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UNRAVELLING THE EFFECTS OF MELT DEPLETION AND SECONDARY INfiltrATION ON MANTLE RE-OS ISOTOPES BENEATH THE FRENCH MASSIF CENTRAL.

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ABSTRACT

Spinel lherzolite xenoliths from Mont Briançon, French Massif Central, retain evidence for multiple episodes of melt depletion and melt/fluid infiltration (metasomatism). Evidence for primary melt depletion is still preserved in the co-variation of bulk-rock major elements (MgO 38.7–46.1 weight %; CaO 0.9–3.6 weight %), and many samples yield unradiogenic bulk-rock Os isotope ratios ($^{187}$Os/$^{188}$Os = 0.11541-0.12626). However, many individual xenoliths contain interstitial glasses and melt inclusions that are not in equilibrium with the major primary minerals. Incompatible trace element mass balance calculations demonstrate that metasomatic components comprise a significant proportion of the bulk-rock budget for these elements in some rocks, ranging to as much as 25% of Nd and 40% of Sr. Critically, for Re-Os geochronology, melt/fluid infiltration is accompanied by the mobilisation of sulfide. Consequently, bulk-rock isotope measurements, whether using lithophile (e.g. Rb-Sr, Sm-Nd) or siderophile (Re-Os) based isotope systems, may only yield a perturbed and/or homogenised average of these multiple events.

Osmium mass balance calculations demonstrate that bulk-rock Os in peridotite is dominated by contributions from 2 populations of sulfide grain: (i) interstitial, metasomatic sulfide with low [Os] and radiogenic $^{187}$Os/$^{188}$Os, and (ii) primary sulfides with high [Os] and unradiogenic $^{187}$Os/$^{188}$Os, which have been preserved within host silicate grains and shielded from interaction with transient melts and fluid. The latter can account for > 97% of bulk rock Os and preserve geochronological information of the melt from which they originally precipitated as an immiscible liquid. The Re-depletion model ages of individual primary sulfide grains preserve
evidence for melt depletion beneath the Massif Central from at least 1.8 Gyr ago despite the more recent metasomatic event(s).

1. INTRODUCTION

Ultramafic xenoliths brought to the surface as accidental inclusions in alkali basalts and kimberlites provide some of the best direct evidence of the composition of the sub-continental lithospheric mantle (SCLM). This is because, unlike orogenic peridotite massifs, xenoliths experience rapid transport to the surface and, where interaction with the host lava is minimal, the composition of the mantle source is largely retained.

Systematic co-variations of major and compatible trace elements in peridotite xenoliths have been used to demonstrate the effects of melt extraction on the composition of fertile mantle (Ross et al., 1954; Kuno and Aoki, 1970; Frey and Green, 1974; Jagoutz et al., 1979) and to estimate the composition of undepleted, i.e. primitive mantle (e.g. Frey and Prinz, 1978; Meisel et al., 1996, 2001). However, incompatible trace element compositions of upper mantle samples are frequently decoupled from the more coherent major element trends. This feature is commonly attributed to re-fertilization processes. Several potential agents of this re-fertilization, or metasomatism, have been suggested, including CO₂-rich fluids (Andersen et al., 1984; O’Reilly and Griffin, 1988), siliceous melts and related fluids (Roden et al., 1984; Menzies et al., 1987; Schiano and Clocchiatti, 1994; Schiano and Bourdon, 1999), carbonated hyper-alkaline melts (Meen et al., 1989), kimberlite-like melts (e.g. Menzies and Wass, 1983), and carbonatitic melts (Yaxley et al., 1991; Dautria et al., 1992; Ionov et al., 1993, Rudnick et al., 1993; O’Reilly and Griffin, 1988; Griffin et
al., 1996). Much of the SCLM has experienced a multi-stage history of melt depletion
and enrichment in lithophile elements since segregation from the homogenising
effects of the convecting mantle (Galer and O’Nions 1988; McKenzie and Bickle,
1988; Zindler and Jagoutz, 1988; Pearson et al., 1995a; Pearson 1999), yet all of the
conventional geochronometers rely upon silicate-hosted lithophile elements (Rb-Sr,
Sm-Nd, U-Th-Pb, Lu-Hf). Critically, perturbation of a primary lithology through
interaction with secondary, lithophile element-rich melts and / or fluids results in the
perturbation of geochronological information relating to the primary melt depletion
event(s).

In contrast, the high temperature behaviour of the siderophile elements Re and Os
potentially offers a means to “see through” recent metasomatic events and provide
geochronological information relating to prior melt depletion. This is because Os
behaves compatibly during melting and is preferentially retained in the mantle,
whereas Re is moderately incompatible and enters the melt (e.g. Pegram and Allègre,
1992; Hauri and Hart, 1997). Thus, melt depletion generates a low Re/Os ratio in the
residue which, with time, through the decay of $^{187}$Re (parent) to $^{187}$Os (daughter),
evolves to unradiogenic $^{187}$Os/$^{188}$Os ratios compared to those of the contemporaneous
undepleted mantle (e.g. Meisel et al., 2001). The high Os concentration of the residual
mantle effectively buffers against the effects of secondary processes as the Os
elemental abundance in silicate metasomatic melts and fluids should be low. Several
recent studies have shown that, in many cases, the Os isotope ratio of lithospheric
mantle peridotite indicates evolution in a low Re environment following ancient
partial melting events i.e. 3.4 Ga in cratonic SCLM (Pearson et al., 1995b; Burton et
al., 2000; Griffin et al., 2004) and > 2 Ga in abyssal peridotites (Alard et al., 2005;
Harvey et al., 2006; Liu et al., 2008).
Exceptionally high sulfide/silicate partition coefficients derived experimentally 
\((K_D^{\text{sulfide/silicate}} \sim 10^4-10^5\) of Bezman et al., 1994; Fleet et al., 1991, 1996) and in natural 
samples \((K_D^{\text{sulfide/silicate}} \sim 4.8 \times 10^4,\) Roy-Barman et al., 1998), coupled with the low 
affinity of Os for silicate minerals (e.g. Righter et al., 2004; Brenan et al., 2005), 
suggest that volumetrically insignificant sulfide will dominate bulk-rock Os budgets. 
This is consistent with studies of natural samples (e.g. Hart and Ravizza, 1996; Burton 
et al., 1999; Alard et al., 2002; Alard et al., 2005; Harvey et al., 2006). However, at 
least two discrete populations of mantle sulfide have been identified; one residual 
after melt depletion and a second, interstitial population attributed to redistribution or 
precipitation by a metasomatising melt or fluid (Alard et al., 2002). This implies that 
the bulk-rock Re and Os budget will be controlled by the relative contributions from 
each of these two sulfide populations. Moreover, as interstitial sulfide may be mobile 
during metasomatism, re-distribution of Re and/or Os has the potential to obscure 
primary geochemical and isotope signatures in a similar way to that of the lithophile-
based isotope systems.

This study presents major and trace element abundances for bulk-rock and mantle 
minerals along with Re-Os isotope and elemental data for bulk-rocks, silicate 
minerals, and individual sulfide grains from a suite of spinel lherzolite and spinel 
harzburgite xenoliths from the Mont Briançon scoria cone, Devès volcanic province, 
French Massif Central. Variations between the major elements indicate that the 
sample suite has experienced differing degrees of melt depletion while bulk-rock and 
mineral trace element measurements reveal that many samples have also experienced 
extensive cryptic metasomatism. Moreover, analysis of trails of melt inclusions 
preserved within silicate minerals and along grain boundaries suggests that sulfide has 
also been added or re-distributed during metasomatism. Despite a significant
metasomatic overprint on many of the Mont Briançon xenoliths, the Re-Os isotopic
measurements of sulfide grains included within silicate minerals demonstrate that
some melt depletion occurred at least 1.8 Gyr ago, and that in many xenoliths the
recent addition of metasomatic sulfide has had little overall effect on the bulk-rock
Re-Os isotope systematics.
2. GEOLOGICAL SETTING AND SAMPLE PETROLOGY

The French Massif Central volcanic province is the largest volcanic area in Europe (Figure 1) and forms part of a much larger igneous province that extends as far east as Poland. Despite the regional thickening imposed during the Variscan orogeny, the Massif Central is characterised by particularly thin crust and a lithospheric thickness as low as 60 km (Hirn, 1980). Elevated heat flow in the region is comparable to other tectono-magmatically active regions (Lucazeau and Bayer, 1982; Granet et al., 1995) and the presence of a transient 20–30 mW m\(^{-2}\) anomaly is considered to be plume-related temperature augmentation (200-250°C) at a depth of 60-70 km (Coisy and Nicolas, 1978; Nicolas et al., 1987; Sobolev et al., 1996), which may indicate a hot mantle body extending to depths of at least 270 km (Granet et al., 1995; Sobolev et al 1996). A lithospheric boundary at 45° 30’ N, which separates northern and southern lithospheric domains, has been suggested based upon several lines of evidence; textural and bulk-rock trace element data (Lenoir et al., 2000); seismic anisotropy (Babuska et al., 2002); clinopyroxene trace elements, Sr and Nd isotope analyses (Downes et al., 2003); Pb and Lu-Hf isotope analyses of clinopyroxene grains from spinel-facies xenoliths (Wittig et al., 2006, 2007). These mantle domains not only possess distinct mantle topographies at the lithosphere / asthenosphere boundary, but also delimit the distribution of the dominant styles of metasomatism in peridotite xenoliths of the Massif Central.
The volcanic activity in the south-eastern Massif Central is characterised by Tertiary to Recent basaltic plateaux, strombolian scoria cones and alkalic plugs. The samples from this study are all spinel peridotite xenoliths from the Mont Briançon scoria cone in the Devès area, located in the southern lithospheric domain. This region was last volcanically active during the Pliocene (c. 1-4 Ma). The 17 samples from this study, like previous studies of Mont Briançon xenoliths (Lorand and Alard, 2001; Meisel et al., 2001), have predominantly protogranular textures (Coisy and Nicholas, 1978; Brown et al., 1980). Grains are separated by curvilinear boundaries and, while strain-free neoblasts are mostly absent, a degree of polygonisation to a coarse substructure is apparent in some samples. Two samples (MBr13 and MBr19) are characterized by both protogranular and porphyroclastic textures, while a single sample (MBr24) has a porphyroclastic texture. This suggests that most samples display a high degree of textural equilibrium.

Large (> 1 kg) xenoliths were preferentially selected so that host basalt could be trimmed and the possible effects of host infiltration minimized while still retaining sufficient material (> 500 g) to represent a heterogeneous sample at the bulk-rock scale. The samples for this study were also selected so as to represent a wide range of clinopyroxene modal abundance, estimated visually in the field and subsequently calculated using a least squares regression method (Tarantola and Valette, 1982). Chrome-diopside modal abundance ranges from 3.3 to 17.2 (± 1.6) %. No discrete metasomatic phases (e.g. phlogopite, amphibole, apatite) were detected during this study, or any prior study of this locality.

The temperatures at which these xenoliths last equilibrated range from 887°C to 1055°C (this study, online supplementary material; Coisy and Nicolas, 1978; Brown et al., 1980; Lorand and Alard, 2001; Sobolev et al., 1996). These calculations
are based upon clinopyroxene-orthopyroxene equilibria (Wells, 1977; Brey and Kohler, 1990). Assuming a geothermal gradient of 13.5 °C kbar⁻¹ (Lucazeau and Bayer, 1982) or 4.2 °C km⁻¹ (Granet et al., 1995), this suggests that they last equilibrated at depths of 55 – 65 km, i.e. near to the local lithosphere / asthenosphere boundary (Hirn, 1980). Entrainment depths are supported by minimum trapping pressures of CO₂-rich inclusions preserved in silicate grains within the xenoliths of 1.3 GPa (Schiano and Clocchiatti, 1994).
3. ANALYTICAL METHODS

Samples were cut and washed to remove host basalt, surface alteration and contamination, and following repeated rinsing in ultra-pure water, were dried and then powdered in an agate mortar. Bulk-rock major elements were measured using a ARL 8420+ dual goniometer wavelength dispersive XRF spectrometer. Major element and Ni data quality was assessed using two rock standards (WS-E and OUG-94). Reproducibility is within 2% of the recommended values (online supplementary material). Bulk-rock trace elements were measured on an Agilent 7500a ICP-MS. Error on LREE measurements ranges from < 1 % for high concentration samples to > 20 % for the lowest concentration sample (MBr1). Errors on Sr measurements are all < 3 %. Errors on Yb measurements are below 7 % with the exception of the low concentration sample MBr16 (11 %). Reproducibility of 3 peridotite reference materials (PCC-1, UB-N and JP-1) can be found in the online supplementary material. Mineral trace elements were obtained by LA ICP-MS using a New Wave Nd:YAG 213 nm laser. Error on clinopyroxene REE measurements ranges from 7 to 18 %, with the exception of MBr4 where the 2 σ error is up to an order of magnitude higher. Errors on Sr measurements range between 7 to 8 % (2 σ). The long-term reproducibility of NIST 612 reference material is presented in the online supplementary material. Mineral and sulfide major elements were determined using a Cameca SX100 electron microprobe at The Open University. Reproducibility of major elements (Mg, Fe, Si, Ca, Al, Ni) of an in-house silicate secondary standard
ranges from 1 to 9% (2σ), and for Mn 28% (2σ). For minor elements (present at < 0.015 weight%; Cl, Na, K, P, S, Ti, Cr) this increases to < 230%. Repeat analyses of an in-house pentlandite standard yields errors on major elements (S, Fe, Ni) of 2, 5 and 6% (2σ) respectively, with minor elements (Co, Cu) errors (2σ) of 22% and 70% respectively. Minerals used for electron microprobe calibration and details of the long term reproducibility of in-house standards are presented in the online supplementary material. Sulfur abundances were measured using a Leco CNS 2000 at The Open University. The uncertainty on the measurements (2σ) is ± 12%. Trapping pressures of CO2 inclusions and major element compositions of silicate melt inclusions were determined at the Laboratoire des Magmas et Volcans, University of Blaise Pascal, Clermont Ferrand, France; the latter using a Cameca SX100 electron microprobe.

