Characterization of hydration in the mantle lithosphere: Peridotite xenoliths from the Ontong Java Plateau as an example

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We report concentrations of hydrogen (H) in upper mantle minerals of peridotites (olivine and pyroxenes) transported by alkaline lavas, which erupted on the southwestern border of the Ontong Java Plateau (Malaita, Solomon Islands, West Pacific). Unpolarized FTIR analyses show that olivine, orthopyroxene, and diopside contain 2–32 ppm, 162–362 ppm and 159–459 ppm wt H2O, respectively. In the studied lherzolites, garnets are anhydrous. The concentration of hydrogen within individual olivine and pyroxene grains is almost homogeneous, indicating no evidence of dehydration or hydration by ionic diffusion. In the lherzolite, the concentration of hydrogen in olivine tends to increase weakly with depth (based on geothermobarometry), consistent with the increase of water solubility with increasing water fugacity as a function of pressure, but concentrations remain well below water-saturation values determined experimentally. The highest concentration of H in olivine (32 ppm wt H2O) is, however, found in refractory spinel harzburgites, which equilibrated at depths of 85 km, while deeper specimens as the high-temperature spinel harzburgites, and some of the garnet lherzolites, contain less hydrogen in olivine. Olivines from pyroxene- or paragasite-rich peridotites have also lower hydrogen concentrations. We interpret the high hydrogen concentrations in olivine from the refractory spinel harzburgites as due to (1) simultaneous hydration and metasomatism of the lithospheric mantle by a water-rich silicate melt/fluid, during which hydrogen follows MREE and where spinel harzburgite have experienced ‘stealth’ metasomatism, and/or (2) to a late ‘fleeting’ hydrogen metasomatism, which would hydrate the rock after this first ‘stealth’ metasomatism event. In the second case, the composition of the ‘fleeting’ percolating fluid (small volume fraction of very evolved fluids, with high volatiles concentration and transient properties) is likely to be linked to the decrease of the plume activity and resulting in the downwelling of the lithospheric asthenosphere boundary. The difference in hydrogen concentration between harzburgites and lherzolites could be linked to variation in trace elements in olivine induced by the first ‘stealth’ metasomatism alone or associated to the late ‘fleeting’ hydrogen metasomatism, triggered by the lack of secondary crystallization of clinopyroxenes in the harzburgite.

To conclude, H concentrations in upper mantle minerals may thus be correlated to metasomatic events, with the later yielding variation of the mineral assemblage and trace element composition that constitute the peridotites.

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1. Introduction

Hydrogen, incorporated as atomic impurities in point defects within the atomic structure of nominally anhydrous minerals (NAMs), is known to affect many physical and chemical properties of mantle minerals and rocks. Indeed, a small amount of hydrogen as atomic impurities (essentially protons in cationic vacancies; e.g., Kohlstedt et al., 1996) increases electrical conductivity (Poe et al., 2010; Yoshino et al., 2006) and enhances ionic diffusion of major elements in mantle minerals (Costa and Chakraborty, 2008; Demouchy et al., 2007; Hier-Majumder et al., 2004). The incorporation of H was also proposed to weaken the strength of both olivine single crystals and aggregates (e.g., Demouchy et al., 2012; Mackwell et al., 1985). The presence of hydrogen in the upper mantle minerals is thus recurrently proposed as a major parameter of Earth dynamics (e.g., Albarède, 2009; Baptiste et al., 2012; Peslier et al., 2010; Regenauer-Lieb and Kohl, 2003; Regenauer-Lieb et al., 2001).

For more than a decade now, experimental studies have therefore tried to identify the mechanisms of hydrogen incorporation and to quantify the ‘water solubility’ in mantle minerals; where ‘water’ designates H incorporated in the mineral structure (as protons) and where its solubility is the maximum amount of hydrogen that a given mineral can incorporate at given thermodynamic conditions (e.g., Férot and Bolfan-Casanova, 2012; Gaetani et al., 2014; Keppler...
and Bolfan-Casanova, 2006; Kohlstedt et al., 1996; Mierdel et al., 2007). The concept of ‘water solubility’ can be extended to the one of ‘water storage capacity’, which takes into account the mineral assemblage, and that partial melting occurs when the water solubility limit of the given mineral assemblage (bulk) has been reached (Hirschmann et al., 2005). Many of these studies focused on the most abundant mineral of Earth’s upper mantle, olivine, and on its pure-end member forsterite. As a complement, proton mobility was characterized and quantified in olivine and forsterite (Demouchy and Mackwell, 2003, 2006; Mackwell and Kohlstedt, 1990; Padron-Navarta et al., 2014). These studies highlighted very fast diffusivities of hydrogen at high temperature with diffusion coefficients around $1.10^{-12} \text{ m}^2 \text{s}^{-1}$ at 1000 °C. However, an extensive data set on natural specimens is needed to confirm, rebut or adjust the current models of mantle hydration/dehydration, which are so far essentially based on experimentation at water-saturated conditions. Yet, the systematic characterization and quantification of hydrogen concentrations in upper mantle minerals from mantle peridotites of different geological settings have only begun recently.

Mantle xenoliths and large phenocrysts from kimberlites in cratonic settings have received most of the attention regarding their hydrogen concentrations (e.g., Baptiste et al., 2012; Bell and Rossman, 1992; Bell et al., 2004; Grant et al., 2007; Kamenetsky et al., 2008; Matveev and Stachel, 2007; Peslier et al., 2008, 2010, 2012). Hydrogen concentrations are also available for mantle-derived xenoliths embedded in alkali basalts and andesites from both intracontinental and subduction settings (Demouchy, 2004; Demouchy et al., 2006; Denis et al., 2013; Falus et al., 2008; Peslier and Luhr, 2006; Soustelle et al., 2010). However, the sub-oceanic mantle, far from any active plate boundaries, remains to be investigated. In this study, we measured by Fourier transform infrared (FTIR) spectroscopy the speciation and concentration of hydrogen in mantle minerals from a suite of peridotite xenoliths, which have sampled the mantle lithosphere beneath the southwestern part of the Ontong Java Plateau (OJP) from 60 km to 110 km depth. Hydrogen quantifications were compiled as a function of the equilibration depth, of the modal composition of the xenoliths, and of the minerals’ composition. Based on these results, we attempt to relate the observed variations in hydrogen concentrations to the petrological processes that might have affected the OJP mantle root.

