Hydrogen incorporation in olivine from 2–12 GPa

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INTRODUCTION

The importance of nominally anhydrous minerals (NAMs) as storage sites for water in the mantle was recognized over thirty years ago by Martin and Donnay (1972). This topic has received increasing attention over the past two decades as a result of new work on natural and synthetic NAMs as well as widespread recognition of the important influence of even trace amounts of OH on geochemical and geophysical processes (e.g., Hirth and Kohlstedt 1996; Asimow and Langmuir 2003; Karato 2003; Dixon et al. 2004). Experimental data suggesting that water storage capacity increases dramatically in the mantle transition zone (Smyth 1994; Kohlstedt et al. 1996; Bolfan-Casanova et al. 2000) has paved the way for an intriguing, though speculative, model to explain the interplay between mantle convection and the geochemical evolution of the Earth (Bercovici and Karato 2003). Seismological and magnetotelluric studies are also providing more evidence for the existence of hydrated zones in the mantle (Karato 2003; Tarits et al. 2004).

Much of the attention on NAMs has been focused on olivine, the predominant mineral of the upper mantle. Pioneering experimental work (Bai and Kohlstedt 1993; Kohlstedt et al. 1996) revealed systematic trends of hydrogen incorporation in Fe-bearing olivine as a function of water and oxygen fugacity, silica activity, and pressure. However, recent developments have raised new questions and controversies. First, Bell et al. (2003) presented a new calibration for H measurement in olivine using Fourier-transform infrared (FTIR) spectroscopy. Their work suggests that previous solubility measurements need to be revised upward by a factor of 2 to 4, but emphasizes that retrospective application of the calibration is complicated by reliance on unpolarized spectra in those studies. Second, experiments by different groups have yielded highly varying estimates for OH solubility in nominally Fe-free forsterite (Kohn 1996; Keppler and Rauch 2000; Locke et al. 2001; Demouchy and Mackwell 2003; Lemaire et al. 2004; Zhao et al. 2004). Third, previous results (Bai and Kohlstedt 1993) with regard to the effect of silica activity on OH speciation have been contradicted by new experimental studies (Matveev et al. 2001; Lemaire et al. 2004). Fourth, another series of experiments suggests that the presence of Ti is critical to incorporation of OH in olivine (Berry et al. 2004). Finally, new data has been brought to bear on the nature of nanometer-scale inclusions of hydrous phases in olivine (Khisina et al. 2001; Khisina and Wirth 2002; Kudoh 2002; Matsuyk and Langer 2004).

In light of these developments, we conducted a new experimental study of hydrogen incorporation in olivine at pressures of 2–12 GPa and temperatures of 1000–1300 °C. In this study, we employ a new technique for FTIR measurement of OH concentrations in olivine that accounts for the correction of polarized spectra on randomly oriented crystals. The full details of this technique are provided in a separate paper (Asimow et al. 2006).

EXPERIMENTAL AND ANALYTICAL METHODS

Details of the experiments, including starting materials, encapsulation, conditions, and final OH concentrations, are summarized in Table 1 and discussed below.

Starting materials

Most of our experiments employed fine (1–10 μm) or coarse (417–495 μm) powders of olivine made by crushing single crystals from Kilbourne Hole (New...
Mosenfelder et al.: Hydrogen Incorporation in Olivine from 2–12 GPa

Mexico, San Carlos (Arizona), and Zabargad Island (Egypt). The forsterite content of all these olivines lies in the range Fo100–x,90 (Beran and Putnis 1983; Kohlstedt et al. 1996; Wasylenki et al. 2003). Olivine from the first two localities is almost hydrogen-free (less than 1 ppm H2O), whereas the Zabargad olivine contains ~5–10 ppm H2O and has unique IR spectra (Beran and Putnis 1983), indicative of the presence of humite-type defects (Miller et al. 1987). Fine-grained (1–10 µm) orthopyroxene powder was also added to some experiments to influence silica activity. The orthopyroxene, from Kilbourne Hole, has Mg no. [100 × mol% Mg/[Mg + Fe]] = 90.7. The full composition is given by Wasylenki et al. (2003). Hydrogen incorporation in nominally pure (Fo100) forsterite was investigated using two different starting materials. The first material was a fine-grained powder synthesized by sintering dried, high-purity MgO and SiO2 weighed, and mixed together in stoichiometric amounts. The mixed oxides were fired at 1500 °C in a Deltech furnace for 24 h, followed by crushing, mixing, and firing again under the same conditions. X-ray diffraction of the final powder showed nearly pure forsterite with minor amounts of excess MgO. For experiment ww331, we used two oriented forsterite crystals cut from a single crystal synthesized by H. Takei using the Czechralski method (Takei and Kobayashi 1974). These crystals are from the same boule used by Zhao et al. (2004). The dimensions of the crystals were 400 µm (along the a- and c-axes) by 600 µm (along the b-axis). Measured impurities include Fe (several hundred ppm) and <20 ppm of Er, Na, and Ca (see Takei and Kobayashi 1974 and Zhao et al. 2004 for details). The crystals were placed within the capsule such that the b-axis of one crystal and the c-axis of the other were vertically aligned with respect to the press.