For the Re-Os chemistry ~ 0.5 g of sample was dissolved by low temperature acid attack (LTAA) as in previous studies (e.g. Birck et al., 1997). To accurately determine Os isotope ratios and abundances of Re and Os a spike solution, enriched in 185Re and 190Os, was added immediately before the addition of 2 ml of Teflon-distilled (TD) 9M HBr and 2 ml of commercial grade 29M HF to prevent the combination of sample and spike forming a button-like clot which prevents effective sample / spike equilibration. Samples were digested in an oven at 140 °C for not less than 72 hours (cf. Meisel et al., 2003). The samples were removed from the oven on a daily basis, shaken vigorously, and placed in an ultrasonic bath for at least 30 minutes each day, before being returned to the oven. At the end of the digestion the sample was examined and if complete dissolution had not occurred, a further quantity of 29M HF was added, and the digestion procedure repeated. The reference material UB-N, a serpentinized peridotite, was digested several times (n = 3) to assess the efficacy of
the LTAA dissolution method for bulk-rock peridotite. Meisel et al., (2003) suggested that LTAA is not an efficient means for peridotite dissolution, often leaving resistant phases undissolved and leading to poor reproducibility of Os concentrations and Os isotope ratios. However, the digestions of UB-N, duplicate digestions of MBr bulk-rock samples (MBr9 and MBr16) by LTAA, and duplicates of an in-house peridotite standard from Kilbourne Hole, dissolved by both LTAA and Carius tubes (Shirey and Walker, 1995), performed during this study suggest that digestions in small vessels (7 mL) yield reproducible results. Osmium concentrations obtained during this study are higher than those derived by LTAA by Meisel et al., (2003) and approach the preferred Os concentration of 3.85 ppb for UB-N. Moreover, the Os isotope ratios are indistinguishable from those derived by high temperature asher (HTA) by Meisel et al., (2003). The results of several dissolutions of UB-N by LTAA and Carius tube, and a comparison with the HTA digestions of Meisel et al., (2003) are presented in the online supplementary material.

After digestion, Osmium was purified using liquid Br₂, CrO₃ in HNO₃ and HNO₃, and recovered in the Br₂, leaving Re in the supernatant fraction. The Os was subsequently recovered from the Br₂ in HBr, micro-distilled for 3 hours at 90 °C and dried. Rhodium was extracted by reducing the CrO₃-rich supernatant with a 1:1 ethanol: ultrapure H₂O solution which was then dried. The residue was redissolved in 2M TD HNO₃ and Re was recovered in isoamylol, cleaned in a wash of 2M TD HNO₃ and finally extracted in ultrapure H₂O (Birck et al., 1997). The total procedural blank for bulk-rock Os during this study was 0.29 ± 0.26 pg, ¹⁸⁷Os/¹⁸⁸Os = 0.2003 ± 0.0639, and for bulk-rock Re 3.13 ± 2.37 pg. The short-term reproducibility of the blank was ± 10 % or better, but the absolute values varied between each batch of prepared reagents. Re-Os chemistry for individual hand-picked sulfide grains was achieved
using a modified microdistillation technique that closely follows previously reported
procedures (Pearson et al., 1998). The total procedural blanks for Os in individual
sulfides were 0.10 ± 0.06 pg, $^{187}\text{Os}/^{188}\text{Os} = 0.36 ± 0.50$, and for Re 3.14 ± 0.61 pg.
Both Re and Os samples were analysed on platinum filaments using negative thermal
ionisation mass spectrometry (N-TIMS) (Volkening et al., 1991; Creaser et al., 1991)
using a Thermo Scientific Triton. Long term reproducibility of a Johnson Matthey
standard solution (14 pg – 8 ng; n=85) is generally within 0.1% of the recommended
values for $^{187}\text{Os}/^{188}\text{Os}$. Details of the analysis of the standard solution is presented in
the online supplementary material.
4. RESULTS

4.1. Bulk-rock analyses

Mineral modal abundances, calculated using a least squares linear regression (Tarantola and Valette, 1982), demonstrate that, with the exception of MBr16, all of the samples studied are spinel lherzolites. MBr16, with clinopyroxene modal abundance of 3.3 (± 1.5) % is the only spinel harzburgite (mineral modal abundances are presented in the online supplementary material). Selected bulk-rock major element abundances are presented Table 1 (full major element analyses are presented in the online supplementary material). Co-variations between MgO/SiO$_2$ and Al$_2$O$_3$/SiO$_2$ (Figure 2) coupled with positive co-variations between bulk rock CaO weight % and Al$_2$O$_3$ weight %, bulk-rock Mg# (Mg# = (MgO/40.32)/(MgO/40.32)+[(Fe$_2$O$_3$/Total)/79.8×0.9])] of 88-91 and the given range of clinopyroxene modal abundance indicates that these xenoliths are compositionally similar to other Massif Central xenoliths (Downes and Dupuy, 1987; Lenoir et al., 2000; Zangana et al., 1999; Figure 2), non-cratonic xenoliths worldwide (e.g. Nixon and Boyd, 1973; Sobolev, 1974; Boyd and Nixon, 1978; Frey and Prinz, 1978; Frey and Green, 1974; Gurney and Hart, 1980; Irving, 1980; Stosch and Seck, 1980; Roden et al., 1988; Kopylova et al., 1999) and massif peridotites (Boudier and Nicolas, 1977; Bodinier, 1988; Menzies and Allen, 1974; Frey et al., 1991; Dickey, 1970; Frey et al., 1985; Hartmann and Wedepohl, 1993; Carswell, 1968; Bodinier and Godard, 2004)
and references therein), but typically less refractory than cratonic xenoliths (e.g. Nixon and Boyd, 1973; Sobolev, 1974; Boyd and Nixon, 1978; Gurney and Hart, 1980; Kopylova et al., 1999) and depleted abyssal peridotites (e.g. Alard et al., 2005; Harvey et al., 2006; Liu et al., 2008). Bulk-rock major element compositions therefore suggest that most of these xenoliths may be related by varying, yet moderate, degrees of melt depletion (< c. 18%) prior to being brought to the surface in the host basalt.

Unlike bulk-rock massif peridotites such as Lherz, which extend to 5 weight % Al$_2$O$_3$ (e.g. Le Roux et al., 2007) no Mont Briançon bulk-rock lherzolites have more than 3.4 weight % Al$_2$O$_3$. Bulk-rock co-variations of Cr$_2$O$_3$, Al$_2$O$_3$ and MgO of Mont Briançon xenoliths all co-vary strongly. This contrasts with similar bivariate plots for re-fertilized massif peridotites which commonly lie on two distinct trends (e.g. Le Roux et al., 2007). Many of the Mont Briançon xenoliths lie within a similar range of bulk-rock Al$_2$O$_3$ weight % (1.6 < Al$_2$O$_3$ <3.5 weight %) to Tok cratonic peridotites (Ionov et al., 2006) ascribed as having melt depletion signatures, and only one sample (MBr16) can be described as particularly refractory (Al$_2$O$_3$ < 1.3%; Ionov et al., 2006) and therefore particularly susceptible to subsequent metasomatism (McDonough and Frey, 1989). Bulk-rock Al$_2$O$_3$ abundances calculated from the product of mineral modes and the Al$_2$O$_3$ weight % of the constituent minerals indicate that a significant proportion of the bulk-rock Al$_2$O$_3$ cannot be accounted for by the silicate minerals and spinel. Taking into account the uncertainties in the modal abundance calculations a grain boundary phase accounts for at least 8 % of the bulk-rock Al$_2$O$_3$ of MBr13 and as much as 33 % of MBr28.

With the exception of MBr13, which have elements of both protogranular and porphyroclastic textures and a relatively high olivine modal abundance for its Mg#, the Mont Briançon xenolith suite demonstrates a broad positive co-variation between
olivine modal abundance and bulk-rock Mg#. This contrasts with many tectonically
emplaced peridotite suites whose longer transit to the surface has led to more
complicated bulk-rock and mineralogical systematics, e.g. ophiolite peridotite from
the Ligurian Alps (Rampone et al., 2005; Rampone and Borghini, 2008). Bulk-rock
abundances of moderately incompatible elements, e.g. Yb, Sc and V all show strongly
covarying trends.

Co-variations between bulk-rock major element indices of melt depletion and
incompatible trace element abundances are notably absent, and many xenoliths have
moderate to strong enrichments in incompatible elements. Three different
incompatible trace element fingerprints are evident (Figure 4). The first group of
xenoliths (n = 7) is depleted in light rare earth elements (LREE), i.e. Ce/Yb ratios
normalized to primitive mantle after Palme and O’Neil, (2004) ([Ce/Yb]N) range from
0.2 to 0.6. REE fractionation is accompanied by a moderate relative enrichment in Zr,
Nb and sometimes Ti, and moderate relative enrichments of Sr. In contrast, a second
group of samples (n = 7) possess ([Ce/Yb]N ≤ 2. In the absence of metasomatic
accessory phases (e.g. phlogopite, amphibole, apatite) this is indicative of cryptic
LREE enrichment. A single sample (MBr6) is characterised by a flat, yet depleted,
MREE – HREE pattern, however ([Ce/Yb]N is highly enriched (4) and is accompanied
by a large Sr enrichment. With the exception of MBr6, which is anomalous, the first
two groups strongly resemble the incompatible trace element patterns for southern
domain FMC xenoliths of Lenoir et al., (2000). Like the other 7 xenoliths with
elevated ([Ce/Yb]N ratios, the trace element composition of MBr6 cannot be attributed
to melt depletion alone and must be due to subsequent interaction with a metasomatic
agent or agents.
4.2. Mineral compositions
4.2.1. Silicates and spinel

Electron microprobe analyses of the constituent phases of 18 xenoliths (online supplementary material) reveal that olivine Mg# ranges from 0.88-0.91 and co-varies with Ni content. Mineral major element compositions are, in most samples, similar to spinel-facies peridotite xenoliths from non-cratonic settings worldwide (e.g. Frey and Green, 1974; Ionov et al., 1993; Frey and Prinz, 1978; Wilshire and Shervais, 1975; Stosch and Seck, 1980). For example, olivine MgO ranges from 46.85 (± 0.18) to 49.70 (± 0.85) weight %; clinopyroxene CaO and Na₂O range from 19.35 (± 0.31) to 20.77 (± 0.15) and 1.29 (± 0.04) to 1.92 (± 0.06) weight % respectively; orthopyroxene Al₂O₃ ranges from 2.76 (± 0.20) to 4.74 (± 0.39) weight %; spinel Cr₂O₃ and Al₂O₃ range from 8.75 (± 0.21) to 23.79 (± 0.98) and 42.71 (± 0.85) to 57.97 (± 0.39) weight % respectively. Bulk-rock TiO₂ strongly co-varies with clinopyroxene TiO₂ abundance; a continuum between the most and least refractory peridotites being evident. Three xenoliths (MBr3, 13 and 24) are exceptional in that they have higher Al₂O₃ in orthopyroxene for a given co-existing spinel Cr# (Cr# = ([Cr₂O₃/151.99]/([Cr₂O₃/151.99]+[Al₂O₃/101.96]))). MBr24 is the only sample with a porphyroclastic texture and MBr13 is one of only two examples transitional between protogranular and porphyroclastic, suggesting that mineral composition may be related to xenolith texture (cf. Brown et al., 1980; Zangana et al., 1999).