2. Geological setting and samples selection

The alnöite-borne xenoliths from Malaita (Solomon Islands, West Pacific, Fig. 1) represent an exceptional and almost continuous sampling with increasing depth of the mantle beneath the south-western border of the OJP (Ishikawa et al., 2004; Neal, 1988; Nixon and Boyd, 1979). The samples selected for hydrogen quantification in this study are part of a larger collection previously studied by Ishikawa et al. (2004, 2005, 2011). They were selected so as to increase the likelihood of probing the entire depth interval (from 60 km down to 110 km of depth) and all lithological types that were sampled as a peridotite xenolith suite. With increasing depth, the different lithologies are as follows: (1) spinel lherzolites (SL), (2) garnet–spinel lherzolites (GSL), (3) spinel harzburgites (SH), (4) high-temperature spinel harzburgites (HTSH), and (5) garnet lherzolites (GL).

The shallow mantle (~85 km) at OJP is thus composed of variably metasomatized spinel lherzolites, garnet–spinel lherzolites and subordinately garnet pyroxenites (Ishikawa et al., 2004). The least metasomatized spinel lherzolites provide an isochron age (Sm-Nd) of melt depletion (~160 Ma), which is older than the plateau magmatism (~120 Ma), marking their origin as part of the Pacific Plate formed in a normal spreading center (Ishikawa et al., 2005).

A particular feature of the OJP mantle is the apparent absence of garnet in the 85 to 100 km depth interval, in which spinel harzburgites have predominantly Fo% ranging from 91 to 92 (i.e., Fo% = 100 × Mg/(Mg + Fe) apfu ratio), indicating the occurrence of an intra-lithospheric refractory layer (Ishikawa et al., 2004, 2005). Some of these spinel harzburgites also show osmium depletion (down to 30 ppt Os), Ti enrichment in clinopyroxene (up to 1.1 wt.% TiO2), and variable amount of amphiboles (Ishikawa et al., 2004, 2011). These geochemical features, presented in previous publications (Ishikawa et al., 2004, 2005), are usually symptomatic of interactions between silicate melt (OIB-like melt) and peridotites at high melt–rock ratio (e.g., Bedini and Bodinier, 1999; Bedini et al., 1997).

The deeper mantle (~100 km) is composed of garnet lherzolites, of high temperature spinel harzburgites, and of a wide variety of garnet pyroxenites. The pyroxenites are not studied here. The strong compositional variability of the Ontong Java mantle lithosphere below 85 km depth was interpreted as resulting from different degrees of partial melting of a mixed peridotite-pyroxenite source in a rising mantle plume (Ishikawa et al., 2004, 2007). Ishikawa et al. (2004) concluded that in all likelihood, the OJP mantle lithosphere is nowadays composed of a weakly genetically related two-layer structure. In the model of Ishikawa et al. (2004), the shallower layer corresponds to a normal oceanic mantle lithosphere partially metasomatized by small volume melts, while the petrology and geochemical features of the deeper layer reflect melt–rock reaction at higher ratio of melt and thus a stronger imprint from the plume, or accretion by the plume (Ishikawa et al., 2004, 2007, 2011).

Fig. 1. Map of Ontong Java Plateau and localisation of the Malaita Island (redrawn from Ishikawa et al., 2005). We also display the magnetic anomaly patterns (thin lines) and fracture zones (thick lines) from Nakanishi et al. (1992).
The pressure and temperature (P-T) of equilibrium of the considered xenoliths were determined and extensively discussed by Ishikawa et al. (2004). Briefly, we recall here the method. Equilibrium P-T estimates of garnet-bearing peridotites and pyroxenites were determined using Al-in-orthopyroxene geobarometer and the two-pyroxenes geothermometer of Brey and Köhler (1990), which yield typical errors of ±3 kbar (0.3 GPa) and ±20 °C, respectively. The equilibrium temperatures of garnet-free peridotites were determined using the two-pyroxenes geothermometer from Brey and Köhler (1990). For these garnet-free peridotites, the equilibrium pressure was derived from a realistic oceanic geotherm calculated by a linear least-squares regression for P-T data of garnet-bearing xenoliths [T (°C) = 311.67 P (kbar) + 252.29, see Ishikawa et al., 2004 for details]. Two groups of peridotites are then identified: the shallow low-temperature samples (SL, SH and GSL) and the deep high-temperature samples (GL and HTSH). The absence of spinel lherzolites equilibrated at similar temperature as the high-temperature spinel harzburgites suggest that the stability of spinel in these harzburgites is a compositional effect, corroborating their deep origin (Ishikawa et al., 2004; Robinson and Wood, 1998). Equilibrium P-T are reported in Table 1.

### 3. Methods

#### 3.1. Sample preparation

The quantification of hydrogen concentration in the nominally anhydrous minerals from the current batch of specimens was performed by unpolarized Fourier transmission infrared spectroscopy (FTIR). Mineral modes were determined by X-ray mapping using an electron microprobe analyzer (EMPA) (Ishikawa et al., 2004) or by electron backscatter diffraction (EBSD) mapping of the entire thin section. For FTIR, grains of olivines, pyroxenes, and garnet with no optically visible alteration or inclusions at macroscale were hand-picked from 24 peridotite xenoliths out of the 50 xenoliths previously studied by Ishikawa et al. (2004). The selection of samples and grains was dictated by the appeal to probe different lithologies and depths and the need for fresh and coarse (>500 μm) grains.

All the grains were double-polished by hand using diamond-lapping films with grid size ranging from 30 μm to 1 μm and a polishing jig equipped with a micrometer. Mineral grains were held in place by thermoplastic adhesive (crystal bond) on a steel plug of 1 in. in diameter. To ensure that the doubly polished slab corresponds to the middle of the initial grain and that core analyses will be accurately located, the grain was ground and polished only by a third of its thickness; then flipped on the other side, where the second third on the other side of the grain was ground and polished. Several successive baths of pure acetone cleaned the samples after polishing and prior to FTIR analysis. Final thickness is usually ~500 μm, but thicker samples were used when possible (Table S2). For mineral mode determination by EBSD, sample slabs were prepared as single polished thin sections (~30 μm of thickness). A final polishing with colloidal silica was done to remove the surface damaged by mechanical polishing to allow EBSD analysis without the need for a carbon coating.

Concentrations of titanium (Ti) in olivine were determined by laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) using the same thick doubly-polished grains of olivine (>500 μm) as for FTIR analyses; these measurements were performed after the FTIR analyses.

