Highly CO₂ -supersaturated melts in the Pannonian lithospheric mantle – A transient carbon reservoir?

Laura Créon, Virgile Rouchon, Souhail Youssef, Elisabeth Rosenberg, Guillaume Delpech, Csaba Szabo, Laurent Remusat, Smaïl Mostefaoui, Paul D Asimow, Paula M Antoshechkina, et al.

► To cite this version:

Laura Créon, Virgile Rouchon, Souhail Youssef, Elisabeth Rosenberg, Guillaume Delpech, et al.. Highly CO₂ -supersaturated melts in the Pannonian lithospheric mantle – A transient carbon reservoir?. *Lithos*, 2017, 286-287, pp.519-533. 10.1016/j.lithos.2016.12.009 . hal-01636579

HAL Id: hal-01636579

<https://hal.science/hal-01636579>

Submitted on 16 Nov 2017

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

Highly CO₂-supersaturated melts in the Pannonian lithospheric mantle – A transient carbon reservoir?

Laura Créon ^a, Virgile Rouchon ^a, Souhail Youssef ^a, Elisabeth Rosenberg ^a, Guillaume Delpech ^b, Csaba Szabó ^c, Laurent Remusat ^d, Smail Mostefaoui ^d, Paul D. Asimow ^e, Paula M. Antoshechkina ^e, Mark S. Ghiorso ^f, Elodie Boller ^g, François Guyot ^d

^a IFP Energies nouvelles, 1 & 4 avenue Bois Préau, 92852 Reuil-Malmaison Cedex, France ; ^b Geosciences Paris Sud department, Paris Sud University, Bâtiment 504, 91405 Orsay Cedex, France ; ^c Lithosphere Fluid Research Lab, Department of Petrology and Geochemistry, Eotvos University, Pazmany Peter setany 1/c, H-1117 Budapest, Hungary; ; ^d IMPMC, Museum National d'histoire Naturelle, Sorbonne Universités, CNRS, UPMC, IRD, 61 Rue Buffon, 75005 Paris, France ; ^e California Institute of Technology, 1200 East California Blvd. Pasadena, California 91125, USA; ^f OFM Research, 7336 24th Avenue NE, Seattle, WA 98115 USA Seattle, Washington-USA; ^g ESRF, 71 Avenue des Martyrs, 38000 Grenoble, France

Corresponding Author: Laura Créon, lauracreon@gmail.com; Present address: Universidad Nacional Autónoma de México, Centro de Geociencias, Campus Juriquilla, A.P. 1-742, Boulevard Juriquilla No. 3001, Juriquilla. Qro., C.P. 76230

Abstract

The carbon budget of the continental lithospheric mantle is poorly known, although it is believed to be substantially fed by subduction zones and a source of the widespread non volcanic CO₂ degassing occurring in active tectonic intracontinental settings. We propose new constraints on the CO₂ budget of the subcontinental lithospheric mantle below the Pannonian

Basin (Central Europe) through present study of a suite of xenoliths from the Bakony-Balaton Highland Volcanic Field. Trails of secondary fluid inclusions, silicate melt inclusions, network of melt veins and melt pockets with large and abundant vesicles, are numerous evidences of mantle metasomatism that affected the lithosphere beneath region. A quantitative estimate of the CO₂ budget of the mantle below the Pannonian Basin was constrained by an innovative approach applied to mantle xenoliths using (1) synchrotron X-ray microtomography, (2) NanoSIMS, Raman spectroscopy and microthermometry, and (3) thermodynamic models (Rhyolite-MELTS). The reconstructed 3D volumes obtained by synchrotron X-ray microtomography allow quantifying the proportions of all petrographic phases in the samples and to visualize their textural relationships. The concentrations of CO₂ and H₂O in glass veins and pockets range between 0.27 ± 0.01 (1 σ) and 0.96 ± 0.02 (1 σ), and 0.54 ± 0.26 (1 σ) and 4.25 ± 0.27 (1 σ) wt. %, respectively, and have higher CO₂ contents than typical arc magmas (0-0.25 wt. % CO₂ and 1.9-6.3 wt. % H₂O). Trapping vesicle pressures were determined based on silicate melt saturation pressures between 0.69 to 1.78 GPa, and are generally in excess compared to fluid inclusions trapping pressures determined by Raman spectroscopy and microthermometry (0.1-1.1 GPa). The CO₂/silicate melt mass ratios in the metasomatic agent that percolated the lithospheric mantle below the Pannonian Basin are estimated between 9.0 and 25.4 wt. %, indicating that silicate melts were CO₂ supersaturated in their source region. We suggest that slab derived melts initially containing up to 25 wt. % of CO₂ migrated congruently into the lithospheric mantle, and were subsequently fractionated by CO₂ segregation in secondary fluid inclusions by fracturing and creeping of the peridotite mineral matrix. A first order estimation of the bulk minimal CO₂ concentration in the lithospheric mantle below the Pannonian Basin is proposed at 1935 ppm. This transient carbon reservoir is believed to be degassed through the Pannonian Basin due to volcanism and

tectonic events, likely drained by the Mid-Hungarian Zone, a regional shear zone of lithospheric scale.

Keywords: X-ray microtomography; Mantle-xenoliths; CO₂-budget; Lithospheric-mantle

1 Introduction

The mantle is thought to be largest carbon reservoir of the Earth (Sleep et al., 2001; Shcheka et al., 2006; Dasgupta and Hirschmann, 2010; Dasgupta, 2013), containing more than 90% of carbon of the Earth (Poinar, 2012). Many geological processes depend on and attest for the existence of deep carbon and its exchange with other major terrestrial reservoirs, such as CO₂ degassing at mid-ocean ridges, CO₂-rich magmatism rooted in the mantle (kimberlites, carbonatites), and diamond or carbonate-bearing xenoliths originating from the deep mantle (Hammouda 2003 and references therein). In order to balance the continuous CO₂ degassing at volcanic centers, carbon must have been recycled into the mantle over geological time scales, most likely at convergent margins where carbonate-altered, sediment-bearing oceanic crust returns into the mantle (Zhang & Zindler 1993). Carbon dioxide is known to be a dominant volatile species in magmatic fluids and gases in subduction zone environments (Anderson, 1975; Fyfe et al., 1978) and its mantle origin has clearly been shown by ³He/⁴He and δ¹³C isotopic ratios of fumaroles and hot spring gases (e.g. Sano & Marty, 1995). There have been several recent reviews of the carbon cycle in subduction environments (Jarrard, 2003; Wallace, 2005; Gorman et al., 2006; Dasgupta and Hirschmann, 2010; Johnston et al., 2011; Dasgupta, 2013; Burton et al., 2013), which estimate that about half of the subducted carbon is removed from the subducting plate beneath fore-arcs and arcs and returns surface of Earth [40% in Gorman et al. (2006); 20–80% in Dasgupta & Hirschmann (2010); and 18–70%

in [Johnston et al. \(2011\)](#)]. A large part of the subducted CO₂ is therefore thought to be stored in the mantle ([Kelemen and Manning, 2015](#)). Kelemen and Manning (2015) recently argued that the carbon input from subduction zones into the overlying plate is larger than output from arc volcanoes plus diffuse venting, and that substantial quantities of carbon are stored in the mantle lithosphere and crust. This recent study suggests that diffuse escape of CO₂ towards the surface and, more importantly, transient storage of CO₂ in the back-arc lithospheric mantle may, when added to the more commonly described fluxes of arc magma and MOR degassing, fully balance the subducted carbon ([Kelemen and Manning, 2015](#)). Therefore, both the diffuse degassing flux in back-arc regions and the lithosphere mantle carbon reservoir need to be further quantified since they have considerable implications for the global carbon cycle. This study aims at constraining the magmatic CO₂ budget of the subcontinental lithospheric mantle below the Pannonian Basin (PB, Central Europe) through the study of a suite of petrographically and geochemically well-studied peridotite xenoliths ([Creon et al., this volume](#)) from the Balaton-Bakony Highland Volcanic Field (BBHVF, [Szabó et al. 2010](#)). The PB is known for its substantial mantle derived CO₂ accumulations ([Ballentine et al., 1991](#); [Sherwood Lollar et al., 1997](#)), complex geodynamic setting ([Horváth et al., 2015](#)) and volcanic activity ([Harangi and Lenkey, 2007](#)) that have developed within a relatively short time scale of 16 My ([Kovács and Szabó, 2008](#)). The PB represents an ideal setting for studying the fluxing of mantle CO₂ through the lithosphere. This study develops an innovative multi-technique approach in order to investigate the abundance and significance of CO₂ fluid inclusions and vesicles within mantle xenoliths by coupling: (1) three-dimensional synchrotron X-ray microtomography, (2) nanoSIMS, Raman spectroscopic and microthermometric characterization and (3) thermodynamic modeling of the magmatic evolution.

2 Geological background

The Pannonian Basin (PB) is located in the Carpathian–Pannonian region of Central Europe. The PB is a typical inter-arc basin ([Embey-Isztin et al, 2001](#)), yet it shows a number of features that are characteristic of rift zones, such as high heat flow, recent alkali volcanism, thin crust and lithosphere, and upwelling asthenosphere ([Embey-Isztin et al., 1990](#); [Praus et al., 1990](#); [Spakman, 1990](#); [Falus et al., 2007](#)). The Pannonian Basin experienced a complex geodynamic evolution during the Miocene to Pleistocene characterized by the succession of extension-compression-extension events ([Bali et al., 2008](#)). The major driving forces that led to the formation of the Pannonian Basin (~20 Ma) were the continuous subduction and roll-back on its eastern boundaries and the synchronous eastward extrusion of ALCAPA blocks ([Kazmer and Kovacs, 1985](#)) from the Alpine compressional belt ([Horvath, 1993](#); [Fodor et al., 1999](#)).

