

Supplementary material

1. Analytical methods

1.1. *Major element compositions*

Whole-rock major element concentrations (Table A.3) were measured by the SARM at the CRPG laboratory (Nancy, France; <http://helium.crpq.cnrs-nancy.fr/SARM/>). Loss on Ignition (LOI) was determined by mass loss after calcination at 1000°C. The anhydrous residue was fused with LiBO₂ flux then digested in acid prior to major element analysis using Inductively Coupled Plasma Optical Emission Spectrometry (ICP-OES iCap6500). Detailed technique description can be found in Carignan et al. (2001).

1.2. *Trace element compositions*

Some of our samples were analyzed at Géosciences Montpellier (AETE-ISO, OSU OREME, Université de Montpellier, France) to determine their whole-rock trace element concentrations using an Agilent 7700x quadrupole ICP-MS (data in Table A.4). We focus here on elements used to trace possible effects of hydrothermal alteration (Cs, Rb, Ba, Th and U) (Tables A.4-5). The behavior of the other trace elements will be discussed in another contribution.

Samples were prepared following the HF/HClO₄ method and measurement protocol described in Godard et al. (2000) and Ionov et al. (1992). They were analyzed after a dilution of 1000. Element concentrations were measured by external calibration, except for Nb and Ta, for which concentrations were determined by surrogate calibration using Zr and Hf, respectively, as internal standards. This technique is an adaptation to ICP-MS analysis of the method described by Jochum et al. (1990), developed to minimize memory effects due to the introduction of concentrated Nb-Ta solutions in the instrument. The Helium cell gas mode of the Agilent 7700X was used to measure Sc, Ti, V, Mn, Co, Ni, Cu, Zn, Ga, As, Sn and Sb while removing polyatomic interferences. Each ICP-MS measurement is an average of three runs and its precision is determined by its standard deviation. The analytical uncertainty was estimated for each sample using the error propagation method described in Godard et al. (2000), which takes into

account the precision of the measurements of (1) the instrumental blank, (2) the procedural blanks and (3) the sample analysis. Analyses (1) below the instrument detection limit, (2) for which the contribution of the procedural blank is >70% or (3) having uncertainties >50 % were eliminated (noted “not determined”). We also analyzed rock reference materials UB-N, DTS-2b, JP-1 to assess the external precision and accuracy of our analyses. The limits of detection of the instrument, the average values of the procedural blanks and rock standards obtained during the two analytical periods of this study are reported in [Table A.5](#).

1.3. Re-Os compositions

Rhenium and osmium concentrations and Os isotopic compositions were analyzed at CRPG. Sample powders (1.5-2 g) were spiked with a mixed tracer solution containing ^{185}Re and ^{190}Os , and digested in a solution of concentrated $\text{HCl}:\text{HNO}_3$ (2:5 for samples and 1.5:5 for blanks) in a High Pressure Asher (HPA-S; Anton Paar) for 15 hours at 100 bars and 285°C. This procedure brings all of the Os to its most oxidized form (OsO_4), assuring spike-sample equilibration. Osmium was chemically extracted from the sample solution into liquid bromine, for which OsO_4 has a strong affinity, and purified by microdistillation, using techniques adapted from Birck et al. (1997). Following Os extraction, Re was separated from the residual acidic solution by anion-exchange column chromatography (Rehkämper and Halliday, 1997).

Rhenium isotopic compositions used for Re concentration determination by isotope dilution were analyzed by MC-ICPMS (ThermoScientific Neptune +). Instrumental mass fractionation was corrected by standard bracketing, assuming a $^{187}\text{Re}/^{185}\text{Re}$ ratio of 1.6738 for the standard. Osmium isotopic compositions were determined by Negative Thermal Ionization Mass Spectrometry (N-TIMS) (Creaser et al., 1991; Völkening et al., 1991) using a Finnigan MAT 262 mass spectrometer. Isotopic ratios of OsO_3^- masses were obtained by peak jumping on an ETP electron multiplier. Corrections for heavy oxides, spike contributions and instrumental mass fractionation were made off-line, assuming a natural $^{192}\text{Os}/^{188}\text{Os}$ ratio of 3.08271.

