in excess of 6 kbar at ~800 °C and 8 kbar at ~900 °C (Wood, 1987). Thus, it is reasonable to infer that minimum pressures of 6 to 8 kbar, equivalent to 20 to 28 km depth, prevailed during crystallization of the gabbroic bodies sampled by these xenoliths. The absence of garnet further suggests pressure was less than 12–14 kbar at 900 °C (Gasparik, 1984; Irving, 1974). The presence of aluminous pyroxenes is consistent with such high pressures. Most Al-Ashkhara gabbroic Cpx have high Al2O3 (6.0–7.8%) and Na2O contents (0.85–0.98%) and high Al1O3 values (0.1–0.14), further indicating pressures compatible with a lower crustal origin. Pressure estimates for gabbroic xenoliths can also be obtained using the clinopyroxene structural geobarometer of Nimis (1995). Pressure estimates using the anhydrous equation of Nimis (1995) for a temperature of 800 °C range between 5.2 and 8.3 kbar (Electronic appendix 8). There is no systematic difference between pyroxene and gabbroinite, although the former cluster more tightly (7.1–7.7 vs. 5.2–8.3 kbar). These results compare well with pressures estimated by comparing gabbroic xenolith mineral assemblages and phase stability fields related to the gabbro to eclogite transition. The P–T estimates and petrographic features indicate that these mafic xenoliths equilibrated in the lower crust (20–30 km). This corresponds to the granulite facies although the rocks themselves show igneous, not metamorphic textures.

P–T estimates for Al-Ashkhara gabbroic and peridotite xenoliths are plotted in Fig. 5, which suggests that the geotherm beneath E Arabia might have been significantly cooler than that beneath W. Arabia.

7. Whole-rock geochemistry

7.1. Ultramafic xenoliths

Major and trace element compositions of Al-Ashkhara xenoliths are given in Table 2. Anhydrous lherzolites are distinguished from hydrous Cpx-rich lherzolites by their higher MgO contents (35–40 vs. 33–34%), coupled with lower concentrations of Al2O3 (2.3–2.5% vs. 2.6–2.8%), alkalis (Na2O + K2O = 0.2–0.3%) and TiO2 (0.04–0.06 vs. 0.11–0.18%). Mg#s of both lherzolite types fall in a narrow range (86–88) that is slightly lower than Mg# 89 expected for the primitive mantle (Lyubetskaya and Korenaga, 2007). The lherzolites are mostly enriched in CaO and Al2O3 relative to global abyssal peridotites and peridotite xenoliths from W Arabia (Fig. 6), and also plot off the olivine-spinel mantle array (OSMA), which defines compositions of mantle peridotite residues (Fig. 7). The altered nature of both lherzolite varieties is demonstrated by loss on ignition (LOI), reflecting calcite and serpentine breakdown (Table 2). In spite of the fact...
that the relative abundances of the major elements are comparable with that of world-wide spinel lherzolites (Maaløe and Aoki, 1977; McDonough, 1990; Nickel and Green, 1984), the studied mantle xenoliths are relatively iron-rich by comparison to typical mantle lherzolites.

Peridotite whole-rock trace element data broadly agree with global mantle xenolith suites (Jagoutz et al., 1979; Jochum et al., 1989). Phlogopite-bearing lherzolites (Table 2) are enriched in elements related to this phase (Ti, K, Na, Rb, Sr, etc.). Hydrous Cpx-rich lherzolites are enriched in Ba, Rb, Sr, Zr, Nb, and LREE. They are slightly enriched in U and Th compared with anhydrous lherzolites. In primitive mantle-normalized trace element diagrams, lherzolites show negative Rb, K, La and Ti and positive Th, Nb, Ce, Sr and Zr anomalies (Fig. 8). Rare earth element (REE) contents of spinel lherzolite xenoliths; (La/Yb)n varies from 4.4 to 6. Hydrous Cpx-rich lherzolite xenoliths have higher REE contents than anhydrous spinel lherzolite xenoliths have higher REE contents than anhydrous spinel (5.6–6.8) garnet pyroxenites of western Arabia (Chang et al., 2011; Stern and Johnson, 2004; Al-Mishwatt and Nasir, 2004) (Fig. 12).