The Bakony–Balaton Highland Volcanic Field (BBHVF), on the northern shore of Lake Balaton, is situated near the center of the Carpathian-Pannonian region, in the western Pannonian Basin (western Hungary). The volcanic centers of the BBHVF were active between 7.96 Ma and 2.61 Ma ([Balogh and Pécskay, 2001](#); [Balogh and Németh, 2005](#); [Wijbrans et al., 2007](#); [Hidas et al., 2010](#)) and produced mostly alkali basaltic volcanic rocks ([Szabó et al., 1992](#); [Embey-isztin et al., 1993](#)).

Ultramafic xenoliths can be found in basanitic lava flows and pyroclastic volcanic products at six locations (Tihany, Bondoró Hill, Füzes-tó, Szentbékkálla, Mindszentkállya, Szigliget). Most of the xenoliths are spinel lherzolites, but harzburgite, clinopyroxenite, orthopyroxenite, wehrlite, websterite and, sometimes, composite xenoliths also occur (See [Creon et al. \(this volume\)](#) for more information).

3 Petrography of the mantle xenoliths

More than 150 ultramafic xenoliths from the BBHVF were collected at the Szentbékállá, Szigliget, Füzes-tó and Mindszentkállya localities (see [Creon et al. \(this volume\)](#) for sampling locations). After microscopic observations, four ultramafic xenoliths from Szentbékállá, one from Szigliget and two from Füzes-tó were selected for further study. Samples selected are 7 to 12 cm long, rounded spinel peridotites and range from lherzolite to harzburgite compositions. A detailed petrographic description of the samples is given in the companion paper by [Creon et al. \(this volume\)](#), and only a short description will be given here. Further petrographic observations obtained by X-ray tomography will be presented in the *Results* section. Peridotite samples show equigranular to mosaic or protogranular microstructures with minor poikilitic features (based on the classification of [Mercier & Nicolas \(1974\)](#)). Samples are composed of olivine (Ol), orthopyroxene (Opx), clinopyroxene (Cpx) and spinels (Sp). Amphibole (Amp) also occurs in some of the xenoliths (Szigliget and Füzes-tó). Microscopic observations show that the majority of these xenoliths contain veins, primary silicate melt inclusions or melt pockets. Melt pocket sizes are variable (from few hundreds microns to few millimeters in diameter) and are usually connected to each other by thin silicate melt veins. The melt pockets are composed of secondary formed clinopyroxene, olivine, spinel, and interstitial glass with vesicles. Melt pocket shapes can be irregular, but some are euhedral pseudomorphs of protominerals (amphibole). Numerous trails of secondary fluid inclusions are observed in Opx, Cpx and Ol. Fluid inclusions display equilibrium and non-equilibrium features depending on the samples ([Creon et al., this volume](#)).

3.1 Analytical methods

3.1.1 Synchrotron X-ray microtomography

The eleven samples analyzed consist of tiny cylinders 5 to 10 mm in length and 4 mm in diameter (Fig. 1). The internal textures of the mantle xenoliths were characterized through synchrotron radiation computed micro-tomography at the European Synchrotron Radiation Facility (ESRF), Grenoble (France) on the ID19 beamline. ESRF offers access to very high X-ray beam intensity. The detector used is a Fast Readout Low Noise FRELON Camera developed at ESRF, with 2048×2048 pixels and 14 bits resolution ([Baruchel et al., 2000](#); [Labiche et al., 2007](#)).

Monochromatic microtomography with 30 keV radiation was done on the 4 mm diameter cylinders, using a multilayer monochromator. The phase contrast effect was obtained with a short propagation distance; the sample-detector distance was of 10 mm for a pixel size equal to 1.27 μm . With this setting (Fig. 1), scanned regions were 4 mm by 2 mm in size when using a half-acquisition mode. Typically, the 3D images obtained are composed of 3639×3639×2048 voxels (elementary image volume) with 32-bit grey-scale resolution. Acquisition time was of the order of 45 min for 5000 projections together with around 625 reference images (blanks). A primary correction of the ring artifacts was done directly after analysis using specific filtering. Tomographic volume reconstructions (example given in Fig. 2) were performed using an algorithm with a single phase retrieval distance in order to optimize the signal/noise ratio ([Paganin et al., 2002](#); [Sanchez et al., 2012](#)). Petrographic data treatments were performed using ImageJ[®] and Avizo[®] softwares. Individual grey levels were identified for each solid (mineral and glass), liquid and gas phase by defining their threshold bracketing on bulk sample grey level histograms calculated using ImageJ[®] (Fig. 3). Then the 3-D volumes of each petrographic phase were defined with the Avizo[®] software. Each phase was individualized by segmentation of the volumes corresponding to their density range.

Uncertainties associated with this segmentation were qualified by manually adjusting minimum and maximum thresholds for each phase (Fig. 4). Once the thresholds well defined, the volumes of the phases were segmented and quantified. The densities of the petrographic phases were calculated from their grey levels using model absorbance compiled with a NIST utility (<http://www.nist.gov/pml/data/xcom/>).

3.1.2 Nano-SIMS microanalyses of C and H

Carbon and water contents of glass were determined using the NanoSIMS 50 installed at the MNHN (Paris) on gold-coated (10 nm thick) samples. Analyses were performed with a 16 keV Cs⁺ primary ion beam rastering the surface of the sample. A primary beam intensity of 40 pA was used to perform analyses with a D1-3 (200 μ m) aperture diaphragm. Secondary ions of ¹²C, ¹⁶O¹H⁻, ²⁸Si⁻, ²⁷Al¹⁶O⁻ and ⁵⁶Fe were collected using electron multipliers over 3 x 3 μ m² square using the beam blanking mode (1.5 x 1.5 μ m²) to reduce surface contamination (Stephant et al., 2014). A presputtering on a 3 x 3 μ m² surface area during 300 s was performed to remove coating and initial surface contamination and to reach the sputtering steady-state. Calibration curves were obtained by measuring tholeiitic glass standards (Fig. 5) provided by the Centre de Recherche en Pétrographie et Géochimie (CRPG), Nancy, France [standard description available in Shishkina et al. (2010)]. Data corrections, using the aforementioned calibration, and error calculations were performed using the R program (Dunn, 2007). Errors combine counting statistics and uncertainty of the calibration curve. However, the errors reported in Table 1 are dominated by the uncertainty of the calibration curve, which corresponds to prediction interval at 68% (1 σ) (Bartoli et al., 2014). During the session, the vacuum in the analysis chamber remained at 5x10⁻¹⁰ Torr.

3.1.3 Raman Spectroscopy and Microthermometry

3.1.3.1 Raman Spectroscopy

The fluid inclusions (trapped by order of abundance in Opx, Cpx and Ol) were analyzed (n=99) by Raman spectroscopy at **Faculty of Science Research and Instrument Core Facility (ELTE FS-RICF, Hungary)**. A Jobin Yvon confocal Labram Raman instrument was used with a frequency-doubled Nd-YAG laser with an excitation wavelength of 532 nm. Laser energy was 50 mW and 20 mW at the laser source and at the sample surface, respectively. 50× and 100× microscope objectives were used to focus the laser onto the sample and to collect the Raman signal. Analytical settings included a 200–500 µm confocal hole, 200 µm spectral slit, 600 or 1800 grooves/mm spectrograph gratings, 2–10 accumulations and 2–150 s acquisition time (all depending on the maximum intensity). The spectral resolution of the measurements varied with wave number but was $\leq 2 \text{ cm}^{-1}$ using the grating with 1800 grooves/mm. Following the method of [Berkesi et al. \(2009\)](#), spectra were collected at room and high (150°C) temperatures using a Linkam THMS600 heating/cooling stage. To characterize the bands in the Raman spectra and to identify the phases present, data were compared to the online Raman database of the French Society of Mineralogy and Crystallography (<http://www.obs.univ-bpclermont.fr/sfmc/ramandb2/index.html>) and to that of [Bonelli & Frezzotti \(2003\)](#), <http://www.dst.unisi.it/geofluids/raman>). Data were processed using LabSpec v5.25.15 software designed for Jobin-Yvon Horiba LabRam instruments. Minimum and maximum values of CO₂ densities in inclusions were calculated from the two Raman bands of CO₂ due to Fermi resonance ([Fall et al., 2011](#)) (Fig. 6).

3.1.3.2 Microthermometry

225

226 A Linkam MDS 600 heating-freezing stage (IFP Energies nouvelles, Rueil-Malmaison,
227 France) was used to measure the phase transitions in CO₂-rich fluid inclusions, which
228 homogenized in the liquid phase in our samples. After freezing the fluid at <-75 °C, the
229 inclusions were slowly heated to the triple point temperature check (-56.6 °C for pure CO₂)
230 with a heating rate of 0.5-2 °C.min⁻¹. Final homogenization occurred to the liquid phase
231 (T_hL). The CO₂ density and pCO₂ were calculated following [Span and Wagner, 1996](#) and
232 [Sterner and Pitzer, 1994](#), respectively, for a magma temperature of 1150 ± 15 °C (estimation
233 from [Creon et al. \(this volume\)](#) from the melting conditions.