Water in most of the experiments was produced by dehydration under the run conditions of a 1:4:1 (by weight) mixture of synthetic talc and brucite (cf. Kohlstedt et al. 1996). The amount of this mixture used in each experiment was chosen so as to provide a nominal total concentration of 1–3 wt% H2O. For experiments with fine-grained olivine, the talc and brucite were intimately mixed with the olivine. In the other experiments, the mixture was packed into the bottom of the capsule below the layer of orthopyroxene. The two experiments conducted on nominally pure forsterite employed only brucite as a water source.

High-pressure experiments

A 1000-ton press and Walker-type multi-anvil module were employed for experiments at 5–12 GPa and 1000–1300 °C. The experimental durations were chosen so as to ensure grain growth (Karato 1989) in runs where fine-grained olivine was used, and to ensure equilibration in experiments using larger crystals, based on the diffusion data of Kohlstedt and Mackwell (1998). Truncated Toshiba “F” grade WC cubes (32 mm edge-length) were used as second-stage anvils to compress octahedral 14/8 (octahedral edge length (OEL)/cube truncation edge length (TEL) in mm) and 18/11 pressure assemblies. The assemblies were comprised of semi-sintered, Cr-doped MgO octahedra containing zirconia insulating sleeves around stepped furnaces made of either LaCrO3 (14/8) or graphite (18/11). We used pyrophyllite gaskets (4.5 mm wide × 2.8 mm thick for the 14/8, and 5 mm wide × 5.24 mm thick for the 18/11), and paper and Teflon tape behind the gaskets for support and electrical insulation. The design of the 18/11 is identical to the 18/8 design illustrated in Frost et al. (2004), but without the high-temperature modification. Temperature was controlled in all experiments using axial W-Re-W,Re thermocouples, with no correction made for the effect of pressure on EMF.

Pressure calibrations for both assemblies are shown in Figure 1. The 14/8 assembly was calibrated at room temperature against the Bi-I-II and Bi-II-III transitions (Getting 1998) and at 1200 °C by bracketing the following transitions: CaGeO3, garnet-perovskite (Susaki et al. 1985), coesite-stishovite (Zhang et al. 1996), and Mg2SiO4 forsterite-wadsleyite (Morishima et al. 1994). The 18/11 assembly was calibrated against the CaGeO3, and coesite-stishovite transitions at 1200 °C. All calibration experiments at high temperature employed a mixture of both low-pressure (50 to 80 wt%) and high-pressure (20–50 wt%) phases, previously synthesized in separate experiments. Sample capsules were made of thin Re foil for the 14/8 and of compound Ni/Au-Pd capsules for the 18/11 (see below). Pressure and temperature uncertainties of multi-anvil experiments are difficult to assess, but are considered to be ±0.5 GPa and ±30 °C, respectively (Rubie 1999).

Initial hydration experiments using welded 1.8 mm O.D. Pt and 1.6 mm O.D. Au-Pd capsules were mostly unsuccessful in retaining excess H2O (Table 1). The best retention was achieved by employing a miniaturized version of the compound capsule technique reported by Ayers et al. (1992). These capsules were comprised of an outer Ni jacket (2.4 mm O.D. /1.6 mm I.D.) with a Au-Pd inner liner (1.6 mm O.D. /1.3 mm I.D.), a 0.3 mm thick Au-Pd gasket, and a 0.5 mm thick Ni lid. The surfaces of the Ni capsule and lid were oxidized by firing at 1000 °C in air for 12 h. The resulting thin layer of NiO serves to buffer oxygen fugacity during the experiment and to retard Ni diffusion through the noble metal liner into the sample (Ayers et al. 1992). The high thermal conductivity and large volume of the capsule also helps to reduce temperature gradients (Ayers et al. 1992). The capsules

**Table 1. Experimental conditions and measured OH concentrations**

<table>
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<tr>
<th>Run no.</th>
<th>Assembly</th>
<th>T (°C)</th>
<th>P (GPa)</th>
<th>Duration</th>
<th>H2O (wt ppm H2O)</th>
<th>Ccal Bell Calibration (wt ppm H2O)</th>
<th>Cpaterson Calibration (wt ppm H2O)</th>
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<td>Pt</td>
<td>1800</td>
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<td>Na</td>
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were insulated from the stepped graphite furnace by a thin MgO sleeve, and the thermocouple was separated from the top of the sample capsule by a 0.3 mm thick disk of fully dense alumina, to prevent thermocouple contamination. Cold-welding of the capsules took place during compression, which typically lasted from 3 to 6 h. All samples were quenched to less than 400 °C within 5–10 seconds by cutting power to the furnace at the completion of each experiment.