Re-Os data are presented in [Tables A.6-8](#). All uncertainties are 2σ and include all significant recognized sources of uncertainty (in-run analytical precision, standard reproducibility, uncertainties related to variability of the amount and isotopic composition of the analytical blank, uncertainties related to the Re/Os ratio and the ensuing age correction). These different uncertainty components are listed for each sample in [Table A.8](#). The total analytical blanks varied between the four main analytical periods of the study: Re = 0.008 ± 0.008 ng, Os = 0.0009 ± 0.0022 ng, $^{187}\text{Os}/^{188}\text{Os}$ = 0.140 ± 0.056 (first period); Re = 0.061 ± 0.087 ng, Os = 0.0086 ± 0.0103 ng, $^{187}\text{Os}/^{188}\text{Os}$ = 0.138 ± 0.220 (second period); Re = 0.011 ± 0.018 ng, Os = 0.0085 ± 0.0052 ng, $^{187}\text{Os}/^{188}\text{Os}$ = 1.454 ± 0.807 (third period); Re = 0.042 ± 0.039 ng, Os = 0.0007 ± 0.0001 ng, $^{187}\text{Os}/^{188}\text{Os}$ = 0.209 ± 0.122 (fourth period); all uncertainties 2σ . Though the blanks of the second and third periods are rather high and variable, the total 2σ uncertainties, including the blank variability, on the initial $^{187}\text{Os}/^{188}\text{Os}$ ratios of 33 of the 42 samples are $<1\%$, and $<2\%$ for the remaining 9 samples. Four analyses of the UB-N reference material ([Tables A.6-8](#)) analyzed concurrently with the samples yielded a mean $^{187}\text{Os}/^{188}\text{Os}$ ratio of 0.1274 ± 0.0006 (2σ), in agreement with the accepted value (Meisel et al., 2003).

2. Seawater alteration of peridotite - mixing model

To exclude a contamination of our samples with seawater, we simulate a simple two component mixture between seawater and a common peridotite (Fig. A.2b), similar to the model used by Sanfilippo et al. (2016), using the following equations:

$$(1) \quad [Os]_{mix} = (f \cdot [Os]_{seawater}) + ((1 - f) \cdot [Os]_{per})$$

$$(2) \quad (^{187}Os/^{188}Os)_{mix} = \frac{1}{[Os]_{mix}} \cdot (f \cdot [Os]_{seawater} \cdot (^{187}Os/^{188}Os)_{seawater}) + ((1 - f) \cdot [Os]_{per} \cdot (^{187}Os/^{188}Os)_{per})$$

Th and U concentrations are calculated in the same manner as Os concentration, using Equation A.1.

where: $[Os]_{per}$: concentration in Os of the peridotite

$[Os]_{seawater}$: concentration in Os of seawater

$[Os]_{mix}$: concentration in Os of the mixture, including seawater and peridotite

$(^{187}Os/^{188}Os)_{per}$: $^{187}Os/^{188}Os$ ratio of the peridotite

$(^{187}Os/^{188}Os)_{melt}$: $^{187}Os/^{188}Os$ ratio of seawater

$(^{187}Os/^{188}Os)_{mix}$: $^{187}Os/^{188}Os$ ratio of the mixture, including seawater and peridotite

f : seawater fraction

We use the following compositions for the peridotite and seawater endmembers:

$$(^{187}Os/^{188}Os)_{per} = 0.124$$

$$(^{187}Os/^{188}Os)_{seawater} = 1$$

$$[Os]_{per} = 0.4 \text{ ppb, } 2 \text{ ppb or } 6 \text{ ppb}$$

$$[Os]_{seawater} = 10^{-5} \text{ ppb}$$

$$[Th]_{per} = 13 \text{ ppb}$$

$$[Th]_{seawater} = 5 \times 10^{-6} \text{ ppb}$$

$$[U]_{per} = 4.5 \text{ ppb}$$

$$[U]_{seawater} = 3.3 \text{ ppb}$$

For the peridotite, the $^{187}Os/^{188}Os$ ratio chosen is typical of Oman harzburgites, as well as of abyssal peridotites. Three different initial Os concentrations were used to cover a compositional range from a typical harzburgite with a quite high Os concentration (6 ppb) to an Os-poor MTZ dunite (0.4 ppb). The peridotite Th and U concentrations are those of the depleted peridotite reference material PCC-1 (Govindaraju, 1994). Seawater Os, Th and U composition is the same as that used in the model of Sanfilippo et al. (2016).

3. Melt-peridotite mixture

As in the previous seawater-peridotite mixture, we simulate a simple two component mixture between a common peridotite and a melt (Fig. 5) using the following equations:

$$(3) \quad [Os]_{mix} = (f \cdot [Os]_{melt}) + ((1 - f) \cdot [Os]_{per})$$

$$(4) \quad ({}^{187}Os/{}^{188}Os)_{mix} = \frac{1}{[Os]_{mix}} \cdot (f \cdot [Os]_{melt} \cdot ({}^{187}Os/{}^{188}Os)_{melt}) + ((1 - f) \cdot [Os]_{per} \cdot ({}^{187}Os/{}^{188}Os)_{per})$$

where: $[Os]_{per}$: concentration in Os of the peridotite

$[Os]_{melt}$: concentration in Os of the melt

$[Os]_{mix}$: concentration in Os of the mixture, including melt and peridotite

$({}^{187}Os/{}^{188}Os)_{per}$: ${}^{187}Os/{}^{188}Os$ ratio of the peridotite

$({}^{187}Os/{}^{188}Os)_{melt}$: ${}^{187}Os/{}^{188}Os$ ratio of the melt

$({}^{187}Os/{}^{188}Os)_{mix}$: ${}^{187}Os/{}^{188}Os$ ratio of the mixture, including melt and peridotite

f : melt fraction

We use the following compositions for the peridotite and the melt:

$$({}^{187}Os/{}^{188}Os)_{per} = 0.124$$

$$({}^{187}Os/{}^{188}Os)_{melt} = 0.150$$

$$[Os]_{per} = 4 \text{ ppb or } 0.5 \text{ ppb}$$

$$[Os]_{melt} = 0.2 \text{ ppb}$$

As in the previous mixture, the ${}^{187}Os/{}^{188}Os$ ratio of the peridotite endmember is typical of Oman harzburgites, as well as of abyssal peridotites, while the ratio of the melt was chosen to be just slightly more radiogenic than that of the MTZ dunites. The two different initial Os concentrations for the peridotite represent those of a typical harzburgite (4 ppb) and a MTZ dunite (0.5 ppb).

4. Melt percolation models

4.1. *Single-cell percolation*

We modeled melt percolation through a volume of peridotite which is flushed incrementally by magma of a constant radiogenic composition (Fig. 6). The peridotite is taken to be harzburgite initially, which is transformed to dunite through melt-rock reaction. We use a melt increment of 2%, and the same melt and peridotite end-member Os concentrations and $^{187}\text{Os}/^{188}\text{Os}$ ratios as in the melt-peridotite mixture. While this peridotite composition is representative of typical abyssal peridotite Os compositions, it should be stressed that these in fact span a substantial range (Becker and Dale, 2016), and cannot be viewed as a single well-defined endmember. When each new batch of melt enters the peridotite, its Os concentration and isotopic ratio equilibrate with those of the peridotite. After equilibration, the melt moves out of the peridotite and is replaced by fresh melt of the original composition ($[\text{Os}]_{\text{melt}_0}$ and $(^{187}\text{Os}/^{188}\text{Os})_{\text{melt}_0}$). On the basis of the experimental results from Lambart et al. (2009), we consider that harzburgite is fully transformed to dunite after having interacted with around 25-30% of melt, equivalent to ~10-15 melt increments. We have included a small amount of sulfide (f_{S_0}) in our model. For simplicity, we assume it to decrease linearly with orthopyroxene content, so that the sulfide content reaches zero at the same time that the transformation to dunite is complete. Complete sulfide exhaustion slightly before or slightly after full transformation to dunite will not significantly influence the results, since it is the loss of sulfide, not the change from harzburgite to dunite that marks the beginning of Os loss from the peridotite during melt percolation. In the example shown in the text, the harzburgite contains 100 ppm sulfide, equivalent to about 40 ppm S, and is assumed to decrease to zero over 10 increments. In other words, during each increment, the peridotite loses a part of its sulfide fraction (10%) until reaching a null value:

$$(5) \quad f_{S_i} = f_{S_0} \cdot (1 - ((i - 1) \cdot 0.1)) \quad i = 1, 2, 3, \dots$$

The global Os partition coefficient of the peridotite (D_{Os_i}) depends on the sulfide fraction in the peridotite (f_{S_i}) and on the Os partition coefficients ($k_{d,\text{Os}}$) of the phases present (e. g. sulfide and olivine):

$$(6) \quad D_{\text{Os}_i} = (f_{S_i} \cdot k_{d,\text{Os in sulfide}}) + ((1 - f_{S_i}) \cdot k_{d,\text{Os in olivine}}) \quad D_{\text{Os}_i} \sim 2 \text{ if dunite}$$

When the peridotite is devoid of sulfide, the dunite has an assumed Os bulk partition coefficient of 2 (see main text for discussion of this choice). For simplicity we assume that this value is constant.

We use the following equations to calculate bulk rock Os composition, including melt and peridotite:

$$(7) \quad [Os]_{bulk_i} = (f \cdot [Os]_{melt_0}) + ((1-f) \cdot [Os]_{per_{i-1}})$$

$$(8) \quad ({}^{187}Os/{}^{188}Os)_{bulk_i} = \frac{1}{[Os]_{bulk_i}} \cdot (f \cdot [Os]_{melt_0} \cdot ({}^{187}Os/{}^{188}Os)_{melt_0}) + ((1-f) \cdot [Os]_{per_{i-1}} \cdot ({}^{187}Os/{}^{188}Os)_{per_{i-1}})$$

$$(9) \quad ({}^{187}Os/{}^{188}Os)_{per_i} = ({}^{187}Os/{}^{188}Os)_{bulk_i} \text{ if } i \neq 1$$

Then, we use the global Os partition coefficient (D_{Os_i}) to determine Os concentration in peridotite and in the new melt moving out of the peridotite:

$$(10) \quad [Os]_{new\ melt_i} = \frac{[Os]_{bulk_i}}{f + ((1-f) \cdot D_{Os_i})}$$

$$(11) \quad [Os]_{per_i} = D_{Os_i} \cdot [Os]_{new\ melt_i}$$

Abbreviations: $[Os]_{per_0}$: initial concentration in Os of the peridotite

$[Os]_{per_i}$: concentration in Os of the peridotite at increment = i

$[Os]_{melt_0}$: initial concentration in Os of the melt

$[Os]_{melt_i}$: concentration in Os of the melt at increment = i

$[Os]_{bulk_i}$: concentration in Os of the bulk rock, including melt and peridotite, at increment = i

$[Os]_{new\ melt_i}$: final concentration in Os of the melt which percolated the peridotite at increment = i and which will migrate out of the peridotite.

${}^{187}Os/{}^{188}Os_{per_0}$: initial ${}^{187}Os/{}^{188}Os$ ratio of the peridotite

${}^{187}Os/{}^{188}Os_{melt_0}$: initial ${}^{187}Os/{}^{188}Os$ ratio of the melt