234

235 3.2 Results

236

237 3.2.1 Synchrotron X-ray tomography

238

239 3.2.1.1 Petrographic observations

240

241 The density resolution of synchrotron X-ray tomography allows clear discrimination among
242 the six major phases of the peridotitic xenoliths: olivine, orthopyroxene, clinopyroxene,
243 spinel, glass and fluid. Figure 3 shows the density spectra of the SZB51 xenolith together with
244 the modal composition corresponding to the pixel count of each phase. In SZB51 (Fig. 3),
245 olivine is the most abundant phase and fluid is the least abundant.

246 Regarding the 3D images obtained by synchrotron X-ray microtomography, we confirm that
247 the rock textures are protogranular, tabular or equigranular-mosaic (according to the
248 classification of [Mercier & Nicolas, 1974](#)) as previously reported by optical microscopy on
249 thin sections of the same samples ([Creon et al., this volume](#)).

The glass phase may be located at grain boundaries as veins or melt pockets (Fig. 6). Fluids are observed as trails of secondary fluid inclusions (FI) inside minerals (Ol, Opx, Cpx) or as vesicles inside the glass (Fig. 6e). Glass containing vesicles forms a complex interconnected network along grain boundaries as shown in figure 7. Trails of FI run across pre-existing melt veins and melt veins cross pre-existing trails of FI. Moreover, trails of secondary FI and the glass (+ vesicles) network intersect each other through multiple grains. Small (~100-200 μm , Fig. 8) newly formed Ol, Cpx and Sp are observed in veins (in low abundance) and in melt pockets (in high abundance).

The glass (+ vesicle) network percolates through the whole volume and is in direct contact with the exterior, indicating that vesicles probably lost all the gas they previously contained during their exposure to surface conditions or during sample preparation.

3.2.1.2 Phase volume quantification

The volumes of minerals, glass and fluid phases have been quantified within the bulk 3-D volumes. Fluid volumes have been calculated on each sample by direct density threshold segmentation; the obtained volumes are between 0.7 and 4.7 vol% (Fig. 9a & Table 1). Vesicles in melt systematically make the majority of the fluid volume, FIs representing at most 5 vol. %. The FIs can be divided according to which mineral they are hosted in. In sample SZB51, for example, FI represent a volume proportion in opx of 0.6%, in cpx of 0.3%, and in ol of 0.18%. These proportions, weighted by the modal composition of SZB51, imply that opx, ol and cpx trapped respectively 57.8%, 39.0% and 3.3% of the total FI volume (Fig. 9b). Glass volume determination was performed in two steps. The first step resulted from the glass volume calculation by direct threshold segmentation (as for fluid volumes) and gave volumes

between 0 and 3.8 vol% (best estimate). The second step was performed for some samples (SZB50, SZG23, FT12 and FT01P) that showed evidence of important secondary crystallization of the melt (quenched as glass) such as newly formed Ol, Cpx and Sp (Fig. 8). In order to correct the glass volumes from this crystallization of new phases, and to consider the pre-crystallized bulk volume of melt pockets, we extracted the X-ray images of silicate melt pockets on five slices of each sample and calculated their (melt+newly formed minerals)/melt volume ratios (Fig. 8). The obtained ratios were used to correct the 3D glass volumes from crystallization of new phases. The corrected glass volumes in these samples range between 0 and 23.4 vol. % of the total volume (Fig. 9a, Table 1). The corrections for crystal fractionation were performed only on silicate melt pockets since veins only show very minor crystallization that should have negligible effects on the melt volumes. After correction for crystallization of newly formed phases in the melt pockets, the glass and the fluid volumes display a linear correlation (Fig. 11).

3.2.1.3 Density determination

The grey level value of each pixel was quantitatively assigned to a density. Densities determined for Ol (3.37 to 3.41 g.cm⁻³), Opx (3.29 to 3.32 g.cm⁻³) and Cpx (3.42 to 3.43 g.cm⁻³) correspond to those described in the literature using the Mg# of the minerals of these samples [Creon et al. \(this volume\)](#) (Fig. 11). Glass densities were then determined using the average density of glass-assigned pixels in each sample. The densities are distributed between 2.45 and 2.74 g.cm⁻³ (Table 1, Figure 11) and are in agreement with andesite to basalt glass densities available in the literature ([Richet et al. 2000 and references therein](#)). Densities calculated with the method of [Dixon et al. \(1995\)](#) (2.41 to 2.59 g.cm⁻³) are also close to those

determined in this study (Fig. 11). Glass densities determined by X-ray microtomography are therefore considered to be accurate and will be used in the following steps of this study. The X-ray microtomography density determinations for the fluid vesicles and inclusions were not conclusive due to the poor X-ray absorbance resolution in the lower density range (typically 0 to 1.5 g.cm⁻³), together with the partial-volume-effects (density artifact due to the presence of a phase boundary within a single voxel) on the smaller segmented volumes, which commonly affect the fluid/solid phase boundaries.

3.2.2 Fluid compositions and densities

Fluid compositions and densities were measured on FI by Raman spectroscopy and microthermometry. As noted above, fluids originally trapped in vesicles in the melt veins and melt pockets were lost from the sample and could not be analyzed; hence the FI were used as representatives of all the fluids of the samples, even though the vesicles dominated the volume fraction assigned to fluid. Following the method developed in [Berkesi et al. \(2009\)](#), we used Raman spectroscopy at room and elevated temperatures (150 °C) in order to discriminate the potential trace volatiles inside fluid inclusions. With this method, only CO₂ was recognized in the fluid inclusions. Independently, the phase transition observed at ~56.6 ± 1 °C by microthermometry measurements is in agreement with pure CO₂ inclusions. The comparison of densities obtained with both Raman spectroscopy and microthermometry methods shows a linear correlation, with a correlation factor of 0.89 (R²). This suggests that the fluid density datasets obtained by the two methods are internally consistent (Fig. 14a). The densities of fluids obtained in SZB, SZG and FT are respectively between 0.30 and 1.11 g.cm⁻³, 0.48 and 0.82 g.cm⁻³ and 0.30 and 1.06 g.cm⁻³ (Fig.13b).

3.2.3 Volatiles dissolved in glass

NanoSIMS analyses have been performed on glass from veins and melt pockets to determine their CO₂ and H₂O contents. H₂O contents vary from 0.54 ± 0.26 (1 σ) to 4.25 ± 0.27 (1 σ) wt. % in glass from veins and from 0.55 ± 0.26 (1 σ) to 2.76 ± 0.28 (1 σ) wt. % in glass from melt pockets. CO₂ contents vary between 0.27 ± 0.01 (1 σ) and 0.48 ± 0.02 (1 σ) wt. % in glass from veins and from 0.38 ± 0.01 (1 σ) to 0.96 ± 0.02 (1 σ) wt. % in glass from melt pockets (Table 1).

4 Discussion

4.1 Magma CO₂-supersaturation conditions

Several petrographic lines of evidence point to CO₂ supersaturation in the melt trapped within the studied xenoliths at lithospheric depths. Such evidence includes 1) the occurrence of CO₂ FI with densities above 1.0 g.cm^{-3} (Table 1) corresponding to pressures higher than the present Moho depth (25-30 km, [Kovacs et al., 2012](#)); 2) the coexistence within veins of vesicles and glass with high dissolved CO₂ and H₂O concentrations in the glass of up to 0.96 and 4.25 wt. %, respectively (Table 1). In order to assess the conditions of magma supersaturation, CO₂ saturation isotherms were calculated for the different melt compositions measured in the present samples over a range of assumed temperatures using the Rhyolite-MELTS software (version 1.2.0, [Gualda et al., 2012](#); [Ghiorso and Gualda, 2015](#)). This model is suitable for the calculation of thermodynamic equilibrium of mafic to felsic systems with H₂O and CO₂ over the pressure range 0-3 GPa, corresponding to the conditions of the lithospheric mantle below the BBHVF ([Kovacs et al., 2012](#)).

Magma CO₂-saturation shows little temperature dependence between 1100 and 1300 °C and for pressure below 0.5 GPa, as already shown by [Shishkina et al. \(2010\)](#) and [Botcharnikov et al. \(2006\)](#). The effect of temperature only tends to increase the solubility of CO₂ in the high pressure range of 0.5-3.0 GPa, as also mentioned by [Papale et al. \(2006\)](#). Based on CO₂-saturation isotherms, vesicle-melt equilibrium pressures were calculated from the CO₂ contents of melts determined by NanoSIMS (Table 1, Figure 13). These saturation pressures at 1100°C (corresponding to approximated mineral equilibrium temperatures determined in [Creon et al. \(this volume\)](#)) vary between 0.69 and 1.78 GPa and correspond to depths of 26 to 62 km (using the geotherm of [Kovacs et al., 2012](#)). Such estimations likely correspond to vesicle-melt equilibrium near the glass transition during rapid cooling of the percolating melt within the melt veins and pockets. Glass transition temperatures (T_g) were estimated between 670 to 854°C by [Giordano et al. \(2008\)](#) at asthenospheric pressures, and between 520 to 930°C by [Morizet et al. \(2007\)](#) at pressures between 1.0 and 2.5 GPa (phonolite melts). CO₂ appears to have almost no effect on the glass transition of silicate melts unlike H₂O, which clearly reduces glass transition temperatures ([Morizet et al. 2007](#)). Furthermore, pressure has a very limited influence on glass transition temperatures as demonstrated by [Bagdassarov et al. \(2004\)](#). These glass transition temperatures would be intersected along the geotherm of [Kovacs et al. \(2012\)](#) at pressures near ~0.5 GPa i.e. at crustal levels (~20 km depth). Because melt pockets were shown to have been quenched under lithospheric mantle pressures while coexisting with fluid vesicles ([Creon et al., this volume](#)), the crustal saturation pressures recorded for most CO₂ depleted melts are likely to reflect the fast ascent of xenoliths after their extended residence time at lithospheric mantle depths.