We also report the result of a hydration experiment conducted at 2 GPa (nominal pressure) and 1300 °C for 48 h. This was performed in an end-loaded piston-cylinder apparatus using a 1.91 cm CaF₂-graphite pressure assembly. Temperature was controlled using a W-Re-W-Re thermocouple and held constant to ±2 °C. Preliminary calibrations at 600 and 800 °C against the equilibria albite-jadeite-quartz (Holland 1980) and quartz-coesite (Bohlen and Boettcher 1982) suggest that a negative pressure correction of ~0.2 GPa should be applied to the nominal pressure. The sample for the hydration experiment was contained in a welded 5 mm O.D. Au₆₆Pd₃₄ capsule.

Following decompression, all samples were pierced to check for excess water, which provides strong evidence for water-saturation in subsolidus experiments. The capsules were then cut in half to prepare polished sections for analysis. For experiment ww331, two thin sections were made to intersect the (010) and (001) planes of the two forsterite crystals. Most capsules were vacuum-impregnated with Petropoxy 151™ to contain friable, loose grains. Sections for FTIR spectroscopy were made using Crystalbond adhesive, which was subsequently dissolved in acetone, followed by cleaning in ethanol and isopropyl alcohol.

**FTIR spectroscopy and electron microprobe analysis**

Infrared spectra were acquired using a Nicolet Magna 860 FTIR spectrometer coupled to a Spectra-Tech Continuum microscope. Polarized spectra were recorded from 650 to 4000 cm⁻¹ by averaging 400 or more scans with 2 cm⁻¹ resolution, using a GLOBAR infrared light source, a KBr beamsplitter, a CaF₂ wire-grid polarizer, and an MCT-A detector. We used square, 30 to 200 μm wide apertures to collect spectra from regions as free as possible from cracks and grain boundaries.

For each sample, polarized spectra were collected for multiple, randomly oriented crystals, with the E-vector (E) parallel to the two extinction positions of each grain. Because the orientations of the crystals were not known a priori, we developed a new method to use these data to synthesize a consistent set of spectra corresponding to Elia, Ell, and Elc (note we use the same convention for these directions in olivine as Bell et al. 2003). The technique is based on comparison of polarized absorbance spectra between unknowns and standards in the silicate overtone region, from ~1500 to 2200 cm⁻¹. The standards were X-ray oriented slabs of single-crystal San Carlos olivine and synthetic forsterite. Full details of the method, including background corrections and other experimental issues, are given by Asimow et al. (this issue). The orientations of crystals close to zone axes were also verified using Raman spectroscopy (see Arredondo and Rossman 2002 for experimental details) and optical microscopy. OH concentrations were determined by applying the calibration of Bell et al. (2003) to the three synthesized, principal-axis spectra over the wavenumber range 2950–3750 cm⁻¹. For the sake of comparison to previous results, we also calculated concentrations using the Paterson (1982) calibration applied to both the synthetic polarized spectra (in three directions) and theoretical “unpolarized” spectra in the (010) plane calculated by averaging (in transmission) polarized Elia and Ellc spectra (see Libowitzky and Rossman 1996). In the latter case we used an “orientation factor” of 0.33.

Wavelength-dispersive chemical analyses were obtained on selected samples using a JEOL 733 microscope operating at 15 kV and 25 nA beam current. Stan-
Hydrogen incorporation in Fe-bearing olivine

In experiments using finely powdered Fe-bearing olivine as the primary starting material, significant grain growth took place, resulting in grains with diameters up to ~300 μm (Fig. 2). This extent of grain growth is consistent with the kinetic data of Karato (1989). Unfortunately, only two of these runs (at 2 GPa and 10 GPa) yielded an apparent excess of fluid after the experiment. This can be attributed to several factors. The amounts of talc and brucite added to these experiments were deliberately chosen so as to yield enough water to saturate the experiment, but not to shift the Mg no. of the olivine by more than one unit. Thus, significantly less water was evolved in these experiments than in those of Kohlstedt et al. (1996), who used similar encapsulation techniques and starting materials, but larger proportions of talc and brucite. Our failure to detect excess fluid in the experiments may simply reflect an inability to detect small amounts of water by the “bubbling” method. Moreover, water loss from the Pt and Au75Pd25 capsules can be significant. Recent studies provide evidence not only for rapid hydrogen loss, but for oxygen diffusion through capsules as well (e.g., Truckenbrodt and Johannes 1999). Experiments that did not evolve visible bubbles upon piercing are taken to provide lower limits to the OH solubility in olivine at the run conditions.

