



**HAL**  
open science

## **C-O-H fluid solubility in Haplo-basalt under reducing conditions: An experimental study**

Yann Morizet, M. Paris, Fabrice Gaillard, Bruno Scaillet

► **To cite this version:**

Yann Morizet, M. Paris, Fabrice Gaillard, Bruno Scaillet. C-O-H fluid solubility in Haplo-basalt under reducing conditions: An experimental study. *Chemical Geology*, 2010, 279, pp.1-16. 10.1016/j.chemgeo.2010.09.011 . insu-00524494

**HAL Id: insu-00524494**

**<https://insu.hal.science/insu-00524494>**

Submitted on 8 Oct 2010

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

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

**C-O-H FLUID SOLUBILITY IN HAPLO-BASALT UNDER REDUCING  
CONDITIONS: AN EXPERIMENTAL STUDY**

Y. MORIZET<sup>1</sup>, M. PARIS<sup>2</sup>, F. GAILLARD<sup>3</sup>, B. SCAILLET<sup>3</sup>

<sup>1</sup> Université de Nantes, Nantes Atlantique Universités, Laboratoire de Planétologie et  
Géodynamique de Nantes (LPGN)

UMR CNRS 6112

2 rue de la Houssinière, 44322 NANTES (France)

<sup>2</sup> Institut des Matériaux Jean Rouxel (IMN), Université de Nantes

UMR CNRS 6502

2 rue de la Houssinière, BP32229, 44322 NANTES Cedex 3 (France)

<sup>3</sup> CNRS/INSU-Université d'Orléans-Université François Rabelais Tours

UMR 6113

Institut des Sciences de la Terre d'Orléans

1a rue de la Férellerie, 45071, Orléans, France

Corresponding author: Dr Yann Morizet

Permanent address:

Laboratoire de Planétologie et Géodynamique de Nantes (LPGN), Université de Nantes

CNRS/INSU – UMR CNRS 6112

2 rue de la Houssinière, BP 92208.

44322 Nantes Cedex 3 (FRANCE).

Email: [yann.morizet@univ-nantes.fr](mailto:yann.morizet@univ-nantes.fr)

Tel: +33 (0) 2 5112 5491

Fax: +33 (0) 2 5112 5268

**Abstract:**

We conducted an experimental study to constrain the C-H-O solubility and speciation in hydrous silicate melts equilibrated under reduced  $fO_2$  conditions. Haplo-basaltic glasses in the NCMAS-C-O-H system were synthesised using IHPV at 1250°C, 200-300 MPa with variable applied  $fH_2$  so as to vary  $fO_2$ . Recovered rapidly quenched glasses were characterized using various spectroscopic methods: Micro-FTIR, Raman and  $^{13}C$ -MAS NMR. Glass  $CO_2$  content changes from 680 to 1320 ppm between  $\Delta FMQ-2.6$  and  $\Delta FMQ+2.6$  independently of  $H_2O$  content changing from 1.3 to 4.0 wt.%. Recent thermodynamic modelling of isobaric  $CO_2$ - $H_2O$  solubility fails to reproduce our  $CO_2$ - $H_2O$  solubility trend under reducing conditions. The lower  $CO_2$  solubility in the melt as compared to more oxidized conditions is directly correlated to the decrease of  $fCO_2$  within the fluid phase under reducing conditions. Carbonate groups ( $CO_3^{2-}$ ),  $OH^-$  and  $H_2O^{mol}$  are the volatile species in the glasses. No evidence for  $CH_4$ , carbides or organic compounds was observed.  $^{13}C$  MAS NMR analysis suggests that several carbonate units are coexisting in the glasses.  $\{^1H\}$   $^{13}C$ -CPMAS NMR suggests that all  $CO_3^{2-}$  units are surrounded by OH groups. Those environments appear to slightly change with changing  $fO_2$  conditions suggesting different degree of hydrogenation in the vicinity of the carbonate groups. Our data show that the presence of a significant amount of dissolved does not increase the solubility of species such as CO or  $CH_4$ . In other words, such species remain insoluble in basaltic melts, as established under dry conditions. Altogether, our  $CO_2$  solubility results show that a wet but reduced basalt will degass more C-species than if oxidized, owing to the lower prevailing  $fCO_2$  and insoluble character of

CO. The presence of an important fraction of CO in the fluid phase will have a large impact on the primitive atmospheric compositions of Mars and the Earth.

Keywords: CO<sub>2</sub>-H<sub>2</sub>O solubility, CO<sub>2</sub> speciation, silicate melt, oxygen fugacity, <sup>13</sup>C NMR.

## **1. Introduction:**

Earth degassing through volcanism is an important process to investigate for understanding the chemical variations of our atmosphere and the impact of volcanism on climate change. Establishing reliable models of the carbon and hydrogen geochemical cycles within planetary bodies such as Mars and the Earth implies understanding the partitioning of carbon and hydrogen species between their internal reservoirs. Earth's mantle is one of the most important carbon and hydrogen reservoir of the Earth (e.g. Javoy et al., 1986; Marty and Jambon, 1987; Ahrens, 1989; Ingrin et al., 1995; Sleep and Zahnle, 2001; Gaillard et al., 2008), being in particular, one of the most important carbon dioxide (CO<sub>2</sub>) long term source for the atmosphere (e.g. Zhang and Zindler, 1989; Azbel and Tolstikhin, 1990).

Under oxidising conditions, CO<sub>2</sub> and H<sub>2</sub>O are the most abundant species to be dissolved in silicate melts. H<sub>2</sub>O and CO<sub>2</sub> solubility, speciation and dissolution mechanisms are fairly well constrained as a function of melt composition, temperature, pressure (e.g. Fogel and Rutherford, 1990; Thibault and Holloway, 1994; Dixon, 1997; Morizet et al., 2002; Behrens et al., 2004; Moore, 2008). CO<sub>2</sub> and H<sub>2</sub>O gases are presently the two most important volatiles released by Earth degassing through volcanism however more reduced species such as CH<sub>4</sub> and CO are thought to have been degassed especially during the Archean period giving rise to a more reduced atmosphere (Kump and Barley, 2007; Kump, 2008). The impact of those reduced species on the global environmental change for the primitive atmospheres is to be considered.

Determining the solubility behaviour of C-H-O volatiles in reduced magmas is essential also for constraining the global geochemical cycle of those volatiles during the evolution

of other terrestrial planets such as Mars (e.g., Hirschmann and Withers, 2008; Gaillard and Scaillet, 2009). There is unfortunately currently little information on the solubility and speciation behaviour of reduced C-H-O species in silicate melts, in particular under H<sub>2</sub>O-rich conditions which may have prevailed during the early magma ocean stage (e.g., Drake and Righter, 2002; Righter et al., 2007; Elkins-Tanton, 2008).

Experimental data were acquired by Pawley et al. (1992) for basaltic glasses: these authors demonstrated that the relationship between CO<sub>2</sub> solubility and CO<sub>2</sub> fugacity ( $f_{\text{CO}_2}$ ) established at high  $f_{\text{O}_2}$  is valid as well under reduced conditions in the low pressure range investigated (<150 MPa). Kadik et al. (2004) presented additional results for the solubility, speciation and dissolution mechanisms of C-O-H fluids in silicate glasses as a function of redox conditions suggesting the presence of more reduced species such as CH<sub>4</sub> dissolved within the melt. At high pressure (1.0-2.5 GPa) and under reducing conditions (iron-wüstite buffer), for glasses in the Na<sub>2</sub>O-SiO<sub>2</sub> system, Mysen et al. (2009) proposed the formation of methyl groups (CH<sub>3</sub>) bound to Non-Bridging Oxygens (NBO) suggesting the depolymerization of the melt through the formation of C-H bonds. Notwithstanding, there is no clear evidence as to whether or not a reduced fluid depolymerizes the melt structure.

We have conducted an experimental study aimed at determining the solubility, speciation and dissolution mechanisms for C-H-O fluid species in equilibrium with a synthetic hydrous basaltic melt synthesised at 200-300 MPa and 1250°C. Experiments were conducted under various redox conditions from  $\Delta\text{FMQ}+2.6$  to  $-2.6$ , i.e. outside the graphite saturation surface. All experiments were performed under hydrous conditions as compared to previous studies (e.g., Pawley et al., 1992). Results on solubility and

speciation for C-H-O species were obtained from spectroscopic analyses. An approach is proposed for the dissolution mechanisms for C-H-O species.

## **2. Experimental and analytical methods:**

### 2.1. Experimental synthesis:

Synthetic Fe-free basaltic starting composition was prepared from a mixture of oxides ( $\text{SiO}_2$ ,  $\text{Al}_2\text{O}_3$  and  $\text{MgO}$ ) and carbonates ( $\text{Na}_2^{13}\text{CO}_3$  and  $\text{Ca}^{12}\text{CO}_3$ ). The absence of iron is justified by the analytical needs for investigating the C speciation with NMR which cannot be used on Fe-bearing materials. The investigated composition is 54.8 wt.%  $\text{SiO}_2$ , 21.1 wt.%  $\text{Al}_2\text{O}_3$ , 8.1 wt.%  $\text{MgO}$ , 12.9 wt.%  $\text{CaO}$  and 3.2 wt.%  $\text{Na}_2\text{O}$ . It corresponds to an  $\text{NBO/T} = 0.42$  (Non-Bridging Oxygen / Tetrahedral charges) according to the definition of Mysen (1988; 1990). The starting composition has an Aluminium Saturation Index  $\text{ASI} = 0.73$  following the definition of London (1992).

The sample nomenclature is explained as follows: the first figure stands for the experiment number, the second figure stands for the capsule number during the experiment and the final part gives some particularity of the sample (e.g. *us*: undersaturated; *ox*: for the most oxidized sample; *red*: for the most reduced sample).

The starting composition was prepared as follows:

- 1)  $\text{SiO}_2$ ,  $\text{Al}_2\text{O}_3$ ,  $\text{MgO}$  and  $\text{Ca}^{12}\text{CO}_3$  were mixed together and held at  $1300^\circ\text{C}$  for few hours in order to dissociate the calcium carbonate.  $^{12}\text{CO}_2$  is degassed during this process, however it appears to be an incomplete decarbonation as  $^{12}\text{C}$  isotope is still present within the glass (see section 4.1).



- 2)  $\text{Na}_2^{13}\text{CO}_3$  (except for run 2-2- $^{12}\text{C}$  and 3-2- $^{12}\text{C}$  in which  $\text{Na}_2^{12}\text{CO}_3$  was added, see Table 1) is added to complete the composition and is the source of both  $\text{CO}_2$  (~1 wt.%) during the experiment and  $^{13}\text{C}$ -source for solid-state NMR analytical needs. Under the studied P,T conditions (200-300 MPa, 1250°C), the 1 wt.%  $^{13}\text{CO}_2$  added to the starting composition is well above the solubility saturation level for basaltic compositions (Fine and Stolper, 1986; Pan et al., 1991; Dixon et al., 1995) and the melt will be in equilibrium with an excess fluid phase in which C-species dominate.
- 3) A small proportion of  $\text{Fe}_2\text{O}_3$  (< 1 wt.%) is added as a source of paramagnetic element to reduce the NMR acquisition time length by reducing the relaxation time (Maekawa et al., 1991) which is usually long for  $^{13}\text{C}$  nuclei (Kohn et al., 1991). The  $\text{FeO}^{\text{tot}}$  was quantified by electron microprobe and is reported in Table 2 along with the NMR analytical conditions. Except for 6-1-*ox*, the  $\text{FeO}^{\text{tot}}$  measured value in the starting materials were below 0.6 wt.% which is recognized not to produce a drastic change in the NMR spectrum profile (Maekawa et al., 1991). For 6-1-*ox*, the  $\text{FeO}^{\text{tot}}$  is relatively high due to high  $f\text{O}_2$  that minimize Fe loss toward the container and is likely to produce changes in the line shape of the NMR spectrum.

A particular case is made for 4-2-*us* which is initially  $\text{CO}_2$ -free. The starting composition was prepared by mixing all oxides and carbonates and firing the mixture at 1300°C for  $^{12}\text{CO}_2$  decarbonation.

Between 0 and 4 wt.% H<sub>2</sub>O is added together with the solid starting materials in the platinum capsules (see Table 1). Starting composition is loaded after water and the platinum capsule is welded shut. Experiments are conducted in an Internally Heated Pressure Vessel (IHPV) for run durations between 2 to 3 hours which is sufficient to ensure chemical equilibrium is achieved throughout the sample charge considering that at 1250°C 1) CO<sub>2</sub> and H<sub>2</sub>O diffusion coefficient are high (Watson, 1994; Zhang et al., 2007) and 2) H<sub>2</sub> diffuses very quickly through the sample charge (Schmidt et al., 1997; Gaillard et al., 2003a).

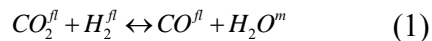
The experimental conditions are conducted between 197 and 320 MPa for pressure, 1250°C for temperature (see Table 1 for detailed experimental conditions). All recovered synthesised samples are listed in Table 1.

The procedure to achieve the experimental conditions is described elsewhere (e.g., Scaillet et al., 1992) and is summarized as follow. The IHPV is first loaded at room temperature with a known partial pressure of pure H<sub>2</sub>, Argon gas is then loaded up to roughly half the desired total pressure (P<sup>tot</sup>) value (either 100 or 150 MPa). The temperature is then brought to the final value and the P<sup>tot</sup> increases up to requested experimental pressure (200 or 300 MPa). P and T are constantly monitored during the experiment using an electronic gauge for P and a type S thermocouple connected to a Eurotherm controller for T. The T change along the capsule is measured with an additional thermocouple and does not exceed 15°C. P, T measurements are accurate within ± 20 bars and ± 5°C.

The estimate of the final P(H<sub>2</sub>) (see Table 1) during the experiment is crucial. The high temperature of the experiment and the short run duration prevent the use of techniques

allowing direct measurements of the  $P(\text{H}_2)$  such as the Shaw membrane ( $\text{Ag}_{40}\text{Pd}_{60}$ , Scaillet et al., 1992) or the solid buffer (Eugster and Wones, 1962). An internal calibration completed over more than 10 years of experiments in the  $\text{H}_2$ -Ar gas mixture with the same autoclave shows that under those P-T conditions the final estimated  $P(\text{H}_2)$  is therefore equivalent to roughly 3-4 times the initial loaded  $P(\text{H}_2)$  (Martel et al., 1999; Gaillard et al., 2003b; Di Carlo et al., 2006). The final  $P(\text{H}_2)$  experienced during the run are shown in Table 1, using an empirical multiplication factor of 3. Changing the multiplication factor to 4 will only slightly affect the final calculated  $f\text{O}_2$  value, by -0.3 log unit in our study.