Melt saturation pressures of 0.69 to 1.78 GPa were converted to CO₂ density using the equations of state of [Span & Wagner \(1996\)](#) and [Sterner & Pitzer \(1994\)](#). The densities obtained vary between 0.93 and 1.33 g.cm⁻³ and are consistent with the density range

determined by Raman spectroscopy and microthermometry in the denser FIs (Table 1). This similarity attests to a coeval formation of FI and vesicles at lithospheric mantle conditions. However, no direct correlation is observed between maximum fluid inclusion and vesicle densities, which likely points to the unstable behavior of highly over-pressurized FIs (Roedder, 1984). In most cases (samples SZB16, SZB50, SZG23, SZG44, FT12), fluid inclusion densities are lower than the associated calculated vesicle densities; which can occur due to the decrepitation of the FIs during xenolith ascent (Bodnar, 2003). In the rest of the xenoliths (SZB51, SZB52, FT01P) the densities of the FI are larger than those of the associated vesicles, which reflects FI trapping in equilibrium with vesicles at pre-ascent lithospheric mantle pressures.

In figure 14, the CO₂ contents dissolved in glass measured by NanoSIMS (0.27-0.96 wt. %) are compared with the data compilation of Wallace (2005). The values measured in this study are in excess of those reported so far for arc magmas and are within the E-MORB range. Moreover, the CO₂ and H₂O contents measured are larger than those described in the literature in silicate melt pockets of the BBHVF (Bali et al., 2002). Hidas et al. (2010) used silicate melt inclusions (SMI) from Szigliget mantle xenoliths to estimate the volatile content of trapped metasomatic melt. Their calculated melt CO₂ concentrations of ~1 wt. % are in agreement with the maximum CO₂ content measured in glass for this study in sample SZG23 from the same xenolith locality (0.8-1.2 wt. % CO₂; Fig. 14). It is thus likely that the SMI analyzed by Hidas et al. (2010) correspond to the trapping of a volatile-rich melt during overgrowth of clinopyroxenes in the mantle lithosphere similar to the most CO₂-rich percolating melts analyzed in this study.

4.2 CO₂ contents of melts in the subcontinental lithosphere beneath the BBHVF

The 3-D images clearly reveal interconnections between trails of secondary fluid inclusions, the glass in veins and in melt pockets and the network of vesicles. This is contrary to observations made in similar samples from the same localities by [Demény et al. \(2010\)](#), who described trails of FI in thin sections that did not extend to grain boundaries. The synchrotron X-ray microtomography used in this study allows 3-D reconstructions that show indisputable intersecting relationships of FI trails with grain boundaries (Figure 7). These 3-D observations are clear petrographic evidence for a common fluid being present in the FI and in the metasomatic melt that percolated these mantle xenoliths. As noted above, fluid in vesicles are not preserved anymore and are, therefore, impossible to identify. Hence, we assumed that vesicles have similar compositions as FIs and that they are dominated by CO₂. This assumption is consistent with the high H₂O and CO₂ contents in the glass determined by nanoSIMS.

The CO₂/glass mass ratio corresponding to the melt + vesicle system has been calculated using equation (1):

$$(1) \quad M(\text{CO}_2)/M(\text{Glass}) = V(\text{CO}_2)/V(\text{Glass}) * \rho(\text{CO}_2)/\rho(\text{Glass})$$

Where: $M(\text{CO}_2)$: mass of CO₂; $M(\text{Glass})$: mass of glass; $V(\text{CO}_2)$: volume of CO₂; $V(\text{Glass})$: volume of glass; $\rho(\text{CO}_2)$: density of CO₂; $\rho(\text{Glass})$: density of glass

$V(\text{CO}_2)$, $V(\text{glass})$ and $\rho(\text{glass})$ were obtained by synchrotron X-ray microtomography and $\rho(\text{CO}_2)$ was derived from the NanoSIMS analyses and thermodynamic modeling (Rhyolite-MELTS, [Gualda et al. 2012](#); [Ghiorso & Gualda 2015](#)) as described in the previous section (Table 1).

The calculated CO₂/melt mass ratios range from 9.0 to 25.4 wt. % (Table 1). These ratios, which can be considered as the CO₂ mass fraction of the entire melt system, would

correspond to a single phase system at pressures of 4 to 9 GPa for basaltic melts following the study of [Guillot & Sator \(2011\)](#). The present lithosphere-asthenosphere boundary below the Pannonian Basin is estimated at depths between 60 and 100 km ([Tašárová et al., 2009](#)). The melts characterized here were therefore likely to be supersaturated in the asthenosphere before reaching lithospheric depths. The range of CO₂/melt mass ratios calculated in this study implies that CO₂ contents in melts were variable either due to magmatic processes or to physical fractionation. Below we review the various processes that might affect the CO₂/melt ratio in lithospheric melts.

4.2.1 Carbonate assimilation and fractionation in the lithospheric mantle

Carbonate minerals have been described in Pannonian Basin xenoliths from the BBHVF. These carbonates are mainly calcitic droplets ([Demény et al., 2004](#)), rounded ([Bali et al. 2002; 2008](#)) or anhedral ([Chalot-Prat and Arnold, 1999](#)) crystals in silicate veins and/or melt pockets of mantle xenoliths. In the samples of this study, carbonate are also observed as veins ([Creon et al., this volume](#)). [Bali et al. \(2002\)](#) and [Demény et al. \(2004\)](#) suggested the infiltration of a mixed carbonate–silicate melt taking place in these mantle xenoliths, which supports the evidence for CO₂-rich magma percolation beneath the BBHVF documented in this study. Carbonates are, therefore, not the source of the melts but instead precipitate from a CO₂-rich metasomatic agent during melt-rock reactions due to carbon saturation and vapor immiscibility in the percolating silicate melts. Thus, the variability in CO₂ mass fractions in melts calculated in this study is not believed to be caused by variable dissolution of carbonated peridotites in the lithospheric mantle. Reversely, the fractionation of carbonate minerals, or phase separation of carbonate and silicate melts may have potentially played a role in fractionating CO₂ mass fractions in

magmas. Since no evidences of carbonate fractionation, nor carbonate melt segregation have been witnessed in our sample set, although they were observed and discussed in comparable samples of xenoliths of Transylvania (Chalot-Prat and Arnold, 1999; Bali et al., 2002; Demény et al., 2004; Bali et al., 2008) we suggest that such processes were not the driving mechanisms of the observed heterogeneity in the CO₂ mass fraction.

4.2.2. Fluid inclusions and physical fractionation

The large abundance of CO₂-rich FIs trapped in minerals along fractures and the absence of comparably abundant silicate melt inclusions indicate that CO₂ was preferentially trapped in grains in comparison with the silicate melt.

Schiano et al. (2006) suggested that the occurrence of fluid inclusion arrays, commonly observed in mantle xenoliths, could be explained by transcrystalline melt migration, contrary to the commonly accepted idea of simple trapping by "crack and seal" processes of individual or multiple mineral grains of a free fluid phase percolating the deep mantle. However, no textural features suggesting transcrystalline melt migration were observed in our samples. Schiano et al. (2006) explained that only the most primitive droplets formed at fertile grain corners (or edges) have time (if melting does not proceed rapidly to higher degrees) to enter the crystals by thermally driven migration instead of escaping by buoyancy-driven flow along grain edges. Melts analyzed in this study have evolved compositions (55.96-62.71 wt. % SiO₂), consistent with the formation of an interconnected wet-edge network and rapid escape to higher levels, where grain scale channels are relayed by fracture networks to form magma bodies (Schiano et al., 2006). The absence of the high thermal gradient (0.03 to 0.06 K/μm) needed for transcrystalline melt migration (Schiano et al., 2006) could also explain the absence of evidence of transcrystalline melt migration in the mantle xenoliths studied here.

The microstructures observed in samples from this study suggest that peridotites are wetted by silicate melt and not by CO₂ as already indicated by several authors (e.g. [Watson & Brenan, 1987](#)). This observation is consistent with the much smaller wetting angles expected for silicate melt in dunite than for CO₂ in dunite ([Watson & Brenan, 1987](#)). Consequently, melt inclusions only rarely trap melts with their primary CO₂ contents (e.g., [Saal et al. 2002](#)). Instead, grain fractures due to mantle dynamic or to local overpressure may much more easily trap volatiles as FI due to preferential viscosity-controlled injection within grain cracks ([Watson and Brenan, 1987](#)). We suggest that cracking and healing processes operating in response to fluid overpressure and creeping of the peridotite matrix induced the preferential entrapment of CO₂ in minerals whereas the silicate melt from the same magma continued its journey at grain boundaries through percolation. Variable amplitudes of this fractionation process may then be expected depending on magma travelling distances, melt/rock ratios and fluid overpressure conditions.