Trace element compositions of the constituent major phases of MBr2, MBr3, MBr6 and MBr8 were obtained by laser ablation ICP-MS (online supplementary material). These samples were selected as being representative of the three contrasting bulk-rock trace element profiles described in section 4.1. Chrome-diopside
incompatible trace element patterns closely match those of the corresponding bulk-rock. Clinopyroxene \((\text{Ce}/\text{Yb})_N\) range from 0.27 to 1.35. The lowest values, \((\text{Ce}/\text{Yb})_N = 0.27\) and 0.34, are derived from MBr1 and MBr8 respectively; two of the xenoliths whose bulk-rock \((\text{Ce}/\text{Yb})_N\) is indicative of melt depletion. Clinopyroxene from MBr2 and MBr3 have \((\text{Ce}/\text{Yb})_N\) of 0.60 and 1.35 respectively, whereas MBr6, the sample with the most elevated bulk-rock \((\text{Ce}/\text{Yb})_N\) has a clinopyroxene \((\text{Ce}/\text{Yb})_N\) of 1.09. This suggests that, in many xenoliths, the majority of the bulk-rock trace element inventory can be accounted for by clinopyroxene. However, the \text{Ce}/\text{Yb} ratio of MBr6 clinopyroxene is much lower than its corresponding bulk-rock which suggests that there is a significant LREE component not accounted for by clinopyroxene. Relative depletions in clinopyroxene HFSE and Ti are most pronounced in MBr6, less so in MBr3, MBr2 and MBr8. These depletions are complemented by enrichments in co-existing orthopyroxene, an antithetic relationship common in peridotite (e.g. Rampone et al., 1991; Rivallenti et al., 1996).

The incompatible trace element budget of several other Mont Briançon xenoliths cannot wholly be accounted for by their constituent minerals. While the majority of MREE and HREE are contained in the major peridotite minerals, deficiencies in the mass balances become more pronounced in the LREE and, with the exception of U and Th, increase with growing incompatibility.

4.2.2. Melt inclusions

Trails of melt inclusions trapped within fractures in silicate minerals are common in Mont Briançon xenoliths (Schiano and Clocchiatti, 1994) and often cross-cut individual silicate grains before terminating at grain boundaries lined with
interstitial glass. Individual melt inclusions rarely exceed 10 μm in diameter, and
comprise silicate glass with rare daughter phases located at the inclusion walls. At
least one fluid bubble is preserved within each melt inclusion. Experiments on a
cryometric stage and laser-Raman analyses reveal that these fluids are almost pure
CO₂ and became trapped within their host phase at pressures of at least 1.3 GPa, i.e. a
depth of at least 40 km (Schiano and Clocchiatti, 1994). Therefore, when the fracture
hosting the melt inclusions annealed, the trapped melt, which now encloses the fluid
inclusions, was CO₂ saturated. Discrete CO₂ inclusions not associated with silicate
glass and trapped at shallower depths are also common. The CO₂ saturated silicate
melt inclusions are also associated with minute sulfide blebs (≤ 10 μm diameter)
within the silicate / CO₂ inclusion trails. In rare examples all three co-genetic phases,
i.e. silicate glass, CO₂ fluid and sulfide, are present within a single inclusion and the
immiscible nature of the co-genetic silicate and sulfide phases is clear (Figure 4).

A total of 54 glassy inclusions in 6 Mont Briançon xenoliths were analysed for
major element abundance, along with S and Cl concentrations (supplementary online
material). The majority (n = 35) are hosted in olivine grains. The remainder are
included in clinopyroxene (n = 15) or orthopyroxene (n = 4). Inclusions were not re-
homogenised prior to analysis as only inclusions with no daughter minerals were
analysed. High Na₂O + K₂O (≤ 7.8 weight %), S abundances (320-2960 ppm) and
Al₂O₃ and SiO₂ (≤ 24.4 and 61.9 weight %, respectively), are consistent with previous
measurements of re-homogenised Mont Briançon melt inclusions, and similar to
inclusions from both oceanic and continental intra-plate settings (Schiano with
Clocchiatti, 1994).

4.2.3. Sulfide grains
Major element analyses (S, Fe, Ni, Cu, Co – online supplementary material) were made on 62 discrete sulfide blebs using a Cameca SX100 electron microprobe. The sulfides can be divided into 5 distinct groups based upon textural constraints, i.e. interstitial or wholly enclosed within a silicate grain, degree of weathering, and proximity to interstitial glass and corroded spinel. The degree of alteration of each sulfide was assessed qualitatively under reflected light and quantitatively on the electron microprobe by assessing measured S/O ratios.

Unweathered included, often rounded, sulfide grains (Group A, Figure 5a) have the most uniform compositions (Figure 6), and are characterized by high Ni/Fe (0.19 – 1.18) and high S/O ratios (14.21 – 166.93). Although the S/O ratio suggests that even these “unweathered” sulfides have, in some instances, experience some alteration, this group of sulfide are the closest to a pentlandite mineralogy as previously observed in other non-cratonic peridotite xenoliths (e.g. Dromgoole and Pasteris, 1987). Often, these seemingly isolated sulfides are associated with trails of micron-scale sulfide blebs which commonly terminate at grain boundaries (Figure 5b). Weathered included sulfides (Group B), by contrast, have lower Ni/Fe and S/O ratios (0.03 – 0.50 and 0.003 – 1.06, respectively) and deviate from a simple pentlandite-like composition. The weathered sulfides in general no longer possess conventional sulfide compositions because of the formation of iron oxy-hydroxides, and perturbed metal : sulfur ratios. Interstitial grains are often less rounded, more ragged and even elongate in shape (Figure 5c with 5d), and occasionally infill cracks in deformed silicate grains (Figure 5d). The Ni/Fe ratios of weathered interstitial sulfides (Group C, Figure 5e) range from 0.03 – 0.57 and S/O ratios 0.003 – 3.01, i.e. a field that wholly encompasses that of weathered included sulfides. Unweathered
interstitial grains (Group D, Figure 5d) form two separate compositional groups in Ni/Fe vs. S/O space. Half of the grains are compositionally indistinguishable from weathered interstitial grains, whereas the remainder have compositions similar to the unweathered included grains with the highest Ni/Fe ratios (~1).

The final group of sulfides (Group E, Figure 5f) are predominantly rounded, although much smaller than included sulfides, are generally unweathered in appearance, and exclusively associated with interstitial glass and corroded spinels. Compositionally, these sulfides are very similar to unweathered interstitial grains and exhibit a similar bimodal distribution in Ni/Fe vs. S/O compositional space.

4.3. Re-Os concentration and isotope ratios

4.3.1. Bulk-rock analyses

Sixteen xenoliths were analysed for Re and Os elemental abundance and isotope ratios (Table 2). Osmium concentrations (0.83 – 2.19 ppb) are similar to other non-cratonic peridotite xenoliths worldwide (e.g. Meisel et al., 2001) but lower than those obtained from orogenic massifs (e.g. Reisberg and Lorand, 1995), cratonic peridotite xenoliths (≤ 9.2 ppb, Pearson et al., 1995a, 1995b; Chesley et al., 1999; Gao et al., 2002; Irving et al., 2003; Wu et al., 2003; Yuan et al., 2007), the average upper mantle concentration of 3.1 (± 0.3) ppb (Morgan, 1986) and the calculated primitive upper mantle (PUM) value of 3.9 (± 0.5) ppb (Becker et al., 2006). Rhenium concentrations range from 0.063 to 0.634 ppb and show a bimodal distribution. Six samples have Re concentrations that are similar to or less than the concentration inferred for fertile mantle, i.e. 0.260 ppb (Morgan, 1986), and significantly below that
estimated for PUM (0.35 ± 0.06; Becker et al., 2006) while the remainder have Re concentrations far in excess of fertile mantle. Consequently, more than half of the xenoliths analysed have suprachondritic Re/Os ratios (> 0.406, Shirey and Walker, 1998).

Bulk-rock $^{187}\text{Os}/^{188}\text{Os}$ ratios range from 0.11541 (± 0.00021) to 0.12897 (± 0.00032) i.e. below the maximum present day estimate of the primitive upper mantle (PUM $^{187}\text{Os}/^{188}\text{Os} \leq 0.1292$, Meisel et al., 2001), and all but one (MBr1) below the present day mean chondrite value i.e. $^{187}\text{Os}/^{188}\text{Os} < 0.127$ (Shirey with Walker, 1998). Bulk-rock $^{187}\text{Re}/^{188}\text{Os}$ ratios range from 0.112 to 3.436, possess a wide range of $^{187}\text{Re}/^{188}\text{Os}$ values for a given Os isotope ratio, show no clear correlation with $^{187}\text{Os}/^{188}\text{Os}$, and therefore yield no meaningful isochron age information (Figure 7a). The scatter between $^{187}\text{Re}/^{188}\text{Os}$ and $^{187}\text{Os}/^{188}\text{Os}$ is, in general, much higher in xenolith suites than massif peridotites (Reisberg and Lorand, 1995) but is particularly high amongst the xenoliths from Mont Briançon. Scatter is usually attributed to Re mobility, as this element, unlike Os, may be mobile under a wide range of conditions (e.g. Reisberg and Lorand, 1995; Meisel et al., 2001; Sun et al., 2003). The use of a suitable immobile proxy for bulk-rock $^{187}\text{Re}/^{188}\text{Os}$ has sometimes yielded more coherent correlations with $^{187}\text{Os}/^{188}\text{Os}$ in instances where Re abundance may have been disturbed. In a suite of peridotites related by a single stage of melt depletion from a homogenous starting composition, one should find a linear correlation between the index of melt depletion (e.g., Al$_2$O$_3$) and $^{187}\text{Os}/^{188}\text{Os}$ (e.g. Rudnick and Walker, 2009). While proxies such as Al$_2$O$_3$ (Reisberg and Lorand 1995), [S], (Burnham et al., 1998) and [Yb], (Reisberg et al., 2005) may reduce scatter on disturbed isochron-type plots no significant improvement is obtained using any of these proxies in this study (Figure 7b, 7c and 7d respectively).
4.3.2. Mineral separates

Optically pure mineral separates of olivine, orthopyroxene and clinopyroxene from MBr2, 3, 6, 8 and 20 yield Os concentrations (Table 3) of 3.7 - 29.3 ppt in olivine, 8.0 - 91.6 ppt in orthopyroxene (the value of 815 ppt for MBr3 is probably derived from an unseen sulfide included within one or more of the grains, cf. Burton et al., 1999), and 28.2 - 185.1 ppt in clinopyroxene. While Os concentrations in spinel range from 11.7 ppt to 450.8 ppt, the opacity of this mineral makes it difficult to assess whether this is due to spinel-hosted Os, sulfide contamination or a combination of the two. Notwithstanding the possibility of sulfide contamination, mean Os abundance increases thus: olivine < orthopyroxene < clinopyroxene < spinel, with absolute values similar to those measured by Hart and Ravizza, (1996) and Burton et al., (1999). The measured Os concentrations do not therefore simply reflect the effects of variable degrees of sulfide incorporation. Rhenium concentrations of the same silicate phases (Table 3) are 21.8 – 94.3 ppt in olivine, 35.3 – 239 ppt in orthopyroxene and 61.2 - 421 ppt in clinopyroxene. Rhenium concentrations in spinel are higher than in the silicate phases (80.2 - 924 ppt), but like Os, the origin of Re in spinel is not unequivocal. Therefore, mean Re concentrations increase thus: olivine < orthopyroxene < clinopyroxene < spinel.

Mineral $^{187}$Os/$^{188}$Os ratios (Table 3) range from 0.120 to 0.136 for olivine, 0.121 to 0.142 for orthopyroxene, 0.120 to 0.203 in clinopyroxene and 0.128 and 0.291 in spinel (Figure 8). As previously reported (Burton et al., 1999) mineral Os isotope ratios are consistently higher than their corresponding bulk-rock. The $^{187}$Re/$^{188}$Os ratios of the mineral phases range from 2.05 to 78.5 for olivine, 0.58 to
125 for orthopyroxene and 5.7 to 187 for clinopyroxene, while spinel $^{187}\text{Re}/^{188}\text{Os}$ ranges from 4.53 to 181. Mean $^{187}\text{Re}/^{188}\text{Os}$ ratios for all of the separated phases are much greater than present day chondrite, i.e. $^{187}\text{Re}/^{188}\text{Os} = 0.406$ (Shirey and Walker, 1998), however there is no individual sample that appears systematically enriched in Re across all of its constituent phases. Moreover, no particular phase across the wide compositional range of xenoliths selected appears to be either particularly susceptible to, or immune from, Re enrichment. Silicates show little variability in $^{187}\text{Os}/^{188}\text{Os}$ ratio for a wide range of $^{187}\text{Re}/^{188}\text{Os}$ and closely resemble the Os isotope composition of the bulk-rock, although each sample has at least one phase significantly more radiogenic than the rest.