Hydrogen acts as a reducing agent for the loaded  $\text{CO}_2$  according to the reaction in the fluid phase:



where  $fl$  stands for the gas in the fluid phase and  $m$  stands for the species in the melt phase. Further reduction might produce more reduced species such as  $\text{CH}_4$  (Holloway, 1987) or even graphite.

The sample is quenched to the ambient temperature by dropping the capsules into the cold part of the IHPV. The quench rate is relatively fast, on the order of  $150^\circ\text{C}/\text{s}$  for that kind of apparatus (Di Carlo et al., 2006; Iacono-Marziano et al., 2008). In all cases, opening of the capsules produced a hiss suggesting that a fluid phase was present in the capsule during the experiment.

Several chips from the recovered glasses were selected for micro-Raman and micro-FTIR analyses. The remaining of the glass sample was then crushed for  $^{13}\text{C}$ -MAS NMR analysis. Optical examination of the glass reveals the presence of randomly distributed round shaped fluid inclusions as shown in Morizet et al. (2009). The recovered glasses are free of crystals. For the most reduced samples, the glass is not totally transparent although not blackened. We carefully scrutinized the sample but did not get evidence for graphite crystalline phase however platinum complexes might have been formed during the experiments. Our observation is also confirmed by the absence of i) Raman signatures for graphite either within the glass or the fluid inclusion walls and ii)  $^{13}\text{C}$ -NMR signatures for crystalline or amorphous graphite.

## 2.2. Confocal Micro-Raman:

Confocal micro-Raman was used to characterize i) the fluid species inside inclusions for determining the  $f\text{O}_2$  experienced during the run (see Table 1 for numerical values,  $\log f\text{O}_2$  (2); and section 3 for calculation description) and ii) the  $\text{H}_2\text{O}$  in the melt via the silicate and water-related Raman signatures (see Table 3 for numerical values,  $\text{H}_2\text{O}^{\text{tot}}$  (2)). The Raman system is a Jobin-Yvon Labram spectrometer (focal distance = 300 mm) equipped with a 2400 gratings/mm CCD detector. The light source is an Ar laser Innova 300-5W from Coherent© operating at 514.5 nm. The typical output laser power was set to 120 mW. The analyses were performed in confocal mode (hole = 500  $\mu\text{m}$ , slit = 200  $\mu\text{m}$ ) with an Olympus (x50) objective. The confocal mode is particularly useful for fluid inclusion analysis as the contribution of the surrounding glass is reduced in the final spectrum.

Spectra were recorded between 200 and 4000  $\text{cm}^{-1}$ . Fluid inclusions were analysed for  $\text{CO}_2$ -CO scattering between 1100 and 2200  $\text{cm}^{-1}$ . Analyses were also done in the  $\text{CH}_4$  region (2900  $\text{cm}^{-1}$ ) to check its presence. The network vibrational modes were characterized in the 200-1200  $\text{cm}^{-1}$  range and the OH stretching band in the 2700-3800  $\text{cm}^{-1}$  range. The spectral frequency position was measured with the emission lines of Ne- and Hg-lamps. The accuracy stays within  $\pm 1 \text{ cm}^{-1}$ .

The acquisition time for the different phases (glass and fluid inclusions) is relatively short and is usually within 20-30 s to obtain a reasonably good signal to noise ratio (S/N). All spectra are normalised to the same acquisition time for relevant quantification.

### 2.3. Micro-FTIR:

Doubly polished plates were prepared from selected glass sections for micro-FTIR analyses in order to quantify the  $\text{H}_2\text{O}^{\text{tot}}$  and  $\text{CO}_2^{\text{tot}}$  contents within the glass. Mid- and Near-IR spectra (between 1000 and 6000  $\text{cm}^{-1}$ ) were obtained for 100x100  $\mu\text{m}$  spot size using a Continuum<sup>®</sup> microscope attached to a Nicolet<sup>®</sup> 5700 Fourier transform infrared spectrometer. A KBr beamsplitter and MCT-B detector were used to collect 128 scans (for sample and background) at a resolution of 4  $\text{cm}^{-1}$ . The atmospheric contribution (in particular water) was reduced by enclosing the sample in a box purged with dry air during at least 20 minutes before analysis. Prior to sample acquisition, an instrumental background was taken to subtract the atmospheric contribution.

Several spectra (>3) are taken on different locations in the glass plate and results show that  $\text{H}_2\text{O}$ - $\text{CO}_2$  are evenly distributed across the sample. A volatile-free FTIR spectrum was taken on a dry glass synthesized at 1400°C, 1 atm. and was subtracted from each

volatile-bearing samples spectrum. Thickness measurements are done with a Mitutoyo<sup>®</sup> digitometer with an accuracy of 1  $\mu\text{m}$ . All spectra are then normalized to 100  $\mu\text{m}$  for quantitative analysis.

#### 2.4. $^{13}\text{C}$ -MAS NMR:

Solid State  $^{13}\text{C}$ -MAS NMR experiments of the partially crushed samples were acquired on a 500 MHz Bruker Avance ( $B_0 = 11.7\text{ T}$ ). A 4 mm Triple channel CP/MAS “Double Bearing” probe was employed and the samples loaded in a  $\text{ZrO}_2$  rotor with Teflon end-cap. Rotors were spun at 10 kHz. The measurements were made at the  $^{13}\text{C}$  Larmor frequency (125.75 MHz). All spectra are referenced against TMS (Tetramethylsilane). Table 2 gives the NMR analytical conditions which we used to analyse our glass samples. For all samples,  $^{13}\text{C}$ -MAS NMR spectra were acquired with single  $\pi/4$  excitation of 2  $\mu\text{s}$  (SPE spectra). For several samples,  $^{13}\text{C}$ -MAS NMR spectra were also acquired by using a rotor-synchronised spin-echo sequence ( $\pi/2 - \tau - \pi - \tau - \text{acq.}$ ) with a  $\pi/2$  pulse of 4  $\mu\text{s}$  and  $\tau$  equal to one rotor period. The second pulse rephases the spin system (Hahn, 1950) due to dephasing of the signal because of external heterogeneity in the magnetic field and dipolar interactions (Taylor, 2004; Xue and Kanzaki, 2004). Thus the loss of collected signal due to the instrumental dead time is reduced. The rotor synchronised echo sequence has also the advantage to suppress probe background signal (Zeng et al., 1999). However, the probe signal does not perturb the  $^{13}\text{C}$  signal acquisition as the concentration in  $^{13}\text{C}$  atoms in the probe is relatively low (rotor  $\text{ZrO}_2$ , stator: BN). These 2 kinds of spectra will be referred hereafter as Direct MAS spectra (see Table 2).

$^{13}\text{C}$ -species can have long spin lattice relaxation time in silicate glasses (several minutes, Kohn et al., 1991). We tested several recycle time D1 (2 to 30 s) values to check for the consistency in the response of the C-species. No noticeable difference in term of line shape and chemical shift signatures was observed (only the peak absolute intensity will change) in the resulting spectra suggesting that all  $^{13}\text{C}$ -species have similar spin-lattice relaxation behaviour. Similar observation was established by Maekawa et al. (1991) for the analysis of  $^{29}\text{Si}$  which is recognized to possess a very long spin-lattice relaxation time.  $\{^1\text{H}\}$   $^{13}\text{C}$ -CPMAS NMR sequence was also employed to investigate  $^1\text{H}$  and  $^{13}\text{C}$  proximity and spatial relationship in the glass (Schmidt et al., 2000; Puls and Eckert, 2006). This method is selective but non quantitative. As mentioned in Robert et al. (2001), CPMAS involves 3 steps: 1) excitation of  $^1\text{H}$ , 2) contact time (*ct*) and 3) acquisition of the  $^{13}\text{C}$  spectrum. The radio-frequency conditions (the so-called Hartmann-Hahn conditions) applied during *ct* (a few ms) induce a through-space transfer of magnetization from the  $^1\text{H}$  to the  $^{13}\text{C}$ . The gain of sensitivity and signal is therefore a function of the time length and the dipolar coupling in between  $^1\text{H}$  and  $^{13}\text{C}$ . The transfer of magnetization is efficient from  $^1\text{H}$  considered as an abundant nucleus compared to  $^{13}\text{C}$  which is less abundant. Structural information can be obtained by comparing CP and non-CP spectra. Considering that the signal of  $^{13}\text{C}$  environment in the proximity of  $^1\text{H}$  will be enhanced, spatial structural information can be obtained on the  $^{13}\text{C}$ - $^1\text{H}$  relationship within the melt structure.

We applied several *ct* time length for investigating the magnetization transfer dynamic. As shown in Table 2, the *ct* time has been varied between 0.5 to 3.5 ms for two samples (7-1-*red* and 7-2). Dynamic CPMAS has proven very useful to probe the hydrogen

proximity to other nuclei (e.g. Xue and Kanzaki, 2004; Cody et al., 2005). At short  $ct$  (0.5 ms), only very close  $^1\text{H}$  atoms to structural  $^{13}\text{C}$  units will be enhanced by the transfer of magnetization whereas structural units with spatially far apart  $^1\text{H}$  atoms will be decrease in intensity. At longer  $ct$  (3.5 ms), all  $^{13}\text{C}$  structural units will undergo the influence of all  $^1\text{H}$  atoms in their vicinity. During CPMAS experiment, the proton repetition time  $D1 = 2$  s was sufficient for total relaxation before applying the next pulse sequence.

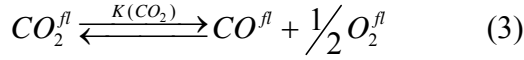
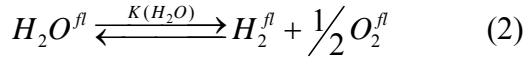
$^1\text{H}$  decoupling during acquisition (SPINAL64 scheme with a rf field of 60 kHz, Fung et al., 2000) was used to prevent from broadening of the  $^{13}\text{C}$  lines due to heteronuclear interactions with protons (Laws et al., 2002).

We collected a large number of scans for obtaining a good S/N ratio with such a low  $^{13}\text{C}$  abundances within the melt ( $\sim 0.1$  wt.%  $^{13}\text{CO}_2$  with  $^{13}\text{C}/^{12}\text{C} = 0.8$  at the best). Thus, depending on  $D1$  values, between 2,000 and 36,000 scans have been co-added. The obtained free-induction-decay signal was treated with a low line broadening exponential function (50 Hz) preventing from loss of structural information (in particular for sharp peak signature) but decreasing the S/N ratio (6-1-ox is one typical example for low S/N ratio).

### **3. Methods for $\log f\text{O}_2$ calculation:**

Two methods were employed to determine the  $\log f\text{O}_2$  experienced by the samples: 1) from  $\text{H}_2\text{O}^{\text{tot}}$  dissolved in the melt and 2) from fluid inclusion composition in the system  $\text{CO}_2\text{-CO}$ . Both methods are fully described in Morizet et al. (2009) and the reader is redirected to that publication for thorough details. The calculation method is based on thermochemical equilibrium involving reactions for  $\text{H}_2\text{O}$  and  $\text{CO}_2$  dissociation:





for each reaction the equilibrium constant is defined as follow:

$$K(H_2O) = \frac{fH_2 \times fO_2^{1/2}}{fH_2O} \quad (4)$$

$$K(CO_2) = \frac{fCO \times fO_2^{1/2}}{fCO_2} = \frac{X(CO)\phi(CO) \times fO_2^{1/2}}{X(CO_2)\phi(CO_2)} \quad (5)$$

Where in Eq. (4) and (5),  $f_i$  is the fugacity,  $X_i$  is the molar fraction and  $\Phi_i$  is the fugacity coefficient for the  $i$  fluid species.

Both methods require constraining the fluid phase composition. In Eq. 4, the  $fH_2$  is determined from experimental conditions (see above);  $K(H_2O)$  is calculated using the thermodynamic modeling from Shi and Saxena (1992); and  $fH_2O$  in the fluid phase is derived from the model of Dixon et al. (1995) which relates the total  $H_2O$  dissolved within the glass to the  $H_2O$  fugacity in the fluid.

In Eq. 5, we used CORK equation from Holland and Powell (1991) to calculate the  $\Phi$ , the fugacity coefficients of the different species. The  $K(CO_2)$  was calculated from the thermodynamic model of Shi and Saxena (1992). The molar fraction ( $X$ ) of the fluid

species was determined from Micro-Raman analysis of the fluid inclusions and using the calibration curve from Morizet et al. (2009).

Results on the fluid phase composition are given in Table 1. The actual  $X(\text{CO}_2)$  and  $X(\text{CO})$  values measured by micro-Raman are calculated taking into account the presence of  $\text{H}_2\text{O}$  within the fluid phase and according to Eq. (6):

$$X(\text{H}_2\text{O}) + X(\text{CO}_2) + X(\text{CO}) = 1 \quad (6)$$

In this equation, the  $X(\text{H}_2)$  is neglected as under the moderately investigated reducing conditions, this species will not be present in significant quantity. Values for  $X(\text{H}_2\text{O})$  were obtained from the thermodynamic equilibrium of Eq. (7):

$$X(\text{H}_2\text{O}) = \frac{f(\text{H}_2\text{O})}{P_{\text{tot}} \cdot \phi(\text{H}_2\text{O})} \quad (7)$$

As can be seen from Table 1, the  $X(\text{H}_2\text{O})$  makes up a considerable part of the fluid inclusion composition and has been included into the thermodynamic calculation in Eq. 5.

The  $\log f\text{O}_2$  determined from the two methods (either fluid inclusion analysis or  $\text{H}_2\text{O}$  dissolved in the melt) compare well to each other as shown in Table 1 and Figure 1. In Figure 1, the  $\log f\text{O}_2$  (1) represents the value determined from  $\text{H}_2\text{O}^{\text{tot}}$  dissolved in the melt and the  $\log f\text{O}_2$  (2) is the value determined from the  $X(\text{CO}_2)$ - $X(\text{CO})$  in the fluid inclusions. We observe a good agreement between the two approaches. The error on the

$\log fO_2$  was calculated considering 1) an upper and lower value for  $X(CO_2)$ - $X(CO)$  for fluid inclusions analyses and 2) an upper and lower value for  $H_2O^{tot}$  dissolved in the glass (see Table 3). The  $\log fO_2$  (2) value shows the least variation consistent with the small variation of fluid inclusions compositions. The maximum error for  $\log fO_2$  (2) is for 3-1 with  $-8.03 \pm 0.24$ . The  $\log fO_2$  (1) variation is slightly higher with a maximum error for 7-1-red with  $-10.29 \pm 0.38$ .