4.3 Trapped lithospheric melts in mantle xenoliths and arc magmas

[Creon et al. \(this volume\)](#) interpreted the metasomatic agent that infiltrated these xenoliths to be a CO₂-rich adakite-like magma derived from slab melting that interacted with the lithosphere mantle before being trapped in the peridotites. [Wallace \(2005\)](#) compiled the CO₂ and H₂O contents of silicate melt inclusions deduced from arc-magmas worldwide, which range between 0 and 2100 ppm and between 1.9 and 6.3 wt. % respectively (Figure 15). The CO₂ concentrations are much lower than the values determined in this study, which range from 0.29 to 0.96 wt. % and are much higher than that of SMI in arc magmas worldwide. The H₂O contents determined in this study range from 0.54 and 4.25 wt. % and are for the majority, in the same range as those reported by [Wallace \(2005\)](#). We suggest that the CO₂

content of arc magmas sampled by SMIs may only represent a negligible fraction of the CO₂ associated with slab-derived melts. For instance, in xenoliths from this study as well as in those from the literature ([Bali et al. \(2008\)](#) and [Hidas et al. \(2010\)](#)), only a very small fraction of the CO₂-rich melt may be trapped as SMIs due to the pre-entrapment supersaturation in CO₂ and the preferential accumulation of most of the CO₂ in vesicles and FIs (95-98 % in samples from this study) rather than in the silicate melt. A large compositional gap seems to exist in subduction-related environments between CO₂/magma mass fractions trapped as melts in the lithospheric mantle and melts characterizing erupting subduction-related magmas. This gap may be the result of physical and/or geochemical processes such as 1) intense physical fractionation of CO₂ vesicles and silicate melt during magma ascent to the crust, 2) important dilution of CO₂-rich parental magmas by assimilation of C-poor lithologies during magma-rock interaction. The investigation of these processes is however beyond the scope of this paper and will not be discussed further here.

Figure 15 illustrates the different potential sources of H₂O and CO₂ expected in a subduction context where 1) E-MORB, 2) altered oceanic basalts and 3) subducted sediments ([Plank and Langmuir, 1998](#); [Wallace, 2005](#)) represent rock-type end-members of slab melting. Partial melting of the slab or of the metasomatized mantle wedge would generate melts enriched in incompatible trace elements and with high CO₂ and H₂O contents like those calculated in this study (0.55-4.25 wt. % H₂O and 9.0-25.4 wt. % CO₂; Figure 15 and Table 1). If the CO₂ content of the melt is converted into carbonate content at the source (which implies congruent melting of the source), the latter could have contained up to 20-40 wt. % of carbonate.

[Dasgupta et al. \(2013\)](#) experimentally showed that carbonated silicate melt with 15-25 wt. % CO₂ and a modest amount of water (1-5 wt. %) can be generated in the upper part of the asthenospheric mantle (1,300-1,750 °C and 2-5 GPa) and is able to percolate upward the lithospheric mantle. We conclude that the high CO₂/melt ratios determined in this study

correspond to deep melts derived from slab melting, which have percolated to lithospheric depths without having been substantially fractionated.

4.4 A lithospheric carbon reservoir?

The mass ratios of CO₂/peridotite have been calculated for each xenoliths which provide values between 2700 and 24,400 ppm by weight of CO₂ with an average of about 9557 ppm. This is considered representative of the average quantity of carbon stored in the samples examined. However, the present study is based on fluid inclusion-rich samples only, which correspond to ~20 % of an initial sample set of collected mantle xenoliths. If we make a conservative assumption that the remaining 80% of the initial sample set contains on average 30 ppm carbon, a typical estimate for the N-MORB mantle (Marty, 2012), then we can deduce a first order estimate of the CO₂ content present in the Pannonian lithospheric mantle. Thus, we estimate that quantity to be 1935 ppm CO₂ ($0.2 \times 9557 \text{ ppm} + 0.8 \times 30 \text{ ppm} = 1935 \text{ ppm}$). We consider this first order CO₂ content as representative of the bulk lithospheric mantle sampled by the alkaline volcanism of the BBHVF. This estimate is within a factor of two of the 1000 ppm estimated by Kelemen & Manning (2015) for back-arc lithospheric mantle based on observed carbon fluxes and reservoirs at subduction zones. For comparison, the carbon content at source regions of MORB is thought to be bound between 50 and 200 ppm, whereas the source mantle of OIB is richer and may have be between 1000 and 4000 ppm CO₂ (Marty and Tolstikhin, 1998; Saal et al., 2002; Pineau et al., 2004; Cartigny et al., 2008). We suggest that the quantitative peridotite + melt + volatile characteristics described here, and found in many other regions of the Earth, may be representative of the carbon reservoir residing in the lithospheric mantle, at least for back-arc areas.

Besides the implications of our findings for the carbon cycle, such a lithospheric carbon reservoir has several consequences for the degassing of mantle volatiles in active geodynamic settings such as the Pannonian Basin, where mantle-derived CO₂ outgassing through the Pannonian Basin sedimentary rocks is observed (Ballentine et al., 1991). This suggests that the carbon reservoir of at least ~2000 ppm quantified in this study is transient, and may be recycled to the atmosphere and to the asthenosphere over geodynamic timescales by passive tectonic degassing, active volcanism or delamination.

5 Conclusions

The magmatic-CO₂ budget of the subcontinental lithospheric mantle below the Pannonian Basin has been constrained by coupling synchrotron 3-D X-ray microtomography, nanoSIMS, Raman spectroscopy, and microthermometry in combination with thermodynamic modeling in order to investigate the significance of abundant CO₂-rich fluid inclusions, CO₂-bearing glass and vesicles within lithospheric peridotite xenoliths.

3-D volume reconstructions allow quantifying the proportions of all mineral phases in the xenoliths and their textural relationships. Interconnection between trails of secondary fluid inclusions and the network of glass + vesicles demonstrates a common fluid composition between fluid inclusions and melt percolation in these mantle xenoliths. The vapor phase volume (vesicles + fluid inclusions) observed requires CO₂-supersaturation of the silicate melts, which allows to calculate melt saturation pressures based on the dissolved CO₂ content of the glass.

The CO₂/glass mass ratio corresponding to the melt + vesicle system is estimated from the microtomography, NanoSIMS analyses and thermodynamic modeling. The CO₂/melt mass ratios obtained range from 9.0 and 25.4 wt. % and are related to a slab-derived melting

source. A first estimate of ~1935 ppm CO₂ for the Pannonian Basin lithospheric mantle is proposed and can be considered as representative of the bulk lithospheric mantle sampled by the alkaline basalt volcanism of the BBHVF. This petrologic and geochemical approach is the first direct quantification of the bulk carbon content in the lithospheric mantle, and is in agreement with independent lines of reasoning suggesting a large transient reservoir of carbon in the lithospheric mantle near subduction-arc systems. Further quantification of the volume, residence time, and general significance of this reservoir requires more systematic regional and global studies using similar approach.

6 Acknowledgements

This project was funded by IFP Energies nouvelles. The European Synchrotron Research Facility of Grenoble, the ID19 beamline team and more specifically Paul Tafforeau are thanked for their financial and technical support. The National NanoSIMS facility at the MNHN was established by funds from the CNRS, Région Ile de France, Ministère délégué à l'Enseignement supérieur et à la Recherche, and the MNHN. We are grateful to Marie-Claude Lynch and Herman Ravelojaona for their precious assistance with X-ray tomography and sample preparations.

7 References

- Anderson A. T. (1975) Some basaltic and andesitic gases. *Rev. Geophys.* **13**, 37.
- Bagdassarov N. S., Maumus J., Poe B., Slutskiy A. B. and Bulatov V. K. (2004) Pressure dependence of T_g in silicate glasses from electrical impedance measurements. *Phys. Chem. Glas. - Eur. J. Glas. Sci. Technol. Part B*, **45**, 197–214.

599 Bali E., Szabó C., Vaselli O. and Torok K. (2002) Significance of silicate melt pockets in
600 upper mantle xenoliths from the Bakony – Balaton Highland Volcanic Field , Western
601 Hungary. *Lithos* **61**, 79–102.

602 Bali E., Zanetti A., Szabó C., Peate D. W. and Waight T. E. (2008) A micro-scale
603 investigation of melt production and extraction in the upper mantle based on silicate melt
604 pockets in ultramafic xenoliths from the Bakony–Balaton Highland Volcanic Field
605 (Western Hungary). *Contrib. to Mineral. Petrol.* **155**, 165–179.

606 Ballentine C. J., O’Nions R. K., Oxburgh E. R., Horvath F. and Deak J. (1991) Rare gas
607 constraints on hydrocarbon accumulation, crustal degassing and groundwater flow in the
608 Pannonian Basin. *Earth Planet. Sci. Lett.* **105**, 229–246.

609 Balogh K. and Németh K. (2005) Evidence for the Neogene small-volume intracontinental
610 volcanism in Western Hungary: K/Ar geochronology of the Tihany Maar Volcanic
611 Complex. *Geol. Carpathica* **56**, 91–99.

612 Balogh K. and Pécskay Z. (2001) K/Ar and Ar/Ar geochronological studies in the Pannonian-
613 Carpathians-Dinarides (PANCARDI) region. *Acta Geol. Hung.* **44**, 281–299.

614 Bartoli O., Cesare B., Remusat L., Acosta-Vigil A. and Poli S. (2014) The H₂O content of
615 granite embryos. *Earth Planet. Sci. Lett.* **395**, 281–290.

616 Baruchel J., Buffiere J.-Y., Maire E., Merle P. and Peix G. (2000) X-ray Tomography in
617 Material Science. *Hermes Sci. Publ.*

618 Berkesi M., Hidas K., Guzmics T., Dubessy J., Bodnar R. J., Szabó C., Vajna B. and
619 Tsunogae T. (2009) Detection of small amounts of H₂O in CO₂-rich fluid inclusions
620 using Raman spectroscopy. *J. Raman Spectrosc.* **40**, 1461–1463.