${}^{187}Os/{}^{188}Os_{bulk_i}$: ${}^{187}Os/{}^{188}Os$ ratio of the bulk rock, including melt and peridotite, at increment = i

f : instantaneous melt fraction (porosity)

f_{S_0} : initial sulfide fraction in the peridotite

D_{Os_i} : global Os partition coefficient at increment = i

$k_{d,Os\ in\ sulfide}$: Os partition coefficient of sulfide

$k_{d,Os\ in\ olivine}$: Os partition coefficient of olivine

In the example in the text, we use the following values, with the partition coefficients based on the discussion in Brenan et al. (2016) and Mungall and Brenan (2014):

$$[Os]_{per_0} = 4 \text{ ppb}$$

$$[Os]_{melt_0} = 0.2 \text{ ppb}$$

$$^{187}Os/^{188}Os_{per_0} = 0.124$$

$$^{187}Os/^{188}Os_{melt_0} = 0.150$$

$$f = 0.02$$

$$f_{S_0} = 0.0001$$

$$k_{d,Os \text{ in sulfide}} = 10^6$$

$$k_{d,Os \text{ in olivine}} = 2 \text{ if sulfide present}$$

$$D_{Os_i} \sim 2 \text{ if dunite}$$

4.2. Percolation through a dunite channel

To consider whether melt percolation by porous flow in dunite channels is consistent with our Os results, we modeled a melt percolation process considering the formation of a dunite channel in the host harzburgite which is flushed incrementally by a magma (Fig. 9). The channel is divided vertically into five reaction cells (Fig. 9a), each similar to that described in the single-cell percolation model. In each step, an increment of fresh radiogenic melt enters the bottom cell of the channel. Melt Os concentration and isotopic ratio equilibrate with those of the peridotite (Fig. 9b-c). Then, the equilibrated melt migrates upwards into the overlying cell and is replaced by melt from the underlying cell. The Os concentration of the melt leaving each peridotite cell ($[Os]_{new\ melt_i}$) is controlled by the Os partition coefficients ($k_{d,Os}$) of the phases present (initially dominated by sulfide, then by PGM and possibly spinel and olivine after sulfide exhaustion) (Becker and Dale, 2016; Brenan et al., 2003, 2005, 2016; Mungall and Brenan, 2014), while the Os isotope ratio is controlled by mass balance. The bottom cell is constantly replenished with magma with the original Os concentration and isotopic composition ($[Os]_{melt_0}$ and $(^{187}Os/^{188}Os)_{melt_0}$).

As in the single-cell percolation model, we consider that harzburgite becomes dunite and loses its sulfide fraction (f_{S_0}) after having interacted with around 25-30% of melt (10-15 melt increments). We also use the same equations to calculate the sulfide fraction of the peridotite (f_{S_i}), the bulk rock Os composition including melt and peridotite ($[Os]_{bulk_i}$ and $(^{187}Os/^{188}Os)_{bulk_i}$), the Os concentration in peridotite ($[Os]_{per_i}$) and in the new melt percolating upwards into the overlying cell ($[Os]_{new\ melt_i}$):

$$(12) \quad f_{S_i} = f_{S_0} \cdot (1 - ((i - 1) \cdot 0.1)) \quad i = 1, 2, 3, \dots$$

$$(13) \quad D_{Os_i} = (f_{S_i} \cdot k_{d,Os\ in\ sulfide}) + ((1 - f_{S_i}) \cdot k_{d,Os\ in\ olivine}) \quad D_{Os_i} \sim 2\ if\ dunite$$

$$(14) \quad [Os]_{bulk_i} = (f \cdot [Os]_{melt_0}) + ((1 - f) \cdot [Os]_{per_{i-1}})$$

$$(15) \quad (^{187}Os/^{188}Os)_{bulk_i} = \frac{1}{[Os]_{bulk_i}} \cdot (f \cdot [Os]_{melt_0} \cdot (^{187}Os/^{188}Os)_{melt_0}) + ((1 - f) \cdot [Os]_{per_{i-1}} \cdot (^{187}Os/^{188}Os)_{per_{i-1}})$$

$$(16) \quad (^{187}Os/^{188}Os)_{per_i} = (^{187}Os/^{188}Os)_{bulk_i}\ if\ i \neq 1$$