For the remaining of the paper and because we do not consider one value more representative than the other, we will refer to an intermediate value between  $\log fO_2$  (1) and (2). Values relative to  $\Delta FMQ$  will be used. Calculations relative to solid buffer were accomplished with the model of Zolotov and Fegley (1999) and with the thermodynamic dataset from Robie and Hemingway (1995).

$\log fO_2$  values experienced in our study cover a range from oxidized for 6-1-ox at  $\Delta FMQ+2.6$  to reduced conditions for 7-1-red at  $\Delta FMQ-2.6$ . Those values can be directly applied to modern Earth interior degassing (e.g. Kadik et al., 2004; Behrens and Gaillard, 2006) but also to Mars interior degassing (Hirschmann and Withers, 2008).

## **4. Results:**

### 4.1 Micro-Raman spectral feature:

#### 4.1.1. Fluid inclusion analysis:

From Micro-Raman analyses, we identified 3 main species present within the fluid inclusions:  $CO_2^{fl}$ ,  $CO^{fl}$  and  $CH_4^{fl}$  ( $H_2O^{fl}$  is also present but not quantified by Micro-Raman). Micro-Raman spectra obtained from fluid inclusions and for 4 different samples are shown in Figure 2. The Raman shift is separated in several windows showing the

gaseous vibrational signature of interest: Figure 2A for CO<sub>2</sub><sup>fl</sup>; Figure 2B for CO<sup>fl</sup> and Figure 2C for CH<sub>4</sub><sup>fl</sup>. We applied an intensity scaling factor of 2 in Figure 2B and 4 in Figure 2C in order to increase the size of the CO peaks and CH<sub>4</sub> peaks. The CO<sub>2</sub> vibration consists in 2 modes: 2ν<sub>2</sub> and ν<sub>1</sub>; those signatures are located at ~1385, 1285 cm<sup>-1</sup> for <sup>12</sup>CO<sub>2</sub> and ~1370, 1260 cm<sup>-1</sup> for <sup>13</sup>CO<sub>2</sub>. The CO vibration has only 1 mode located at 2138 cm<sup>-1</sup> for <sup>12</sup>CO and 2093 cm<sup>-1</sup> for <sup>13</sup>CO. A small quantity of CH<sub>4</sub> has been identified in the most reduced samples and is shown in Figure 2C with a signature at 2909 cm<sup>-1</sup> for <sup>12</sup>CH<sub>4</sub> and 2914 cm<sup>-1</sup> for <sup>13</sup>CH<sub>4</sub>.

From the fluid inclusion analysis, we can see that in most of the cases both <sup>12</sup>C and <sup>13</sup>C isotopes are present (except for 2-2-<sup>12</sup>C ΔFMQ-2.0 which is a <sup>12</sup>C natural sample). This is consistent with the incomplete decarbonation of the starting material. We have calculated the <sup>13</sup>C enrichment following the method Morizet et al. (2009). The values for the %<sup>13</sup>C within the fluid are reported alongside the spectra in Figure 2A and for all studied samples in Table 1.

The method to determine the concentration of each species in the fluid inclusions is fully described in Morizet et al. (2009). The area of each peak is directly proportional to the concentration of the species. For CO<sub>2</sub> the best calibration was obtained when the 2ν<sub>2</sub> mode was considered. We only use this mode in the X(CO<sub>2</sub>) calculation reported in Table 1 and Figure 2B. The error on the measured molar fraction was calculated from standard deviation based on replicated analyses on different fluid inclusions (>3) within each sample. The maximum error is observed for 4-1 with X(CO<sub>2</sub>) = 0.628 ± 0.049 which is less than 10%. The reported error is consistent with error given in Morizet et al. (2009) and suggests that fluid inclusions are well-equilibrated across the sample showing little

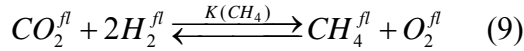
variation in composition. The X(CO<sub>2</sub>) values shown in Figure 2B are the corrected values which take into account the presence of H<sub>2</sub>O within the fluid inclusions (see Table 1).

Thus, X(CO<sub>2</sub>) is roughly constant (~0.52) for 7-2 (ΔFMQ-1.2), 2-2-<sup>12</sup>C (ΔFMQ-2.0) and 7-1-*red* (ΔFMQ-2.6) but the X(CO) is increasing progressively from 0.11 to 0.41 as shown in Table 1 accompanying a decrease in *f*O<sub>2</sub> conditions from ΔFMQ-1.2 to -2.6.

A peak at 2900 cm<sup>-1</sup> and corresponding to CH<sub>4</sub> is observed in the most reduced sample at ΔFMQ-2.0 and ΔFMQ-2.6 and is totally absent for the most oxidised samples. CH<sub>4</sub> only corresponds to a few percent compared to the other species. We choose to compare CH<sub>4</sub> and CO to derive the %CH<sub>4</sub> as follow:

$$\frac{\%CH_4}{\%CO} = \frac{A_{CH_4}}{\zeta\sigma_{CH_4}} \times \frac{\zeta\sigma_{CO}}{A_{CO}} \quad (8)$$

In Eq. (8), A is the area of the peak species, σ is the cross-section factor for the species and ζ is the efficiency factor which is defined for a given spectrometer. We have quantified the %CH<sub>4</sub> reported in Figure 2C using 7.5 as the cross-section calibration factor (Burke, 2001). The ratio σCH<sub>4</sub>/σCO between the cross-section factors is roughly ~8, hence for similar CH<sub>4</sub> and CO peak area the concentration of CH<sub>4</sub> will be 8 times lower than CO. The resulting %CH<sub>4</sub> is small. In the most reduced sample 7-1-*red*, the %CH<sub>4</sub> estimated is 3% and thus the molar fraction of CH<sub>4</sub> within the fluid phase is roughly 0.03 at best. The CH<sub>4</sub> can be related to other species by the following equilibrium:



We calculated the  $fO_2$  from this reaction and for the most reduced sample in which the  $CH_4$  concentration is the most significant. The equilibrium constant for reaction (9) reads as follow:

$$K(CH_4) = \frac{fCH_4 \times fO_2}{fCO_2 \times P(H_2)^2} = \frac{X(CH_4)\phi(CH_4) \times fO_2}{X(CO_2)\phi(CO_2) \times P(H_2)^2} \quad (10)$$

Using  $K(CH_4)$  from the thermodynamic model of Shi and Saxena (1992),  $\Phi$  from the CORK equation of Holland and Powell (1991),  $P(H_2)$  from the experimental conditions (7-1-*red* is equal to 68 bars), and  $X(CO_2) = 0.53$ , give an  $fO_2$  equal to  $\Delta FMQ-3.8$ . This is considerably lower than the values calculated from the two other methods ( $\Delta FMQ-2.6$ , Table 1). To increase the  $fO_2$ , would require to decrease the concentration of  $CH_4$ . For instance a calculation done with  $X(CH_4) = 0.004$  (7-2 sample) gives an  $fO_2$  of  $\Delta FMQ-2.9$  compared to  $\Delta FMQ-1.2$  reported in Table 1 and calculated from the other methods. Another source of error comes from the experimental  $P(H_2)$  which may be slightly different from the one given in Table 1. Anyhow, it appears that the observed  $CH_4$  in fluid inclusions is difficult to quantify and therefore is not suitable for  $fO_2$  calculation.

#### 4.1.2. $H_2O^{tot}$ quantification:

We tested the possibility to determine the  $H_2O^{tot}$  through Raman acquisition and following the method described by Behrens et al. (2006). We show in Figure 3 the Micro-

Raman spectra of different glasses between 200-1250  $\text{cm}^{-1}$  and 2700-3900  $\text{cm}^{-1}$ . Several vibrational signatures are observed within each spectrum. The broad asymmetric peak with maximum at 3500  $\text{cm}^{-1}$  is attributed to OH stretching (e.g. Ihinger et al., 1994; Thomas, 2000; Zajacz et al., 2005; Mercier et al., 2009). Two pronounced signatures are observed for the glass structure. Bands at 500 and 1000  $\text{cm}^{-1}$  are assigned to T-O-T bending and rocking and T-O stretching respectively (Frantz and Mysen, 1995; Behrens et al., 2006). The 1000  $\text{cm}^{-1}$  high intensity is consistent with basaltic composition as mentioned by Behrens et al. (2006). More silicic composition will show a more intense vibrational signature at 500  $\text{cm}^{-1}$  compared to T-O stretching at 1000  $\text{cm}^{-1}$ .

A typical non-linear baseline is represented underneath the spectra and was used to derive the peak area for integration. The baseline is arbitrary however it mimics the continuum of the Raman spectra adequately. Moreover, the good agreement with the FTIR quantification (see Figure 9 and Table 3) suggests that our approach is satisfactory. Spectra shown here are not corrected for black-body radiation (Long, 2002). The correction procedure does not seem to improve the baseline in the case of our composition, although it tends to complicate the positioning of the non-linear baseline for water quantification. The water content derived from this method ( $\text{H}_2\text{O}^{\text{tot}}$  (2)) is given next to each spectrum and in Table 3 for all samples. The results obtained using both methods FTIR and Micro-Raman will be further discussed later in section 4.4.

#### 4.2 FTIR spectral features:

##### 4.2.1. Carbonate environment:

Carbonate species ( $\text{CO}_3^{2-}$ ) are represented in Figure 4 by a complex “doublet” assigned to the carbonate antisymmetric stretch ( $\nu_3$ ) located between 1300 and 1600  $\text{cm}^{-1}$  consistent with the description given by Fine and Stolper (1986). Only carbonate units are present in our haplobasaltic composition. No evidence of  $\text{CO}_2^{\text{mol}}$  at 2350  $\text{cm}^{-1}$  (Brooker et al., 1999; Behrens et al., 2004) has been identified and also consistent with NMR results. This observation is in agreement with currently known  $\text{CO}_2$  dissolution mechanisms:  $\text{CO}_2$  speciation changes from  $\text{CO}_2^{\text{mol}}$  towards  $\text{CO}_3^{2-}$  units progressively from rhyolitic to basaltic composition (Blank and Brooker, 1994; King and Holloway, 2002; Behrens et al., 2004). However, it should be mentioned that the speciation observed is the  $\text{CO}_2$  speciation preserved at the glass transition temperature ( $T_g$ ) as demonstrated by recent work (e.g. Morizet et al., 2001; Nowak et al., 2003; Spickenbom et al., 2010). Therefore, the measured  $\text{CO}_2$  speciation does not represent the true  $\text{CO}_2$  speciation present in the melting conditions at which  $\text{CO}_2^{\text{mol}}$  is expected.

No vibrationnal mode for C-O stretching is seen around 2150  $\text{cm}^{-1}$  which would have attested for the presence of  $\text{CO}^{\text{mol}}$  (Brooker et al., 1999). The absence of  $\text{CO}^{\text{mol}}$  is consistent with previous findings from Pawley et al. (1992) who suggested that the  $\text{CO}^{\text{mol}}$  species is not soluble in basaltic glass at least under its molecular form. However, it is not excluded that  $\text{CO}^{\text{mol}}$  would be present under those reduced conditions at high temperature within the melt consistent with C speciation behaviour above  $T_g$ . The signature visible at 1630  $\text{cm}^{-1}$  is attributed to water vibration and corresponds to total  $\text{H}_2\text{O}$  bending mode (Ihinger et al., 1994).

We have stated that due to incomplete decarbonation the synthesized glasses have different  $^{13}\text{C}$  enrichment. We show the FTIR spectra for ~80%, 40% and 1%  $^{13}\text{C}$



enrichment within the fluid phase in Figure 4A, B and C, respectively. In all cases, the carbonate peak maxima are shifted progressively to higher frequency when the %<sup>13</sup>C decreases. In Figure 4A (80% <sup>13</sup>C in fluid) the peak maxima are located at ~1404 and ~1482 cm<sup>-1</sup> giving a centre point at 1443 cm<sup>-1</sup> and a  $\Delta\nu_3 = 78$  cm<sup>-1</sup>. In Figure 4C (1% <sup>13</sup>C in fluid) the peak maxima are displaced at ~1438 and 1520 cm<sup>-1</sup> with a centre point at 1479 cm<sup>-1</sup> and  $\Delta\nu_3 = 82$  cm<sup>-1</sup>. The shift in peak position is consistent with the change in isotopic enrichment in <sup>13</sup>C. According to Hooke's law, it is possible to predict the peak position ( $\nu$  in cm<sup>-1</sup>) of a vibration as a function of the reduced mass (Ihinger et al., 1994). The  $\nu$  value is inversely proportional to the square root of the reduced mass of the atoms implied into the vibration. As such, a vibration involving a <sup>13</sup>C atom will have a lower frequency position compared to the same vibration with a <sup>12</sup>C atom involved which is in agreement with the results shown in Figure 4. We estimated the change in reduced mass between <sup>12</sup>C and <sup>13</sup>C for a C-O vibration. We obtain a scaling factor (<sup>13</sup>C/<sup>12</sup>C  $\nu$ (C-O)) of 1.0460 and its square root 1.0227. The value of 1.0227 is directly correlated to the frequency shift between the two isotopes. It is in excellent agreement with the change in peak position between <sup>13</sup>C-enriched and -poor samples observed in Figure 4: 1438 / 1404 ~ 1.0242; 1520 / 1482 ~ 1.0256.

For a given enrichment, the peak maximum does not change as reduction of the samples proceeds as indicated by the dashed lines in Figure 4. Similarly, the  $\Delta\nu_3$  remains constant and oscillates between ~65 and 82 cm<sup>-1</sup> for 40% and 1% <sup>13</sup>C respectively. The lowest  $\Delta\nu_3$  value for 40% <sup>13</sup>C reflects the fact that those samples are a mixture of both <sup>12</sup>C-<sup>13</sup>C isotopes and therefore the peak positions are intermediate in between the one observed for <sup>13</sup>C-poor glasses (Figure 4C, 1% <sup>13</sup>C) and <sup>13</sup>C-rich glasses (Figure 4A, 80% <sup>13</sup>C). The

very small value for  $\Delta\nu_3$  suggests a low distortion from the ideal trigonal planar symmetry ( $D_{3h}$ ) (Brooker et al., 2001b). Such a value is often observed for natural basaltic or similar compositions (Fine and Stolper, 1986; Taylor, 1990).