621 Bodnar R. J. (2003) Re-equilibration of fluid inclusions. *Fluid Inclusions Anal. Interpret.*
622 *Short Course Ser. Samson I, Anderson A, Marshall D Miner. Assoc Can, Vancouver* **32**,
623 213–231.

624 Bonelli R. and Frezzotti M. . (2003) Raman Spectra Database.
625 <http://www.dst.unisi.it/geofluidslab/Raman%20intro.html>.
626 Botcharnikov R., Behrens H. and Holtz F. (2006) Solubility and speciation of C–O–H fluids
627 in andesitic melt at T=1100–1300°C and P=200 and 500MPa. *Chem. Geol.* **229**, 125–
628 143.

629 Bruce Watson E. and Brenan J. M. (1987) Fluids in the lithosphere, 1. Experimentally-
630 determined wetting characteristics of CO₂H₂O fluids and their implications for fluid
631 transport, host-rock physical properties, and fluid inclusion formation. *Earth Planet. Sci.*
632 *Lett.* **85**, 497–515.

633 Burton M. R., Sawyer G. M. and Granieri D. (2013) Deep Carbon Emissions from Volcanoes.
634 *Rev. Mineral. Geochemistry* **75**, 323–354.

635 Cartigny P., Pineau F., Aubaud C. and Javoy M. (2008) Towards a consistent mantle carbon
636 flux estimate: Insights from volatile systematics (H₂O/Ce, δD, CO₂/Nb) in the North
637 Atlantic mantle (14° N and 34° N). *Earth Planet. Sci. Lett.* **265**, 672–685.

638 Chalot-Prat F. and Arnold M. (1999) Immiscibility between calciocarbonatitic and silicate
639 melts and related wall rock reactions in the upper mantle: a natural case study from
640 Romanian mantle xenoliths. *Lithos* **46**, 627–659.

641 Creon L., Delpech G., Rouchon V. and Guyot F. Melt-peridotite interaction in the lithospheric
642 mantle - Genesis of the Pannonian calc-alkaline suites revealed by Alkali basalt hosted
643 xenoliths. *this Vol.*

644 Dasgupta R. (2013) Ingassing, Storage, and Outgassing of Terrestrial Carbon through
645 Geologic Time. *Rev. Mineral. Geochemistry* **75**, 183–229. Dasgupta R. and Hirschmann
646 M. M. (2010) The deep carbon cycle and melting in Earth’s interior. *Earth Planet. Sci.*
647 *Lett.* **298**, 1–13.

648 Dasgupta R., Mallik A., Tsuno K., Withers A. C., Hirth G. and Hirschmann M. M. (2013)

649 Carbon-dioxide-rich silicate melt in the Earth's upper mantle. *Nature* **493**, 211–215.

650 Demény A., Dallai L., Frezzotti M.-L., Vennemann T. W., Embey-Isztin A., Dobosi G. and

651 Nagy G. (2010) Origin of CO₂ and carbonate veins in mantle-derived xenoliths in the

652 Pannonian Basin. *Lithos* **117**, 172–182.

653 Demény A., Vennemann T. ., Hegner E., Nagy G., Milton J. ., Embey-Isztin A., Homonnay Z.

654 and Dobosi G. (2004) Trace element and C–O–Sr–Nd isotope evidence for subduction-

655 related carbonate–silicate melts in mantle xenoliths (Pannonian Basin, Hungary). *Lithos*

656 **75**, 89–113.

657 Dixon J. E., Stolper E. M., Holloway J. R. and Science A. (1995) An Experimental Study of

658 Water and Carbon Dioxide Solubilities in Mid-Ocean Ridge Basaltic Liquids . Part II :

659 Applications to Degassing. *J. Petrol.* **36**, 1607–1631.

660 Dunn P. K. (2007) tweedie: Tweedie exponential family models.

661 Embey-Isztin a, Dobosi G., Altherr R. and Meyer H.-P. (2001) Thermal evolution of the

662 lithosphere beneath the western Pannonian Basin: evidence from deep-seated xenoliths.

663 *Tectonophysics* **331**, 285–306.

664 Embey-isztin A., Downes H., James D. E. E., Upton B. G. J. G. J., Dobosi G., Ingram G. A.

665 a., Harmon R. S. S. and Scharbert H. G. G. (1993) The petrogenesis of Pliocene alkaline

666 volcanic rocks from the Pannonian Basin, Eastern central Europe. *J. Petrol.* **34**, 317–343.

667 Embey-Isztin A., Sharbert H. G., Dietrich H. and Poultidis H. (1990) Mafic granulites and

668 clinopyroxenite xenoliths from the Transda- nubian Volcanic Region (Hungary):

669 implications for the deep structure of the Pannonian Basin. *Mineral. Mag.* **54**, 463–483.

670 Fall A., Tattitch B. and Bodnar R. J. (2011) Combined microthermometric and Raman

671 spectroscopic technique to determine the salinity of H₂O–CO₂–NaCl fluid inclusions

672 based on clathrate melting. *Geochim. Cosmochim. Acta* **75**, 951–964.

673 Falus G., Szabó C., Kovács I., Zajacz Z. and Halter W. (2007) Symplectite in spinel lherzolite

674 xenoliths from the Little Hungarian Plain, Western Hungary: A key for understanding
 675 the complex history of the upper mantle of the Pannonian Basin. *Lithos* **94**, 230–247.

676 Fodor L., Csontos L., Bada G., Györfi I. and Benkovics L. (1999) Tertiary tectonic evolution
 677 of the Pannonian basin and neighbouring orogenes: a new synthesis of palaeostress data.
 678 In: Durand, B., Jolivet, L., Horváth, F., Séranne, M. (Eds.), The Mediterranean Basins:
 679 Tertiary Extension with Alpine Orogen. *Geol. Soc.* **156**, 295–334.

680 Fyfe W. S., Price N. J. and Hompson A. B. (1978) Fluids in the Earth's Crust: Their
 681 Significance in Metamorphic, Tectonic, and Chemical Transport Processes. *Elsevier*,
 682 *Amsterdam*, 384 pp.

683 Ghiorso M. S. and Gualda G. A. R. (2015) An H₂O-CO₂ mixed fluid saturation model
 684 compatible with rhyolite-MELTS. *Contrib. to Mineral. Petrol.*

685 Giordano D., Russell J. K. and Dingwell D. B. (2008) Viscosity of magmatic liquids: A
 686 model. *Earth Planet. Sci. Lett.* **271**, 123–134.

687 Gorman P. J., Kerrick D. M. and Connolly J. a D. (2006) Modeling open system metamorphic
 688 decarbonation of subducting slabs. *Geochemistry, Geophys. Geosystems* **7**.

689 Gualda G. A. R., Ghiorso M. S., Lemons R. V. and Carley T. L. (2012) Rhyolite-MELTS: a
 690 Modified Calibration of MELTS Optimized for Silica-rich, Fluid-bearing Magmatic
 691 Systems. *J. Petrol.* **53**, 875–890.

692 Guillot B. and Sator N. (2011) Carbon dioxide in silicate melts: A molecular dynamics
 693 simulation study. *Geochim. Cosmochim. Acta* **75**, 1829–1857.

694 Hammouda T. (2003) High-pressure melting of carbonated eclogite and experimental
 695 constraints on carbon recycling and storage in the mantle. *Earth Planet. Sci. Lett.* **214**,
 696 357–368.

697 Harangi S. and Lenkey L. (2007) Genesis of the Neogene to Quaternary volcanism in the
 698 Carpatian Pannonian region : Role of subduction , extension , and mantle plume. *Geol.*

699 *Soc. Am. Special Pa*, 67–92.

700 Hidas K., Guzmics T., Szabó C., Kovács I., Bodnar R. J., Zajacz Z., Nédli Z., Vaccari L. and

701 Perucchi A. (2010) Coexisting silicate melt inclusions and H₂O-bearing, CO₂-rich fluid

702 inclusions in mantle peridotite xenoliths from the Carpathian–Pannonian region (central

703 Hungary). *Chem. Geol.* **274**, 1–18.

704 Horvath F. (1993) Towards a mechanical model for the formation of the Pannonian Basin.

705 *Tectonophysics* **226**, 333–357.

706 Horváth F., Musitz B., Balázs a., Végh A., Uhrin a., Nádor A., Koroknai B., Pap N., Tóth T.

707 and Wórum G. (2015) Evolution of the Pannonian basin and its geothermal resources.

708 *Geothermics* **53**, 328–352.

709 Jarrard R. D. (2003) Subduction fluxes of water, carbon dioxide, chlorine, and potassium.

710 *Geochemistry, Geophys. Geosystems* **4**.

711 Johnston F. K. B., Turchyn A. V. and Edmonds M. (2011) Decarbonation efficiency in

712 subduction zones: Implications for warm Cretaceous climates. *Earth Planet. Sci. Lett.*

713 **303**, 143–152.

714 Kazmer M. and Kovacs S. (1985) erman–Paleogene paleogeography along the Eastern part

715 of the Insubric–Periadriatic Lineament system: evidence for continental escape of the

716 Bakony- Drauzug unit. *Acta Geol. Hung.* **1-2**, 71–84.

717 Kelemen P. B. and Manning C. E. (2015) Reevaluating carbon fluxes in subduction zones,

718 what goes down, mostly comes up. *Proc. Natl. Acad. Sci.*, 201507889.