#### 3.2. Fourier transform infrared analysis

The hydrogen concentrations of separated polished grains have been qualified and quantified by FTIR spectroscopy in transmission. The infrared spectra were acquired using a Bruker Vertex 70 spectrometer, equipped with an MCT detector (Mercury Cadmium Telluride) cooled with liquid nitrogen, using a Globar light source and a KR/Ge beam splitter, coupled to a microscope (type Hyperion). Analyses were made at ambient temperature and pressure, under a stream of H₂O- and CO₂-free purified air flowing through the entire microscope. FTIR analyses were done in the mid-infrared frequency range (between 3000 and 600 cm⁻¹) using the MCT detector equipped with an MCT detector (Mercury Cadmium Telluride) cooled and quantified by FTIR spectroscopy in transmission. The infrared spectra were acquired using a Bruker Vertex 70 spectrometer, equipped with an MCT detector (Mercury Cadmium Telluride) cooled with liquid nitrogen, using a Globar light source and a KR/Ge beam splitter, coupled to a microscope (type Hyperion). Analyses were made at ambient temperature and pressure, under a stream of H₂O- and CO₂-free purified air flowing through the entire microscope. FTIR analyses were done in the mid-infrared frequency range (between 3000 and 600 cm⁻¹).
5000 and 700 cm\(^{-1}\) by accumulating 200 scans at a resolution of 4 cm\(^{-1}\), and with a squared beam spot of 200 or 100 \(\mu\)m. Analyses were performed for at least three different random orientations of each grain (by rotating the grain) and the spectra were averaged to compensate for any significant anisotropy (pleochroism) of the IR bands and its effect on the hydrogen concentration determination (average values per xenolith are reported in Table 1, details are given in the supplementary Tables S1 and S2).

To compare our results with previous studies, we used the calibration of Paterson (1982), which quantifies the OH concentration as:

\[
C_{\text{OH}} = \frac{Q_i}{1500} \int k(v) \frac{dv}{v},
\]

where \(C_{\text{OH}}\) is the concentration of hydroxyl, \(\xi\) is the orientation factor, which equals 1/3 for unpolarized measurements, and \(k(v)\) is the absorption coefficient for a given wavenumber \(v\). The density factor \(Q_i\) is chemistry dependent; it equals 2695 ppm wt H\(_2\)O for olivine that has a composition of Fo\(_{90}\), 2752 ppm wt H\(_2\)O for clinopyroxene (cpx), and 2812 ppm wt H\(_2\)O for orthopyroxene (opx). The method for calculation of \(Q_i\) can be found in Bolfan-Casanova et al. (2000). The concentration of hydrogen was calculated by integration of the spectrum between 3610 and 3150 cm\(^{-1}\) for olivine; 3770 and 3000 cm\(^{-1}\) for cpx; and between 3670 and 2800 cm\(^{-1}\) for opx. After background correction, each FTIR spectrum was baseline corrected and normalized to 1 cm of thickness. The estimated error on the resulting water content is 30% (± 15%); this error is due essentially to a combination of the empirical calibration equation, error on the sample thickness, the background spectrum, and the baseline subtraction. Olivine and several pyroxenes were analyzed, but analyses of both types of pyroxenes for each peridotite could not be achieved. The analyses of amphibole grains could not be performed since their FTIR measurements are extremely challenging as detailed by Skogby and Rossman (1991) mostly due to the fragility of this silicate, the strong anisotropy of its crystalline structure and its relative small size in mantle-derived peridotite (20–100 \(\mu\)m).

The calibration from Paterson (1982) for unpolarized IR light on non-oriented crystals used here is likely to underestimate the absolute hydrogen concentration in olivine, whereas the calibration from Bell et al. (2003) using polarized IR light on oriented crystals is likely to overestimate it (Wither et al., 2012). In the present study, we chose the calibration of Paterson (1982) since it allows a larger number of analyses on un-oriented samples and because the use of this calibration facilitates comparison between the concentrations from different mineral phases (same calculation method for all phases) and with results from previous studies. As indicated in Férot and Bolfan-Casanova (2012) and Bolfan-Casanova et al. (2014), a factor of 3 can be used with confidence to convert the hydrogen concentrations in olivine reported here by unpolarized IR method to values equivalent to those obtained with the polarized IR method of Bell et al. (2003); and a factor between 1.8 and 2 can be used to convert them to values equivalent to those obtained with the recent calibration by Wither et al. (2012).

The homogeneity of hydrogen concentration within olivine crystals has been tested by unpolarized FTIR analyses along profiles (over 20 points, along transect up to 3.5 mm), across two large olivine grains, which have their approximated spherical center at the center of the doubly polished slab (which is the case for every polished grains in this study).

3.3. Mineral compositions and modes

Modal and mineral compositions for most studied specimens are published in Ishikawa et al. (2004); a compilation is provided in Table S1. Minerals modes for these specimens were obtained by processing of automatic X-ray concentration images on electron microprobe for each thin section with the following assumptions: (i) late veins filled with carbonate were not included in the calculation, (ii) carbonates and serpentines replacing olivine were calculated as olivine, (iii) kelyphititized margins of garnet were calculated as garnet.

Mineral modes for the few samples which were not analyzed previously were obtained from electron backscattered diffraction (EBSD) mapping of the entire thin sections in a JEOL JSM 5600 scanning electron microscope using an acceleration voltage of 17 kV and a working distance of 23 mm at the SEM-EBSD facility in Geosciences Montpellier. The thin sections were tilted at 70° to the horizontal. Depending on the grain size, maps, which covered almost entirely each thin section, were performed using sampling steps of 50 or 100 \(\mu\)m. Indexation rates range from 50 to 85% depending on the fracturing and alteration state (i.e., reaction with the alnoite magma) of the xenoliths. This method allows determining not only the modal compositions (with an accuracy of ±2%), but also the grain sizes and crystal-preferred orientations of all mineral phases; the microstructural features are the focus of another study (Tomasni and Ishikawa, 2014). The same assumptions as for X-ray data were used in the calculation of the modes from EBSD data.

3.4. Laser ablation ICP-MS

Ti content in selected olivine grains was obtained by laser ablation inductively coupled mass spectrometry (LA-ICP-MS) at Géosciences Montpellier (AETE platform). The LA-ICP-MS system is composed of a Geolas 193 excimer laser coupled with an XR-element 2 ICP-MS. Ablation took place in He, with a beam size of 80 \(\mu\)m in diameter, the laser repetition rate was of 10 Hz, and an energy density of ca. 12 J/cm\(^2\) was used. The results were normalized to NIST 612 glass standard and internally to SiO\(_2\) concentrations obtained by electron microprobe and processed with the GLITTER software package. The standard deviation does not exceed 3 at. ppm for Ti (see Table S3) and the detection limit is 0.1 at. ppm Ti.