4.3.3. Individual sulfide grains

In addition to the optically pure mineral separates, 14 individual sulfide grains were recovered from the crushate of MBr3 (n = 3), MBr6 (n = 4), MBr8 (n = 3), and MBr20 (n = 4). Both Re and Os concentrations vary widely (0.01 – 412 ppm, and 0.047 – 40.84 ppm respectively), and a wide range of Os isotope ratios was also obtained, ranging from $^{187}\text{Os}/^{188}\text{Os} = 0.115$ to 0.171 (Figure 9) (Table 3). Many of the sulfides (n=9) have subchondritic Os isotope ratios (0.115 – 0.126) that yield $T_{\text{RD}}$ model ages of up to 1.78 Ga (Figure 10), that is, similar to the oldest $T_{\text{RD}}$ obtained from bulk-rock Os isotope analyses (1.71 Ga).
5. DISCUSSION

Melt depletion and subsequent secondary melt and/or fluid infiltration, either as individual or multiple events, are thought to be the main processes responsible for the present-day composition of peridotite xenoliths. Depletion of SCLM by removal of a basaltic melt results in systematic trends in bulk-rock major and trace element composition, modal mineralogy and, crucially for dating melt depletion events in the SCLM, the systematic fractionation of parent-daughter elements that comprise the traditional radiogenic isotope systems. However, metasomatism systematically obscures the elemental and isotopic fingerprints of ancient melt depletion, overprinting or completely erasing evidence for earlier events. Early studies suggested that the Re-Os isotope system is robust against the effects of metasomatism (e.g. Carlson and Irving, 1994; Pearson et al., 1995a) as bulk-rock Re and Os abundance and isotope ratios are controlled almost entirely by volumetrically insignificant sulfide (e.g. Hart and Ravizza, 1996; Burton et al., 1999; Alard et al., 2002; Harvey et al., 2006). However, interaction with sulfur-saturated metasomatic melts or fluids can result in the precipitation of secondary sulfides which in turn make bulk-rock Re-Os isotope analyses simply the weighted mean of the two generations of sulfide, yielding complex geochronological information. Despite the interaction with metasomatic agents, geochemical evidence of early melt depletion may still be preserved in sulfide grains trapped within host silicates and thus protected from interaction with metasomatic melts or fluids. By systematically dis-assembling individual xenoliths
and examining their constituent components it is still possible to extract information regarding the relative timing of melt depletion and subsequent, secondary processes.

5.1. Petrological and major element evidence for melt depletion and subsequent metasomatism

5.1.1. Melt depletion

The SCLM underlying the Massif Central is considered to have experienced multiple melt depletion events; (i) subduction related volcanism prior to the Variscan orogeny (Pin and Marini, 1993; Alexandrov et al., 2000; Lardeaux et al., 2001; Wittig et al., 2007), (ii) at 257 Myr ago (Femenias et al., 2004), and (iii) voluminous Pliocene volcanism (c.1 – 4 Ma). Strong co-variations between traditional indices of melt depletion (e.g. CaO weight %, Al₂O₃ weight % and MgO weight %) suggest that evidence for partial melting is, at least in part, preserved. While clinopyroxene modal abundances (3.3 ± 1.5 to 17.2 ± 1.6 modal %) are consistent with variable degrees of melt depletion, the ranges of modal abundance of all the major silicate phases are not consistent with melting experiments of synthetic peridotites, natural lherzolites or pyrolites (Baker and Stolper, 1994; Baker et al., 1995). Repeated extraction of smaller proportions of melt, i.e. incremental batch melting (Robinson et al. 1998), produces ranges of mineral modal abundance consistent with the multiple episodes of melt depletion known to have occurred in the southern French Massif Central.

Many peridotites are re-fertilized after an earlier period of melt depletion, and the precipitation of secondary clinopyroxene is an almost ubiquitous process. Abyssal peridotite (Seyler et al., 2007) and orogenic massifs such as Lherz (Le Roux et al.,
2007) and Totalp (van Acken et al., 2008) all preserve evidence for such re-fertilization events. However, evidence is scant for the addition of secondary 
clinopyroxene in the Mont Briançon suite. Firstly, the range of clinopyroxene modal 
abundance at Mont Briançon is consistent with melt depletion; re-fertilization such as 
that observed at Lherz (Le Roux, 2007) would have the potential for extending the 
compositional range of moderately melt depleted lherzolites into that of pyroxenites, 
i.e. if re-fertilization had occurred, there is no reason why it should stop at c. 17 modal 
% clinopyroxene. Moreover, there is no textural evidence for re-fertilization. All of 
the Mont Briançon peridotites possess protogranular to porphyroclastic textures; and 
no evidence of mechanical mixing of peridotite and pyroxenite at the centimetre scale 
(cf. Le Roux, 2007; van Acken et al., 2008) is evident. The chemical composition of 
the Mont Briançon clinopyroxenes, discussed below, also suggests that they are part 
of an equilibrium melt depleted residue and that evidence for a secondary origin for 
some or all of the clinopyroxene is scant.

Strong correlations between bulk-rock TiO₂ and clinopyroxene TiO₂ 
abundances suggests that the more fertile MBr lherzolites are linked to the less fertile 
lherzolites and harzburgite by melt depletion. These bulk-rock characteristics contrast 
with arched patterns seen in re-fertilized massif peridotites (Bodinier and Godard, 
2004, and references therein). Moreover, no evidence for wholesale olivine 
precipitation or pyroxene dissolution is evident from an examination of the xenoliths 
in hand specimen and thin section or from calculated modal abundances. Apart from 
MBr13, with elements of both protogranular and porphyroclastic textures and has a 
relatively high olivine modal abundance for its Mg#, the co-variation between olivine 
modal abundance and bulk-rock Mg# is consistent with melt depletion. There is no 
petrological and very little major element evidence for significant degrees of melt /
rock interaction, i.e. the formation of melt channels such as those described by e.g. Kelemen et al., (1992) Rampone et al (2005) Rampone and Borghini (2008) and Buchl et al., (2002) can be discounted. The broad inverse co-variation of clinopyroxene modal abundance with olivine Mg# also suggests that Mont Briançon clinopyroxene is primary.

Bulk-rock Al$_2$O$_3$ abundances of 0.8 – 1.0 weight % are consistent with a peridotite residue containing 3-5 modal % clinopyroxene (Canil, 2004), the result of partial melting at significantly less than F = 0.20 which is consistent with the most depleted Mont Briançon peridotites. In addition, evidence for the likely depth of at least some of the melt extraction and the redox conditions within the mantle at that depth can be obtained from the systematic behaviour of moderately incompatible elements. Bulk-rock [Yb] vs Al$_2$O$_3$ suggests that the Mont Briançon xenoliths lost at least some of their melt at < 3.0 GPa. The behaviour of V during mantle melting is sensitive to fO$_2$ and variations of this element in peridotite can provide a paleoredox indicator (Canil and Fedortchouk, 2000). Bulk-rock [V] vs. Al$_2$O$_3$ suggests an fO$_2$ of between NNO-2 and NNO (NNO = nickel-nickel-oxide) at 1.5 GPa (Canil, 2004).

This is consistent with melt loss in a mantle environment more oxidizing than that of abyssal peridotites and similar to peridotites whose melt depletion signature has been attributed to subduction related melting, typically at fO$_2$ of NNO-3 to NNO (Canil, 2002). However, bulk-rock [V] vs. [Yb] is a better discriminator between the effects of pressure and fO$_2$ (Canil, 2004) and places the Mont Briançon peridotite suite closer to an fO$_2$ of NNO at 1.5 GPa than bulk-rock [V] vs. Al$_2$O$_3$ weight %.

Only one xenolith (MBr16) can be described as particularly refractory. Siberian cratonic peridotites from Tok (Ionov et al., 2006) and Udachnaya (Boyd et al., 1997) with < 1.6 weight % Al$_2$O$_3$ retain evidence for melt extraction in excess of
20\%, and show the greatest degree of metasomatism. Cratonic xenoliths (e.g. Ionov et al., 2006) with intermediate bulk-rock Al$_2$O$_3$ abundances (1.6 < Al$_2$O$_3$ < 3.5 weight \%) preserve the best melt depletion fingerprint. This range encompasses all but 3 samples from this study (MBr1, 6, 19).

5.1.2. Metasomatism

Given the prior melt depletion history of the Massif Central region, it is unlikely that local segments of the SCLM escaped melt depletion entirely. However, several samples (n = 7) have bulk-rock Al$_2$O$_3$ and CaO abundances that approach primitive upper mantle values (Hart and Zindler, 1986; Jagoutz et al., 1979; Palme and Nickel, 1985; McDonough, 1990; McDonough and Sun, 1995), which could suggest that these xenoliths have been re-fertilized through the addition of metasomatic pyroxene. In addition, because of the similar incompatibility of Na and Al in clinopyroxene, which dominates bulk-rock Na and Al budgets, the bulk-rock abundances of these elements should co-vary systematically if only partial melting had influenced their composition. Mont Briançon clinopyroxenes have a wide range of Na$_2$O abundance for a given Al$_2$O$_3$ weight \% which is difficult to reconcile solely with melt depletion. Metasomatic clinopyroxene addition would have little discernable effect on bulk-rock composition unless, as discussed above, the degree of re-fertilization was such that > c.17 modal \% clinopyroxene resulted. A scenario could be envisaged where all of the clinopyroxene in the Mont Briançon suite is secondary, but there is no evidence for all of the xenoliths being previously depleted to the point of clinopyroxene exhaustion, followed by the subsequent addition of varying amounts of secondary clinopyroxene.
No discrete metasomatic phases (e.g. apatite, phlogopite, amphibole) have been observed at this locality (Rosenbaum et al., 1997; Alard et al., 2002; this study). Petrographic evidence for metasomatism is however provided by the presence of trails of melt inclusions preserved in annealed fractures and texturally linked to interstitial glass. Highly silicic, alkali-enriched melts found as inclusions or as interstitial melts have been observed in anhydrous peridotite xenoliths worldwide (e.g. Frey and Green, 1974; Stosch and Seck, 1980, Schiano and Clocchiatti, 1994; Yaxley et al., 1999) and have been variously attributed to decompression melting of pre-existing metasomatic phases, the migration of volatile-rich melts, deeply sourced mantle fluids, and small volume partial melts from a peridotite source. Experimentally, highly silicic, alkali-rich melts can be generated by very low degrees of melting of a primitive mantle-like precursor at pressures around 1 GPa (Baker and Stolper, 1994; Baker et al., 1995; Hirschmann et al., 1998, Schiano et al., 1998). However, based upon their major-element compositions that differ drastically from upper mantle lithologies, their systematic association with CO₂-fluid inclusions and the presence of multi-phase, multiply volatile-saturated inclusion assemblages, it has also been proposed that they could act as metasomatic agents of the upper mantle. Similarly, P-rich grain boundary phases (Rosenbaum et al., 1997) cannot be attributed to basaltic melt.

Moreover, the discrepancy between bulk-rock Al₂O₃ weight % and that calculated from the sum of the main silicate minerals and spinel cannot be ignored. A significant reservoir of Al₂O₃ can only be accounted for by a grain boundary phase. Figure 4 illustrates that melt inclusions that terminate in areas of interstitial and intergranular glass are clearly the product of a secondary, metasomatic process. Irrespective of their mode of generation, discussed further in section 5.4 below, they comprise 20 – 24 weight % Al₂O₃ (online supplementary material). Combined with
the contribution from the interstitial glasses they therefore contribute a significant
proportion of bulk-rock Al₂O₃. Unlike other metasomatized European spinel
lherzolites brought to the surface during Cenozoic volcanism, e.g. Ackerman et al.,
(2009), large melt / rock ratios are not responsible for metasomatism of Mont
Briançon peridotites. In addition to the elevation of bulk-rock Al₂O₃, 5 samples, with
either porphyroclastic textures or textures transitional towards porphyroclastic (MBr1,
2, 13, 15 and 24), have elevated bulk-rock FeO (> 8.5 weight %) which cannot be
attributed to shallow (1-2 GPa) polybaric melting (Walter, 2003; Herzberg, 2004).
Ionov and Hofmann, (2007) attribute elevated FeO (> 8.5 weight %) to post-melting
interaction of melt depleted peridotite with percolating Fe-rich melt. However, the
majority of Mont Briançon peridotites have lower bulk-rock FeO (< 8.5 weight %)
with bulk-rock Al₂O₃ that ranges from 0.8 to 2.2 weight %, consistent with melt
depletion conditions outlined above. Other possible indicators of re-fertilization in
peridotites are discussed by van Acken et al., (2008) and Walter, (2003) and
references therein, particularly the generation of linear trends between bulk-rock
Na₂O, MgO and Al₂O₃. In instances where up to 10 % basaltic melt has infiltrated the
peridotite, e.g. at Lherz (Le Roux et al., 2007, 2008) or in other massif peridotites
(Bodinier and Godard, 2004) this re-fertilization is obvious and consistent with
precipitation of clinopyroxene from melts during porous melt flow and melt–rock
reaction (Müntener et al., 2004) but has not occurred at Mont Briançon where there
are no such co-variations and an absence of textural evidence.