7.2. Mafic lower crustal xenoliths

All Al-Askkhara lower crustal xenoliths are mafic, with SiO2 ranging from 46.1 to 48.9% (Table 2), thus falling in the range of lower crustal xenolith compositions from Saudi Arabia, Syria, and Jordan (43 to 57% SiO2; Al-Mishwatt and Nasir, 2004; Al-Safarjalani et al., 2009; Förster et al., 2010; McGuire and Stern, 1993; Stern and Johnson, 2010). Mg# ranges from 74 to 79. Gabbronorite xenoliths have higher Al2O3 (15.6–18.9%), Na2O (1.9–2.49%) and lower total FeO (5.6–7.1 wt.%), MgO (9.5–12.8%), CaO (10.5–12.8%) and TiO2 (0.28–0.43) than pyroxenites (8.55–9.46% Al2O3, 0.62–0.81% Na2O, 8.14–9.21% total FeO, 16.89–17.92% MgO, 13.53–14.11% CaO and 0.48–0.62% TiO2), reflecting their respective mineralogies. Also reflecting their mineralogies, pyroxenites contain much higher Cr (2880–2995 ppm) and Ni (710–810 ppm) abundances than do gabbronorites (330–580 ppm Cr; 130–220 ppm Ni). Pyroxenites also contain significantly more Sc (53–63 ppm) and Y (16–18 ppm) than do gabbronorites (26–47 ppm Sc, 6–13 ppm Y; Table 2).

Lower crustal xenolith REE contents are 2–7 times primitive mantle abundances. Most have similar mantle-normalized trace element patterns (except for Ba and Th), characterized by enrichments in Sr, Rb, K, Hf, and Sm (Fig. 10). They are relatively depleted in Ti, Zr, Nd and La. Gabbronorite xenoliths are depleted in Th whereas plagioclase pyroxenite xenoliths are depleted in Ba relative to Sr and Rb. All of the mafic xenoliths contain very low REE (ΣREE = 14–30 ppm). The whole-rock chondrite-normalized REE patterns of all samples are similarly flat ((La/Sm)n = 0.4–2.0; and (La/Yb)n = 0.34–2.2) (Fig. 11), characterized by a relatively strong positive Eu anomaly (Eu/Eu* = 5.6–7.6). Pyroxenites are depleted in LREE relative to the heavy REE (HREE) and middle REE (MREE), and contrast with other Arabian lower crust xenoliths which have LREE-enriched patterns (Al-Mishwatt and Nasir, 2004) (Fig. 12).

7.3. Isotopic compositions

Nd, Sr, and Pb isotope data are listed in Table 3. Isotopic compositions need to be corrected for in situ radiogenic growth, which is done for all samples for Sr and Nd and for those samples with U, Th, and Pb concentrations for Pb. We do not know when peridotite and lower crust formed, but upper crust of E. Arabia formed ~700–800 Ma (Stern and Johnson, 2010) so we correct to 700 Ma. Lherzolite and lower crustal xenoliths are distinctly in terms of parent/daughter relationships, with peridotites being relatively enriched in the more incompatible elements. This is clear for 87Sr/86Sr (lherzolites = 0.707–0.715; lower crust = 0.703–0.707) at 700 Ma. Lherzolite and lower crustal xenoliths are enriched in Ba, Rb, Sr, Zr, Nb, and LREE. They are characterized by relatively strong positive Eu anomalies (Eu/Eu* = 5.6–7.6). Pyroxenites are depleted in LREE relative to the heavy REE (HREE) and middle REE (MREE), and contrast with other Arabian lower crust xenoliths which have LREE-enriched patterns (Al-Mishwatt and Nasir, 2004) (Fig. 12).