4. Results

4.1. FTIR spectra

Over 124 grains from 24 samples were analyzed by FTIR, providing estimates of hydrogen distribution and concentration in nominally anhydrous minerals in all five groups of Malaita peridotite xenoliths (Table 1, Table S2) reported by Ishikawa et al. (2004). In some xenoliths, in particular the deepest specimens, despite optical clearness, many crystals were altered, and occurrence of hydrous minerals (e.g., serpentine and amphibole) along fractures or cleavages prevented accurate quantification of hydrogen concentrations. Examples of contamination by lamellae of hydrous minerals are shown in Fig. S1, where the strong absorption band at 3686 cm\(^{-1}\) corresponds to structural OH groups in serpentines (Baptiste et al., 2012; Miller et al., 1987; Post and Borer, 2000). This problem was not encountered in the xenoliths equilibrated at shallower depths and hydrogen quantification in olivine was achieved for 21 samples (Table 1). Nevertheless, the analysis of pyroxenes remains challenging due to their smaller size and frequent alteration along cleavages; hydrogen quantification in pyroxenes was only successful for 8 samples (Table 1).

The unpolarized absorption spectra of Malaita olivine are shown in Fig. 2. There are three groups of absorption bands: group 1 from 3598 to 3420 cm\(^{-1}\), group 2 from 3420 to 3280 cm\(^{-1}\), and group 3 from 3280 to 3195 cm\(^{-1}\). The band at 3572 cm\(^{-1}\) of group 1 has, in the vast majority of olivine grains, the highest absorbance, except for SAS21 where group 2 bands are dominant. Group 1 is always expressed, SAS21 where group 2 bands are dominant. Group 1 is always expressed, whereas the calibration from Paterson (1982) for unpolarized IR light on oriented crystals is likely to overestimate it (Wither et al., 2012). In the present study, we chose the calibration of Paterson (1982) since it allows a larger number of analyses on un-oriented samples and because the use of this calibration facilitates comparison between the concentrations from different mineral phases (same calculation method for all phases) and with results from previous studies. As indicated in Férot and Bolfan-Casanova (2012) and Bolfan-Casanova et al. (2014), a factor of 3 can be used with confidence to convert the hydrogen concentrations in olivine reported here by unpolarized IR method to values equivalent to those obtained with the polarized IR method of Bell et al. (2003); and a factor between 1.8 and 2 can be used to convert them to values equivalent to those obtained with the recent calibration by Wither et al. (2012).

The homogeneity of hydrogen concentration within olivine crystals has been tested by unpolarized FTIR analyses along profiles (over 20 points, along transect up to 3.5 mm), across two large olivine grains, which have their approximated spherical center at the center of the doubly polished slab (which is the case for every polished grains in this study).
The spectra of Malaita orthopyroxenes display five major absorption bands at 3592, 3523, 3416, 3318, and 3066 cm\(^{-1}\) (Fig. 3A), whereas clinopyroxenes show three dominant absorption bands at 3634, 3540, and 3460 cm\(^{-1}\) as illustrated in Fig. 3B. Spinel, due to their opacity, could not be analyzed. Garnets, which have an exceptional clarity and large grain size (>3 mm diameter, thickness >2.5 mm), do not show hydroxyl absorption bands (i.e., detection limit of 0.5 ppm wt H\(_2\)O). They are hence considered anhydrous (Fig. S1).

The infrared spectra are typical of mantle-derived olivine, orthopyroxenes and clinopyroxenes (Bell and Rossman, 1992; Demouchy, 2004; Demouchy et al., 2006; Denis et al., 2013; Miller et al., 1987; Peslier and Luhr, 2006; Peslier et al., 2002; Soustelle et al., 2010, 2013). In contrast, the absence of hydrogen in garnet from samples equilibrated at depths greater than 80 km is unusual (Bell et al., 1995; Peslier, 2010; Peslier et al., 2010, 2012; Withers et al., 1998).

4.2. Hydrogen distribution

The distribution of hydrogen concentration in olivine has been tested by FTIR analyses profiles across two very large olivine grains from garnet-spinel lherzolite SAG4 (two profiles perpendicular to each other) and garnet herzolite SAG30 (one profile), respectively. In both samples, olivine borders do not show extensive depletion in hydrogen (Fig. 4A and B), but only minor erratic decrease at the very edge of the grains. This can be an edge or crack effect (reflection of the IR beam) as also reported by Denis et al. (2013). Further, when hydrogen concentrations are plotted as a function of the longest grain length (reported in Table S2, and plotted in Fig. S2), no positive correlation is seen (Fig. S2), in contrast to data reported from samples showing clear dehydration of olivine due to the fast ascent of the xenoliths (Denis et al., 2013). Therefore, we conclude that the present samples did not experience major dehydration or hydration (i.e., variation by more than 60% between rim and core hydrogen concentration) during ascent toward the surface within the alnöite lavas.

The distribution of hydrogen concentration in pyroxenes has been tested by analyses at both core and rim of grains, when it was possible (Fig. 4). No variation in IR peaks intensity was observed. Therefore, profile analyses were not performed.

4.3. Hydrogen concentration in NAMs

Hydrogen concentrations, expressed in ppm by weight of H\(_2\)O and quantified using the calibration of Paterson (1982), range between 2 and 32 ppm wt H\(_2\)O in olivine, 162 and 362 ppm wt H\(_2\)O in opx, and
159 and 459 ppm wt H₂O in cpx (Table 1). Conversion to atomic ratio H/10⁶ Si (= ppm H/Si), which is commonly used in mineral physics, is achieved by multiplying the concentration in ppm wt H₂O by a factor of 16.35 for olivine (Fo₉₀), of 11.49 for orthopyroxenes, and of 11.61 for clinopyroxenes.