Nine experiments were conducted using coarser San Carlos olivine crystals and compound Ni/Au75Pd25 capsules. Excess fluid was clearly observed in six of these experiments. Thus, this technique was apparently more successful for retaining water than using noble metal capsules alone. Pre-oxidation of a surface layer on the thick Ni outer capsule ensures effective buffering of oxygen fugacity throughout the experiment (Watson and Cherniak 1997). In contrast, in experiments that employ thin Ni liners, the oxygen buffer might become exhausted as a consequence of reaction with the sample (Kohlstedt et al. 1996). Ni diffusion through the inner capsule in experiments at 1000 °C was minimal, but in long duration experiments at 1100 °C (25 h) and at 1200 °C, Ni diffusion into the sample resulted in the formation of Ni-rich rims on the olivine single crystals (Fig. 2c).

Polarized infrared spectra taken through clear portions of crystals in all these experiments show consistent bands for given orientations. No significant zoning of OH concentrations in the crystals was detected (using 50 μm apertures) in any of the experiments. The expected mechanism of incorporation could be different in the experiments using fine- and coarse-grained materials, because in the former case hydrogen is taken up during grain growth, while in the latter case incorporation proceeds by volume diffusion. Simultaneously, in the experiments with coarse single crystals, Mg-rich olivine forms by reaction of talc and brucite at the bottom of the capsule, and hydrogen incorporation in these new olivine crystals must occur during grain growth. Electron microprobe analyses show that this reaction-product

![Figure 2](https://example.com/figure2.png)

**FIGURE 2.** Examples of experimental products. All scale bars are 500 μm. (a) Experiment ww227 (8 GPa, 1150 °C, 24 h). Olivine grains up to ~300 μm wide grew from a fine-grained powder of Kilbourne Hole olivine mixed with talc and brucite. The dark rind at the bottom of the capsule is the epoxy with which the sample was impregnated. Photograph in reflected light. (b) Sample ww317 (8 GPa, 1000 °C, 24 h). San Carlos olivine contained in a Ni/Au75Pd25 capsule. The dark band near the bottom of the capsule is a layer of orthopyroxene (Opx) added to buffer silica activity. Mg-rich olivine crystals grew from talc and brucite packed into the bottom of the capsule. Photograph in cross-polarized transmitted light. (c) Back-scattered electron micrograph of sample ww325 (8 GPa, 1200 °C, 4 h). Ni-rich rims (lighter gray color) formed on olivine crystals as a result of diffusion of Ni through the inner Au75Pd25 capsule. (d) Large grains of Mg2SiO4 forsterite formed from a partial melt (ww276, 5.65 GPa). The experiment was held at 1300 °C for 15 h, and then cooled rapidly to 1175 °C and held for an additional 15 h. Arrows point to a layer of small MgO crystals at the bottom of the capsule. Photograph in cross-polarized transmitted light.
olivine exchanges Mg and Fe with the rest of the charge (Table 2). Despite these fundamentally different processes of hydrogen incorporation and differences in Mg no. of the product olivine, the IR spectra do not significantly differ between the olivine loaded as large starting crystals and the olivine resulting from tale-brucite breakdown.

Figure 3 shows the synthesized spectra from experiment ww227 (at 8 GPa and 1150 °C) corresponding to Ela, Elb, and Elle. The measured spectra that were used to synthesize these principal-axis spectra are shown in Asimow et al. (this issue). This figure also shows spectra from a natural crystal of olivine from Buell Park, Arizona, which contains ~50 ppm H2O. The comparison reveals a remarkable consistency in band positions and relative peak heights even though the spectra differ by a factor of ~40 in integrated absorption. The spectra from this Buell Park olivine, previously unpublished, are different from the spectra of an orange-brown colored olivine from this locality, which contains planar titanoclinohumite defects (Kitamura et al. 1987; Mosenfelder et al. 2004).

The Ela spectra in our hydrothermally grown olivines are dominated by the same high-wavenumber bands seen in the study of Kohlstedt et al. (1996), in which H diffused into relatively large crystals of San Carlos olivine. Different starting olivines yielded spectra with nearly identical band structure (Fig. 4), despite the difference in initial speciation (i.e., the bands attributed to humite in Zabargad olivine). The strongest bands are at 3613, 3598, 3579, and 3567 cm⁻¹, and the heights of the peaks increase with increasing water fugacity (Fig. 5). Note that the height of the most intense peaks may be underestimated in some samples, because the extinction ratio of the polarizer becomes increasingly important as the absorbance approaches a high value (close to 2) (Libowitzky and Rossman 1996). This could explain, for instance, the apparent change in relative peak intensities near 3613 and 3598 cm⁻¹ as water fugacity increases (Fig. 5). However, this should have a minimal effect on the integrated absorbance over the entire OH-stretching range.