8. Discussion

Omani xenoliths in general are of interest because these provide a chance to study the poorly-known lithosphere of eastern Arabia, which differs significantly from that of western Arabia (Chang et al., 2011;
There is no obvious difference in crustal thickness between these changes; coincides with a prominent magnetic anomaly (Central Arabia from which samples of the lower crust are available for study. The xenoliths from Al-Ashkhara that we studied are of particular interest because this is the only site known in eastern Arabia for which samples of the lower crust are available for study. The crust of both E and W Arabia contains significant proportions of Cenozoic crust and igneous and metamorphic activity in E Arabia was largely finished by 750 Ma, in contrast to W Arabia where deformation and magmatism continued for another 200 m.y. Following stabilization, E Arabia became the site of virtually continuous sedimentation from 725 Ma on and into Phanerozoic time. This implies that strong lithosphere was in place beneath eastern Arabia by 700 Ma in contrast to lithospheric instability that persisted until ~550 Ma in the west. Lithospheric differentiation is further indicated by Phanerozoic sedimentation patterns with steady subsidence and accumulation of 5–14 km thick sediments in the east (Konert et al., 2001) and a consistent high-standing and thin to no Phanerozoic accumulation over the Arabian Shield in the west (Stern and Johnson, 2010). There is no obvious difference in crustal thickness between E and W Arabia but the lithospheric mantle thickens from ~80 km beneath the Shield to ~120 km beneath E Arabia. The locus of these changes coincides with a prominent magnetic anomaly (Central Arabian Magnetic Anomaly or CAMA). CAMA can be traced from the Al Amar suture in the extreme eastern part of the Arabian Shield north across north-central Arabia (Stern and Johnson, 2010). Thus, a first-order motivation for studying Oman xenoliths is to compare their compositions to those of much more abundant and well-studied xenoliths of W Arabia for the purpose of determining if these can reveal any differences between the composition of E and W Arabian lithospheres. We explore these similarities and differences below. It is noteworthy that the Al-Ashkhara locality is the only place in E Arabia where lower crustal xenoliths are reported, so generalizations about E Arabia are tentative. Below, we first consider the lower crustal xenoliths before considering the mantle xenoliths. Finally, we consider the relationship between lower crust and upper mantle beneath Al-Ashkhara, as represented by these xenoliths.

8.1. Lower crust xenoliths

A division of the Arabian xenoliths into lower crustal and upper-mantle xenoliths reflects whether or not plagioclase is present (presence...
of plagioclase indicates lower-crustal rock), but a further distinction can be made using whole-rock compositions and the T conditions that the xenoliths record. Lower-crustal xenoliths generally have lower Mg# (100Mg/Mg+Fe), Ni, and Cr contents, and yield lower mineral equilibrium temperatures (~700–900 °C) than do upper-mantle xenoliths (~800–1050 °C; Nasir and Safarjalani, 2000; Stern and Johnson, 2010).

Omani lower crustal xenoliths are undeformed mafic cumulates (Fig. 4) that are divisible into pyroxene-rich and plagioclase-rich varieties, similar to those from western Arabia (Al-Mishwat and Nasir, 2004; Nasir, 1995). Their origin as mafic cumulates explains the limited variations in their whole-rock SiO2 (46–49%) which is accompanied by large variations in Al2O3 (8.5–18.9%) and MgO (9.5–17.9%) as well as positive anomalies of Eu and Sr in chondrite-normalized REE patterns (Fig. 11) and on primitive mantle-normalized spidergrams (Fig. 10). These chemical variations reflect variable abundances in modal plagioclase and clinopyroxene (up to 60% and 40%, respectively). The origin of Omani lower crustal xenoliths as mafic cumulates rather than residues after melting is similar to the inferred origin of W. Arabian lower crustal xenoliths (e.g., Al-Mishwat and Nasir, 2004; Krienitz and Haase, 2010; McGuire and Stern, 1993; Nasir, 1995). We know of no lower crustal xenoliths from Arabia that are examples of felsic residues after melting that prevail in the lower crust beneath other continental regions (Rudnick and Gao, 2003) or felsic relaminates as recently proposed by Hacker et al. (2011).