The average hydrogen concentration in olivines from lherzolites slightly increases with increasing depth and temperature (Fig. 5A), in agreement with the increase in water solubility with increasing pressure (Férot and Bolfan-Casanova, 2012; Kohlstedt et al., 1996). The measured values in lherzolites are in good agreement with previously reported hydrogen concentrations from mantle-derived xenoliths (Baptiste et al., 2012; 2014; Demouchy et al., 2006; Denis et al., 2013; Falus et al., 2008; Peslier et al., 2010; Soustelle et al., 2010). These values remain, however, significantly lower than the most recent experimentally determined water solubility law of olivine (Férot and Bolfan-Casanova, 2012), which has been calculated for a hot geotherm (i.e., spreading ridge-like geotherm, Katsura et al., 2004) and is represented in Fig. 5A. Since water solubility is likely to increase with decreasing temperature (Bali et al., 2008, for T > 700 °C), the solubility curve in Fig. 5A defines minimum water solubility values for each depth.

Among harzburgites within the garnet depleted layer, the single deep high-temperature spinel harzburgites have olivine with low hydrogen concentrations (4–6 ppm wt H₂O), in agreement with shallower spinel lherzolite (Fig. 5A). In contrast, the deep spinel harzburgites have olivines with highly variable hydrogen concentrations (Table 1), which can reach 32 ppm wt H₂O. Notably, two spinel harzburgite samples (SAS20, and SAS58) show relatively higher hydrogen concentrations in olivine (i.e., between 10 to 32 ppm wt H₂O) for their equilibrium depth and mineralogical assemblage. These concentrations are markedly higher than those from olivines in spinel lherzolites equilibrated at similar depths (Fig. 5A).

The scarcity and small size (< 100 μm) of pyroxenes in numerous samples, in particular the harzburgites, and common alteration along cleavages did not allow for high quality FTIR analyses on many grains. We could not therefore establish the variation of hydrogen concentration in the pyroxenes as a function of depth. However, hydrogen

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**Fig. 4.** Hydrogen concentrations (H/10⁶ Si and ppm wt. H₂O) in Malaita olivine as a function of the distance along the grain of olivine (in mm) from unpolarized FTIR analyses; (A) across sample SAG30, and (B) across sample SAG4. The circles represent the hydrogen concentration of the each absorption spectrum. The error on the hydrogen concentration is about 30%, the error bars are not shown for clarity.
concentrations in pyroxenes are, as usual, higher than in olivine with diopside containing more hydrogen than orthopyroxene (e.g., Bell et al., 1995). Both pyroxenes have concentrations (Fig. 5B) within the normal concentration range for mantle peridotite (e.g., Ingrin and Skogby, 2000; Peslier, 2010; Skogby, 2006).

The studied xenoliths have olivines with Fo% varying between 89.5 and 92.3; the highest Fo% values are observed in the harzburgites, but not all harzburgites have Mg-rich olivines. Yet, the highest hydrogen concentrations are observed in refractory olivines (Fo% = 92) as illustrated in Fig. 6.

Mineral modes for the studied samples are reported in Table 1. Positive correlations are not observed between hydrogen concentration in olivine and modal content of pyroxene or amphibole (i.e., pargasite, for which the precise water contents have not been determined) as shown in Fig. 7. In contrast, a slight increase in hydrogen concentration in olivine is observed with increasing olivine content and with decreasing orthopyroxene + clinopyroxene modal content. Spinel harzburgites are notably very poor in clinopyroxene ($b_0.5\%$).

When possible (only for four samples), the bulk hydrogen concentrations were calculated. They range from 84 to 96 ppm wt H$_2$O (Table 1). When hydrogen concentration was not available for one or two mineral phases, we used partition coefficients from Aubaud et al. (2004) to calculate the bulk concentration (Table S1). We wish to emphasize here that partition coefficients of H between olivine and pyroxenes must be used with extreme caution, since hydrogen partition coefficient is very sensitive to pressure, temperature, and composition of the system (Keppler and Bolfan-Casanova, 2006). No partition coefficients are available for the exact pressure, temperature range and composition of the studied mantle specimens (i.e., transport by alnoite, and for pressure below 4 GPa). The partition coefficients from Aubaud et al. (2004), used in the present calculation, are the best approximation to the equilibrium conditions to the studied samples. These coefficients of $D_{\text{ol(opx)}} = 0.11$ and $D_{\text{ol(cpx)}} = 0.08$, $D_{\text{cpx(opx)}} = 1.4$ are based on secondary ion mass spectrometry (SIMS) measurements on experimental charges at 1–2 GPa, but at high temperature (1230–1380 °C), and wherein crystalline silicates co-exist with a basaltic melt (with some carbonates in the system). Other experimental studies providing empirical equations to estimate ol-opx-cpx partition coefficients are based

Fig. 5. (A) Hydrogen concentrations in olivine (in both H/10$^6$ Si and ppm wt. H$_2$O) as a function of depth and pressure (in km and GPa). Small dots are from previous studies: small black dots from Demouchy (2004), Demouchy et al. (2006) and Peslier et al. (2010); small gray dots from Peslier et al. (2008), small white dots from Baptiste et al. (2012), narrow diamond from Demouchy et al. (2006). Diamond symbols for megacrysts from Bell et al. (2004). The recent hydrogen solubility curve for olivine from Férot and Bolfan-Casanova (2012) for a hot oceanic geotherm, and corrected for the calibration of Paterson, is also reported as a dotted line. (B) Hydrogen concentration (in both H/10$^6$ Si and ppm wt. H$_2$O) in pyroxenes as a function of depth and pressure (in km and GPa).

Fig. 6. Hydrogen concentrations (in both H/10$^6$ Si and ppm wt. H$_2$O) in Malaita olivine as a function of Fo% (i.e., Fo% = Mg/(Mg + Fe) * 100). Small black and white dots are from previous studies (Baptiste et al., 2012; Denis et al., 2013; Peslier et al., 2010).
Hydrogen incorporation as atomic impurities in the structure of a NAM is strongly a function of pressure (Kohlstedt et al., 1996), temperature (Férot and Bolfan-Casanova, 2012), but also of the other atomic impurities present in the crystalline structure (Keppler and Smyth, 2006). Experimental studies (Berry et al., 2005, 2007; Padron-Navarta et al., 2014), ab initio calculations (Walker et al., 2007), and limited data from mantle specimens (Peslier et al., 2010; Schmädicke et al., 2013) have suggested that Ti enhances H incorporation in both olivine and iron-free forsterite. However, our samples do not show a well-defined positive correlation between hydrogen concentrations and Ti content in olivine. This is true whether the Ti content in olivine is derived from EMPA data using experimentally determined olivine/cpx partition coefficient (McDade et al., 2003) or from direct measurements by laser ablation-ICP-MS (Fig. 8A). One sample (garnet-lherzolite SAG30-3, Table S2) shows an anomalously high concentration of Ti and in other trace elements in olivine (Table S3). The Ti content in this olivine (~200 ppm) is significantly higher than typical Ti concentrations in mantle olivine (<75 ppm; cf. De Hoog et al., 2010, 2014; Foley et al., 2013), but its hydrogen concentration (20 ± 4 ppm wt. \( H_2O \)) is only slightly higher than those from spinel harzburgites and garnet-spinel lherzolites with much lower Ti contents (Fig. 8A). The lack of correlation between Ti and hydrogen concentrations in olivine observed in the Malaita peridotite xenoliths corroborates the recent experimental results from Gaetani et al. (2014). Together, these data suggest that H incorporation in olivine only depends on the Ti concentration for very high Ti concentrations as reported by De Hoog et al. (2014).