In addition to the strong bands discussed above, most spectra in the present study show a triplet of absorption bands at 3502, 3483, and 3456 cm⁻¹. These bands, also observed in the Buell Park olivine shown in Figure 3, are prominently observed only in one unpolarized spectrum (at 12 GPa, 1100 °C) presented by

![Figure 3](https://via.placeholder.com/150)

**Figure 3.** Polarized IR spectra from an experimental sample (ww227, 8 GPa, 1150 °C, 24 h) and a natural crystal of olivine from the Buell Park diatreme in Arizona (no. GRR1629). Experimental (black) and natural (gray) sample spectra are normalized to 1 mm and 1 cm absorbance, respectively, and offset arbitrarily for the sake of comparison. The spectra for ww227 are synthesized using the technique outlined in the text, and are baseline corrected. The spectra from the Buell Park sample were acquired on an oriented single crystal, and are not corrected for baseline. The E-vector of each spectrum is noted in the figure.

![Figure 4](https://via.placeholder.com/150)

**Figure 4.** Synthetic Ela IR spectra for olivine from three different experiments that used three different starting materials. The starting olivine material and conditions of the experiments are annotated in the figure. The run numbers (see Table 1), starting from the top of the figure going down, are ww256, ww325, and ww227.

![Figure 5](https://via.placeholder.com/150)

**Figure 5.** Ela IR spectra (synthetic) for olivine from five experiments, showing the increase in absorption with increasing water fugacity. The conditions and water fugacity (in parentheses) for each experiment are annotated in the figure. Water fugacity was calculated according to the equation of state for water of Pitzer and Sterner (1994), as discussed in the text. The peak height for the 3613 cm⁻¹ band in the uppermost spectrum may be underestimated, as discussed in the text.
Kohlstedt et al. (1996), and were not observed in lower pressure studies (Bai and Kohlstedt 1993; Zhao et al. 2004). Olivine in the latter studies also exhibits prominent bands at 3525 and 3573 cm⁻¹ that are not observed in this study. These differences may reflect fundamental changes in the incorporation mechanism of hydrogen in olivine between low pressure (<1 GPa) and high-pressure experimental studies.

Another notable feature of the spectra in this study is the low intensity or complete lack of strong absorption bands in the region between 3300 and 3400 cm⁻¹. These bands were generally seen in samples annealed at higher oxygen fugacities in the study of Bai and Kohlstedt (1993). Matveev et al. (2001, 2005) also observed these bands in samples annealed in the presence of orthopyroxene, and attributed them to high silica activity rather than to high oxygen fugacity.

Measured OH concentrations in the present study are given in Table 1, and compared to the measurements of Bai and Kohlstedt (1993) and Kohlstedt et al. (1996) in a log-log plot of concentration against water fugacity in Figure 6. The water fugacity was calculated using the equation of state for water of Pitzer and Sterner (1994). Our values for water fugacity differ by about 13–14% from those calculated by Kohlstedt et al. (1996), and we have replotted their data with our recalculated values; the conclusions of our study are not dependent on this discrepancy, however. The results show a difference of 2.5 to 4 in absolute values between our study and that of Kohlstedt et al. (1996).

### Hydrogen incorporation in Fo100

Results of the two experiments we performed in the nominally pure MgO-SiO₂-H₂O system were surprisingly different, despite a similarity in P-T conditions. In experiment ww276 at 5.65 GPa, the temperature was first held above the solidus (Inoue 1994) at 1300 °C for 15 h. Subsequently, the temperature was decreased at a rate of 50 °C/min to 1175 °C and held for 15 h. The end product (Fig. 2d) consisted of forsterite crystals with diameters up to ~500 μm, with a small amount of fine-grained (5–10 μm) crystals of MgO concentrated on the bottom of the capsule. Measured FeO concentrations were below the detection limit of 0.02 wt% (Table 2). IR spectra (Fig. 7a) reveal the...
same features as in the experiments of Lemaire et al. (2004) at low silica activity. The concentration of OH in these olivines is comparable to that of Fo90 at similar conditions (Table 1).

In contrast, in experiment ww331, conducted at 6 GPa using oriented single crystals, OH concentrations were much lower (Fig. 7b). Polarized IR spectra of these crystals are similar to those of forsterite in the studies of Demouchy and Mackwell (2003) and Zhao et al. (2004). The duration of this experiment was chosen so as to limit water loss from the sample, while ensuring homogenization of OH according to the diffusion data of Demouchy and Mackwell (2003). Nevertheless, distinct zoning profiles of OH were observed in the sample. The spectra shown in Figure 7b represent the highest absorption measured in the samples. The large difference in OH concentrations between the two Fo100 experiments is discussed below.