#### 4.2.2. Water band assignment:

Infrared spectra in Figure 5A show that  $H_2O$  dissolves in the melt as two main species: i) hydroxyl unit ( $OH^-$ ) and ii) molecular water ( $H_2O^{mol}$ ). The peak located at  $4500\text{ cm}^{-1}$  is attributed to OH groups and peak at  $5200\text{ cm}^{-1}$  is assigned to  $H_2O^{mol}$  as suggested by previous studies (e.g. Stolper, 1982a; King et al., 2002; Stuke et al., 2006). An additional signature is given by the fundamental OH stretching vibration band at  $3500\text{ cm}^{-1}$  and represented for several spectra in Figure 5B (except for sample 7-2 for which no spectrum was resolved in this frequency range due to the high water content saturating the FTIR detector). For the clarity of Figure 5 we did not represent the whole set of glass spectra as it turn out to present the same line shape for all and only differs in the  $H_2O^{tot}$  measured within the glass.

We fitted a baseline (dashed line) to mimic the background for the FTIR continuum underneath the OH and  $H_2O^{mol}$  peaks in Figure 5A and the peak for  $H_2O^{tot}$  in Figure 5B. In most cases, this procedure is easily performed and the integration of the peaks for quantification is straightforward. This is true in particular for the peak at  $3500\text{ cm}^{-1}$  for which a tangential linear baseline has been fitted as shown for 6-1-ox, 4-1 and 7-1-red. The quantification of each individual species (OH and  $H_2O^{mol}$  peaks) is sometime complicated by the presence of interference lines which are due to the low thickness ( $<60\text{ }\mu\text{m}$ ) of the glass doubly polished plate.

The different concentrations ( $\text{OH}$ ,  $\text{H}_2\text{O}^{\text{mol}}$  and  $\text{H}_2\text{O}^{\text{tot}}$ ) as well as the  $\Delta\text{FMQ}$  value are indicated in Figure 5 and listed in Table 3. At low water content ( $< 2 \text{ wt.}\% \text{ H}_2\text{O}^{\text{tot}}$ ), the water speciation is mainly represented by the  $\text{OH}$  groups at  $4500 \text{ cm}^{-1}$ . With further increase in  $\text{H}_2\text{O}^{\text{tot}}$  as for 6-1-ox and 7-2, there is a progressive increase in the peak intensity for  $\text{H}_2\text{O}^{\text{mol}}$  vibration. This behavior is in agreement with known  $\text{H}_2\text{O}$  speciation trends showing a crossover between the two species at  $\sim 4 \text{ wt.}\% \text{ H}_2\text{O}^{\text{tot}}$  (e.g. Dixon et al., 1995; Withers and Behrens, 1999; Botcharnikov et al., 2006).

### 4.3 $^{13}\text{C}$ -MAS NMR:

#### 4.3.1. Carbonates environments from Direct $^{13}\text{C}$ MAS NMR:

$^{13}\text{C}$  MAS NMR spectra for several samples (with increasing  $\log f\text{O}_2$ ) are represented in Figure 6. The  $\log f\text{O}_2$  relative to  $\Delta\text{FMQ}$  and the ppm  $\text{CO}_2$  is also indicated next to each spectrum.

Several species are identified from NMR spectra.  $^{13}\text{CO}_2$  and  $^{13}\text{CO}$  in fluid are observed and are located 125 and 183.5 ppm respectively (Kohn et al., 1991; Morizet et al., 2009). Those species are entrapped within fluid inclusions. The sharp peaks are consistent with free motion for “gaseous” molecules and therefore cannot be attributed to dissolved species within the glass. When present, fluid inclusions are mainly constituted by  $^{13}\text{CO}_2$  and, for the most reduced sample, by additional  $^{13}\text{CO}$ . No signature for  $^{13}\text{CH}_4$  at 0 ppm (Ripmeester and Ratcliffe, 1988; Subramanian et al., 2000) is seen in the different spectra which is consistent with the low concentration of  $\text{CH}_4$  measured by Micro-Raman analysis ( $< 3\% \text{ CH}_4$ ).

Dissolved C species are represented by carbonate units with signature located at chemical shift between ~160 and 170 ppm. The peak does not show a significant asymmetric shape and has its maximum at ~167 ppm regardless of the redox conditions.

There is no evidence for other species than carbonates to be dissolved within the glass. NMR results are consistent with the FTIR measurements showing only the carbonate peak doublet. The  $^{13}\text{C}$  NMR spectra do not reveal the presence of more reduced species such as  $\text{CH}_3$  at ~15 ppm (Khimyak and Klinowski, 2001). Mysen et al. (2009) suggested the presence of methyl groups dissolved within  $\text{Na}_2\text{O-SiO}_2$  glasses synthesized under high pressure (1.0-2.5 GPa) and highly reducing conditions (IW buffer). The formation of  $\text{CH}_3$  groups thus seems to be restricted to severe reducing conditions and is not possible under our moderate reducing conditions.

Several samples with different redox conditions were analyzed. In Figure 6  $^{13}\text{C}$  MAS NMR spectra for oxidized ( $\Delta\text{FMQ}+2.6$ ) and reduced ( $\Delta\text{FMQ}-0.1$  to  $-2.6$ ) glasses are almost identical in lineshape. It appears there is a slight broadening of the 6-1-ox peak line shape which is attributed to the high concentration of paramagnetic impurities (1.17 wt.%  $\text{FeO}^{\text{tot}}$ ). NMR and FTIR spectra give consistent results as no dramatic change is observed in the carbonate environments as a function of  $f\text{O}_2$ .

$^{13}\text{C}$  NMR spectra obtained for haplophonolite (Morizet et al., 2002) and sodamelilite (Kohn et al., 1991) have been added for comparison with our haplobasaltic glass samples. Both spectra have a better S/N ratio compared to our glass spectra. The phonolite and sodamelilite glasses from Morizet et al. (2002) and Kohn et al. (1991) were synthesized at high pressure, 2.0 and 1.0 GPa respectively and under moderately oxidizing conditions; hence the total  $\text{CO}_2$  solubility is very high (several wt.%).

Our  $^{13}\text{C}$ -NMR spectra for haplobasaltic glasses is similar to haplophonolite spectrum from Morizet et al. (2002). The peak maximum is identical at 167 ppm; the full width at half maximum (FWHM) is almost equal with 7 and 8 ppm for the basalt and phonolite respectively. One major difference is the clear asymmetric shape observed in haplophonolite glass at lower frequency suggesting the existence of several carbonate units.

Our haplobasaltic glass is different from sodamelilite glass spectrum obtained by Kohn et al. (1991). For sodamelilite, the peak maximum is slightly shifted to higher frequency at 168 ppm, the peak is symmetrical (one unique carbonate environment) but has a lower FWHM of  $\sim 4.5$  ppm.

For haplobasaltic glass, the poor S/N ratio renders difficult the interpretation of the spectra. However, the results obtained by  $\{^1\text{H}\}^{13}\text{C}$  CPMAS experiments will comfort the existence of several carbonate groups within our haplo-basaltic glasses.

#### 4.3.2. Carbon and hydrogen proximity revealed from $\{^1\text{H}\}^{13}\text{C}$ CPMAS

##### NMR:

We performed  $\{^1\text{H}\}^{13}\text{C}$  CPMAS to investigate the possibility for the carbonate environments to be spatially close to hydrogen atoms.  $\{^1\text{H}\}^{13}\text{C}$ -CPMAS NMR spectra ( $ct = 3.5$  ms) are shown in Figure 7 for 5 different samples (see Table 2).

In Figure 7 the  $\Delta\text{FMQ}$  and the  $\text{H}_2\text{O}^{\text{tot}}$  are indicated for the different samples.

Unfortunately, we did not obtain a reliable signal from the most oxidized sample, 6-1-ox ( $\Delta\text{FMQ}+2.6$ ) possibly due to the high  $\text{FeO}^{\text{tot}}$  of this sample.

The line shape of the spectra is varying in between the samples. A peak at 167 ppm is observed for all samples consistent with  $^{13}\text{C}$  Direct MAS NMR spectra (see Figure 8). A second component is appearing in, 4-1 and 7-1-*red* and possibly 7-2 at 177 ppm. This component is invisible in Direct MAS spectra as shown in Figure 8. The corresponding carbonate group is probably below the detection limit and can only be observed when the  $^{13}\text{C}$  signal is enhanced by the transfer of magnetization from neighboring hydrogen atoms. The other carbonate group is always present and contributes to the majority of the carbonate environments within the glass. Although the concentration of the 177 ppm component is low, it cannot be ignored and is clearly implied in the reduction process of the basaltic liquid during the experiment.

The 177 ppm component is not correlated to the  $\text{H}_2\text{O}^{\text{tot}}$  as observed from the spectra reported in Figure 7. On the other hand, we can see that its appearance or intensity for a given contact time in 4-1 and 7-1-*red* seems to be directly correlated to the degree of reduction experienced by the samples. When reduction is maximum (at high  $f\text{H}_2$ ), the additional component is revealed in 4-1 and even more in 7-1-*red* by CPMAS.

The advantage of  $\{^1\text{H}\}^{13}\text{C}$  CPMAS is to provide indication on the  $^1\text{H}$  environment of the different  $^{13}\text{C}$  units. The fact that 177 ppm appears in CPMAS whereas it is totally absent in Direct MAS suggests that this carbonate unit possesses a higher hydrogenated (higher OH concentrations) environment than the carbonate groups at 167 ppm.

We made the comparison in Figure 8 between  $^{13}\text{C}$  Direct MAS and  $\{^1\text{H}\}^{13}\text{C}$  CPMAS spectra acquired with different  $ct = 0.5, 1.0$  and  $3.5$  ms. The spectra are plotted normalized to the same number of scans (8,192) so one can easily compare the change in intensity of the different unit depending on  $ct$ . The dynamic  $\{^1\text{H}\}^{13}\text{C}$  CPMAS reveals

that the carbonate group at 177 ppm is preserved and increases as the  $ct$  is decreased from 3.5 to 0.5 ms. Conversely, the peak at 167 ppm decreases in intensity as the  $ct$  is decreased from 3.5 to 0.5 ms. This result implies that hydrogen atoms in the 177 ppm carbonate group are spatially closer than for the 167 ppm carbonate environment. We can also see that the 177 ppm component has a maximum intensity at 0.5 ms in the case of 7-2 suggesting a strong dipolar coupling between hydrogen atoms which could be explained by the higher water content (3.66 wt.% H<sub>2</sub>O).

#### 4.4 CO<sub>2</sub> and H<sub>2</sub>O concentrations measured in glasses:

H<sub>2</sub>O and CO<sub>2</sub> quantifications from FTIR were achieved using Beer-Lambert law (e.g. Ihinger et al., 1994; King et al., 2002; Morizet et al., 2002; Okomura et al., 2003).

$$C_i = \frac{Abs_i \cdot MM_i}{\rho \cdot d \cdot \varepsilon_i} \quad (11)$$

where  $C_i$ , the concentration of  $i$ -species, is a function of the integrated absorbance of the peak vibration ( $Abs_i$ ) times the molecular mass ( $MM_i$ , 18.02 and 44.01 g.mol<sup>-1</sup> for H<sub>2</sub>O and CO<sub>2</sub> respectively) and divided by the product between the density ( $\rho$ , g.L<sup>-1</sup>), the thickness of the sample ( $d$ , cm) and the integrated molar extinction coefficient for the considered specie ( $\varepsilon_i$ , L.mol<sup>-1</sup>.cm<sup>-2</sup>). A density value of 2537 g.L<sup>-1</sup> was used and determined from the Lange and Carmichael (1987) density model. This density value does not take into account the potential effect of H<sub>2</sub>O on the glass density (Ohlhorst et al.,

2001), however in the range of water content studied here (up to 4 wt.% H<sub>2</sub>O) the expected change in density is roughly ~2.5% of the value and therefore will be neglected.

#### 4.4.1. H<sub>2</sub>O concentration determined from Micro-FTIR and Micro-Raman:

The H<sub>2</sub>O<sup>tot</sup> was determined using either the H<sub>2</sub>O<sup>tot</sup> peak at 3500 cm<sup>-1</sup> (for most of the samples) or the sum of the OH + H<sub>2</sub>O<sup>mol</sup> contribution at 4500 and 5200 cm<sup>-1</sup> respectively (see Table 3). FTIR quantification can only be achieved if suitable values for  $\epsilon$  are available; however those values are depending on glass composition (Ohlhorst et al., 2002; Mandeville et al., 2002).

For OH and H<sub>2</sub>O<sup>mol</sup>, we used an  $\epsilon$  value derived by Stolper (1982) and valid for a basaltic composition: 200 L.mol<sup>-1</sup>.cm<sup>-2</sup> for OH groups and 300 L.mol<sup>-1</sup>.cm<sup>-2</sup> for H<sub>2</sub>O<sup>mol</sup>. We summed both concentrations of OH and H<sub>2</sub>O<sup>mol</sup> to determine the total water content, H<sub>2</sub>O<sup>tot</sup> (1). The error reported in Table 3 for OH and H<sub>2</sub>O<sup>mol</sup> is obtained from the standard deviation analysis and the error on H<sub>2</sub>O<sup>tot</sup> (1) is then calculated. The maximum error for the H<sub>2</sub>O<sup>tot</sup> (1) observed is  $\pm 0.30$  wt.% H<sub>2</sub>O for 6-1-ox sample.

For the total water content with the peak located at 3500 cm<sup>-1</sup> (H<sub>2</sub>O<sup>tot</sup> (3500 cm<sup>-1</sup>), see Table 3), we tested several extinction coefficient values. H<sub>2</sub>O<sup>tot</sup> (3500 cm<sup>-1</sup>) was calculated using either linear or integrated extinction coefficient. The associated error was derived from standard deviation analysis. Using the integrated (32000 L.mol<sup>-1</sup>.cm<sup>-2</sup>) or linear value (67 L.mol<sup>-1</sup>.cm<sup>-1</sup>) given by Stolper et al. (1982) gives results in relatively good agreement with the results obtained from H<sub>2</sub>O<sup>tot</sup> (1) as the sum of OH + H<sub>2</sub>O<sup>mol</sup>. However with high water content, the agreement between the two determined values disappears (see Figure 9). It is particularly true for 7-2 and 3-1 samples for which we



obtained a high water content from  $\text{H}_2\text{O}^{\text{tot}}$  (1) ( $> 3$  wt.%) and a totally different value for  $\text{H}_2\text{O}^{\text{tot}}$  ( $3500\text{ cm}^{-1}$ ). The obvious lack of consistency for those two samples can be explained by the fact that at high water content, the quantification with the peak at  $3500\text{ cm}^{-1}$  becomes uncertain. The  $3500\text{ cm}^{-1}$  peak cannot be resolved and the intensity turned out to be too high and saturates the IR detector leading to wrong  $\text{H}_2\text{O}^{\text{tot}}$  ( $3500\text{ cm}^{-1}$ ) value. As such, we could not determine the water content with this method for sample 5-1 due to incomplete resolution of the FTIR peak at  $3500\text{ cm}^{-1}$ .