Al-Ashkhara pyroxene-rich xenoliths have labradoritic–bytownitic plagioclase (mean~An70) and relatively Mg-rich pyroxenes (mean OPX~En79; mean CPX~Wo49 En42 Fs8) indicating that these crystallized from moderately fractionated mafic magmas. Plagioclase-rich varieties are slightly more fractionated, with labradoritic–bytownitic plagioclase (mean ~An63) and slightly more Fe-rich pyroxene (mean OPX ~En77; mean CPX ~Wo49 En42 Fs9). This indicates a compositional continuum between the two lower crust xenolith suites, from the more primitive pyroxene-rich to the more fractionated plagioclase-rich lower-crustal suites. We speculate that density variations among these lithologies may reflect plagioclase-rich lower-crustal types concentrated in the shallower lower crust, pyroxene-rich lower-crustal types concentrated in the deeper part of the lower crust.

The lower crusts of E. Arabia (as sampled at Al-Ashkhara) and that of Arabia west of CAMA (as sampled at many localities, summarized by Stern and Johnson, 2010) are compositionally distinct, as summarized in Electronic appendix 9. Both are mafic but Omani

Fig. 6. Comparisons of whole-rock abundances of CaO and Al2O3 for Al-Ashkhara, other Omani, and W. Arabian mantle xenoliths. PM = composition of primitive (undepleted) mantle (McDonough and Sun, 1995). Peridotites from W. Arabia are from Stern and Johnson (2010). Compositional fields for forearc peridotites and mid-ocean ridge peridotites are from Azer and Stern (2007). Also shown is a melting-depletion trajectory, showing expected CaO and Al2O3 contents for residual peridotites after 1%, 5%, 10% and 20% melting to generate basalt; composition of the “Mean Arabian Peridotite” corresponds to ~15% melt depletion (Stern and Johnson, 2010).

Fig. 7. Olivine–spinel mantle array (OSMA; from Arai, 1994) for Al-Ashkhara, other Omani, and W. Arabian mantle xenoliths. Fields for forearc and abyssal peridotites are from Peslier et al. (2002).

Fig. 8. Primitive mantle normalized trace element abundances in Al-Ashkhara spinel lherzolite. Normalizing values from Sun and McDonough (1989).
lower crustal xenoliths are more MgO-rich (10–18% MgO), i.e. generally much higher than the 7–11% MgO characteristic of W. Arabian lower crust xenoliths, which are similar to the global lower crustal average of 7.1% MgO (Rudnick and Fountain, 1995). Correspondingly, Al-Ashkhara xenoliths have much higher Mg# (76–78), Cr (500–3000 ppm) and Ni (200–800 ppm) relative to those of Arabia west of CAMA (Mg# = 55–67; 120–780 ppm Cr, 70–170 ppm Ni).

Temperature estimates for Al-Ashkhara lower crustal xenoliths yielded 840–865 °C for the plagioclase-rich suite, and 810–825 °C for the pyroxene-rich suite using the Wells (1977) 2-pyroxene thermometer. Slightly lower ranges were obtained using the Brey and Koehler (1990a, 1990b) thermometer: 765 °C (Brey and Koehler, 1990a, 1990b) whereas 11 plagioclase-rich granulites yield 857±78 °C (Wells, 1977) and 776±75 °C (Brey and Koehler, 1990a, 1990b) thermometers: 765–795 °C for the plagioclase-rich suite, and 615–740 °C for the pyroxene-rich suite. Temperature estimates of lower crustal xenoliths from western Arabia are reported by Al-Mishwat and Nasir (2004), Ghent et al. (1980), Nasir (1992), Nasir (1995), and Nasir and Safarjalani (2000). Seven pyroxene-rich granite xenoliths from W. Arabia gave mean T of 786 ± 25 °C (Wells, 1977) and 710 ± 76 °C (Brey and Koehler, 1990a, 1990b) whereas 11 plagioclase-rich granulites yield 857 ± 78 °C (Wells, 1977) and 776 ± 75 °C (Brey and Koehler, 1990a, 1990b). The higher temperatures for the plagioclase-rich group relative to the pyroxene-rich group in both Al-Ashkhara and W. Arabian lower crustal suites are surprising, given the likelihood referred to above of a density-stratified lower crust and the more fractionated nature of the plagioclase-rich lower crustal suite.

There are also subtle mineralogical differences between the two suites of lower crustal xenoliths. Those from Arabia west of CAMA also contain abundant plagioclase and pyroxene, but a few contain garnet and rare amphibole, and lack olivine and spinel (Al-Mishwat and Nasir, 2004). Omani lower crustal xenoliths (as sampled at Al-Ashkhara) have no garnet or hydrous minerals and contain spinel.