**Discussion**

**Mechanisms of hydrogen incorporation**

In the following discussion, we refer to three groups of absorption bands in olivine. Following Bai and Kohlstedt (1993), Group I bands occur in the range 3450–3630 cm⁻¹, and Group II bands are in the range 3200–3450 cm⁻¹. We also distinguish a subset of Group I, hereafter referred to as “Group I-H,” represented by strong bands at 3525 and 3573 cm⁻¹. All of these bands are strongly anisotropic, with the greatest absorption along c-axis.

Many hypotheses have been put forward to explain the incorporation of hydrogen in olivine. Point defects have been linked to hydrogen incorporation, but the question of which point defect predominates under what conditions is controversial, and it is not possible to obtain unique assignments for olivine based on dipole directions and the wavenumber dependence of IR bands (e.g., Kusová et al. 2004). Despite attempts to correlate trace element concentrations in olivine with H content (e.g., Kusová et al. 1997), boron is the only element clearly demonstrated to play a role in hydrogen incorporation via coupled substitution (Kent and Rossman 2002), and is unlikely to play an important role at high OH concentrations. Studies of natural olivines using spectroscopy and electron microscopy have demonstrated that OH can be incorporated into olivine via nm- to μm-scale inclusions or planar defects of hydrous phases, including talc, serpentine, and various humite-series minerals (Kitamura et al. 1987; Miller et al. 1987; Khisina et al. 2001; Matsuoy and Langer 2004; Mosenfelder et al. 2004). A structure of “hydrous modified olivine” has also been proposed based on electron microscopy (Khisina et al. 2001) and modeling (Kudoh 2002). This defect locally has the structure of clinohumite, and can also be considered as an ordered array of point defects.

The hydrous phases that may be included in olivine exhibit distinct IR bands (Miller et al. 1987; Matsuoy and Langer 2004). None of the bands corresponding to talc or serpentine are observed in our experimentally hydrated olivines. Group I-H bands, which are common in the majority of natural, H-bearing olivines, were attributed to humite by Miller et al. (1987), and may also represent a humite-related defect that is not detectable optically (for instance, the hydrous modified olivine structure proposed by Kudoh 2002). We also do not observe these bands in our experimental study, but note that they are the dominant bands in lower-pressure experiments (Bai and Kohlstedt 1993; Zhao et al. 2004). Berry et al. (2004) also produced olivines with these bands, and attributed them to the presence of Ti. Trace amounts of Ti may stabilize a defect related to titanoclinohumite, which exhibits Type I-H bands as well as a prominent band near 3400 cm⁻¹ (Kitamura et al. 1987), which is apparently missing in the experimental studies.

We conclude that incorporation of hydrogen via inclusions of hydrous phases is unlikely to be an important mechanism in our experiments or those of Kohlstedt et al. (1996), although electron microscopy is needed to verify this hypothesis. This may reflect a change in incorporation mechanism between olivines hydrated at low pressures, where “hydrated modified olivine” or another humite-related defect dominates, and high pressures, where disordered point defects dominate. Ancillary evidence for this hypothesis comes from the close correspondence in band positions between our samples and the natural olivine from Buell Park (Fig. 3); the exact P-T conditions of formation for this xenocryst are unknown, but derivation from conditions in the garnet peridotite stability field is likely based on association with pyropic garnets and chromian diopside (Smith and Levy 1976).

Bai and Kohlstedt (1993) and Kohlstedt et al. (1996) promoted incorporation via oxygen interstitials (balanced by two protons), but this idea was later dropped in favor of metal vacancies (Kohlstedt and Mackwell 1998). Brodholt and Refson (2000) delineated a more important role for Si vacancies on the basis of ab initio calculations showing that the concentration of Si vacancies is enhanced relative to metal vacancies under wet conditions. More recently, Matveev et al. (2001, 2005) argued that silica activity plays a crucial, previously unrecognized role, resulting in fundamentally different IR band structures at different conditions. They ascribed Group I bands to Si vacancies formed under conditions of low silica activity, and Group II bands to metal vacancies formed at higher silica activity in equilibrium with orthopyroxene. A surprising conclusion of their study is that olivines from orthopyroxene-bearing peridotite sources (Miller et al. 1987) are suggested to have been re-equilibrated at lower silica activity (based on the predominance of Group I bands). Their experimental results contradict those of Bai and Kohlstedt (1993), who showed no evidence for a significant effect of silica activity on incorporation mechanism. Matveev et al. (2001) also suggested that the olivines in experiments by Bai and Kohlstedt (1993) and Kohlstedt et al. (1996) were not in equilibrium because of an apparent lack of Fe-Mg exchange between the olivine and the silica buffer.