Mandeville et al. (2002) recently investigated the change in the extinction coefficient value as a function of the glass composition and for the peak at  $3500\text{ cm}^{-1}$  for water quantification. In particular, they derived a value of  $69.21\text{ L}\cdot\text{mol}^{-1}\cdot\text{cm}^{-1}$  calibrated for an Fe-free andesite composition. This value is very close to the one derived by Stolper et al. (1982). In a more general point of view, the change in the extinction coefficient especially for the peak at  $3500\text{ cm}^{-1}$  is relatively small (Mandeville et al., 2002) for the type of glass composition investigated here and the impact on the  $\text{H}_2\text{O}^{\text{tot}}$  variation is negligible as shown in Figure 9. As a result, we are confident when using the  $\epsilon$  values proposed by Stolper et al. (1982) to determine the  $\text{H}_2\text{O}^{\text{tot}}$  (1) from the sum of OH and  $\text{H}_2\text{O}^{\text{mol}}$  concentrations.

The  $\text{H}_2\text{O}^{\text{tot}}$  was also estimated using micro-Raman measurements ( $\text{H}_2\text{O}^{\text{tot}}$  (2) column in Table 3) and following the method developed by Behrens et al. (2006). The procedure has been successfully applied by those authors for various compositions (ranging from granite to basalt) and a large range of water content (from 0.5 to 11 wt.%  $\text{H}_2\text{O}^{\text{tot}}$ ).

Although the method is proposed to be suitable only for one type of spectrometer (T64000 Jobin Yvon), we show that the obtained results with our equipment (LabRam

Jobin Yvon) are in good agreement with the FTIR results suggesting that the method can also be applied for LabRam spectrometer.

The integrated area of water broad vibrational frequency at  $3500\text{ cm}^{-1}$  is normalized to the aluminosilicate vibration mode at  $1000\text{ cm}^{-1}$  (see Figure 3). The  $\text{H}_2\text{O}^{\text{tot}}$  is then obtained directly using the linear regression proposed for basaltic composition by Behrens et al. (2006, their Eq. 3). As shown in Table 3 and Figure 9, results obtained from both methods FTIR and micro-Raman are in good agreement. A typical error of  $\pm 0.20$  as given in Behrens et al. (2006) was taken.

The final retained value for  $\text{H}_2\text{O}^{\text{tot}}$  is given in the last column in Table 3 and represents an intermediate value between the  $\text{H}_2\text{O}^{\text{tot}}$  (1) and  $\text{H}_2\text{O}^{\text{tot}}$  (2) and ranges from 1.24 to 4.01 wt.%  $\text{H}_2\text{O}$ . The error associated with the values is estimated to be  $\pm 0.20$  which is i) the average error observed for  $\text{H}_2\text{O}^{\text{tot}}$  (1) and ii) the assumed error value for  $\text{H}_2\text{O}^{\text{tot}}$  (2).

Though, this error value is rather typical when estimating volatile concentrations from spectroscopic method.

#### 4.4.2. $\text{CO}_2$ concentration determined from Micro-FTIR:

Carbonate doublet were integrated for  $\text{CO}_2$  quantification. The procedure is different as compared to water quantification as there is complication due to the presence of the strong vibrational signature for  $\nu_2\text{ H}_2\text{O}$  at  $1630\text{ cm}^{-1}$ . Subsequently, we used a simulation of the spectrum with 3 Gaussian peaks: 1)  $1630$ , 2)  $1430$  and 3)  $1530\text{ cm}^{-1}$ . We fitted and subtracted a flexicurve baseline prior to the simulation. This approach is suitable for basalt-like compositions (e.g. Silver et al., 1990; Zhang et al., 1995; Olhorst et al., 2001). We used the sum for both  $\text{CO}_3^{2-}$  peaks to obtain the absolute integrated area of the

carbonate doublet. The main advantage of the simulation method is to remove the possible contribution of the  $\nu_2$  H<sub>2</sub>O tail at 1630 cm<sup>-1</sup> to the carbonate doublet. The simulation of the spectrum is not represented, yet the procedure is relatively robust with a correlation factor  $R^2 \sim 0.98$ . We applied the Beer-Lambert approximation with a value for  $\epsilon$  of 69500 L.mol<sup>-1</sup>.cm<sup>-2</sup> as determined by Fine and Stolper (1986) for natural basaltic and Ca-Mg silicate glasses. The total CO<sub>2</sub> content is given in Table 3 and ranges from 223 to 1320 ppm.

There is a clear lack of appropriate extinction coefficient data which are known to be a strong function of composition (Ihinger et al., 1994). Our studied composition is not identical to that of Fine and Stolper (1986) and the  $\epsilon(\text{CO}_3^{2-})$  for our haplobasaltic glass is expected to be slightly different from this value of 69500 L.mol<sup>-1</sup>.cm<sup>-2</sup>. Moreover, it has been applied for <sup>12</sup>C carbonate groups and may not be valid for <sup>13</sup>C isotopes carbonate band. However, there is up to now no calibration available for <sup>13</sup>CO<sub>2</sub>-bearing glasses investigated by FTIR.

#### 4.4.3. CO<sub>2</sub> versus H<sub>2</sub>O concentrations

The CO<sub>2</sub><sup>tot</sup> versus H<sub>2</sub>O<sup>tot</sup> dissolved in the quenched glasses is shown in Figure 10. As mentioned above, CO<sub>2</sub> ranges from 223 ppm for 4-2-*us* (initially CO<sub>2</sub>-free sample) to 1320 ppm for 6-1-*ox*. H<sub>2</sub>O<sup>tot</sup> is changing from ~1.24 wt.% for 1-1-200 to 4.01 wt.% for 5-1. 1-1-200 has a lower CO<sub>2</sub> content because it was synthesized at lower pressure. 4-2-*us* was initially a CO<sub>2</sub>-free sample but because of incomplete decarbonation we measured a non negligible amount of CO<sub>2</sub>, 223 ppm.

Other samples were synthesized at 300 MPa total pressure and at temperature of 1250°C. Apart from the 2 particular samples mentioned above, the  $\text{H}_2\text{O}^{\text{tot}}$  dissolved in the melt (corresponding to the sum of the  $\text{H}_2\text{O}$  initially loaded in the capsule and the  $\text{H}_2\text{O}$  formed from the  $\text{H}_2$  diffusion inside the capsule) is relatively large ranging from ~1.3 up to 4.0 wt.%, for 3-2- $^{12}\text{C}$  and 5-1 respectively. The change in  $\text{H}_2\text{O}^{\text{tot}}$  dissolved in the glass will influence Tg position (Dingwell et al., 1998). Between 1 and 4 wt.%  $\text{H}_2\text{O}$ , the expected decrease in Tg could be up to 100°C according to the model of Giordano et al. (2008). The influence of different Tg values in the present study is difficult to apprehend however it is possible that the C and H speciation could be affected by such changes. In Figure 10, we distinguish 2 groups of data points: 1) one corresponding to severe reducing conditions at roughly  $\Delta\text{FMQ} < -1.5$  and 2) one to more oxidized conditions at roughly  $\Delta\text{FMQ} > -1.2$ . One important difference noticed between the 2 groups is the water content. The first group with strongly reducing conditions has lower  $\text{H}_2\text{O}$  content whereas the second is constituted by 3 points with at least 3 wt.%  $\text{H}_2\text{O}$ . Although the reason for those to be particular is unclear, Behrens et al. (2009) suggested that high  $\text{H}_2\text{O}$  content could produce an increase in  $\text{CO}_2$  solubility.

At 300 MPa, the  $\text{CO}_2$  solubility data show a maximum at 1320 ppm for 6-1-*ox* and 681 ppm for 7-1-*red*. All data points are located within those 2 values.  $\text{CO}_2$  solubility data measured for reduced samples are systematically lower than the  $\text{CO}_2$  solubility measured for oxidized conditions for 6-1-*ox*. The change in  $\text{CO}_2$  solubility is likely to be related to the changes in fluid phase composition in response to  $f\text{O}_2$  decrease (e.g., Pawley et al., 1992).

Isobaric CO<sub>2</sub>-H<sub>2</sub>O solubility trends derived from Dixon et al. (1995) thermochemical equilibrium model at total pressure of 150, 200, 250 and 300 MPa have been represented by solid lines in Figure 10 and will be discussed in section 5.2.

## **5. Discussion:**

### 5.1 Carbonate environments assignment:

We have attempted to simulate the spectra for the carbonate environments with several Gaussian peaks for 7-1-*red* and 7-2 shown in Figure 8. The simulation was performed with 2 peaks considering that 1) the main component at 167 ppm is symmetric in shape and can be the result of only one type of carbonate unit and 2) the presence of the additional peak at 177 ppm.

The respective positions for the 2 Gaussian peaks are 167 and 176.5 ppm. The full width at half maximum (FWHM) is ~7.5 ppm for the component at 167 and ~9.3 ppm for the peak at 176.5 ppm. The assignment for both CO<sub>3</sub><sup>2-</sup> components is complicated by 1) the complex character of the studied composition (5 oxides), 2) the fact that the S/N ratio is often poor and 3) the little information available in the literature.

Infrared spectra suggest that the carbonate environments observed in the haplobasaltic glasses have a low distortion or, in other words, are very symmetrical molecules. This assertion is based on the small value for  $\Delta\nu_3$  ( $< 80 \text{ cm}^{-1}$ ). Therefore, it is unlikely that such low  $\Delta\nu_3$  would correspond to a so-called network carbonate. According to Brooker et al. (2001b), network carbonates such as T-carb.-T (with T = Si or Al) exhibit a large  $\Delta\nu_3$  ( $> 200 \text{ cm}^{-1}$ ) inconsistent with our results. Moreover such network unit has been

mainly identified in fully polymerized melts which is not the case for the studied haplobasaltic composition in which a non negligible amount of NBO is expected. Following the work of Brooker et al. (2001b), such a low  $\Delta\nu_3$  at  $80\text{ cm}^{-1}$  could be attributed to the presence of NBO-carb.  $\text{Ca}^{2+}$ . This assignment is also concordant with the fact that i) NBOs should exist in our composition ( $\text{NBO}/\text{T} = 0.42$ ), ii) the composition is Ca-rich and  $\text{Ca}^{2+}$  is one important charge balancing cation and iii) the  $\text{CO}_2$  solubility is suspected to be in large part controlled by the Ca amount within the melt but less influenced by Na and even less by Mg (Dixon, 1997). This point is also supported by recent experimental work by Iacono-Marziano et al. (2008) showing a dramatic  $\text{CO}_2$  solubility increase with increasing  $\text{Ca}^{2+}$  content.

We have a small amount of Na present in our glass (3.2 wt.%  $\text{Na}_2\text{O}$ ) and thus the infrared peak responsible for this can be perfectly hidden underneath the main infrared doublet since NBO-carb.  $\text{Na}^+$  has a lower distortion with  $\Delta\nu_3 \sim 35\text{-}45\text{ cm}^{-1}$ . Nevertheless, the isotropic chemical shift for  $\text{Na}_2\text{CO}_3$  is 171 ppm according to Jones et al. (2005). Thus, it is unlikely that NBO-carb.  $\text{Na}^+$  unit exists on the basis of the NMR results. However, it should be mentioned that hydrated crystalline carbonate have lower chemical shift than dry carbonates. For example, Nebel et al. (2008) showed that the  $^{13}\text{C}$  NMR for  $\text{NaHCO}_3$  is observed at 164 ppm.

In the light of infrared observations, we propose to assign the main NMR resonance at 167 ppm to NBO-carb.  $\text{Ca}^{2+}$ . Morizet et al. (2002) suggested a similar assignment for an individual NMR component simulated at 168 ppm in haplo-phonolite glasses.

Our assignment of 167 ppm resonance to NBO-carb.  $\text{Ca}^{2+}$  is also consistent with recent investigation on the structure of amorphous and crystalline calcium carbonates by Nebel

et al. (2008) and using  $^{13}\text{C}$  Solid-State NMR spectroscopy. In their study, calcite ( $\text{CaCO}_3$ ), ikaite ( $\text{CaCO}_3 \cdot 6\text{H}_2\text{O}$ ) and Amorphous Calcium Carbonate (ACC) show an NMR signature at 168 ppm. For other crystalline materials (aragonite, vaterite,  $\text{CaCO}_3$ ; monohydrocalcite,  $\text{CaCO}_3 \cdot \text{H}_2\text{O}$ ), the resonance is observed above that value at ppm shifts from 169 to 171 ppm.

The assignment for the component at 177 ppm which is only observed with  $\{^1\text{H}\}^{13}\text{C}$  CPMAS experiments is much more difficult to apprehend as only little information is currently available for  $^{13}\text{C}$  chemical shift above 170 ppm. We know this carbonate unit i) is in very low quantity as it is not observed in  $^{13}\text{C}$  Direct NMR, ii) is a high hydrogenated environment as it is greatly enhanced in  $\{^1\text{H}\}^{13}\text{C}$  CPMAS experiments, iii) has a distance between the carbonate molecule and the surrounding hydrogen atoms very short compared to the carbonate unit at 167 ppm for which the peak intensity decreases as the *ct* for CPMAS experiment is decreased, and iv) is directly correlated to the reduction process as it is mainly enhanced in the most reduced samples. Our assignment is supported by the following lines of evidence.

$\{^1\text{H}\}^{13}\text{C}$  CPMAS experiments from Nebel et al. (2008) did not detect  $^{13}\text{C}$  chemical shift above 171 ppm which is the maximum noticed for vaterite. Thus, it is likely that 177 ppm component is not correlated to calcium.