5.2. Trace element evidence for melt depletion and subsequent metasomatism

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Trace element analyses of Mont Briançon bulk-rocks and corresponding clinopyroxenes demonstrate decreasing LREE/HREE consistent with melt depletion (Figure 3a). Ophiolitic clinopyroxenes also commonly preserve depleted LREE profiles (Rampone et al., 2005; Rampone and Borghini, 2008), although in these peridotites the incompatible trace element patterns are ascribed to the precipitation of secondary clinopyroxenes from impregnating melts derived from depleted melt fractions. However, in addition to LREE depletion, ophiolite clinopyroxenes often possess convex-upwards REE patterns or high HREE/MREE which suggests equilibration with evolved silicate melts at high melt-rock ratios, for which no evidence is preserved at Mont Briançon. But, partial melting cannot account for the trace element abundances of the majority of this xenolith suite (Figures 3b and 3c).

Globally, LREE enrichment is a ubiquitous feature of harzburgites and depleted lherzolites (McDonough and Frey, 1989). Interaction with a transitory metasomatising melt or fluid has the greatest effect on the most depleted peridotites due to their lower bulk distribution coefficients. Three samples from this study (MBr6, 13, and 19), characterised by high Ce/Yb ratios for a given Ce concentration, are indicative of such an enrichment process. The low Yb concentrations, coupled with low modal clinopyroxene in two of the three samples also suggests that they experienced a high degree of prior melt depletion.

MBr13 and MBr19, the samples with transitional porphyroclastic textures, also have higher Ce/Yb ratios than the remainder of the suite. While MBr13 has a higher clinopyroxene content (14.3 modal %), its anomalously low Mg# (bulk-rock = 0.863; clinopyroxene = 0.868) is clearly not attributable to melt depletion (Pearson et al., 2003 and references therein). It is conceivable that this xenolith has experienced some re-fertilization through the addition of clinopyroxene and that its bulk-rock
characteristics have been substantially modified through interaction with a
metasomatic melt or fluid, although major element and petrographic evidence is
lacking. Elsewhere in the southern Massif Central the effects of metasomatism are
also generally restricted to the most refractory peridotites (Downes and Dupuy, 1987;
Xu et al., 1998), and to deformed facies (Downes, 1990). A feedback relationship
between percolation and deformation that concentrates metasomatic fluids in
lithospheric shear zones has been postulated as the tectonic mechanism controlling
melt / fluid mobility (Downes, 1990). In both brittle (Wilshire and Kirby, 1989) and
ductile stress regimes (Downes, 1990), LREE enrichment accompanies the
deformation of peridotite which could account for the enriched characteristics of the
most deformed samples from the Mont Briançon suite (cf. Zangana et al., 1997).
Moreover, the xenoliths with little or no LREE enrichment have lower equilibration
temperatures (887 °C – 964 °C) than the xenoliths with the most enriched composition
(962 °C – 1019 °C) (Wells, 1977). Griffin et al. (2004) suggested that metasomatized
cratonic peridotite xenoliths are more likely to have equilibrated at greater depths in
the lithosphere and are therefore more susceptible to interaction with asthenospheric
melts at the base of the non-cratonic lithosphere. Mont Briançon xenoliths
demonstrate a similar relationship between depth of final equilibration and the degree
of metasomatism, suggesting that enrichment is ubiquitous at or near the base of the
lithosphere, consistent with the hypotheses of previous studies (Eggerl, 1987; Zindler
and Jagoutz, 1988; McKenzie 1989, Chesley et al., 1999; Ackerman et al., 2009).
However, one of the most depleted samples studied here (MBr6, 5.1 modal %
clinopyroxene) possesses the highest bulk-rock and clinopyroxene Ce/Yb ratios, and
yet has one of the lowest equilibration temperatures of the entire suite (903°C),
implying that either not all enrichment occurs at the base of the lithosphere, or
additional processes have modified the composition of this xenolith, e.g. fluid flow through shear zones (Downes, 1990).

It is possible to account for the bulk-rock budget of peridotite xenoliths from the summed contribution of their major constituent phases (e.g. Eggins et al., 1998). Occasionally samples with trace element budgets entirely attributable to the major upper mantle phases can be found in the same locality as xenoliths that require an additional trace element reservoir in order to effect a mass balance (e.g. Condie et al., 2004). More commonly, it is not possible to account for bulk-rock incompatible trace element abundances simply from their concentrations in the major silicate phases. The majority of Mont Briançon xenoliths contain melt inclusions, but there is no unequivocal relationship between the presence or absence of inclusions and trace element evidence of metasomatism. While the effect of the volumetrically insignificant melt inclusions on bulk-rock trace element abundances is most likely small, their presence is indicative of a larger reservoir of incompatible trace elements derived from interstitial metasomatic melt nearby in the SCLM. Figure 11 illustrates the composition of the melt needed to balance the incompatible trace element budget of MBr9. The most incompatible REEs and fluid mobile elements, such as Rb and Ba, are the most deficient. The magnitude of the deficiency co-varies inversely with the partition coefficient of the elements in mantle silicates. Previous work (Bedini and Bodinier, 1999) demonstrated the presence of a LREE and Ba-rich pervasive interstitial component, with fluid inclusions and a reaction layer around spinel grains accounting for the deficit in Rb and HFSE. Wittig et al., (2009) have also demonstrated that a significant proportion of bulk-rock U and Th can be accounted for by grain boundary components. Furthermore, Yaxley and Kamenetsky, (1999) demonstrated that interstitial melts derived from decompression breakdown of
metasomatic amphibole provide silicic, alkali-enriched melts that are highly enriched in LREE, Nb, Ta and Ba, the most deficient elements at Mont Briançon. While this mechanism is a possible source for the Mont Briançon glasses it (i) fails to account for the large reservoir of “missing” Rb, (ii) requires a pre-existing metasomatic event to generate amphibole, for which no evidence is present and (iii) produces different textural relations between the major silicate phases and the glass to those actually observed. Small percentage melting of wet, fertile mantle produces a similar melt composition, but is also enriched in Rb (Chazot et al., 1996). Fluid fluxing during Variscan subduction of oceanic lithosphere beneath the Massif Central could provide the mechanism for deep, wet partial melting of the lithospheric mantle and a small contribution to the melt from volatiles fluxed from oceanic sediment could also account for the high P content of some intergranular glasses at Mont Briançon (Rosenbaum et al., 1997).

5.3. The origins of Re and Os elemental abundances and isotope ratios

Mont Briançon xenoliths, like all non-cratonic xenoliths, are characterized by significantly less Os than other peridotites. Mont Briançon xenoliths contain < 30 ppm sulfur (this study; Lorand and Alard, 2001), an order of magnitude lower than fertile mantle ([S] = 300 ppm; McDonough, 2004) and only 25% of that of depleted mantle ([S] = 119 ppm; Salters and Stracke, 2004). Supergene weathering, evident in the low S/O ratios of many sulfide grains in this study, may account for some sulfide loss (e.g. Lorand et al., 2003) although it is unclear whether Os would also be mobilised during weathering of sulfide grains. While interstitial sulfide grains may be vulnerable to the effects of surface processes, sulfide grains preserved within the
interior of silicate grains should not be affected. Since these primary grains characteristically have much higher Os concentrations than interstitial sulfides, supergene weathering seems unlikely to be entirely responsible for the low bulk-rock Os concentrations.

The Re concentrations of half of the Mont Briançon xenoliths are consistent with varying degrees of melt depletion (< 260 ppt), but the remainder are enriched by as much as 235%. Moreover, the absence of a correlation between Re/Os and major element indices of melt depletion suggests that Re has become mobile after melt depletion, and / or the bulk-rock budget of otherwise immobile proxies of melt depletion have been disturbed. Alternatively, in addition to Re mobility, the Os budget may have also been perturbed, either by the addition of radiogenic Os, or through Re addition and subsequent ingrowth of $^{187}$Os.

With one exception (MBr1) the bulk-rock sub-chondritic Os isotope ratios are broadly consistent with melt depletion ($^{187}$Os/$^{188}$Os$_{chondrite} = 0.127$, Shirey and Walker, 1998). Bulk-rock Mont Briançon $^{187}$Os/$^{188}$Os ranges from 0.11737 to 12897, similar to other melt depleted peridotite xenoliths associated with European Cenozoic volcanism (e.g. Schmidt and Snow, 2002; Witt-Eickschen and Kramm, 1998; Witt-Eickschen et al., 2003; Ackerman et al., 2009) However, the lack of a coherent correlation between $^{187}$Os/$^{188}$Os and bulk-rock Al$_2$O$_3$, clinopyroxene Al$_2$O$_3$ and clinopyroxene modal abundance suggests that these factors are not simply linked by the prior extraction of basaltic melt. Bulk-rock Re depletion ages ($T_{RD} = 1/\lambda \times \ln \{ [(^{187}\text{Os}/^{188}\text{Os}_{chondrite}} -

^{187}\text{Os}/^{188}\text{Os}_{sample})/^{187}\text{Re}/^{188}\text{Os}_{chondrite} + 1] \}$, i.e. model ages which assume no change in $^{187}$Os/$^{188}$Os after melt depletion, are widely scattered, ranging from 0.11 to 1.71 Ga. While the oldest $T_{RD}$ may represent a minimum possible age for melt depletion, bulk-rock $T_{RD}$s should be treated with caution, especially where the possibility exists that
both Re and Os have experienced open system behaviour since the time of melt
depletion.

Despite sulfide representing an insignificant modal proportion of peridotite
xenoliths (0.001 – 0.006 weight %, e.g. Luguet et al., 2003) Os sulfide / silicate
partition coefficients (Os $K_d^{sulfide/silicate}$) are sufficiently high ($10^4 – 10^5$) that the
silicate phases and spinel of several Mont Briançon xenoliths contribute no more than
3% of the Os budget (cf. Hart and Ravizza, 1996; Burton et al., 2000). Thus sulfide
will dominate bulk-rock Re-Os systematics. Any perturbation of sulfide compositions
subsequent to melt depletion will potentially have a significant effect on bulk-rock
Re-Os systematics. Figure 12 demonstrates that bulk-rock Os isotope ratios are
dominated by “primary” sulfides, i.e. shielded from interaction with metasomatic
melts and fluids by their host silicate grain. Hence, the contribution to the bulk-rock
$^{187}\text{Os}/^{188}\text{Os}$ from the metasomatic sulfides will be negligible.

While many of the sulfide grains from this study possess sub-chondritic Os
isotope ratios, Re and Os concentration data suggest that many sulfides no longer
retain their original Re and Os abundances. With the exceptions of MBr3_1 and
MBr6_2, all of the sulfides with subchondritic $^{187}\text{Os}/^{188}\text{Os}$ ratios possess high Re
concentrations (e.g. MBr8_3 and MBr20_4, [Re] = 112 ppm and 412 ppm
respectively). For these sulfides to retain subchondritic Os isotope ratios, practically
all of the Re must have been added to the sulfide shortly before the host xenolith was
brought to the surface. Moreover, not all of the sulfide grains with suprachondritic Os
isotope ratios have a high Re abundance. MBr3_4 ([Re] = 0.01 ppm) morphologically
resembles a primary sulfide, yet retains a $^{187}\text{Os}/^{188}\text{Os}$ ratio of 0.12983. Such a
radiogenic Os isotope ratio is inconsistent with melt depletion alone, yet cannot be
accounted for simply by radiogenic ingrowth of Re. Many peridotite xenoliths and
orogenic massif peridotites have experienced Re mobility after initial melt depletion (Meisel et al., 2001; Reisberg and Lorand, 1995). Loss of increasing amounts of primary sulfide (e.g. Handler et al., 1999; Reisberg et al., 2005) would increase the influence of metasomatic sulfide on the bulk-rock Os budget, resulting in higher and more variable Os isotope ratios. The lack of co-variance between bulk-rock Os isotope ratio and Re/Os, or any of its proxies supports the notion that, in addition to Re mobility, the Os budget may have been disturbed.