The εNd and 87Sr/86Sr calculated at 700 Ma for Al-Ashkhara lower crustal xenoliths (+6.4 to +6.6; 87Sr/86Sr=0.7028 to 0.7039) are consistent with the interpretation that juvenile, mafic melts were added to the lower crust and that there was no discernible contribution from pre-Neoproterozoic crust. Consistent with available zircon ages for E. Arabia (Stern and Johnson, 2010), this implies that all of Arabia is Neoproterozoic juvenile crust (excluding the Khida terrane in the Arabian Shield and parts of Yemen).

Major element compositions of Omani lower crustal xenoliths show tholeiitic affinities. Tholeiites form by relatively high extents of mantle melting (5–20%) of spinel peridotite. Plagioclase-rich Omani xenoliths have major element compositions – especially Al2O3 abundance – that suggest these approximate the composition of the original tholeiitic magmas. If so, these melts were strongly depleted in TiO2 and K2O, indicating high extents of melting and/or derivation from very depleted mantle. Tholeiitic mafic melts may form either in an intraplate (hotspot) setting or above a subduction zone (e.g., Condle, 1999), a set of possibilities that mirrors models for the origin of Arabian–Nubian Shield crust as either (or both) from accreted oceanic plateaus (Stein and Goldstein, 1996) or from accretion of island arcs (Stern, 1994). The trace element spider-diagrams for Al-Ashkhara lower crust xenoliths show no negative Nb anomalies (Fig. 10), as would be expected for convergent margins magmas. Condle (1999) found that La/Nb in mafic rocks discriminates between hotspot (low La/Nb) and acclerated (high La/Nb) basalts, with the boundary between them at La/Nb = 1.4. Omani lower crust xenoliths have low La/Nb (0.24–1.0; mean = 0.59±0.28 (1 std. dev.)) indicating that they probably did not form at a convergent plate margin; instead they may have formed above a mantle hotspot. In contrast, W. Arabian lower crust xenoliths have relatively high La/Nb (mean La/Nb = 2.6 ± 1.6 (1 std. dev.)) indicating that this lower crust mostly formed in association with a convergent plate margin. We emphasize that this conclusion is based on only 8 lower crust samples from a single locality in E Arabia and so must be tested with more samples at more localities, but the data at hand suggest that the lower crusts of E and W Arabia formed in different tectonic environments.

8.2. Mantle xenoliths

In contrast to lower crustal xenoliths, which are only known in E. Arabia from Al-Ashkhara, Oman has several mantle xenolith localities
Sr, Nd, and Pb isotopic compositions of eastern Oman xenoliths.

Table 3

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<th>Sample Type</th>
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<th>KG1</th>
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Al-Asshkhara mantle xenolith mineral assemblages indicate a pressure range of 12–18 kbar (35–55 km depth). The lower end of this range is close to estimates for the crust/mantle boundary beneath Oman (38–40 km) (Al-Hashmi et al., 2011; Al-Lazki et al., 2002). Western Arabian peridotite xenoliths yield similar temperature and pressure estimates (900–1050 °C and 13–19 kbar) (Ghen et al., 2008; Ismail et al., 2008; Kaliwoda et al., 2007; McGuire, 1988a,b; Medaris and Syada, 1998; Medaris and Syada, 1999; Nasir, 1992; Stein and Katz, 1989). Petrological and geochemical characteristics of hydrous Al-Asshkhara lherzolites – especially the presence of phlogopite, carbonate veins, LREE, U, and Sr enrichments (Figs. 8, 9), and disturbance of the Pb isotopic system – provide unequivocal evidence for metasomatism. LREE enrichments, especially spoon-shaped REE patterns (Fig. 9) suggest chromatographic metasomatism by melt infiltration (Navon and Stolper, 1987). Most Al-Asshkhara spinel lherzolites are enriched in incompatible elements, including LREE, Th and U, implying that the mantle lithosphere represented by these rocks was metasomatically re-enriched after melt depletion (Fig. 9). The εNd calculated for samples with 147Sm/144Nd = 0.165.