One aspect of the experiments by Matveev et al. (2001) that has not been discussed is that they were conducted at higher oxygen fugacity (buffered at Re-ReO₂; see Pownceby and O’Neill 1994) than the experiments of Kohlstedt et al. (1996), which were buffered at Ni-NiO. Bai and Kohlstedt (1993) also conducted high-pressure experiments using olivines pre-equilibrated over a wide range of oxygen fugacities at 1 atm, and demonstrated that the intensity of the Group II bands increased with increasing pre-annealing oxygen fugacity. More recently, Zhao et al. (2004) demonstrated a trend of increasing intensity of Group II bands with increasing temperature, which could also be related to increasing oxygen fugacity (either increasing with temperature, or increasing due to exhaustion of the buffer).

Lemaire et al. (2004) also examined the effect of varying silica activity on hydrogen incorporation, in nominally pure
Mg$_2$SiO$_4$ olivine. They demonstrated that, with increasing silica activity, the intensity of Group I bands decreases, but that the bands do not disappear, in contrast with the results of Matveev et al. (2001). The Group II bands seen in Fe-bearing samples were not observed, but two additional bands were seen at even lower wavenumbers, at 3160 and 3220 cm$^{-1}$. These bands exhibit a broader FWHM than other bands, and also exhibit distinctly different anisotropy, appearing in Et of spectra only. The band at 3160 cm$^{-1}$ has not been seen in natural olivines, to our knowledge, but the band at 3220 cm$^{-1}$ nearly corresponds to a band seen in Zabargad olivine, with the same anisotropy (Beran and Putnis 1983). This band was attributed by Lemaire et al. (2004) to metal vacancies at high silica activity. However, it can also be attributed to a humite-type defect, as it is seen in the spectra of humite-type minerals (Matsyuk and Langer 2004).

In our experiments, which were nominally either buffered at high silica activity (in the presence of orthopyroxene) or unbuffered, we only see Group I bands. We therefore suggest that the band assignments of Matveev et al. (2001) are either incorrect or that there is a fundamental change in incorporation mechanism under high silica activity conditions between their experiments, at pressures of 2 GPa and below, and those at higher pressures (this study and Kohlstedt et al. 1996).

**Calibration of OH in olivine**

Bell et al. (2003) showed that their new calibration for OH in olivine differed systematically from that of Paterson (1982) by a factor of either 2.3 or 3.5, depending on whether the analysis was carried out using polarized or unpolarized IR spectra, respectively. They reiterated the warning of Libowitzky and Rossman (1996) that the use of unpolarized radiation can introduce non-systematic errors into the analysis, due to partial polarization of the beam absorption intensities, which are not a linear function of thickness. Nevertheless, several authors have adopted the 3.5 correction factor (Koga et al. 2003; Aubaule et al. 2004; Bolfan-Casanova 2005; Katayama et al. 2005).

In Figure 8, we show a comparison for our samples indicating that OH concentrations derived using the Bell et al. (2003) calibration are, on average, 2.95 and 4.12 times the amount calculated using the Paterson (1982) calibration for polarized and unpolarized radiation, respectively. This disparity with Bell et al. (2003) may result from either the 2.3 or 3.5 correction factor, or other factors.

Unfortunately, the OH concentrations of olivines in the present study exhibit too much scatter to derive a meaningful function of OH solubility with temperature or Fe-content at high pressure. Therefore, we restrict ourselves to an analysis similar to that of Kohlstedt et al. (1996), using their Equation 16:

$$ C_{OH} = A f_{H_2O} \exp(-P \Delta V / RT) $$

where $C_{OH}$ is the concentration of OH in olivine expressed as H/10$^6$ Si, $A$ is a constant, $f_{H_2O}$ is the fugacity of water, $\Delta V$ is the change in molar volume of the olivine lattice due to hydrogen incorporation, $R$ is the gas constant, and $P$ and $T$ are pressure and temperature. The exponent $n$ for water fugacity is 1 whether the mechanism of incorporation involves two hydrogen atoms substituted in a metal vacancy or bonded to an oxygen interstitial (see Kohlstedt et al. 1996). For simplicity we assume $T = 1373$ K even though our experiments vary in temperature from 1273 to 1573 K. We fit the equation only to experiments in which excess water was present (Table 1). The analysis yields $A = 2.45$ H/10$^6$ Si/GPa and $\Delta V = 10.2 \times 10^{-6}$ m$^3$/mol. The value for $\Delta V$ is thus indistinguishable from the values of 10.6 $\times 10^{-6}$ m$^3$/mol and 10.0 $\times 10^{-6}$ m$^3$/mol calculated by Kohlstedt et al. (1996) and Zhao et al. (2004), respectively. The fit to all the data with these parameters is shown in Figure 6 as a dashed line. The new values indicate an OH solubility 2.2–3.4 times higher than that of Kohlstedt et al. (1996). Clearly, more work is needed to accurately assess the effects of temperature, Fe content, silica
activity and oxygen fugacity at the high pressures (up to 13 GPa) relevant to the entire upper mantle.