Variscan subduction could have resulted in the fluxing of radiogenic marine Os (e.g. Brandon et al., 1996) at the same time that P-rich fluids were also added to the SCLM beneath the Massif Central. Such a fluid could also be expected to be rich in Ba and Rb (Chesley et al., 2004). However, it is difficult to account for bulk-rock $^{187}$Os/$^{188}$Os purely from the addition of a metasomatic, radiogenic source of Os that is characterised by such low Os concentrations. While a contribution to the bulk-rock Os concentration of MBr1 from slab-related fluid cannot be eliminated, the modal abundance of metasomatic sulfide necessary to raise the $^{187}$Os/$^{188}$Os to sufficiently radiogenic levels is not supported by (i) the sulfur concentration, or (ii) the observed sulfide modal abundance.

Sulfides preserved within silicate grains should be resistant to mobilisation or alteration by transient metasomatic melts or fluid. Only interstitial grains would normally be vulnerable to this mechanism. Interstitial sulfide grains typically have high Re/Os ratios and low Os concentrations (Burton et al., 1999; Alard et al., 2002) and the removal of these grains while lowering Re content would have little effect on bulk-rock Os. In order to sufficiently lower bulk-rock Os to approximately half that of the fertile mantle, the high [Os] enclosed sulfide grains, must also, at least in part, be removed. Many silicate grains are cross cut by planes of secondary melt inclusions
where silicate minerals have fractured during the through-flux of a metasomatic melt. Since sulfides possess melting temperature well below the peridotite solidus (Hsieh et al., 1987; Egger and Lorand, 1993), Cu-Fe-Ni sulfides may be highly mobile at subcontinental lithosphere temperatures (e.g., Chesley et al., 1999; Gaetani and Grove, 1999), but require dissolution or mobilization in a lower viscosity liquid as molten sulfide itself is immobile at low degrees of porosity (Minarik et al., 1996). As a result of the high Os $K_{0}$sulfide/silicate and the low modal abundance of sulfide in fertile mantle, relatively few sulfide grains need be affected in order to significantly lower bulk-rock Os abundance. Therefore the mechanism by which the melt inclusions are emplaced may also be capable of mobilising enclosed sulfide grains with high Os concentrations, thus lowering bulk-rock Os content. The trails of micron-scale sulfide blebs linking seemingly enclosed sulfides to grain boundaries (Figure 5a with 5b) also provide compelling evidence that these “enclosed” sulfides may not have the high [Os], low [Re] and unradiogenic $^{187}$Os/$^{188}$Os as previously thought (e.g. Alard et al., 2002).

The removal of sulfide cannot, by itself, account for both low Re bulk-rock concentrations in half of the xenoliths and enriched Re concentrations in the remainder. An additional process, capable of enriching bulk-rock peridotite in Re, while not materially increasing Os abundance must be sought. Silicate magmas are characterized by suprachondritic Re/Os ratios (Shirey and Walker, 1998). Experimental evidence also suggests that Re is highly incompatible in silicate minerals (Righter et al., 2004; Walker et al., 1991; Watson et al., 1987) but may (Righter and Hauri, 1998) or may not (Mallman and O’Neil, 2007) be moderately compatible in garnet. Therefore, during melting of a sulfide-free mantle source, Re should be moderately to strongly enriched in the melt, relative to Os, with the extent
of enrichment dependent on the effect of residual garnet. Even in the presence of 
sulfide, Re $K_D^{\text{sulfide/silicate}}$ is at least an order of magnitude lower than Os $K_D^{\text{sulfide/silicate}}$ 
(Sattari et al., 2002; Fonseca et al., 2006; Roy-Barman et al., 1998). Thus, irrespective 
of the presence or absence of sulfide, melts generated in the mantle will have higher 
Re/Os than their source.

Metasomatic re-enrichment could be responsible for the Re and Os elemental 
abundances of those xenoliths with supra-chondritic Re abundances. Melting of fertile 
peridotite (e.g. Meisel et al., 2001) by 1% will produce a magma containing > 4 ppb 
Re. The same melt derived from an enriched peridotite would have an even higher Re 
concentration. Figure 13 illustrates the effect of mixing residual peridotite with 
basaltic magmas of varying compositions. Ocean island basalts (Reisberg et al., 1993; 
Marcantonio et al., 1995), MORB (Schiano et al., 1997; Gannoun et al., 2004, 2007; 
Escrig et al., 2005) and melt generated from a fertile or enriched mantle precursor 
(Meisel et al., 2001) all contain an order of magnitude more Re than lithospheric 
mantle and it is conceivable that mixing between these two reservoirs could produce 
the elevated Re concentrations measured in some Mont Briançon xenoliths. The 
composition of the basalt that brought the Mont Briançon xenoliths to the surface is 
also shown on Figure 13. Although there is unlikely to be a genetic relationship 
between the peridotite xenoliths and the Mont Briançon host basalt, the addition of a 
significant volume of host basalt would have a similar compositional effect to the 
mixing of residual peridotite and a primary magma generated at depth.

Raising the Re concentration of a xenolith to twice the mean mantle 
concentration requires a melt/rock ratio of > 0.5, for which there is no textural 
evidence. Even the addition of Re generated by the melting of pyroxenite with [Re] 
of ≤ 3.2 ppb Re (e.g. van Acken et al., 2008) would require a greater melt / rock ratio
than can be accounted for with the textural evidence preserved in Mont Briançon peridotites. The addition of 0.0015 – 0.0020 weight % (15 – 20 ppm) of metasomatic sulfide from an infiltrating melt (e.g. Rehkamper et al., 1999) could sufficiently raise bulk-rock Re concentrations and is supported by textural observations, that is, the presence of sulfide-bearing interstitial glasses and melt inclusions. However, the mechanism responsible for such an addition at Mont Briançon, i.e. precipitation from a sulfur-saturated melts versus the addition of locally re-mobilized secondary sulfide is difficult to identify. The effect of a migrating mafic melt will be strongly influenced by the degree of sulfur saturation in the melt and its interaction with pre-existing residual mantle sulfide grains (Mavrogenes and O'Neill, 1999; Bockrath et al., 2004; Brenan et al., 2005). Sulfur undersaturated melts may remove sulfides and HSE, whereas sulfur saturated melts may precipitate sulfides (Reisberg et al., 2005). While the compatibility of Re in sulfide is dependent upon $f_O^2$ and $f_S^2$ (Fonseca et al., 2007), in effect, interstitial sulfide, when exposed to a high temperature melt or fluid behaves physically incompatibly and the entire high [Re] sulfide is mobilised (Ballhaus et al., 2006). Furthermore, Re crystal/silicate melt partition coefficients, controlled by $f_O^2$, vary by 4-5 orders of magnitude (Mallmann et al., 2007), making generalisations about Re incompatibility during partial melting of peridotite unreliable. In contrast, high Os sulfide grains, residual after an early melt depletion event and shielded from subsequent interaction from melts and fluids are retained in the mantle, resulting in Os being physically compatible in the residue (Ballhaus et al., 2006).

Rhenium-enrichment would also be accompanied by a moderate addition of Os, despite the high Re/Os ratio of metasomatic sulfides (Alard et al., 2002) and inconsistent with the low Os concentrations of the Re-enriched xenoliths here. It is, therefore difficult to account for the Re and Os abundances of the enriched xenoliths
with a simple, single-stage process. Wholesale removal of both primary sulfides (high
[Os], low [Re], unradiogenic $^{187}$Os/$^{188}$Os; Alard et al., 2002) and pre-existing
secondary sulfide (low [Os], high [Re] and radiogenic $^{187}$Os/$^{188}$Os; Alard et al., 2002)
would systematically lower bulk-rock Re and Os abundances. This could be achieved
by the breakdown of sulfide as proposed by Handler et al., (1999), but the addition of
Re to half of the xenoliths must be derived from an additional process. The
subsequent, or synchronous, addition of metasomatic sulfides with low Os and high
Re abundance and variably radiogenic $^{187}$Os/$^{188}$Os to some, but not all of the
xenoliths, would not only raise Re abundances to those measured in the Re-enriched
xenoliths, but also elevate Os abundances by a moderate amount from their depleted
condition after large-scale sulfide removal (Figure 13b). Evidence for a mechanism
capable of the removal of at least some primary sulfides can be seen in the trails of
melt inclusions across silicate grains. That some primary sulfides survived this
removal and re-enrichment process is evident in the relatively high (< 3 ppb) bulk-
rock Os abundance, and the presence of individual sulfide grains that possess high Os
concentrations with unradiogenic Os isotope ratios. However, their “included” nature
is uncertain. Examples of seemingly enclosed sulfides are joined to grain boundaries
by trails of micron-scale sulfide blebs. Several other “included” sulfides are situated
very close to grain boundaries, i.e. in the third dimension may not be “included”
sulfides. The mis-identification of wholly included sulfides could also account for the
presence of weathered “included” sulfides that should have been protected from
oxidation by the host grain, yet have experienced significant oxidation. This could
also partly account for the compositional, if not textural, similarities between
interstitial grains with high Ni/Fe ratios and weathered grains that would otherwise
have been described as included sulfides. The similarities between these texturally
variable groups of sulfides suggest that their compositions are the result of a common
process. Griffin et al., (2004) also observed a similar relationship in cratonic sulfides.
Only those Mont Briançon sulfides identified unequivocally as unweathered and
entirely enclosed within silicate grains have a discrete and narrowly defined
compositional range. Further deformation and / or comminution of host grains would
likely result in an even more comprehensive loss of primary sulfides, and thus lower
bulk-rock Os abundances and more radiogenic Os isotope ratios.

5.4. The origin and nature of metasomatism beneath the southern FMC.
Several processes have been proposed to account for the geochemical
signature of peridotite xenoliths brought to the surface in Cenozoic European basalts,
and more specifically those found within the basalts of the southern French Massif
Central. For example, evidence for the existence of phosphorous-rich melts has been
recorded in mantle xenoliths from the Massif Central (Rosenbaum et al., 1997) and
the influence of carbonate-rich melt or supercritical fluids produced at the base of the
lithosphere by separation from mafic alkaline magmas also been suggested (Wilson
and Downes 1991; Wilson et al. 1995).

Bedini et al., (1997) proposed a model for the chemical evolution of a basaltic
melt infiltrating the SCLM underlying the East African rift during ascent and
concluded that the lithospheric mantle near the asthenosphere has higher porosity than
mantle just below the Moho. This would permit infiltration of asthenospheric melts
into the lower portions of the SCLM, although high melt/rock ratios would lead to a
partial to complete re-equilibration of host peridotite with the ascending basaltic
melts. A similar process has been proposed to explain the stratified trace element
enrichments observed beneath the Bohemian peridotite massif (Ackerman et al., 2009) and is also consistent with the varying degrees of metasomatism and textural variability with depth in other localities within the southern FMC (e.g. Zangana et al., 1997). In addition, this is supported by the relationship between xenolith equilibration temperatures and degree of metasomatic overprinting suggested by this study.

Elsewhere in the European Cenozoic volcanic province, Witt-Eickschen and Kramm, (1998) and Witt-Eickschen et al., (2003) have described strong correlations between Pb-Sr-Nd and trace elements in clinopyroxenes from Eifel anhydrous xenoliths that resemble depleted MORB mantle, while in the West Eiffel region xenoliths metasomatized with an EM-1 – like component have also been reported. The Eifel xenoliths share some similarities with highly deformed southern FMC peridotite xenoliths (Zangana et al., 1997), but no previous studies of protogranular peridotites from the southern FMC have reported EM-1-like metasomatism. However, the Sr and Nd isotope compositions of protogranular and porphyroclastic FMC peridotite xenoliths reported in previous studies (Downes with Dupuy, 1987; Zangana et al., 1997; Touron et al., 2008; Yoshikawa et al., 2009), show a pronounced relationship between textural type and lithophile element isotope systematics. Xenoliths with relatively enriched isotopic compositions, protogranular and protogranular-to-porphyroclastic textures, define an array trending away from depleted MORB mantle towards Bulk Earth values and towards the range of values calculated for the European Asthenospheric Reservoir (EAR).

Several studies (e.g. Zangana et al., 1997; Lenoir et al., 2000; Touron et al., 2008; Yoshikawa et al., 2009) have also suggested that the southern FMC has been enriched by recent infiltration of small degree partial melts or fluids from the asthenospheric mantle, possibly related to the “low velocity component” of Hoernle et
al. (1995) in European Neogene alkaline magmas. One possible reason for this style of enrichment is that focussing of upwelling mantle beneath the younger and thinner southern SCLM domain may have favoured decompression melting in this area, as suggested by seismic tomography modelling (Sobolev et al., 1996; Lenoir et al., 2000). This is consistent with the accumulation of plume-derived melts at asthenosphere / lithosphere topographic highs as recorded by the high-temperature poikiloblastic xenoliths from Boree, Velay (Xu et al., 1998), caused by the bounding effects, and differing sub-SCLM architecture, of the juxtaposed northern and southern FMC as described by Wittig et al., (2007).