There is an apparent correlation between the distortion of the carbonate environment and the  $^{13}\text{C}$  chemical shift. The more distorted is the carbonate group, the lower the  $^{13}\text{C}$  chemical shift is (Brooker et al., 1999; Kohn et al., 1991). We find the highly distorted Si-carb.-Si at 155 ppm then followed by Si-carb.-Al at 160 ppm and Al-carb.-Al at 165

ppm. The identified component at 177 ppm should therefore be less distorted than those network units and even less distorted than the NBO-carb.  $\text{Ca}^{2+}$  proposed at 167 ppm. Ab initio molecular orbital calculations conducted by Tossell (1995) suggest that  $^{13}\text{C}$   $\delta_{\text{iso}}$  above 170 ppm can be attributed to very peculiar carbonate unit: “ionic” clusters. In particular, a  $\delta_{\text{iso}}$  at 177 ppm has been calculated for a “ionic” cluster composed of a  $\text{Na}^+$  cation coordinated to a  $\text{CO}_2$  molecule which is weakly bonded to an oxygen bridging 2  $\text{AlO}_4$  tetrahedron.

Based on the above evidence, one possible assignment for the 177 ppm component is to a “ionic” carbonate cluster such as the one described by Tossell (1995) or Kubicki and Stolper (1995):  $\text{BO-CO}_2 \text{M}^{n+}$ . We do not currently know the kind of charge compensating cation ( $\text{M}^{n+}$ ). We state that  $\text{Ca}^{2+}$  is probably not a good candidate and instead  $\text{Na}^+$  or  $\text{Mg}^{2+}$  should be considered. There is also a large uncertainty on the aluminosilicate network units connected by the BO. The assignment proposed for the 177 ppm component is questionable. Therefore, it appears important to carry out more work in order to better constrain the carbonate environment in silicate melts.

### 5.2 Modeling isobaric $\text{CO}_2$ - $\text{H}_2\text{O}$ solubility trend lines as a function of $f\text{O}_2$ :

$\text{CO}_2$ - $\text{H}_2\text{O}$  solubility trend lines are shown in Figure 10 along with the  $\text{CO}_2$  and  $\text{H}_2\text{O}$  analytical results we have obtained. The solid lines represent the isobaric  $\text{CO}_2$ - $\text{H}_2\text{O}$  solubility trends for 150, 200, 250 and 300 MPa. The positions of the solid lines were calculated from solubility model established by Dixon et al (1995).

In the most oxidized sample, 6-1-ox ( $\Delta\text{FMQ}+2.6$ ), the solubility for  $\text{CO}_2$  and  $\text{H}_2\text{O}$  is in excellent agreement with the calculated solubility model at 300 MPa. This is consistent



with the fact that in 6-1-*ox* the basaltic melt equilibrated with a fluid phase only constituted by a binary mixture of CO<sub>2</sub> and H<sub>2</sub>O (i.e. no other species are present). On the contrary, the other samples solubility results depart significantly from the calculated trends either at 200 MPa for 1-1-200 or at 300 MPa for the other samples. For samples synthesized at 300 MPa and under reducing conditions at roughly  $\Delta\text{FMQ}-1.8$ , the CO<sub>2</sub>-H<sub>2</sub>O solubility data are equivalent to the isobaric CO<sub>2</sub>-H<sub>2</sub>O trend at 250 MPa according to the model of Dixon et al. (1995). Under more severe reducing conditions at  $\Delta\text{FMQ}-2.6$ , the CO<sub>2</sub>-H<sub>2</sub>O solubility falls on the isobaric line at 150 MPa from Dixon et al. (1995).

The reducing conditions experienced by the samples are obviously responsible for this downward displacement of dissolved CO<sub>2</sub> content. As stressed above, under reducing conditions, the fluid phase during the experiment cannot be considered anymore as a binary mixture of CO<sub>2</sub> and H<sub>2</sub>O, instead, CO<sub>2</sub> is progressively replaced by CO (40 mol% in 7-1-*red*, see Table 1). The system becomes at least ternary (CO<sub>2</sub>-H<sub>2</sub>O-CO) but can be even more complex if we consider the presence of other reduced species such as CH<sub>4</sub> or H<sub>2</sub>. The obtained datum for reduced samples is systematically lower in CO<sub>2</sub> solubility compared to 6-1-*ox* synthesized under oxidizing conditions. This result suggests that 1) the addition of CO reduces the CO<sub>2</sub> solubility but also 2) that CO is almost not soluble into basaltic melt which is consistent with the previous study of Pawley et al. (1992) under drier conditions (<0.7 wt.% H<sub>2</sub>O). Most of the data are consistent with the described approach: however 3 points with  $f\text{O}_2$  at  $\Delta\text{FMQ}-1.2$ ,  $-0.7$  and  $-0.1$  fall out of such a general trend. This discrepancy can in large part be accounted for by 1) moderate changes in experimental conditions, in particular the pressure of synthesis, 5-1 at  $\Delta\text{FMQ}-$

0.7 has an equilibrium pressure of 320 MPa which explains its slightly higher CO<sub>2</sub> content relative to other data; 2) possible analytical uncertainties using FTIR spectroscopy for CO<sub>2</sub> quantification; and 3) the 3 charges are all characterized by elevated H<sub>2</sub>O content, hence it is possible that another mechanism of CO<sub>2</sub> solution plays a role. For example, the high water content of those 3 particular samples may change the CO<sub>2</sub> solubility trends as proposed by Behrens et al. (2009) for ultra-potassic magmas. Aside from those second order effects, our results thus show that, as in H<sub>2</sub>O-poor basaltic liquids (Pawley et al., 1992), reduced hydrous basaltic liquids also display a decrease in their CO<sub>2</sub> solubility compared to more oxidizing conditions. This decrease could be even more dramatic if  $fO_2$  conditions more reducing were to be applied.

### 5.3 CO<sub>2</sub> solubility versus CO<sub>2</sub> fugacity ( $fCO_2$ ):

The evolution of CO<sub>2</sub> solubility as a function of  $fCO_2$  is represented in Figure 11. Our results (*open circle*) are compared to several other authors' data and for different compositions ranging from rhyolite to basalt compositions. Our data points are also additionally reported in a wider scale plot placed within the Figure 11 for additional details. The H<sub>2</sub>O<sup>tot</sup> is given for each point alongside.

Altogether, our study is in agreement with the general trend showing the increase in CO<sub>2</sub> solubility with increasing  $fCO_2$  in the fluid phase, as can be predicted from thermodynamic arguments (e.g. Fogel and Rutherford, 1990; Dixon, 1997; Papale, 1999; Behrens et al., 2004; Papale et al., 2006). Yet, at a given  $fCO_2$ , our results appear to be slightly but systematically lower than in other studies.

We observe different trends depending on composition. The dataset for andesite is significantly higher than the other compositions represented here. The data on alkali-basalt (not represented here for clarity) from Botcharnikov et al. (2005) is even higher than andesite.

The results for our Fe-free synthetic glass are lower than the CO<sub>2</sub> solubility for rhyolite glass for example. This observation might reflect differences in the used extinction coefficients for CO<sub>2</sub> quantification by FTIR. We used a value of 69500 L.mol<sup>-1</sup>.cm<sup>-2</sup> from Fine and Stolper (1986) which is valid for basaltic glass. However, our composition is Fe-free therefore the extinction coefficient value is only approximate in CO<sub>2</sub> quantification for our synthetic composition.

In the detailed plot, despite some scatter, H<sub>2</sub>O rich melts appears to be those enriched in CO<sub>2</sub> relative to those H<sub>2</sub>O poor, in qualitative agreement with early suggestions that an increase in melt H<sub>2</sub>O content leads to an increase in CO<sub>2</sub> solubility (see for instance Behrens et al., 2009).

All data points are represented in Figure 11 regardless of  $fO_2$ . The lowest value is given by 1-1-200 (FMQ-2.3). 6-1-ox (FMQ+2.6) lies in the highest part of the trend. At first sight, lowering the  $fO_2$  will act primarily on the fluid phase composition increasing the ratio between CO and CO<sub>2</sub> and thus leading to  $fCO_2$  decrease.

#### 5.4 Implications of $fO_2$ reducing conditions for magmatic degassing in primitive planetary bodies, the Earth and Mars:

Considering that the Earth and Mars interiors may have experienced very reducing conditions, in particular during the first stages of their geological evolution (e.g. Righter

et al., 2007), our data confirm that, for comparable bulk C content of the mantle, at low  $fO_2$  (i.e. at IW), larger amount of C-species may have degassed due to the combined effects of lower  $fCO_2$  and insoluble character of CO, regardless of the dry or wet status of the magma ocean (see Gaillard and Scaillet, 2009).

We also show that under the reducing conditions investigated (from  $\Delta IW$ -2 to +1) other species than  $CO_2$  are present in the fluid phase such as CO or even  $H_2$ . The presence of those reduced species will influence greatly for the modeling of the magma ocean degassing. Recent work by Elkins-Tanton (2008) modeled the outgassed flux of  $CO_2$  produced by the solidification of the primitive magma ocean with application to Mars (and the early Earth), but such an estimate is based on the assumption only  $CO_2$  is degassed during the crystallization of the magma ocean. However, the impact of CO release in the atmosphere cannot be neglected. For instance, at  $\Delta FMQ$ -2.6 (equivalent to  $\Delta IW$ +1.1), the X(CO) in the fluid phase represents 43 mol% (see Table 1). Prather (1996) and Daniel and Solomon (1998) showed that an intense release of CO in the atmosphere can induce a severe greenhouse effect through photochemical reaction producing on a short time scale  $CH_4$  and  $O_3$  gases. In fact, the indirect radiative forcing of CO on global warming appears more important than the  $CO_2$  effect. Recent work by Kawaragi et al. (2009) showed that CO will produce an increase in the surface temperature which is probably 2 times the effect of  $CO_2$  hence provoking major environmental changes.

## **6. Summary:**

We have conducted an experimental investigation on the C-H-O fluid solubility and speciation in silicate melts as a function of redox conditions from moderately oxidized

( $\Delta\text{FMQ}+2.6$ ) to reduced ( $\Delta\text{FMQ}-2.6$ ) conditions which are considered relevant to the Earth and Mars degassing.

Spectroscopic results suggest that C-H-O is dissolved as 3 main species:  $\text{CO}_3^{2-}$ ,  $\text{OH}^-$  and  $\text{H}_2\text{O}^{\text{mol}}$ . The presence of carbonate units is consistent with dissolution mechanism model for basaltic glasses. We do not observe the presence of more reduced species such as  $\text{CH}_4$ .

Volatile species were quantified by FTIR. The  $\text{CO}_2$  solubility is directly proportional to the  $f\text{CO}_2$  in agreement with thermodynamic modeling. On the other hand, the  $\text{CO}_2$  solubility is correlated to the fluid phase composition. On the contrary, the reduction of the aluminosilicate melt does not seem to play a major role into the  $\text{CO}_2$  solubility trend.  $\text{CO}_2$  solubility is strongly affected by redox conditions.  $\text{CO}_2$  solubility measured in oxidized glass is in agreement with thermochemical modeling for isobaric  $\text{CO}_2\text{-H}_2\text{O}$  solubilities. Under reducing conditions,  $\text{CO}_2$  solubility departs from thermochemical equilibrium and the  $\text{CO}_2\text{-H}_2\text{O}$  solubility trends as established at high  $f\text{O}_2$ . Our study shows that the presence of water does not lead to an increase in the solubility of “insoluble” species such as CO or  $\text{CH}_4$ , as compared to their behaviour under nominally dry conditions. Overall, our experimental results provide additional constraints for modeling the degassing of primitive interiors for Mars or the Earth. Future modeling should take into account the presence of reduced species (i.e., CO) into the fluid phase as an important part of the degassing processes which is likely to modify estimates so far done on the primitive atmosphere compositions.

Examination of carbonate environments presumes the existence of one main unit determined by  $^{13}\text{C}$ -MAS NMR but an additional second component is revealed when

acquisition is performed with  $\{^1\text{H}\} \text{ }^{13}\text{C}$ -CPMAS sequence. This unit is strongly enhanced in the most reduced samples. Dynamic CPMAS measurements also suggest a very close proximity of hydrogen atoms to this carbonate unit compared to the main carbonate unit. A possible assignment is proposed for the two components but remains hypothetical. The chemical shift values are in favor to carbonate groups NBO-carb.  $\text{Ca}^{2+}$  for 167 ppm and “ionic” carbonate groups  $\text{BO-CO}_2 \text{M}^{n+}$  for 177 ppm where M is a charge compensating cation such as  $\text{Mg}^{2+}$  or  $\text{Na}^+$ . More work is however to be conducted to confirm this point. Based on the study of the  $\text{CO}_2$  solubility and speciation data, we suggest that applying reducing conditions to the fluid phase does not change dramatically the dissolution mechanisms for C species in hydrous melts; hence under moderately reducing conditions we do not possess evidence for a depolymerization or polymerization process of the melt with C-H-O species dissolution mechanisms.

*Acknowledgements:*

The authors are grateful to the University of Orléans, University of Nantes, the CNRS and INSU for their financial support and access to analytical facilities. We also thank Dr. R. Champallier and Mr. E. Lemenn for their technical help. Constructive reviews by Gordon Moore and an anonymous reviewer greatly improved the quality of the manuscript.

**References list:**

Ahrens, T.J., 1989. Planetary origins – Water storage in the mantle. *Nature* 342, 122-123.

Azbel I.Y., and Tolstikhin, I.N., 1990. Geodynamics, magmatism and degassing of the Earth. *Geochimica Cosmochimica Acta* 54, 139-154.

Behrens, H., and Gaillard, F., 2006. Geochemical aspects of melts: Volatiles and redox behavior. *Elements* 2, 275-280.

Behrens, H., Ohlhorst, S., Holtz, F., and Champenois, M., 2004. CO<sub>2</sub> solubility in dacitic melts equilibrated with H<sub>2</sub>O-CO<sub>2</sub> fluids: Implications for modeling the solubility of CO<sub>2</sub> in silicic melts. *Geochimica Cosmochimica Acta* 68, 4687-4703.

Behrens, H., Roux, J., Neuville, D.R., and Siemann, M., 2006. Quantification of dissolved H<sub>2</sub>O in silicate glasses using confocal microRaman spectroscopy. *Chemical Geology* 229, 96-112.

Behrens, H., Misiti, V., Freda, C., Vetere, F., Botcharnikov, R.E., and Scarlato, P., 2009. Solubility of H<sub>2</sub>O and CO<sub>2</sub> in ultrapotassic melts at 1200 and 1250°C and pressure from 50 to 500 MPa. *American Mineralogist* 94, 105-120.

Blank, J.G., and Brooker, R.A., 1994. Experimental studies of carbon dioxide in silicate melts: solubility, speciation and stable carbon isotope behaviour. In: *Volatiles in magmas. Review in mineralogy* (eds. Carroll M.R., Holloway, J.R.). Mineralogical Society of America, Washington, DC, 30, pp 157-186.