4.4. Hydrogen incorporation: effect of trace elements in olivine

Hydrogen incorporation as atomic impurities in the structure of a NAM is strongly a function of pressure (e.g. Ardia et al., 2012) or for minerals co-existing with large amounts of melt (e.g., SIMS analyses of various experimental studies, see e.g., Hauri et al., 2006) or for model systems (e.g., iron-free and/or alumina-free, Grant et al., 2006; O’Leary et al., 2010).
Hydrogen concentration (H/10⁶ Si) in Malaita olivine as a function of Ti in at. ppm, derived from TiO₂ wt.% content in cpx combined with Ti partition coefficient olivine/clinopyroxene from McDade et al. (2007) and from measurements by laser ablation ICP-MS for four selected olivines (spinel harzburgites) of this study. (B) Hydrogen concentration (in both H/10⁶ Si and ppm wt. H₂O) in Malaita olivine as a function of (Sm/Yb)N in co-existing clinopyroxene. Dashed line represents values for primitive mantle. (C) Hydrogen concentration (H/10⁶ Si and ppm wt. H₂O) in Malaita olivine as a function of Sm/Yb (normalized to Chondrite) in the co-existing clinopyroxene for all peridotites. Dashed line represents values for primitive mantle.

Fig. 8. (A) Hydrogen concentration (in both H/10⁶ Si and ppm wt. H₂O) in Malaita olivine as a function of Ti in at. ppm, derived from TiO₂ wt.% content in cpx combined with Ti partition coefficient olivine/clinopyroxene from McDade et al. (2007) and from measurements by laser ablation ICP-MS for four selected olivines (spinel harzburgites) of this study. (B) Hydrogen concentration (in both H/10⁶ Si and ppm wt. H₂O) in Malaita olivine as a function of Sm in at. ppm, in the co-existing clinopyroxene for the spinel-bearing peridotites. Dashed line represents values for primitive mantle. (C) Hydrogen concentration (in both H/10⁶ Si and ppm wt. H₂O) in Malaita olivine as a function of Sm/Yb (normalized to Chondrite) in the co-existing clinopyroxene for all peridotites. Dashed line represents values for primitive mantle.

5. Discussion

The primary outcome from the present study is that the lherzolites and harzburgites xenoliths from Malaita, which sampled the Ontong Java Plateau, an oceanic mantle lithosphere affected by extensive partial melting and by important volumes of percolating melt in response to the impact of the Ontong Java plume (Ishikawa et al., 2004), are not completely dehydrated. Here, we first discuss the preservation of hydrogen concentration in the studied olivine grains during the xenoliths ascent towards the surface. Then, we propose two hypotheses explaining the relative abundance of hydrogen in olivine in the lherzolites and harzburgites.

5.1. Hydrogen preservation

Hydrogen concentration in olivine is almost homogeneous within crystals (Fig. 4); decrease in the hydrogen concentration is limited to a few erratic points within 150 μm from the crystal edges and does not have (Sm/Yb)N ratios ranging >1 from 1.5 to 8 (Fig. 8C). Garnet-bearing peridotites (GSL and GL) have (Sm/Yb)N ratios ranging from 0.6 to 11 (Fig. 8C) and High hydrogen contents (>10 ppm wt%) in olivine from harzburgites are systematically associated with REE enrichment ([Sm/Yb]N >1, SH and HTSH in Fig. 8C). These relatively high hydrogen concentrations in olivine are, however, not accompanied by concomitant enrichment in light rare earth elements (LREE) in clinopyroxenes, but are positively correlated to the medium REE (MREE) contents as illustrated in Figs. 8B-C and 9. This correlation between hydrogen incorporation in olivine and MREE enrichment in co-existing clinopyroxene is also illustrated by the positive correlation between the hydrogen content in olivine of the harzburgites and the Sm concentration (Fig. 8B) and the Sm/Yb ratio (Fig. 8C) of the co-existing clinopyroxene. An exception is the high-temperature (very opx rich) spinel harzburgite, which shows a high Sm/Yb ratio in clinopyroxene, but low hydrogen concentration in olivine. The relation between hydrogen concentration in olivine and REE composition of clinopyroxene is also less clear for the garnet-bearing lherzolites (GSL and GL), where the REE composition of clinopyroxene is perturbed by the presence of garnet (yielding (Sm/Yb)N >1, see Harte and Krirkley, 1997) and where hydrogen incorporation is enhanced by increasing pressure (i.e., water solubility law, Férrot and Bolfan-Casanova, 2012).
match the characteristic bell-shape profiles as expected for significant H-loss by ionic diffusion (see for instance Demouchy et al., 2006; Denis et al., 2013; Thoraval and Demouchy, 2014). Thus, we conclude that extensive dehydration did not occur during the xenoliths extraction and transport by the alnöite magma and assume no or only very minor hydrogen disequilibrium. We interpret the preservation of the relative high hydrogen concentration in olivine as a consequence of a lack of concentration gradient between the alnöitic magmas, which have potentially high volatile contents, and the mantle xenoliths (Nixon et al., 1980). Indeed, xenolithic olivines carried by other volatile-rich magmas such as kimberlites also show no dehydration profiles (Baptiste et al., 2012; Kamenetsky et al., 2008; Peslier et al., 2008). In the present study, hydrogen concentrations in olivine are also rather homogeneous at the xenolith scale: crystals from the same xenolith have similar hydrogen concentrations regardless of their position in the enclave (see also Demouchy et al., 2006; Denis et al., 2013). This observation corroborates the lack of extensive ‘catch-in-the-act’ dehydration or re-hydration during the xenoliths transport by the alnöite magma. We therefore consider that the measured hydrogen concentrations are representative of mantle processes at depth.