719 Kovacs I., Falus G., Stuart G., Hidas K., Szabó C., Flower M. F. J., Hegedűs E., Posgay K.

720 and Zilahi-Sebess L. (2012) Seismic anisotropy and deformation patterns in upper

721 mantle xenoliths from the central Carpathian–Pannonian region: Asthenospheric flow as

722 a driving force for Cenozoic extension and extrusion? *Tectonophysics* **514-517**, 168–179.

723 Kovács I. and Szabó C. (2008) Middle Miocene volcanism in the vicinity of the Middle

724 Hungarian zone: Evidence for an inherited enriched mantle source. *J. Geodyn.* **45**, 1–17.

725 Labiche J.-C., Mathon O., Pascarelli S., Newton M., Guilera Ferre G., Curfs C., Vaughan G.,

726 Homs A. and D. F. C. (2007) The fast readout low noise camera as a versatile X-ray

727 detector for time resolved dispersive extended X-ray absorption fine structure and

728 diffraction studies of dynamic problems in materials science, chemistry, and catalysis.

729 *Rev. Sci. Instrum.* **78**, 091301–091311.

730 Marty B. (2012) The origins and concentrations of water, carbon, nitrogen and noble gases on

731 Earth. *Earth Planet. Sci. Lett.* **313-314**, 56–66.

732 Marty B. and Tolstikhin I. N. (1998) CO₂ fluxes from mid-ocean ridges, arcs and plumes.

733 *Chem. Geol.* **145**, 233–248.

734 Mercier J. C. and Nicolas A. (1975) Textures and Fabrics of Upper-Mantle Peridotites as

735 Illustrated by Xenoliths from Basalts. *J. Petrol.* **16**.

736 Morizet Y., Nichols A. R. L., Kohn S. C., Brooker R. a. and Dingwell D. B. (2007) The

737 influence of H₂O and CO₂ on the glass transition temperature: insights into the effects of

738 volatiles on magma viscosity. *Eur. J. Mineral.* **19**, 657–669.

739 Paganin D., Mayo S. C., Gureyev T. E., Miller P. R. and Wilkins S. W. (2002) Simultaneous

740 phase and amplitude extraction from a single defocused image of a homogeneous object.

741 *J. Microsc.* **206**, 33–40.

742 Papale P., Moretti R. and Barbato D. (2006) The compositional dependence of the saturation

743 surface of H₂O+CO₂ fluids in silicate melts. *Chem. Geol.* **229**, 78–95.

744 Pineau F., Shilobreeva S., Hekinian R., Bideau D. and Javoy M. (2004) Deep-sea explosive

745 activity on the Mid-Atlantic Ridge near 34°50'N: a stable isotope (C, H, O) study. *Chem.*

746 *Geol.* **211**, 159–175.

747 Plank T. and Langmuir C. H. (1998) The chemical composition of subducting sediment and

748 its consequences for the crust and mantle. *Chem. Geol.* **145**, 325–394.

749 Poinar K. (2012) The Carbon Cycle in the Mantle. , 1–11.

750 Praus O., Pěčová J., Petr V., Babuška V. and Plomerová J. (1990) Magnetotelluric and
 751 seismological determination of the lithosphere-asthenosphere transition in Central
 752 Europe. *Phys. Earth Planet. Inter.* **60**, 212–228.

753 Richet P., Whittington A., Holtz F., Behrens H., Ohlhorst S. and Wilke M. (2000) Water and
 754 the density of silicate glasses. *Contrib. to Mineral. Petrol.* **138**, 337–347.

755 Roedder E. (1984) Fluid inclusions. *Rev. Mineral.* **12**.

756 Saal A. E., Hauri E. H., Langmuir C. H. and Perfit M. R. (2002) Vapour undersaturation in
 757 primitive mid-ocean-ridge basalt and the volatile content of Earth's upper mantle. *Nature*
 758 **419**, 451–5.

759 Sanchez S., Ahlberg P. E., Trinajstić K. M., Mirone A. and Tafforeau P. (2012) Three
 760 Dimensional Synchrotron Virtual Paleohistology: A New Insight into the World of
 761 Fossil Bone Microstructures. *Microsc. Microanal.* **18**, 1095–1105.

762 Sano Y. and Marty B. (1995) Origin of carbon in fumarolic gas from island arcs. *Chem. Geol.*
 763 **119**, 265–274.

764 Schiano P., Provost A., Clocchiatti R., Faure F., Schiano P., Provost A., Clocchiatti R. and
 765 Faure F. (2006) Transcrystalline Melt Migration and Earth's Mantle. *Science* (80-.).
 766 **314**, 970–974.

767 Shcheka S. S., Wiedenbeck M., Frost D. J. and Keppler H. (2006) Carbon solubility in mantle
 768 minerals. *Earth Planet. Sci. Lett.* **245**, 730–742.

769 Sherwood Lollar B., Ballentine C. J. and Onions R. K. (1997) The fate of mantle-derived
 770 carbon in a continental sedimentary basin: Integration of relationships and stable isotope
 771 signatures. *Geochim. Cosmochim. Acta* **61**, 2295–2307.

772 Shishkina T. a., Botcharnikov R. E., Holtz F., Almeev R. R. and Portnyagin M. V. (2010)
 773 Solubility of H₂O- and CO₂-bearing fluids in tholeiitic basalts at pressures up to

774 500MPa. *Chem. Geol.* **277**, 115–125.

775 Sleep N. H., Zahnle K. J. and Neuhoff P. S. (2001) Initiation of clement surface conditions on
 776 the earliest Earth. *Proc. Nat. Acad. Sci.* **98**, 3666–3672.

777 Spakman W. (1990) Tomographic images on the upper mantle below Europe and the
 778 Mediterranean. *Ter. Nov.* **2**, 542–553.

779 Span R. and Wagner W. (1996) A New Equation of State for Carbon Dioxide Covering the
 780 Fluid Region from the Triple-Point Temperature to 1100 K at Pressures up to 800 MPa.
 781 *J. Phys. Chem.* **25**, 1509–1596.

782 Stephant A., Remusat L., Thomen A. and Robert F. (2014) Reduction of OH contamination in
 783 quantification of water contents using NanoSIMS imaging. *Chem. Geol.* **380**, 20–26.

784 Sterner S. M. and Pitzer K. S. (1994) An equation of state for carbon dioxide valid from zero
 785 to extreme pressures. *Contrib. to Mineral. Petrol.* **117**, 362–374.

786 Szabó C., Harangi S. and Csontos L. (1992) Review of Neogene and Quaternary volcanism of
 787 the Carpathian-Pannonian region. *Tectonophysics* **208**, 243–256.

788 Szabó C., Kovacs I., Dégi J., Kothay K., Torok K., Hidas K. K., Konya P., Berkesi M., Degi
 789 J., Kothay K., Torok K., Hidas K. K., Konya P. and Berkesi M. (2010) From maars to
 790 lava lakes: Ultramafic and granulite xenoliths associated with the alkaline basaltic
 791 volcanism of the Pannonian Basin. In *Mineralogica-petrographica field guide series* pp.
 792 1–32.

793 Tašárová A., Afonso J. C., Bielik M., Götze H.-J. and Hók J. (2009) The lithospheric structure
 794 of the Western Carpathian–Pannonian Basin region based on the CELEBRATION 2000
 795 seismic experiment and gravity modelling. *Tectonophysics* **475**, 454–469.

796 Wallace P. J. (2005) Volatiles in subduction zone magmas: concentrations and fluxes based
 797 on melt inclusion and volcanic gas data. *J. Volcanol. Geotherm. Res.* **140**, 217–240.

798 Wijbrans J., Németh K., Martin U. and Balogh K. (2007) $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology of

799 Neogene phreatomagmatic volcanism in the western Pannonian Basin, Hungary. *J.*
800 *Volcanol. Geotherm. Res.* **164**, 193–204. Youxue Zhang and Zindler A. (1993)
801 Distribution and evolution of carbon and nitrogen in Earth. *Earth Planet. Sci. Lett.* **117**,
802 331–345.
803

Figures captions

Figure 1: Principles of synchrotron X-ray microtomography. a- Sample is placed into the beam in different angular positions; the transmitted beam is saved by the detector. b- All radiographs obtained are compiled and a 3D image is reconstructed mathematically. c- Each voxel of the 3D image is assigned to a grey level proportional to the local attenuation coefficient of the incident beam. d- Example of drilled peridotite sample.

Figure 2: Reconstructed X-ray tomography 3D image of the SZB51 sample (4 mm diameter). The dark grey phase is orthopyroxene, the medium-grey grey phase is olivine and the white is clinopyroxene. We can observe in black the grain boundaries, cracks and some trails of secondary fluid inclusions.

Figure 3: Density histograms of xenolith SZB51 (A, B) and xenolith FT12 (C, D) acquired by X-ray tomography. Left: full histograms. Right: deconvolved histogram with density peak assignment (B, D).

Figure 4: Example of phase density threshold defining for fluid vesicle segmentation of xenolith SZB51. Three cases of density threshold assigning were defined for each sample. The “Chosen fluid threshold” is considered the best estimated threshold. The minimum (“Min. fluid threshold”) and maximum (“Max. fluid threshold”) thresholds represent the upper and lower acceptable values near the best estimate. These provide a sense of the uncertainty associated with the user defined density threshold assigning.

Figure 5: NanoSIMS CO₂ and H₂O calibration curves (dashed curves) obtained with the tholeiitic glass standards (black dots) provided by CRPG, Nancy, France [standard description available in [Shishkina et al. \(2010\)](#)]. CO₂ and H₂O were normalized to SiO₂.