$$(17) \quad [Os]_{new\ melt\ i} = \frac{[Os]_{bulk\ i}}{f + (1-f) \cdot D_{Os\ i}}$$

$$(18) \quad [Os]_{per\ i} = D_{Os\ i} \cdot [Os]_{new\ melt\ i}$$

Abbreviations: $[Os]_{per\ 0}$: initial concentration in Os of the peridotite

$[Os]_{per\ i}$: concentration in Os of the peridotite at increment = i

$[Os]_{melt\ 0}$: initial concentration in Os of the melt

$[Os]_{melt\ i}$: concentration in Os of the melt at increment = i

$[Os]_{bulk\ i}$: concentration in Os of the bulk rock, including melt and peridotite, at increment = i

$[Os]_{new\ melt\ i}$: final concentration in Os of the melt which percolated the peridotite at increment = i and which will migrate into the overlying peridotite.

$^{187}Os/^{188}Os_{per\ 0}$: initial $^{187}Os/^{188}Os$ ratio of the peridotite

$^{187}Os/^{188}Os_{melt\ 0}$: initial $^{187}Os/^{188}Os$ ratio of the melt

$^{187}Os/^{188}Os_{bulk\ i}$: $^{187}Os/^{188}Os$ ratio of the bulk rock, including melt and peridotite, at increment = i

f : instantaneous melt fraction (porosity)

$f_{s\ 0}$: initial sulfide fraction in the peridotite

$D_{Os\ i}$: global Os partition coefficient at increment = i

$k_{d,Os\ in\ sulfide}$: Os partition coefficient of sulfide

$k_{d,Os\ in\ olivine}$: Os partition coefficient of olivine

In the example in the text, we use the same values as in the single-cell percolation model.

Figures

Fig. A.1. Stratigraphic cross-sections of the mantle section in the Nakhl **(a)**, Sumail (SE to NW cross-section) **(b)** and Wadi Tayin **(c-e)** massifs of the southeastern Oman ophiolite showing the distribution of the three types of dunite in the ubiquitous harzburgite and the sample locations. Mylonites, plagioclase impregnations, gabbroic and pyroxenite veins or bands are also represented.

Fig. A.2. Osmium isotope composition vs. tracers of seawater contamination (LOI and whole-rock highly mobile trace elements) for harzburgites and the three types of dunite. $^{187}\text{Os}/^{188}\text{Os}_{95\text{Ma}}$ vs. **(a)** LOI, **(b)** Th/U ratio, **(c)** U, **(d)** Rb, **(e)** Ba, **(f)** Cs. In panel **(b)**, three (dark, medium and light) grey curves represent the compositions of a mixture of Os, Th and U from seawater ($^{187}\text{Os}/^{188}\text{Os} = 1$; $[\text{Os}] = 10^{-5}$ ppb; $[\text{Th}] = 5 \times 10^{-6}$ ppb; $[\text{U}] = 3.3$ ppb) and peridotite ($^{187}\text{Os}/^{188}\text{Os} = 0.124$; $[\text{Th}] = 13$ ppb; $[\text{U}] = 4.5$ ppb) with different starting Os concentrations (6, 2 and 0.4 ppb respectively) (cf. [Supplementary Material](#)). Black dotted lines indicate seawater/rock ratios of 1, 10, 100, 1000 and 10 000. In all panels, published Oman peridotite data are also shown for comparison: MTZ dunites (Godard et al., 2000; Xiong et al., 2020a,b), MMS dunites (Godard et al., 2000; Hanghøj et al., 2010; Xiong et al., 2020a,b) and MMS harzburgites (Hanghøj et al., 2010) (symbols as in [Fig. 3](#)). Attribution of literature data to different dunite types is based on their geographic positions. No correlation is observed between initial Os isotope ratios, LOI, whole-rock highly mobile trace elements and the Th/U model curves. These curves show that seawater/rock ratios >1000 would be required to produce $^{187}\text{Os}/^{188}\text{Os}$ ratios comparable to those of the MTZ dunites, and these would be associated with Th/U ratios much lower than the observed values. Thus, seawater contamination cannot explain the radiogenic initial $^{187}\text{Os}/^{188}\text{Os}$ ratios of the MTZ dunites.