5.2. Hydrogen concentration in olivine vs. lithology

In the lherzolites, the measured hydrogen concentrations in olivine agree with the range of concentrations reported by previous studies in mantle peridotites as illustrated in Fig. 5. The general trend for the lherzolites shows an increase in hydrogen concentration in olivine with increasing equilibrium depth. This trend can be explained by the increase of water fugacity (even below water saturation level) with increasing pressure as reported in experiments performed at high pressures and temperatures (e.g., Bali et al., 2008; Féraud and Bolfan-Casanova, 2012; Kohlstedt et al., 1996).

The harzburgites (SH and HTSH) do not follow this trend (Fig. 5A). The deepest, high-temperature harzburgites (HTSH) have olivines with low hydrogen concentrations, whereas the spinel harzburgites (SH) equilibrated at intermediate depths have olivines with similar or higher hydrogen concentration (>10 ppm wt H2O) than the deepest garnet lherzolites (Fig. 5A and Table 1). Furthermore, among these spinel harzburgites, the most refractory specimens (SAS20 and SAS59) contain the highest amount of hydrogen in olivine (Figs. 6 and 8B-C). These results are in contradiction with the conventional wisdom derived from a study on mantle melting (Dixon et al., 2002), which infers that hydrogen has an incompatibility level between lanthanum (La) and cerium (Ce) and, hence, that refractory rocks should be very dehydrated.

To account for these observations, we propose two hypotheses, which are detailed in the subsequent paragraphs. In hypothesis (1), hydrogen incorporation in olivine results from a metasomatic event during which hydrogen incorporation in olivine follows enrichment in MREE in olivine and cpx. In hypothesis (2), hydrogen incorporation in olivine occurs during a late “fleeting” hydrogen metasomatism by a very hydrous fluid. Both episodes of metasomatism, are associated with the Ontong Java plume. The change in the composition of the metasomatic agents through time (and space) would result from the decrease of plume activity, which lead to progressive downwelling of the lithosphere-asthenosphere boundary (see Ishikawa et al., 2004) and the evolution of the associated fluids (from melt, to hydrous melts, to hydrous vapors) as illustrated in Fig. 10. The late “fleeting” hydrogen metasomatism would hence be delayed in time compared to the first metasomatism affecting REE concentration in all peridotites. In both cases, the final hydration level in olivine would be function of the composition of the metasomatic agent, the point defect population of olivine (i.e., including trace element composition) and abundance of the NAMs co-existing with olivine in the hydrated peridotites. We detail below the two hypotheses and their type of metasomatism.

The mantle beneath the OJP has experienced a high degree of melting and melt extraction during the plume activity, leading to a compositional layering as a function of depth. The highly refractory, garnet-free interval between 85 and 100 km depth probably represents the topmost domain of plume melting (Ishikawa et al., 2004). In hypothesis

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**Fig. 10.** Schematic model of the evolution of the upper mantle lithosphere beneath the Ontong Java Plateau (modified from Ishikawa et al., 2004) and the overlaying heterogeneous hydrogen incorporation (blue smearing). Column C and D correspond to the two hypotheses of genesis of high hydrogen concentration in olivine (see text for details).
(1) as the plume activity starts to subside, the peridotites have been metasomatized by water-rich and silicate-bearing fluids/melts produced by continued melting in the plume tail. Variations in the melt composition, in particular in the volatiles content may be explained by an association of chromatographic effects and melt–rock reactions as these melts percolate the mantle lithosphere (e.g., Bedini et al., 1997). This metasomatic event is supported by the REE patterns of the studied peridotites, which show variable enrichment in LREE and MREE (Fig. 9). However, the lack of positive correlation between the hydrogen concentration in olivine and the modal abundance of amphibole and clinopyroxene (Fig. 7), excludes that hydration of olivine might result from a typical modal metasomatism (O’Reilly and Griffin, 2012). The REE patterns and low Sm/Yb (0.1 < Sm/Ybnorm < 1) of spinel lherzolites are consistent with a light cryptic metasomatism. In contrast, the high Sm/Yb (Sm/Ybnorm > 1) of the harzburgites cannot be explained by cryptic metasomatism (Fig. 8C). In harzburgites, the MREE concentrations are similar or higher than the LREE ones, which is consistent with a ‘stealth’ metasomatism as defined by O’Reilly and Griffin (2012). The ‘stealth’ metasomatism implies interactions with larger volumes of melts/liquids and crystallization of secondary minerals (ol, opx and cpx) identical in major element composition to the primary assemblage, but having higher concentrations in MREE than LREE (O’Reilly and Griffin, 2012). The observation that the high hydrogen concentrations in olivine from spinel-bearing harzburgites correlate with the MREE enrichment instead of LREE (Figs. 8B–C and 9) indicates that hydrogen would have a different behavior during ‘stealth’ metasomatism than during cryptic metasomatism (Fig. 8C) or mantle melting. This correlation between high hydrogen concentration in olivine and high Sm/Yb ratio in the harzburgites (Sm/Ybnorm > 1, Fig. 8C) is in fact puzzling, since high Sm/Yb ratio implies interaction with large volumes of melt, which is not supposed to favor hydrogen incorporation, but to the contrary to efficiently dehydrate NAMs. Hydration may, however, occur if the fluid/melt responsible for ‘stealth’ metasomatism is water-saturated. At greater depth, the relatively high concentrations of hydrogen in olivine from garnet-bearing peridotites would be primarily controlled by the increase in water solubility and fugacity with pressure, being less sensitive to the nature of the metasomatism. An alternative explanation is that the water-saturation of the metasomatic fluids increases as they percolate the lithospheric column, due to interactions with the peridotites and partial crystallization. Finally, the variations in hydrogen content in olivine among the same type of spinel- or garnet-bearing peridotites may result from a spatially heterogeneous intensity of this simultaneous hydration and metasomatism (Figs. 8B–C and 9).