Botcharnikov, R.E., Freise, M., Holtz, F., and Behrens, H., 2005. Solubility of C-O-H mixtures in natural melts: new experimental data and application range of recent models. *Annals of Geophysics* 48, 634-646.

Botcharnikov, R.E., Behrens, H., and Holtz, F., 2006. Solubility and speciation of C-O-H fluids in andesitic melt at T=1100-1300°C and P=200 and 500MPa. *Chemical Geology* 229, 125-143.

Brooker, R.A., Kohn, S.C., Holloway, J.R., McMillan, P.F., and Carroll, M.R., 1999. Solubility, speciation and dissolution mechanisms for CO<sub>2</sub> in melts on the NaAlO<sub>2</sub>-SiO<sub>2</sub> join. *Geochimica Cosmochimica Acta* 63, 3549-3565.

Brooker, R.A., Kohn, S.C., Holloway, J.R., and McMillan, P.F., 2001b. Structural controls on the solubility of CO<sub>2</sub> in silicate melts. Part II: IR characteristics of carbonate groups in silicate glasses. *Chemical Geology* 174, 241-254.

Burke E.A.J., 2001. Raman microspectrometry of fluid inclusions. *Lithos* 55, 139-158.

Cody, G.D., Mysen, B.O., and Lee, S.K., 2005. Structure vs. composition: A solid-state <sup>1</sup>H and <sup>29</sup>Si NMR study of quenched glasses along the Na<sub>2</sub>O-SiO<sub>2</sub>-H<sub>2</sub>O join. *Geochimica Cosmochimica Acta* 69, 2373-2384.



Daniel, J.S., and Solomon, S., 1998. On the climate forcing of carbon monoxide. *Journal of Geophysical Research* 103 (D11), 13249-13260.

Di Carlo, I., Pichavant, M., Rotolo, S.G., and Scaillet, B., 2006. Experimental crystallization of a high-K arc basalt: The golden pumice, Stromboli volcano (Italy). *Journal of Petrology* 47, 1317-1343.

Dingwell, D.B., Hess, K.-U., and Romano, C., 1998. Viscosity data for hydrous peraluminous granitic melts: Comparison with a metaluminous model. *American Mineralogist* 83, 236-239.

Dixon, J.E., 1997. Degassing of alkalic basalts. *American Mineralogist* 82, 368-378.

Dixon, J.E., Stolper, E.M., and Holloway, J.R., 1995. An experimental study of water and carbon dioxide solubilities in Mid-Ocean Ridge Basaltic liquids. Part I: Calibration and solubility models. *Journal of Petrology* 36, 1607-1631.

Drake, M.J., and Righter, K., 2002. Determining the composition of the Earth. *Nature* 416, 39-44.

Elkins-Tanton, L. 2008. Linked magma ocean solidification and atmospheric growth Earth and Mars. *Earth Planetary Science Letter* 271, 181-191.

- Eugster, H.P., and Wones, D.R., 1962. Stability relations of the ferruginous biotite, annite. *Journal of Petrology* 3, 82-125.
- Fine, G., and Stolper, E., 1986. Dissolved carbon dioxide in basaltic glasses: concentrations and speciation. *Earth Planetary Science Letter* 76, 263-278.
- Fogel, R.A., and Rutherford, M.J., 1990. The solubility of carbon dioxide in rhyolitic melts: A quantitative FTIR study. *American Mineralogist* 75, 1311-1326.
- Frantz, J.D., and Mysen, B.O., 1995. Raman spectra and structure of BaO-SiO<sub>2</sub>, SrO-SiO<sub>2</sub> and CaO-SiO<sub>2</sub> melts to 1600°C. *Chemical Geology* 121, 155-176.
- Fung, B.M., Khitrin, A.K., and Ermolaev, K., 2000. An improved broadband decoupling sequence for liquid crystals and solids. *Journal of Magnetic Resonance* 142, 97-101.
- Gaillard, F., Schmidt, B.C., Mackwell, S., and McCammon C., 2003a. Rate of hydrogen-iron redox exchange in silicate melts and glasses. *Geochimica Cosmochimica Acta* 67, 2427-2441.
- Gaillard, F., Pichavant, M., and Scaillet, B., 2003b. Experimental determination of activities of FeO and Fe<sub>2</sub>O<sub>3</sub> components in hydrous silicic melts under oxidizing conditions. *Geochimica Cosmochimica Acta* 67, 4389-4409.

- Gaillard, F., Malki, M., Iacono-Marziano, G., Pichavant, M., and Scaillet, B., 2008. Carbonatite melts and electrical conductivity of the asthenosphere. *Science* 322, 1363-1365.
- Gaillard, F., and Scaillet, B., 2009. The sulfur content of volcanic gases on Mars. *Earth Planetary Science Letter* 279, 34-43.
- Giordano, D., Russell, J.K., and Dingwell, D.B., 2008. Viscosity of magmatic liquids: A model. *Earth Planetary Science Letter* 271, 123-134.
- Hahn, E.L., 1950. Spin Echoes. *Physical Review* 80, 580-594.
- Hirschmann, M.M., and Withers, A.C., 2008. Ventilation of CO<sub>2</sub> from a reduced mantle and consequences for the early Martian greenhouse. *Earth Planetary Science Letter* 270, 147-155.
- Holland, T., and Powell, R., 1991. A Compensated-Redlich-Kwong (CORK) equation for volumes and fugacities of CO<sub>2</sub> and H<sub>2</sub>O in the range 1bar-50kbar and 100-1600°C. *Contribution to Mineralogy and Petrology* 109, 265-273.
- Iacono-Marziano, G., Gaillard, F., and Pichavant, M., 2008. Limestone assimilation by basaltic magmas: an experimental re-assessment and application to Italian volcanoes. *Contribution to Mineralogy and Petrology* 155, 719-738.

Ihinger, P.D., Hervig, R.L., and McMillan, P.F., 1994. Analytical methods for volatiles in glasses. In: Volatiles in magmas. Review in mineralogy (Carroll M.R., Holloway J.R., Eds.) Mineral. Soc. Am., Washington, DC, 30, pp 67-121.

Ingrin, J., Hercule, S. and Charton, T., 1995. Diffusion of hydrogen in diopside: Results of dehydration experiments. *Journal of Geophysical Research* 100, 15489-15499.

Javoy, M., Stillman, C.J., and Pineau, F., 1986. Oxygen and hydrogen isotope studies on the basal complexes of the Canary-Islands – implications on the conditions of their genesis. *Contribution to Mineralogy and Petrology* 92, 225-235.

Jones, A.R., Winter, R., Greaves, G.N., and Smith, I.H., 2005.  $^{23}\text{Na}$ ,  $^{29}\text{Si}$ , and  $^{13}\text{C}$  MAS NMR investigation of glass-forming reactions between  $\text{Na}_2\text{CO}_3$  and  $\text{SiO}_2$ . *Journal of Physical Chemistry B*, 109, 23154-23161.

Kadik, A., Pineau, F., Litvin, Y., Jendrzewski, N., Martinez, I., and Javoy, M., 2004. Formation of Carbon and Hydrogen species in magmas at low oxygen fugacity. *Journal of Petrology* 45, 1297-1310.

Kawaragi, K., Sekine, Y., Kadono, T., Sugita, S., Ohno, S., Ishibashi, K., Kurosawa, K., Matsui, T., and Ikeda, S., 2009. Direct measurements of chemical composition of shock-induced gases from calcite: an intense global warming after Chicxulub impact due to the

indirect greenhouse effect of carbon monoxide. *Earth Planetary Science Letter* 282, 56-64.

King, P.L., and Holloway, J.R., 2002. CO<sub>2</sub> solubility and speciation in intermediate (andesitic) melts: The role of H<sub>2</sub>O and composition. *Geochimica Cosmochimica Acta* 66, 1627-1640.

King, P.L., Vennemann, T.W., Holloway, J.R., Hervig, R.L., Lowenstern, J.B., and Forneris, J.F., 2002. Analytical techniques for volatiles: A case study using intermediate (andesitic) glasses. *American Mineralogist* 87, 1077-1089.

Khimyak, Y.Z., and Klinowski, J., 2001. Solid-state NMR studies of the organic template in mesostructured aluminophosphates. *Physical Chemistry and Chemical Physics* 3, 616-626.

Kohn, S.C., Brooker, R.A., and Dupree, R., 1991. <sup>13</sup>C MAS NMR: A method for studying CO<sub>2</sub> speciation in glasses. *Geochimica Cosmochimica Acta* 55, 3879-3884.

Kubicki, J.D., and Stolper, E.M., 1995. Structural roles of CO<sub>2</sub> and [CO<sub>3</sub>]<sup>2-</sup> in fully polymerized sodium aluminosilicate melts and glasses. *Geochimica Cosmochimica Acta* 59, 683-698.

Kump, L.R., 2008. The rise of atmospheric oxygen. *Nature* 451, 277-278.

Kump, L.R., and Barley, M. E., 2007. Increased subaerial volcanism and the rise of atmospheric oxygen 2.5 billion years ago. *Nature* 448, 1033-1036.

Lange, R.A., and Carmichael, I.S.E., 1987. Densities of Na<sub>2</sub>O-K<sub>2</sub>O-CaO-MgO-FeO-Fe<sub>2</sub>O<sub>3</sub>-Al<sub>2</sub>O<sub>3</sub>-TiO<sub>2</sub>-SiO<sub>2</sub> liquids: new measurements and derived partial molar properties. *Geochimica Cosmochimica Acta* 53, 2195-2204.

Laws, D.D., Bitter, H.M.L., and Jerschow, A., 2002. Solid-state NMR spectroscopic methods in chemistry. *Angewandte Chemie International Edition* 41, 3096-3129

Lesne, P., Scaillet, B., Pichavant, M., and Beny, J.-M., *subm.* The carbon dioxide solubility in alkali basalts: and experimental study. *Contribution to Mineralogy and Petrology*.

London, D., 1992. Phosphorous in S-type magmas: The P<sub>2</sub>O<sub>5</sub> content of feldspars from peraluminous granites, pegmatites, and rhyolites. *American Mineralogist* 77, 126-145.

Long, D.A., 2002 *The Raman effect: A unified treatment of the theory of Raman scattering by molecules.* (John Wiley and Sons Eds) pp. 598.

Maekawa, H., Maekawa, T., Kawamura, K., and Yokokawa, T., 1991. The structural groups of alkali silicate glasses determined from  $^{29}\text{Si}$  MAS NMR. *Journal of Non-Crystalline Solids* 127, 53-64.

Mandeville, C.W., Webster, J.D., Rutherford, M.J., Taylor, B.E., Timbal, A., and Faure, K., 2002. Determination of molar absorptivities for infrared absorption bands of  $\text{H}_2\text{O}$  in andesitic glasses. *American Mineralogist* 87, 813-821.

Martel, C., Pichavant, M., Holtz, F., Scaillet, B., Bourdier, J.-L., and Traineau, H., 1999. Effects of  $f(\text{O}_2)$  and  $\text{H}_2\text{O}$  on andesite phase relations between 2 and 4 kbar. *Journal of Geophysical Research Solid Earth* 104, 29453-29470.

Marty, B., and Jambon, A., 1987.  $\text{C}/^{3}\text{He}$  in volatile fluxes from the solid Earth – Implications for carbon geodynamic. *Earth Planetary Science Letter* 83, 16-26.

Mercier, M., Di Muro, A., Giordano, D., Métrich, N., Lesne, P., Pichavant, M., Scaillet, B., and Clocchiatti, R., 2009. Influence of glass polymerisation and oxidation on micro-Raman water analysis in alumino-silicate glasses. *Geochimica Cosmochimica Acta* 73, 197-217.

Moore, G., 2008. Interpreting  $\text{H}_2\text{O}$  and  $\text{CO}_2$  contents in melt inclusions: Constraints from solubility experiments and modeling. In: *Minerals, Inclusions and Volcanic processes*.

Review in mineralogy and geochemistry (Putirka K.D. and Tepley III F.J., Eds.) Mineral. Soc. Am., Washington, DC, 69, pp 333-361.

Morizet, Y., Kohn, S.C., and Brooker, R.A., 2001. Annealing experiments on CO<sub>2</sub>-bearing jadeite glass: an insight into the true temperature dependence of CO<sub>2</sub> speciation in silicate melts. *Mineralogical Magazine* 65, 701-707.

Morizet, Y., Brooker, R.A., and Kohn, S.C., 2002. CO<sub>2</sub> in haplo-phonolite Melt: Solubility, speciation and carbonate complexation. *Geochimica Cosmochimica Acta* 66, 1809-1820.

Morizet, Y., Paris, M., Gaillard, F., and Scaillet, B., 2009. Raman quantification factor calibration for CO-CO<sub>2</sub> gas mixture in synthetic fluid inclusions: Application to oxygen fugacity calculation in magmatic systems. *Chemical Geology* 264, 58-70.

Mysen, B.O., 1988. Structure and properties of silicate melts. In: *Development in Geochemistry*, vol. 4. Elsevier, Amsterdam, 354 pp.

Mysen, B.O., 1990. Effect of pressure, temperature and bulk composition on the structure and species distribution in depolymerised alkali aluminosilicate melts and quenched melts. *Journal of Geophysical Research B* 95, 15733-15744.



Mysen, B.O., Fogel, M.L., Morrill, P.L., and Cody, G.D., 2009. Solution behavior of reduced C-O-H volatiles in silicate melts at high pressure and temperature. *Geochimica Cosmochimica Acta* 73, 1696-1710.

Nebel, H.N.M., Mayer, C., and Epple, M., 2008. On the structure of amorphous calcium carbonate - A detailed study by solid-state NMR spectroscopy. *Inorganic Chemistry* 47, 7874-7879.

Nowak, M., Probatzki, D., Spickenbom, K., and Diedrich, O., 2003. Carbon dioxide speciation in silicate melts: a restart. *Earth and Planetary Science Letters* 207, 131-139.

Okomura, S., Nakamura, M., and Nakashima, S., 2003. Determination of molar absorptivity of IR fundamental OH-stretching vibration in rhyolitic glasses. *American Mineralogist* 88, 1657-1662.

Olhorst, S., Behrens, H., and Holtz, F. 2001. Compositional dependence of molar absorptivities of near-infrared OH- and H<sub>2</sub>O bands in rhyolitic to basaltic glasses. *Chemical Geology* 174, 5-20.

Pan, V., Holloway, J.R., and Hervig, R., 1991. The pressure and temperature dependence of carbon dioxide solubility in tholeiitic basalt melts. *Geochimica Cosmochimica Acta* 55, 1587-1595.