Fig. A.3. Effect of varying instantaneous melt fraction (f) on $[\text{Os}]$ and $^{187}\text{Os}/^{188}\text{Os}$ in the melt percolation model for a single peridotite cell (black curve) and for a dunite channel composed of five cells (black to light grey curves upwards). Results for the composition of each peridotite cell (cf. [Fig.](#)

6, 9 caption in main text) with melt fractions of 0.5% **(a-b)**, 2% **(c-d)** and 4% **(e-f)** are shown in **(a,c,e)** Os vs. increment and **(b,d,f)** $^{187}\text{Os}/^{188}\text{Os}$ vs. increment diagrams. To attain a $^{187}\text{Os}/^{188}\text{Os}$ ratio of 0.135 for the single peridotite cell or the bottom cell of the channel, melt/rock ratios of 4.28, 4.41 and 4.58 are needed with the considered porosities respectively. Here and in [Figures A.4-6](#), all other parameters are held constant at the values given above.

Fig. A.4 Effect of varying Os bulk partition coefficient for dunite (D_{Os}) on [Os] and $^{187}\text{Os}/^{188}\text{Os}$ in the melt percolation model for a single peridotite cell (black curve) and for a dunite channel composed of five cells (black to light grey curves upwards). Results for the composition of each peridotite cell (cf. [Fig. 6, 9](#) caption) with a $D_{Os} = 0.5$ **(a-b)**, 2 **(c-d)** and 5 **(e-f)** are shown in **(a,c,e)** Os vs. increment and **(b,d,f)** $^{187}\text{Os}/^{188}\text{Os}$ vs. increment diagrams.

Fig. A.5. Effect of varying melt Os concentration on [Os] and $^{187}\text{Os}/^{188}\text{Os}$ in the melt percolation model for a single peridotite cell (black curve) and for a dunite channel composed of five cells (black to light grey curves upwards). Results for the composition of each peridotite cell (cf. [Fig. 6, 9](#) caption) during the percolation of a melt which has an original Os concentration of 0.02 **(a-b)**, 0.2 **(c-d)** and 0.5 ppb **(e-f)** are shown in **(a,c,e)** Os vs. increment and **(b,d,f)** $^{187}\text{Os}/^{188}\text{Os}$ vs. increment diagrams.

Fig. A.6. Effect of varying melt $^{187}\text{Os}/^{188}\text{Os}$ on [Os] and $^{187}\text{Os}/^{188}\text{Os}$ in the melt percolation model for a single peridotite cell (black curve) and for a dunite channel composed of five cells (black to light grey curves upwards). Results for the composition of each peridotite cell (cf. [Fig. 6, 9](#) caption) during the percolation of a melt which has an original $^{187}\text{Os}/^{188}\text{Os}$ ratio of 0.13 **(a-b)**, 0.15 **(c-d)** and 0.2 **(e-f)** are shown in **(a,c,e)** Os vs. increment and **(b,d,f)** $^{187}\text{Os}/^{188}\text{Os}$ vs. increment diagrams.

Tables

Table A.1. Structural data of the studied Oman peridotites

Table A.2. Locations and petrographic data of the studied Oman peridotites

Table A.3. Locations and major element compositions of the studied Oman peridotites

Table A.4. Ba, Rb, Cs, Th and U concentrations of the studied Oman peridotites

Table A.5. ICP-MS analytical parameters and Ba, Rb, Cs, Th and U concentrations of the standards

Table A.6. Re and Os concentrations and Os isotope compositions of the studied Oman peridotites and of the UB-N standard

Table A.7. Re and Os concentrations and Os isotope compositions of the harzburgite/dunite pairs

Table A.8. Different components of the uncertainties on Re and Os concentrations and Os isotope compositions entering into the total uncertainties

Table A.9. Petrostructural data and locations of additional (not analyzed) Oman peridotites

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