Figure 6: X-ray microtomography images illustrating the variety of textures and the different evidences of glass and fluid inclusions in mantle xenoliths. a - Protogranular texture. b- Equigranular mosaic texture. c- Equigranular tabular texture. d- Melt pocket with large vesicles (Vs), glass (Gl) and secondary crystallized minerals. e- A melt vein (glass + vesicles) and trails of secondary fluid inclusions (FI). Opx – orthopyroxene, Ol – olivine, Opx – orthopyroxene, SP- spinel, Gl – glass, Vs – vesicle.

Figure 7: 3D representation of the segmented volumes of glass (blue) and fluid (pink and black) obtained by X-ray tomography in xenolith SZB51 (4 mm diameter). This image clearly illustrates the crosscutting relationship between trails of fluid inclusions and glass+vesicles along grain boundaries. In this sample 5 Vol. % of the fluid phase is in fluid inclusions (black) and 95 Vol. % is in vesicles (pink) trapped in glass.

Figure 8: Example of melt pocket volume segmentation based on X-ray tomography. Pink, blue and brown phases refer to vesicles, glass and newly formed minerals, respectively.

Figure 9: a- Abundances of glass, fluid (FI + vesicles) and mineral assemblages for each studied xenoliths ([Creon et al., this volume](#)); b- AbundanceFluid inclusion abundances by host mineral in SZB51.

Figure 10: Glass and fluid volume correlation. Arrows indicate the crystal fractionation correction of the glass volume. Diamonds represent xenoliths with melt pockets + melt veins; circles represent xenoliths containing only melt veins. Black, dark-grey and light-grey symbols represent respectively SZB (Szentbékállá), SZG (Szigliget) and FT (Füzes-tó) xenoliths.

Figure 11: Comparison between calculated densities based on EPMA chemical analyses (Dixon et al., 1995) and measured densities by X-ray micro-CT for olivine, orthopyroxene, clinopyroxene and glass. The shaded areas correspond to density range for melt of basalt, andesite and trachyte ([Richet et al. 2000 and references therein](#))

Figure 12: A- Correlation between density determination methods (Raman spectroscopy vs. microthermometry); B- Histogram of FI density for locations of the studied xenoliths (SZB - Szentbékállá, SZG - Szigliget & FT - Füzes-tó)

Figure 13: CO₂ saturation curves calculated with rhyolite-MELTS (Gualda et al., 2012; Ghiorso and Gualda, 2015) at 1000 °C using glass compositions obtained by EMPA and H₂O contents obtained by Nano-SIMS measurements. Saturation pressure estimates of the melt prior to quenching are defined by the intercept of the CO₂ contents in glass (NanoSIMS data) and the CO₂ saturation curves (grey lines). Black diamonds: SZB xenoliths), dark-grey diamonds: SZG xenoliths) and light-grey diamonds: FT xenoliths)

Figure 14: H₂O vs. CO₂ for glass trapped as vein or melt pockets in SZB (Szentbékállá); SZG (Szigliget) and FT (Füzes-tó) mantle xenoliths. For comparison melt inclusions from arc

basalts are shown (Grey fields, Modified from [Wallace, 2005](#)). Literature data are shown for Cerro Negro (Roggensack et al., 1997), Fuego (Fg; [Roggensack 2001](#)), Guatemalan volcanoes behind the volcanic front (Guat BVF; [Walker et al. 2003](#)), the Trans-Mexican Volcanic Belt (Luhr, 2001; Cervantes and Wallace, 2003), Stromboli (Metrich et al., 2001) and Cayuse Crater, a cinder cone with OIB-like trace element composition from the Oregon Cascades (Cas OIB; [M. Rowe, unpublished data](#)). For Cerro Negro, only inclusions from the 1992 explosive eruption are plotted. For comparison primary CO₂ contents of N-MORB ([Marty and Tolstikhin, 1998](#)), E-MORB ([Hekinian et al., 2000](#)), Kilauea basalt ([Gerlach et al., 2002](#)) and popping-rock ([Graham and Sarda, 1991](#)) are shown. Also shown is the CO₂ content of basaltic magma from Reunion based on rare high-pressure melt inclusions ([Bureau et al., 1999](#)). SZB, Szentbékállá; SZG, Szigliget; FT, Füzes-tó; Fuego; Guat (BVF), Guatemalan volcanoes behind the volcanic front; Mexico, the Trans-volcanic Belt; Stromb., Stromboli; Cas OIB, Cayuse Crater from the Oregon Cascades.

Figure 15: H₂O vs. CO₂ diagram illustrating the NanoSIMS data from peridotite hosted glass, and the reconstructed CO₂ contents of the melt+fluid system. For comparison, subducted sediments from [Plank and Langmuir \(1998\)](#), altered MORB from [Staudigel et al. \(1989\)](#), N-MORB from [Marty and Tolstikhin \(1998\)](#), E-MORB from [Hekinian et al. \(2000\)](#) are shown. Green area corresponds to the range of compositions of the subducting crust. Pink area corresponds to plausible compositions for magmas produced from incongruent melting of the subducted crust. Blue area corresponds to melts relatively depleted in volatiles issued from slab melting and modified by subsequent volatile depletion, such as the assimilation of volatile poor material and/or some physical fractionation during magma decompression.

Figure 1
[Click here to download high resolution image](#)

Figure 2
[Click here to download high resolution image](#)

Figure 3
[Click here to download high resolution image](#)

Figure 4
[Click here to download high resolution image](#)

Figure 5
[Click here to download high resolution image](#)

Figure 6
[Click here to download high resolution image](#)

Figure 7
[Click here to download high resolution image](#)

Figure 8
[Click here to download high resolution image](#)

Figure 9
[Click here to download high resolution image](#)

Figure 10
[Click here to download high resolution image](#)

Figure 11
[Click here to download high resolution image](#)

Figure 12
[Click here to download high resolution image](#)

Figure 13
[Click here to download high resolution image](#)

Figure 14

[Click here to download high resolution image](#)

SZB, Szentbékállá; SZG, Szigliget; FT, Füzes-Tó; Fg, Fuego; Guat (BVF), Guatemalan volcanoes behind the volcanic front, Mexico, the Trans-Mexican Volcanic Belt; Stromb., Stromboli; Cas OIB, Cayuse Crater from the Oregon Cascades

Figure 15
[Click here to download high resolution image](#)

Table 1

[Click here to download Table: Table 1.doc](#)

							Vol. (cm ³)	Glass volume (cm ³)			Fluid volume (cm ³)			Density (g/cm ³)			Mass (g)				
Sample	MP	SiO ₂	H ₂ O	Error (1σ)	CO ₂	Error (1σ)	Minerals	Best estimate	Min (%)	Max (%)	Best estimate	Min (%)	Max (%)	Glass	Vesicles	FI (Max)	Glass	CO ₂ (vesicle)	CO ₂ (dissolved)	CO ₂ /glass (wt. %)	CO ₂ /π (ppm)
SZB16	no	56.6	0.54	0.26	0.29	0.14	4.6E-03	9.1E-05	3.7E+01	1.3E+01	5.5E-05	1.5E+01	3.3E+01	2.599	1.022	0.573	2.4E-04	5.6E-05	6.8E-07	24.1	3723
SZB50	no	55.4	0.59	0.26	0.39	0.14	3.1E-02	6.4E-03	6.6E+00	6.4E+00	1.6E-03	1.3E+00	1.4E+01	2.740	1.295	0.617	1.8E-02	2.1E-03	6.8E-05	12.2	21287
SZB51-v	no	61.9	3.1	0.26	0.33	0.16	6.0E-03	2.1E-04	1.4E+01	2.6E-04	1.1E-04	3.2E+01	2.0E+01	2.496	0.970	1.024	5.3E-04	1.0E-04	2.1E-06	19.9	5409
SZB51-MP	yes	56.6	2.76	0.28	0.46	0.15															
SZB51-Av	yes	59.6	2.95	0.28	0.39	0.16															
SZB52	no	57.9	2.35	0.26	0.34	0.15	2.3E-02	7.8E-04	1.3E+01	1.0E-03	3.4E-04	5.8E+00	3.4E+01	2.452	0.953	1.114	1.9E-03	3.3E-04	6.5E-06	17.4	4422
SZG23	yes	56.3	2.38	0.26	0.96	0.20	3.5E-02	6.1E-03	2.3E+00	1.6E-03	2.0E-03	3.2E+00	1.3E+01	2.744	1.330	0.703	1.7E-02	2.6E-03	1.6E-04	16.5	24383
SZG44	yes	51.7	3.45	0.24	0.53	0.14	nd		0.0E+00	0.0E+00	1.3E-04	nd	nd	0.000	1.045	0.824	nd	1.4E-04	nd	nd	nd
FT12	yes	54.8	0.55	0.26	0.38	0.14	3.1E-02	7.8E-04	4.2E+00	9.7E-04	4.0E-04	1.0E+01	1.5E+01	2.542	1.228	0.661	2.0E-03	5.0E-04	7.5E-06	25.4	4946
FT01P	no	57.8	4.25	0.27	0.48	0.16	3.6E-02	1.4E-03	9.3E+00	1.7E-03	3.3E-04	4.9E+01	3.9E+01	2.611	0.930	1.057	3.6E-03	3.0E-04	1.7E-05	9.0	2727

Table 1: Summary of petrographic observations, chemical compositions and phase volumes, densities and masses. V – vein; MP – melt pocket; Av – average composition; Vol. – volume; Min & Max – Relative determined errors for each best estimate; vs - vesicle; FI – fluid inclusion; gl – glass; π - peridotite