8.3. Relationship between lower crust and mantle lithosphere

Crustal and mantle xenoliths provide a random sample of the lithosphere beneath a given site, such as Al-Asshkhara. Especially where juvenile crust and mafic lower crust can be demonstrated, as is the case for E. Arabia, it is reasonable to consider the possibility that the three compositionally distinct lithospheric layers – upper crust, lower crust, and mantle – bear a complementary relationship. In simplest terms, this relationship might reflect the formation of unidentifiable crust by partial melting of the mantle, with continued magmatic addition causing compositional and density stratification of the crust itself (e.g., Annen et al., 2006; Glazner, 1994). This might result from the basaltic melts comprising the lower crust being derived by partial melting of the upper mantle, represented by the peridotite xenoliths. Such a mantle source would have been very depleted as a result of such large volumes of melt extraction.
The Al-Ashkhara xenolith suite reveals no complementary relationship between lower crust and upper mantle xenoliths. Isotopic compositions for 700 Ma of radiogenic growth are distinct. εNd and δ7Srl/86Sr for both suites are juvenile but show no overlap: lherzolite εNd = +6.8 to +7.8 vs. lower crust εNd = +6.4 to +6.6; lherzolite δ7Srl/86Sr = 0.7016 to 0.7025 vs. lower crust δ7Srl/86Sr = 0.7028 to 0.7039. Trace element patterns and ratios (e.g., Rb/Sr) are also inconsistent with a complementary relationship, the lower crustal xenoliths being depleted relative to the peridotites. Some of these differences may reflect metasomatic additions to the peridotite, but the low Cr# of spinels further indicates the relatively undepleted nature of the peridotites.

The lack of a complementary relationship between Al-Ashkhara lower crust and upper mantle as sampled by the xenolith suite that we have studied can be explained if the upper mantle that was depleted to form the crust was removed. This could have happened as a result of delamination, as has been argued to have happened beneath W. Arabia at the end of Neoproterozoic time (Avigad and Gvitzman, 2009). Alternatively, replacement of melt-depleted upper mantle by the enriched mantle sampled by Al-Ashkhara peridotite xenoliths could have been accomplished by convection. However if occurred, replacement of mantle and stabilization of the lithospheric mantle beneath Al-Ashkhara must have happened earlier in Neoproterozoic time, perhaps at the end of the crust-forming event, which ended no later than ~700 Ma in Eastern Arabia.

9. Conclusions

The Al-Ashkhara xenolith suite provides a rare opportunity to study the lower crust and upper mantle beneath E. Arabia. The lower crust sampled here is composed of pyroxene- and plagioclase-rich mafic cumulates as indicated by xenoliths with high MgO and Al2O3 contents and weakly positive Eu anomalies and LREE enrichment. Mafic lower crust formed from underplated, deplated tholeiites with trace element compositions (lack of Nb anomaly, low La/Nb) indicating formation above a mantle hotspot, probably in Neoproterozoic time.

The low εNd of 750 Ma indicates that the upper crust of Arabia formed (~750 Ma). Their high δ7Srl/86Sr (~6.4 to ~6.6) and low 87Sr/86Sr (0.7028 to 0.7039) are inconsistent with a complementary relationship, the lower crustal xenoliths being depleted relative to the peridotites. Some of these differences may reflect metasomatic additions to the peridotite, but the low Cr# of spinels further indicates the relatively undepleted nature of the peridotites.

The lack of a complementary relationship between Al-Ashkhara lower crust and upper mantle as sampled by the xenolith suite that we have studied can be explained if the upper mantle that was depleted to form the crust was removed. This could have happened as a result of delamination, as has been argued to have happened beneath W. Arabia at the end of Neoproterozoic time (Avigad and Gvitzman, 2009). Alternatively, replacement of melt-depleted upper mantle by the enriched mantle sampled by Al-Ashkhara peridotite xenoliths could have been accomplished by convection. However if occurred, replacement of mantle and stabilization of the lithospheric mantle beneath Al-Ashkhara must have happened earlier in Neoproterozoic time, perhaps at the end of the crust-forming event, which ended no later than ~700 Ma in Eastern Arabia.

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