Thus, the southern FMC xenoliths preserve evidence for interaction between asthenospheric mantle-derived melts/fluids and depleted lithospheric mantle. This is supported by the U-Th-Pb and Lu-Hf systematics of clinopyroxenes from southern FMC lherzolites, including two samples from Mont Briançon (Wittig, et al., 2007), which demonstrated that southern FMC peridotites have Hf isotope ratios consistent with melt depletion followed by metasomatic overprinting of a mainly silicate melt, calculated by Yoshikawa et al., (2009) to be consistent with an EAR-derived melt. This contrasts with the hydrous/carbonatitic metasomatic agent that overprinted the incompatible trace element abundances of the northern FMC (Wittig et al., 2007). The source of metasomatism, whether classified as metasomatism in the strictest sense, or low degree partial melting, is therefore at or near the base of the lithosphere and not at shallower depths, consistent with the composition and depth of equilibration of the melt inclusions from this study, the previous work of Schiano with Clocchiatti, (1994) and the recent work of Yoshikawa et al., (2009). The timing of at least the most recent metasomatic episode must be at or near to the eruption age of the host basalts because of the implications for ingrowth of radiogenic Os in sulfides to which Re may have
been added. This is consistent with the perturbations of highly siderophile elements in
the Bohemian massif that were probably linked to Tertiary rifting and volcanism (5 –
25 Ma) rather than during the Variscan orogeny (cf. Ackerman et al., 2009).
Critically, the influx of incompatible trace elements makes it difficult to
extract meaningful geochronological information pertaining to melt depletion that
predates later metasomatic and tectonic events using silicate-hosted lithophile isotope
systems. Evidence for melt depletion events extends as far back as 480 Myr ago (Pin
and Marini, 1993) and multiple incidences of melt depletion have been confirmed by
Nd isotope systematics in clinopyroxene from an unmetasomatized Mont Briançon
xenolith, and at 257 Myr ago by Femenias et al., (2004). However, the longer scale
evolution of the SCLM beneath the southern FMC remains elusive as the overprinting
of subsequent events only serves to obscure the earliest melt depletion. Fortunately,
the recent metasomatizing silicic melt itself does not possess a significant Os budget.
As Figure 12 illustrates, although the metasomatic melt is S-saturated, the absolute
amount of Re and Os added to the Mont Briançon xenoliths will be negligible. While
the silicate metasomatic melt may well contribute to the incompatible lithophile
element budget, perturbation to Re-Os isotope systematics may only be slight, thus
permitting the chronology of earlier melt depletion events to be investigated.

5.5. “Seeing through” the addition of metasomatic sulfide

Osmium isotope disequilibrium likely exists in all the Mont Briançon
xenoliths of this study. As with previous work (e.g. Burton et al., 1999), bulk-rock Os
isotope ratios are somewhat lower than those of the constituent silicate phases. In
addition, at Mont Briançon these silicate phases do not yield any useful
geochronological information as the Re-Os systematics of these minerals have been
perturbed since the first episode of melt depletion. Consequently, any “ages” obtained
from the silicate phases have large errors and are indistinguishable from the eruption
of host basalt and the entrainment of the xenolith suite (c.1 – 4 Ma). Mont Briançon
sulfide Os isotope ratios, while commonly unradiogenic, more often than not are not
representative of those of the corresponding bulk-rock. This is not uncommon in
cratonic peridotites (e.g. Griffin et al., 2004) or abyssal peridotites (e.g. Harvey et al.,
2006) and bulk-rock Os isotope ratios should not therefore be (i) taken as representing
any particular individual event and (ii) be indiscriminately used for calculating $T_{RD}$
ages. In addition, Ackerman et al., (2009) noted that for the Bohemian Massif, the
range of bulk-rock Os isotope ratios is similar to the distribution of $^{187}$Os/$^{188}$Os ratios
reported for samples from the modern convecting upper mantle (e.g., Liu et al., 2008).
This is also true for the Mont Briançon xenolith suite. Consequently, the bulk-rock Os
isotopic variations in this suite may reflect the formation of lithospheric peridotites by
variable degrees of partial melting of a mantle precursor whose $^{187}$Os/$^{188}$Os was not
perfectly uniform at the time of melting, or a mixed age population (Rudnick and
Walker, 2009), although this could also be due to sulfide breakdown (e.g. Handler et
al., 1999), removal of sulfide by S-undersaturated melts (e.g. Reisberg et al., 2005) or
the time integrated addition of radiogenic Os following the addition of Re sometime
in the past. This variability in the upper mantle means that Os isotopes are not useful
for discriminating among relatively young, e.g., Phanerozoic events. So, bulk-rock
$T_{RD}$ ages of young events should be treated with caution.

Individual sulfide blebs that separated as an immiscible fluid during an ancient
partial melting event and were subsequently shielded from interaction with later melts
and fluids should provide a better means for dating individual melt depletion events.

Primary sulfides associated with melt depletion should be easily distinguished from secondary metasomatic sulfides based upon their major element composition (e.g. Dromgoole and Pasteris, 1987, Luguet et al., 2003), their morphology and relationship with surrounding silicates (e.g. Burton et al., 1999) and their Re and Os abundance and isotope ratios (e.g. Alard et al., 2002). However, it is not possible to subdivide the Mont Briançon sulfide grains into primary and secondary populations according to their major element composition; similarly, they cannot be subdivided on the basis of their Re/Os ratios, their Os isotope ratio or Re-Os abundances. Even the most unradiogenic sulfides (e.g. MBr20_3) have high Re abundances suggesting that long-term evolution in a low Re environment, to preserve an unradiogenic Os isotope ratio, must have been followed by an event which added a significant amount of Re to the sulfide grain. Unradiogenic sulfides from this study appear to have been exposed to a metasomatic agent that has significantly altered the Re/Os ratio. In fact no single sulfide measured here exhibits the chemical characteristics expected of a primary sulfide i.e. high [Os] and low [Re], and an unradiogenic Os isotope ratio. It seems unlikely that the diffusion of Re through a silicate mineral and into the low [Re] enclosed sulfide is responsible for the elevated [Re] in some seemingly primary sulfides. Diffusion would occur over a protracted period, over timescales that would allow significant ingrowth of $^{187}\text{Os}$. While this could be a reasonable mechanism to account for some of the seemingly enclosed sulfides from this study, such diffusion of Re would be expected to be a universal process and should affect all enclosed sulfides in xenoliths worldwide. What seems more likely from the textural evidence of Figure 4 is that the two populations of sulfide have experienced limited physical mixing when interstitial melts and melt inclusions were added.
Mixing has therefore occurred between originally high [Os], unradiogenic sulfides and a metasomatic sulfides with high Re/Os and low [Os] (cf. interstitial, metasomatic sulfides of Alard et al., 2002). Alternatively, additional Re may have partitioned into a pre-existing sulfide from a transient silicate melt, but it seems unlikely that a previously enclosed sulfide could exchange Re with a large volume of metasomatic melt or fluid and not be removed from its within-grain location. As long as the hybrid sulfide grain is not removed during interaction with the metasomatic agent, the Os isotope ratio will only be materially affected over longer intervals through radiogenic ingrowth of $^{187}$Os. Figure 13b illustrates that for a narrow range of Os isotope ratios, the variation in sulfide Re/Os ratios can be accounted for by varying degrees of mixing between pre-existing sulfide and a metasomatic sulfide with a high Re/Os ratio. Even if wholesale Re addition did occur, because the timing was so close to that of the eruption of the host basalt there has been little measurable ingrowth of $^{187}$Os. Thus, the characteristic high [Os] and unradiogenic Os isotope ratio of primary sulfides is still evident and dominates the bulk-rock Os isotope ratios.

However, it is interesting that, if the addition of significant proportions of Re to primary sulfides is recent, and that the unradiogenic sulfides preserve primary or near primary Os isotope ratios, that the “ages” derived from sulfide $T_{RD}$ calculations do not cluster to signify a small number of discrete events. Only two sulfides (MBr3_1 with MBr6_2) yield $T_{RD}$ ages that are within error of each other, yet do not correspond with any discrete event previously reported. Similarly, none of the sulfides yield $T_{RD}$ ages that correspond with ages derived by Lu-Hf geochronology for melt depletion during the Variscan orogeny (Wittig, et al., 2006) or subsequent melt melting at 257 Myr ago (Femenias et al., 2004). However, this is not to say that Re-Os geochronology of individual sulfides is not without merit. As previously discussed
with respect to bulk-rock $T_{\text{RDS}}$, sulfide $T_{\text{RDS}}$ should not be wielded indiscriminately, especially when attempting to unravel the cumulative effects of several Phanerozoic events. The utility of sulfide Re-Os isotope systematics lies in the ability of the system to preserve evidence for melt depletion that pre-dates Phanerozoic events, i.e. pre-Phanerozoic $T_{\text{RDS}}$ have, by definition, survived the effects of all subsequent processes, metastatic or otherwise. Seven of the nine sulfide $T_{\text{RDS}}$ presented in this study preserve evidence for melt depletion events before the Variscan orogeny, and six of those pre-Variscan $T_{\text{RDS}}$ yield ages that span the Neo-proterozoic to the early Paleoproterozoic. This suggests that the SCLM that now resides beneath the southern FMC experienced a complex melt depletion history before its incorporation into continental Europe during the Variscan orogeny, Phanerozoic melt depletion, and the extraction of further melt and subsequent or synchronous metasomatism during the Cenozoic.

Hence, the sulfide $T_{\text{RD}}$ age of 1.78 Ga of MBr20_4, although only a minimum age, is likely the best available estimate of the earliest melt depletion events that this portion of the SCLM experienced, although the limited number of sulfides analysed here does not preclude melt depletion that extends still further into the past, making its history of melt depletion similar to that experienced by abyssal peridotites (Harvey et al., 2006; Alard et al., 2005) or off-cratonic SCLM in Europe (e.g. Schmidt and Snow, 2002) and worldwide (Meisel et al., 2001).
6. CONCLUDING REMARKS

In the absence of a coherent bulk-rock Re-Os isochron, alumichron or co-
variation between Os isotope ratios and any immobile proxies of Re,
geochronological information pertaining to the earliest melt depletion of the SCLM
from bulk-rock compositions will often be erroneous, whether determined by silicate
hosted lithophile element isotope systems (e.g. Rb-Sr, Sm-Nd, U-Th-Pb) or
siderophile Re-Os isotopes. However, perturbation of [Re] and [Os] in Mont Briançon
xenoliths likely occurred only a short time before, or contemporaneously with, the
entrainment of the xenolith in the host basalt. Therefore, Re-depletion model ages of
individual sulfide grains preserve evidence for older melt depletion that extends back
to at least 1.8 Ga. Despite multiple episodes of melt depletion, perturbations of trace
element budgets, infiltration of a highly silicic melt, the mobilisation of a significant
proportion of the xenoliths sulfide budget and possible Re addition to the remainder,
at least some Mont Briançon xenoliths preserve sulfide Re-Os isotope systematics
which are still capable of seeing through this range of mantle processes.