The second hypothesis (2) is sustained by the extremely fast diffusion of H in olivine at high temperature (10⁻¹² m² s⁻¹ at 1000 °C, Demouchy, 2010; Demouchy and Mackwell, 2003, 2006; Mackwell and Kohlstedt, 1990; Padron-Navarta et al., 2014). Indeed, diffusion of hydrogen in olivine at high temperature is several orders of magnitude faster than REE (i.e., see Cherniak, 2010; Demouchy and Mackwell, 2006; Spandler and O’Neill, 2010; Qian et al., 2010). Thus, dehydration or rehydration of olivine may occur during a late rapid ‘fleeting’ hydrogen metasomatism (i.e., very small fluid volume, very rich in water, potentially water super-saturated), where REE do not have time to re-equilibrate in the minerals. In this scenario, the NAMs (olivine and pyroxenes) of the anhydrous ‘stealth’ metasomatized harzburgites may have REE compositions and then point defects populations (in olivine and opx) different than the NAMs in the lherzolites, as indicated by the REE compositions in the clinopyroxenes (Figs. 8B–C and 9). The higher concentration in REE of NAMs in the stealth metasomatized harzburgites would enhance the amount of point defects that might potentially store hydrogen (i.e., of the type [REE] = ([OH]⁺[V⁴⁺]) [H₂O]), allowing for a more effective H incorporation in the harzburgitic olivine during the late ‘fleeting’ hydrogen metasomatism. In other words, the H partition coefficient between olivine (D_H²O[ol]) and the late ‘fleeting’ hydrogen metasomatism agent would be function of the REE composition and amount of the co-existing NAMs (D_H²O[ol] < D_H²O[NAM]) in the peridotite. It will therefore depend on its petrological history.

Up to now, point defect theory of NAMs in general and in olivine, in particular (e.g. Nakamura and Schmalzlried, 1983), has reported that hydrogen incorporation tends to increase with increasing number of chemical components present in a NAM. This theoretical approach was confirmed by a relatively large number of experimental studies in olivine and pyroxenes (e.g., Berry et al., 2007; Férot and Bolfancasanova, 2012; Gaetani et al., 2014; Mierdel et al., 2007; Stalder, 2004; Zhao et al., 2004). Variations of H partition coefficients as a function of pressure, temperature, and compositions of the co-existing phases, such as the one proposed here, have been suggested previously (see Fig. 3 in Hirschmann et al., 2005). However, no theory or experimental data is available for predicting the relative hydrogen incorporation in three- to four-solid phases assemblages (cpx-free harzburgites versus lherzolites), such for real mantle assemblies, as those that compose the mantle lithosphere below Malaita Island. Experimental studies have so far focused on simplified 1- or 2-phase systems at significantly higher pressure (e.g., 8 GPa, Ardila et al., 2012) or with co-existing silicate-rich melt (e.g., Aubaud et al., 2004; Hauri et al., 2006).

The range of hydrogen concentrations in olivine in the spinel harzburgite group might be explained by local small variation of the H partition coefficients, as a function of pressure, temperature, and chemical composition of the metasomatic fluid, which is in turn affected by the composition of the percolated peridotites. Unfortunately, the lack of a complete data set of hydrogen concentrations for co-existing orthopyroxene does not permit to test the second hypothesis by comparing the hydrogen concentrations in pyroxenes of the spinel harzburgites with the ones of the lherzolites (Figs. 5B and 7).

Whatever the hypothesis considered for the genesis of the hydrated olivines in the harzburgites, we must also notice that the absence of correlation of H concentration with the amphibole modal content (Fig. 7) is in contradiction with the general assumption that the crystallization of hydrous minerals is associated with water-saturation of the NAMs. Our data corroborates similar recent observations reported by Schmädicke et al. (2013).

At last, the lack of hydrogen in garnet from the deep lherzolites is not in agreement with partition coefficients determined by experimental mineralogy and petrology using FTIR or SIMS (D_H²O[ol] – 0.5–2, e.g. Hirschmann et al., 2009). Their anhydrous state could be explained by their secondary character (i.e., metamorphosed from anhydrous spinel, Ishikawa et al., 2004) or also by the slower decrease of the partial molar volume of water in hydro-garnet compared to the rapid decrease of the water molar volume in the fluid as previously reported experimentally for pure pyrope between 5 and 7 GPa (Withers et al., 1998). For mantle garnets (not pure pyrope), the exact pressure of dehydration for mantle garnets (not pure pyrope), the exact pressure of dehydration at high temperature (> 1000 °C) remains uncertain (e.g., Lu and Keppler, 1997; Mookherjee and Karato, 2010).

In summary, the OJP mantle root is presently characterized by a heterogeneous hydrogen concentration in olivine, with a maximum hydrogen concentration in the refractory layer below 85 and 100 km depths. This layer marks the transition between the old oceanic mantle lithosphere modified by the plume activity and the deep part of the OJP mantle, essentially composed by frozen plume-derived material (Ishikawa et al., 2004). We propose that the lithological layering created by the plume activity controlled the subsequent re-hydration of the OJP mantle lithosphere by metasomatic agent, which percolated the cooling lithosphere as the plume activity decreased, as illustrated in Fig. 10. The high water contents in olivine in this layer may hence result from either hydration due to an initial metasomatism by a water-rich silicate melt/fluid (scenario 1) and/or by a subsequent late ‘fleeting’ hydrogen metasomatism by small volumes fraction of evolved water-supersaturated fluids (scenario 2). In the second scenario, the hydration of olivine is influenced by the changes of point defect population in the co-existing
NAMs and also by their modal compositions produced by the preceding metasomatic events.

6. Conclusion

Hydrogen concentration in olivine, orthopyroxene, clinopyroxene, and garnet from Malaita xenoliths were measured with unpolarized FTIR. Hydrogen concentrations for olivine, orthopyroxene, and clinopyroxene range around 2–32 ppm, 150–270 ppm, and 150–450 ppm wt H2O, respectively, but garnets are anhydrous. Measurements show that even refractions mantle peridotites can contain hydrogen and suggest that the hydration of the OJP mantle lithosphere is controlled by one or multiple metasomatic events associated to late stages of the plume activity and subsequent cooling and thickening of the lithosphere.

One of the lithologies that characterizes the mantle root below the Ontong Java Plateau, the spinel harzburgite, which equilibrated between 85 and 100 km depth, have olivines that are relatively rich in hydrogen, with concentrations up to 32 ppm wt H2O. These relatively high hydrogen concentrations in olivine may be explained by (1) a simultaneous stealth and hydrous metasomatism with the hydration correlated with MREE enrichment or by (2) a metasomatism that lead to enrichment in REE, followed by a late ‘fleeting’ hydrogen metasomatism related to the decrease of plume activity and retreat of the Coulomb, at Université de Montpellier 2 (D. Maurin). The LA-ICPMS Louis Bodinier, Marguerite Godard, and Nathalie Bolfan-Casanova. This manuscript has been enriched from informal discussions with Jean-Louis Bodinier, Marguerite Godard, and Nathalie Bolfan-Casanova. The application to mantle petrogenesis and geothermobarometry. Chemical Geology

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