**Attainment of equilibrium in studies of hydrogen incorporation**

The question of whether hydrogen incorporation experiments have reached equilibrium has rarely been addressed (cf. Bromiley et al. 2004). In our experiments on Fe-bearing olivine, we are encouraged by the fact that the IR spectra of the products are similar regardless of starting material or mode of incorporation (i.e., during grain growth vs. via diffusion into larger crystals), and that the crystals do not exhibit zoning profiles in OH concentration. The discrepancy in OH concentrations in our two experiments on Mg$_2$SiO$_4$ forsterite is disturbing, however. There is no obvious difference in composition of the two starting materials that could account for this discrepancy.

The inferred solubility of OH in Mg$_2$SiO$_4$, forsterite varies by orders of magnitude in the literature, from very low values (Keppler and Rauch 2000; Demouchy et al. 2003; Zhao et al. 2004) to values equal to or higher than that for Fe-bearing olivine at comparable conditions (Kohn 1996; Locke et al. 2001; Lemaire et al. 2004). This discrepancy can even be observed within a single study. Lemaire et al. (2004) produced forsterites with either 60 or 750 ppm H$_2$O in two separate experiments, even though both runs were saturated with H$_2$O, and the composition, temperature and cooling rate from above the liquidus were all nominally identical. Lemaire et al. (2004) inferred that cooling rate was an important factor in OH solubility, but the details of this relationship are not well constrained. We are left with the troubling conclusion that disequilibrium effects may attend hydrogen incorporation during growth of crystals from a melt or partial melt, and inferring OH solubility from such experiments is dangerous. Kinetic effects may also play a complex role in incorporation of OH by diffusion into large crystals, even when diffusion rates are nominally constrained (Bromiley et al. 2004).

**Geophysical and geochemical implications**

Our new results indicate that models need to be revised for processes that are affected by a high solubility of OH in olivine, such as partitioning of hydrogen between olivine and other phases during melting (Hirth and Kohlstedt 1996; Asimow and Langmuir 2003), and viscosity of hydrous zones in the mantle (Hirth and Kohlstedt 1996; Dixon et al. 2004). Although the change in estimated bulk partition coefficient for H$_2$O between melt and fertile peridotite due to our revised estimate of OH solubility is small (about 30%), the difference would increase during partial melting as the modal abundance of olivine in the residue increases. On a broader scale, Bercovici and Karato (2003) recently presented an intriguing model that attempts to reconcile geochemical and geophysical constraints on mantle convection. One of the postulates of their “transition zone water filter” hypothesis is that a sharp contrast in OH solubility between minerals of the transition zone (with high solubility represented by wadsleyite) and the upper mantle (represented by olivine) can result in the formation of hydrous melts at the 410 km discontinuity that sequester trace elements. They also propose that hot mantle plumes can “bypass” this filter because the solubility in olivine increases with temperature while that of wadsleyite decreases with temperature. These legs of the hypothesis hinge on experimental constraints on OH solubility such as those presented here.

Our updated estimate for OH solubility in olivine reduces the contrast in solubility inferred by Bercovici and Karato (2003), because estimates of hydrogen solubility in wadsleyite are based on different methods that remain robust at high hydrogen content: SIMS analyses (Demouchy et al. 2005) and X-ray diffraction (Smyth 1994). This conclusion was also reached by Hirschmann et al. (2005), on the basis of SIMS measurements and the work of Bell et al. (2003). The simple thermodynamic model of Bercovici and Karato (2003) for the dependence of OH solubility on pressure and temperature hinges on an extreme extrapolation of the low-pressure data (Zhao et al. 2004), which may be complicated by changes in incorporation mechanism between low- and high-pressure experiments as discussed above. Furthermore, Hirschmann et al. (2005) speculated that changes in fluid chemistry at high pressures might reverse the temperature dependence of OH solubility in olivine. Bercovici and Karato (2003) also ignored the contribution of pyroxenes to the water storage capacity of the upper mantle. Finally, the inference that OH solubility in wadsleyite and ringwoodite decreases with temperature was based at the time on only a few experiments, many of which were conducted at water-undersaturated conditions in the presence of a melt (Ohnai et al. 2000). Although subsequent experiments have borne out this trend (Demouchy et al. 2005), they demonstrate that the decrease in OH content of wadsleyite above 1200 °C results from decreasing H$_2$O activity in the coexisting fluid as it becomes, either continuously or discontinuously, more melt-like. Extrapolation of these results to Fe-bearing compositions is highly uncertain. Clearly, more mineral physics constraints are needed to evaluate the transition zone water filter hypothesis.

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