Moreover, even in xenolith suites that have experienced greater degrees of
metasomatism or addition of much larger quantities of metasomatic sulfide, as long as
the metasomatism was recent, or near contemporaneous with a known exhumation
age, the effects of mixing metasomatic Re-rich sulfide with primary Os-rich sulfide
are negligible, or easily corrected for, and minimum melt depletion ages can still be
obtained in all but the most tectonised SCLM, where formerly enclosed sulfides
subsequently experience open system behaviour.
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1980  


1999  

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2003
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2008
2010 Geochemical variation in peridotite xenoliths and their constituent clinopyroxenes
2011 from Ray Pic (French Massif Central): implications for the composition of the
2013
2015 319-333.
2016
2017
2018
2019
2020
2021
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Table 1. Representative major and trace element data for peridotite xenoliths from Mont Briançon, Massif Central, France. SiO$_2$, Al$_2$O$_3$ and CaO abundances expressed in wt. %. Major element and Ni data quality assessed using two rock standards (WS-E and OUG-94). Reproducibility is within 2% of recommended values. All other abundances shown in ppm. Error on LREE measurements ranges from < 1 % for high concentration samples to >20% for the lowest concentration sample (MBr1). Errors on Sr measurements are all <3%. Errors on Yb measurements range are below 7% with the exception of the low concentration sample MBr16 (11%). All errors as 2 S.D. Complete major and trace element abundances for all of the samples are available in the online supplementary information.
Table 2. Re–Os isotope data for bulk-rock peridotite xenoliths from Mont Briançon, Massif Central, France. [Re] and [Os] expressed in parts per billion (ppb). Errors shown as 2σ mean. T_{RD} = (minimum) time of rhenium depletion calculated using a mean present-day chondritic value $^{187}\text{Os}/^{188}\text{Os} = 0.1278$ from Walker et al. (2002) and assumes that $^{187}\text{Re}/^{188}\text{Os}_{\text{blank}} = 0$. Given isotope ratios blank corrected. $^{187}\text{Os}/^{188}\text{Os}$ ratios normalised to $^{192}\text{Os}/^{188}\text{Os} = 3.08271$ and corrected using measured $^{18}\text{O}/^{16}\text{O}$ and $^{17}\text{O}/^{16}\text{O}$ of 0.002047 and 0.00037 respectively. Johnson Matthey (n=26) 2.75ng Os standard $^{187}\text{Os}/^{188}\text{Os}=0.17373\pm12$ (2σ). “*” denotes duplicate measurement on separate powder split using the same digestion method.
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Table 3. Re–Os isotope data for mineral separates and sulfides from MBr2, 3, 6, 8 with 20.

Errors shown are 2σ mean. Re and Os concentrations expressed as parts per trillion (ppt) for silicates and spinel, and parts per million (ppm) for sulfides. Standards and corrections for $^{18}$O/$^{16}$O and $^{17}$O/$^{16}$O as in Table 2.
FIGURE CAPTIONS

Figure 1. Simplified geological sketch map of the eastern French Massif Central (FMC) modified after Coisy and Nicholas (1978) and Lorand (2003) showing the location of Mont Briançon scoria cone, Devèrs area, last active during the Pliocene (c. 1 – 4 Ma). Also shown is the boundary between the northern and southern FMC (dashed line at 45°30’’) (Lenoir et al., 2000; Downes et al., 2003).

Figure 2. The range of melt depletion experienced by Mont Briançon peridotite xenoliths. Polybaric melt extraction curves of Herzberg (2004) are also shown for 2 GPa melt depletion (solid line) and 1 GPa (dashed line) with indications of expected residual composition after 20 % melt extraction. Northern domain FMC xenoliths from Lenoir et al., (2000), southern domain FMC from Downes and Dupuy, (1987); Lenoir et al., (2000) and Zangana et al., (1997).

Figure 3. Incompatible trace element abundances for 15 Mont Briançon peridotite xenoliths normalised to primitive mantle (Palme with O’Neil, 2004). (a) Seven peridotites with low primitive mantle-normalised Ce/Yb ratios i.e. (Ce/Yb)N = 0.24-0.51 are inferred to have experienced the least cryptic metasomatism - their trace element abundances being due to varying degrees of melt depletion. (b) Seven peridotites with (Ce/Yb)N > 0.9 ≤ 2.4. Elevated light rare earth element ratios are unlikely to be attributable simply to melt depletion and suggest the influence of more intense cryptic metasomatism than in (a). (c) (Ce/Yb)N of MBr6 is the highest of all the samples in this study and is accompanied by a large Sr enrichment. This sample has experienced the most intense cryptic metasomatism.
Figure 4. Phase relations between silicate, CO$_2$ and sulfide melt and fluid inclusions. (a) Glassy silicate melt inclusions trapped within a re-annealed, formerly fractured olivine grain from MBr12. The trail of inclusions terminates in the intergranular glass to the left of the image. Plane polarized light (PPL). (b) Plane of melt inclusions trapped within an annealed olivine grain from MBr12 (PPL). (c) CO$_2$ fluid inclusion (top) with co-genetic, silicate melt inclusion beneath. The two phases are joined by a narrow neck approximately 1 μm wide (PPL). (d) Trail of silicate melt inclusions trapped within an olivine grain from MBr27. In reflected light the presence of co-genetic sulfide grains can be seen. (e) Co-genetic silicate and sulfide grains preserved as immiscible liquids within the same inclusion (PPL). (f) CO$_2$ fluid inclusion with co-genetic sulfide bleb within olivine from MBr12.

Figure 5. Sulfides from Mont Briançon peridotite xenoliths and their relationships with silicate phases and spinel. (a) Rounded sulfide included within olivine in MBr7. A trail of micron-scale sulfide blebs leads away from the main bleb (bottom right). Scale bar 100 μm. (b) Rounded sulfide blebs within olivine from MBr3. The three largest sulfides are joined by a network of micron sized blebs within the re-annealed olivine grain. Scale bar 100 μm. (c) Interstitial sulfide from MBr7. Scale bar 200 μm. Secondary sulfide has been precipitated and / or (re)mobilized along grain boundaries, or has exploited fracture in mineral grains as sulfide was precipitated / (re)mobilized during deformation (d) (Scale bar 100 μm). (e) Large weathered intergranular sulfide from MBr2. Small areas of fresh sulfide remain, however extensive oxidation has resulted in the formation of iron oxy-hydroxides. Scale bar 100 μm. (f) Small rounded sulfides preserved in metasomatic glass in MBr2. Despite the morphological similarity with
“enclosed” rounded sulfides, above, these blebs are smaller and associated with interstitial glass in the vicinity of corroded spinel. Scale bar 50 μm.

Figure 6. Major element composition of 63 individual sulfide grains. Sulfides classification is based upon textural constraints, degree of weathering, and proximity to interstitial glass and corroded spinel. Only unweathered sulfides, unequivocally entirely enclosed within silicate grains, have a narrowly defined compositional range. All other sulfides, have much wider, and possibly disturbed, major element compositions.

Figure 7. (a) Scatter between \(^{187}\text{Re}^{188}\text{Os}\) and \(^{187}\text{Os}^{188}\text{Os}\) is particularly high amongst the xenoliths from Mont Briançon. No significant improvement is obtained in this study using previously utilised proxies such as (b) \(\text{Al}_2\text{O}_3\), (c) S, or (d) Yb. Unless elements such as Al or Yb, conventionally considered to be immobile, have also been disturbed, the Os isotope ratio may also have been perturbed, either by the addition of radiogenic Os, or through Re addition and subsequent ingrowth of \(^{187}\text{Os}\).

Figure 8. Re-Os isotope systematics of silicate minerals, spinel and corresponding bulk-rocks for MBr 2, 3, 6, 8 with 20. Host basalt is shown for reference. Mineral \(^{187}\text{Os}^{188}\text{Os}\) ratios are consistently higher than their corresponding bulk-rock. Mean \(^{187}\text{Re}^{188}\text{Os}\) ratios for all of the separated phases are much greater than present day chondrite (\(^{187}\text{Re}^{188}\text{Os}\) = 0.406) however no individual sample is systematically enriched in Re across all of its constituent phases.

Figure 9. [Os] and \(^{187}\text{Os}^{188}\text{Os}\) of individual sulfides from MBr3 (n=3), MBr6 (n=4), MBr8 (n=3), and MBr20 (n=4). As in cratonic mantle sulfides (e.g Griffin et al.,
2004), the most unradiogenic sulfides possess the greatest [Os]. With the exception of MBr3_1, all of the sulfides with subchondritic $^{187}\text{Os}/^{188}\text{Os}$ ratios have high Re/Os ratios. For these sulfides to retain subchondritic Os isotope ratios, Re must be added to the sulfide shortly before the host xenolith was brought to the surface. Moreover, not all of the sulfide grains with suprachondritic Os isotope ratios have a high Re abundance.

Figure 10. Re-Os isotope systematics of bulk-rock (open symbols) for MBr3, 6, 8 and 20 (circles, rectangles, triangles, diamonds respectively) and individual sulfide grains (n = 14, grey symbols) handpicked from the bulk-rock samples. Outlined white area denotes range of bulk-rock compositions of the entire Mont Briançon suite. Many of the sulfides (n=9) possess subchondritic Os isotope ratios that yield $T_{\text{RD}}$ model ages of up to 1.8 Ga, i.e. similar to the oldest $T_{\text{RD}}$ obtained from the most unradiogenic bulk-rock Os isotope analysis (1.7 Ga).

Figure 11. Incompatible trace element mass balance for MBr9. The most incompatible REEs and fluid mobile elements are the most deficient. The deficiency, in most cases, co-varies inversely with the partition coefficient of the elements in mantle silicates. Mass balance calculations performed using modal abundances of olivine, orthopyroxene, clinopyroxene and spinel (online supplementary information) and incompatible trace element compositions obtained by LA ICP MS (online supplementary information).

Figure 12. Os mass balance calculations for MBr3, MBr8 and MBr20. Silicates and spinel account for only < 3 % of Mont Briançon bulk-rock Os budgets, the remainder being accounted for by a balance between the two populations of sulphides. The
difference in Os concentration between enclosed and interstitial sulphides is sufficiently large in MBr 3 that interstitial sulphide contributes no more than 1% to the whole rock budget whilst maintaining the measured whole rock isotopic ratio. Similarly, in MBr 8 interstitial sulphide can contribute no more than 11%, the remainder being accounted for by enclosed sulphide.

Figure 13. (a) OIB, MORB and PUM-derived magma have an order of magnitude more Re than lithospheric mantle. However, raising bulk xenolith [Re] to 2x PUM requires a melt/rock ratio of > 0.5, for which there is no textural evidence in this study. OIB (Reisberg et al., 1993; Marcantonio et al., 1995), MORB (Schiano et al., 1997; Gannoun et al., 2004, 2007; Escrig et al., 2005), PUM (Meisel et al., 2001). (b) Precipitation of 15–20 ppm metasomatic, i.e. secondary sulfide raises bulk xenolith [Re] to twice the value of fertile mantle and is consistent with textural observations (Figure 3 with Figure 4). Wholesale removal of both primary sulfides (high [Os], low [Re], unradiogenic $^{187}$Os/$^{188}$Os) and pre-existing secondary sulfide (low [Os], high [Re] and radiogenic $^{187}$Os/$^{188}$Os; Alard et al., 2002) would systematically lower bulk-rock Re and Os abundances. Subsequent, or synchronous, addition of metasomatic sulfides with low Os and high Re abundance to some, but not all of the xenoliths, would raise Re abundances to those measured in the Re-enriched xenoliths.
Figure 1
This study

Southern domain FMC

Northern domain FMC

Figure 2
Figure 3. 

(a) 

(b) 

(c)
Figure 4
Figure 5
Figure 6

Ni/Fe vs. S/O for different conditions:
- Fresh, included
- Fresh, interstitial
- Weathered, included
- Weathered interstitial
- Associated with glass and corroded spinel
Figure 7
Figure 8

The graph shows the relationship between $^{187}$Re/$^{188}$Os and $^{187}$Os/$^{188}$Os for different rock components: host lava, olivine, orthopyroxene, clinopyroxene, and spinel. The logarithmic scale on the x-axis represents the ratio of $^{187}$Re/$^{188}$Os, while the y-axis represents the ratio of $^{187}$Os/$^{188}$Os.
Figure 9

\[
\frac{^{187}\text{Os}}{^{188}\text{Os}} \quad \text{vs} \quad [\text{Os}] \text{ ppm}
\]

- MBr bulk rock
- MBr sulfide
- Cratonic sulfide (Griffin et al., 2004)
Figure 10

The graph shows a plot of $^{187}$Os/$^{188}$Os on the y-axis against $^{187}$Re/$^{188}$Os on the x-axis. The data points are labeled with ages: 0 age, 0.5 Gyr, 1.0 Gyr, 1.5 Gyr, and 2.0 Gyr. The error bars indicate the uncertainty in the measurements.
Figure 11

- Figure 11 displays a bar chart showing the distribution of various elements (Rb, Ba, Th, U, Nb, Ta, La, Ce, Pr, Sr, Nd, Zr, Hf, Sm, Eu, Ti, Gd, Tb, Dy, Ho, Er, Yb, Lu) in different components of the mineralogical study, including olivine, orthopyroxene, clinopyroxene, spinel, and grain boundary component.
Figure 13

(a) Diagram showing the relationship between [Os] ppb and [Re] ppb. The graph compares various basalt types, including OIB and MORB, with data points indicating 1% melt enriched precursor and metasomatic sulphide addition. The fertile mantle is marked with dashed lines.

(b) Additional graph with annotations for metasomatic sulphide addition and sulphide loss. The [Re] ppb axis is expanded with a scale from $10^{-1}$ to 1000.

- OIB
- MORB
- MBr host basalt
- 1% melt, enriched precursor
- Metasomatic sulphide addition
- Sulphide loss
- Fertile mantle