Papale, P., 1999. Modeling of the solubility of a two-component H<sub>2</sub>O + CO<sub>2</sub> fluid in silicate liquids. *American Mineralogist* 84, 477-492.

Papale, P., Moretti, R., and Barbato, D., 2006. The compositional dependence of the saturation surface of H<sub>2</sub>O+CO<sub>2</sub> fluids in silicate melts. *Chemical Geology* 229, 78-95.

Papenguth, H.W., Kirkpatrick, R.J., Montez, B., and Sandberg, P.A., 1989. <sup>13</sup>C MAS NMR spectroscopy of inorganic and biogenic carbonates. *American Mineralogist* 74, 1152-1158.

Pawley, A.R., Holloway, J.R., and McMillan, P.F., 1992. The effect of oxygen fugacity on the solubility of carbon-oxygen fluids in basaltic melt. *Earth Planetary Science Letter* 110, 213-225.

Prather, M.J., 1996. Time scales in atmospheric chemistry: Theory, GWPs for CH<sub>4</sub> and CO, and runaway growth. *Geophysical Research Letters* 23, 2597-2600.

Puls, S.P., and Eckert, H., 2006. Site discrimination in mixed-alkali glasses studied by Cross-Polarization NMR. *Journal of Physical Chemistry B* 110, 14253-14261.

Righter, K., Drake, M.J., and Scott, E., 2007. Compositional relationships between meteorites and terrestrial planets. In: *Meteorites and the Early Solar system II*, (Lauretta D., Leshin L., McSween H.Y., Eds). Univ. Arizona Press, Tucson.

Ripmeester, J.A., and Ratcliffe, C.I., 1988. Low-temperature cross-polarisation/magic angle spinning  $^{13}\text{C}$  NMR of solid methane hydrates: structure, cage occupancy, and hydration number. *Journal of Physical Chemistry* 92, 337-339.

Robert, E., Whittington, A., Fayon, F., Pichavant, M., and Massiot, D., 2001. Structural characterization of water-bearing silicate and aluminosilicate glasses by high-resolution solid-state NMR. *Chemical Geology* 174, 291-305.

Robie, R.A., and Hemingway, B.S., 1995. *Thermodynamic Properties of Minerals and Related Substances at 298.15 K and 1 Bar (105 Pascals) Pressure and at Higher Temperatures*, U.S. Geological Survey Bulletin 2131. U.S. GPO, Washington, DC.

Scaillet, B., Pichavant, M., Roux, J., Humbert, G., and Lefèvre, A., 1992. Improvements of the Shaw membrane technique for measurement and control of  $f\text{H}_2$  at high temperatures and pressures. *American Mineralogist* 77, 647-655.

Scaillet, B., and Pichavant, M., 2004. Role of  $f\text{O}_2$  on fluid saturation in oceanic basalt. *Nature* 430, DOI: 10.1038/nature02814|ISSN 0028-0836.

Schmidt, B.C., Holtz, F., Scaillet, B., and Pichavant, M., 1997. The influence of  $\text{H}_2\text{O}$ - $\text{H}_2$  fluids and redox conditions on melting temperatures in the haplogranite system. *Contribution to Mineralogy and Petrology* 126, 386-400.

Schmidt, B.C., Riemer, T., Kohn, S.C., Behrens, H., and Dupree, R., 2000. Different water solubility mechanisms in hydrous glasses along the Qz-Ab Join: Evidence from NMR Spectroscopy. *Geochimica Cosmochimica Acta* 64, 513-526.

Shi, P., and Saxena, S.K., 1992. Thermodynamic modeling of the C-H-O-S fluid system. *American Mineralogist* 77, 1038-1049.

Silver, L.A., Ihinger, P.D., and Stolper, E., 1990. The influence of bulk composition on the speciation of water in silicate glasses. *Contribution to Mineralogy and Petrology* 104, 142-162.

Sleep, N.H., and Zahnle, K., 2001. Carbon dioxide cycling and implications for climate on ancient Earth. *Journal of Geophysical Research – Planets* 106, 1373-1399.

Spickenbom, K., Serrialta, M., and Nowak, M., 2010. Carbon dioxide and argon diffusion in silicate melts: insights into the CO<sub>2</sub> speciation in magmas. *Geochimica Cosmochimica Acta*, doi: 10.1016/j.gca.2010.08.022.

Stolper, E.M., (1982). Water in silicate glasses: an infrared spectroscopic study. *Contribution to Mineralogy and Petrology* 81, 1-17.

Stuke, A., Behrens, H., Schmidt, B.C., and Dupree, R., 2006. H<sub>2</sub>O speciation in float glass and soda lime glass. *Chemical Geology* 229, 64-77.

Subramanian, S., Kini, R.A., Dec, S.F., and Sloan, E.D., 2000. Evidence of structure II hydrate formation from methane + ethane mixtures. *Chemical Engineering. Science* 55, 1981-1999.

Tamic, N., Behrens, H., and Holtz, F., 2001. The solubility of H<sub>2</sub>O and CO<sub>2</sub> in rhyolitic melts in equilibrium with a mixed CO<sub>2</sub>-H<sub>2</sub>O fluid phase. *Chemical Geology* 174, 333-347.

Taylor, W.R., 1990. The dissolution mechanism of CO<sub>2</sub> in aluminosilicate melts – infrared spectroscopic constraints on the cationic environment of dissolved [CO<sub>3</sub>]<sup>2-</sup>. *European Journal of Mineralogy* 2, 547-563.

Taylor, R.E., 2004. Setting up <sup>13</sup>C CP/MAS experiments. *Concepts in Magnetic Resonance Part A* 22, 37–49

Thibault, Y., and Holloway, J.R., 1994. Solubility of CO<sub>2</sub> in a Ca-rich leucitite: effects of pressure, temperature and oxygen fugacity. *Contribution to Mineralogy and Petrology* 116, 216-224.

Thomas, R., Webster, J.D., and Heinrich, W., 2000. Melt inclusions in pegmatite quartz: complete miscibility between silicate melts and hydrous fluids at low pressure.

Contribution to Mineralogy and Petrology 139, 394-401.

Tossell, J.A., 1995. Calculation of the  $^{13}\text{C}$  NMR shieldings of the  $\text{CO}_2$  complexes of aluminosilicates. *Geochimica Cosmochimica Acta* 59, 1299-1305.

Watson, E.B., 1994. Diffusion in volatile-bearing magmas. In: Volatiles in magmas.

Review in mineralogy (Carroll M.R., Holloway J.R., Eds.) Mineral. Soc. Am.,

Washington, DC, 30, pp 371-411.

Withers, A.C., and Behrens, H., 1999. Temperature-induced changes in the NIR spectra of hydrous albitic and rhyolitic glasses between 300 and 100 K. *Physics and Chemistry of Minerals* 27, 119-132.

Xue, X., and Kanzaki, M., 2004. Dissolution mechanisms of water in depolymerised silicate melts: constraints from  $^1\text{H}$  and  $^{29}\text{Si}$  NMR spectroscopy and ab initio calculations.

*Geochimica Cosmochimica Acta* 68, 5027-5057.

Yamashita, S., Behrens, H., Schmidt, B.C., and Dupree, R., 2008. Water speciation in sodium silicate glasses based on NIR and NMR spectroscopy. *Chemical Geology* 256,

231-241.

Zajacz, Z., Halter, W., Malfait, W.J., Muntener, O., Bodnar, R.J., Bahcman, O., Webster, J.D., Ulmer, P., Mandeville, C.W., Hirschmann, M.M. and Morizet, Y., 2005. A composition independent quantitative determination of the water content in silicate glasses and silicate melt inclusions by confocal Raman-spectroscopy. *Contribution to Mineralogy and Petrology* 150, 631-642.

Zeng, Q., Nekvasil, H., and Grey, C.P., 1999. Proton environments in hydrous aluminosilicate glasses: a  $^1\text{H}$  MAS,  $^1\text{H}/^{27}\text{Al}$ , and  $^1\text{H}/^{23}\text{Na}$  TRAPDOR NMR study. *Journal of Physical Chemistry B* 103, 7406-7415.

Zhang, Y., and Zindler, A., 1989. Noble-gas constraints on the evolution of the Earth's atmosphere. *Journal of Geophysical Research Solid Earth* 94, 13719-13737.

Zhang, Y., Stolper, E.M., and Ihinger, P.D., 1995. Kinetics of the reaction  $\text{H}_2\text{O} + \text{O} = 2\text{OH}$  in rhyolitic and albitic glasses: Preliminary results. *American Mineralogist* 80, 593-612.

Zhang, Y., Xu, Z., Mengfan, Z., and Wang H., 2007. Silicate melt properties and volcanic eruptions. *Reviews of Geophysics* 45, 216-242.

Zolotov, M.Y., and Fegley, B.Jr., 1999. Oxidation state of volcanic gases and the interior of Io. *Icarus* 141, 40-52.

**Figure caption:**

Figure 1: Log  $fO_2$  (1) calculated from the  $H_2O^{tot}$  in the melt versus the log  $fO_2$  (2) calculated from  $X(CO_2)$ - $X(CO)$  measured in fluid inclusions. The error on log  $fO_2$  (1) was calculated considering an upper and lower value for  $H_2O^{tot}$  dissolved in the glass ( $\pm 0.20$  wt.%  $H_2O$  as given in Table 3). The error on the log  $fO_2$  (2) was 1) an upper and lower value for  $X(CO_2)$ - $X(CO)$  for fluid inclusions analyses. The 1:1 line is represented in dashed line.

Figure 2: Micro-Raman spectra for fluid inclusion analysis. A,  $CO_2$  vibrational signatures for  $2\nu_2$  and  $\nu_1$  and for both  $^{12}C$  and  $^{13}C$  isotopes between  $1200$  and  $1500\text{ cm}^{-1}$ . The  $\Delta FMQ$  conditions and the %  $^{13}C$  calculated from fluid inclusions and reported in Table 1 are indicated next to each spectrum. B,  $^{12}CO$  and  $^{13}CO$  vibrational signatures between  $2000$  and  $2200\text{ cm}^{-1}$ . An intensity scaling factor of 2 was applied for better viewing. The  $CO_2$  molar fraction calculated from fluid inclusions is also reported. C, Evidence for  $CH_4$  in fluid inclusions. A scaling factor of 4 is applied to show the very low intensity of this species (calculated value indicated alongside).

Figure 3: Raman spectra obtained in the water vibrational region ( $2700$ - $3900\text{ cm}^{-1}$ ) and in the glass structure vibrational region ( $200$ - $1250\text{ cm}^{-1}$ ) showing 2 main vibrations: T-O stretching ( $800$ - $1200\text{ cm}^{-1}$ ) and T-O-T bending and rocking ( $300$ - $700\text{ cm}^{-1}$ ). A typical baseline is shown as dashed line underneath 6-1-ox. Spectra are uncorrected for black



body radiation. The  $\text{H}_2\text{O}^{\text{tot}}$  (2) is indicated next to each spectrum and has been calculated with Behrens et al. (2006) method.

Figure 4: FTIR spectra for glasses showing different  $^{13}\text{C}$  enrichment calculated from fluid inclusions analysis: A, 80%; B, 40%; C, 1%  $^{13}\text{C}$ . In all case the carbonates are represented by a peak doublet between 1400 and 1600  $\text{cm}^{-1}$  corresponding to the antisymmetric stretching ( $\nu_3$ ). The  $\Delta\nu_3$  remains constant with a value between 65 and 80  $\text{cm}^{-1}$ . We reported next to each spectrum the  $\Delta\text{FMQ}$  and ppm  $\text{CO}_2$ .

Figure 5: A, FTIR vibrational signature for  $\text{OH}^-$  groups (4500  $\text{cm}^{-1}$ ) and  $\text{H}_2\text{O}^{\text{mol}}$  (5200  $\text{cm}^{-1}$ ); B, FTIR vibrational signature for  $\text{H}_2\text{O}^{\text{tot}}$  (3500  $\text{cm}^{-1}$ ). The concentrations for each species is given next to each spectrum and reported in Table 3. All values indicated on A and B were calculated from Beer-Lambert law and using the extinction coefficients proposed by Stolper et al. (1982). The error on each concentration was determined from standard deviation on replicated FTIR analyses. A dashed line underneath each spectrum represents the baseline.

Figure 6:  $^{13}\text{C}$ -MAS NMR spectra acquired with spin-echo sequence. Spectra show 3 main signatures: 125, 184 ppm for  $\text{CO}_2$  and CO in the gaseous phase entrapped in fluid inclusions; broad signature between 155 and 170 ppm for carbonate groups dissolved in the glass.  $\Delta\text{FMQ}$  and  $\text{CO}_2$  content have been indicated for comparison. As a comparison, we included spectra from other studies: haplophonolite from Morizet et al. (2002) and sodamelilite from Kohn et al. (1991).

Figure 7:  $\{^1\text{H}\}^{13}\text{C}$ -CPMAS NMR spectra acquired with 3.5 ms contact time for different samples.  $\Delta\text{FMQ}$  and wt.%  $\text{H}_2\text{O}$  are given. One symmetric component is visible at 167 ppm for all samples. An additional broad carbonate group appears in the most reduced sample at 177 ppm.

Figure 8:  $\{^1\text{H}\}^{13}\text{C}$  CPMAS spectra for 7-1-*red* ( $\Delta\text{FMQ}$ -2.6) and 7-2 ( $\Delta\text{FMQ}$ -1.2) obtained for different contact time (0.5, 1.0 and 3.5 ms) showing the change in carbonate environments within the glass. This dynamic study suggests a close proximity between hydrogen and carbon for the component at 177 ppm.

Figure 9: wt.%  $\text{H}_2\text{O}^{\text{tot}}$  determined from Micro-Raman analyses ( $\text{H}_2\text{O}^{\text{tot}}$  (2)) versus wt.%  $\text{H}_2\text{O}^{\text{tot}}$  determined from FTIR (see Table 3): ■,  $\text{H}_2\text{O}^{\text{tot}}$  (1) determined from the sum  $\text{OH} + \text{H}_2\text{O}^{\text{mol}}$ ; ○ and ●,  $\text{H}_2\text{O}^{\text{tot}}$  determined from respectively the height and the area of the peak at  $3500\text{ cm}^{-1}$ ; Those data points were calculated using the extinction coefficients ( $\epsilon$ ) of Stolper et al. (1982). ▽,  $\text{H}_2\text{O}^{\text{tot}}$  determined from the height of the peak at  $3500\text{ cm}^{-1}$  and using the extinction coefficient derived by Mandeville et al. (2002). 1:1 slope is represented with dashed line. Error on each point is given in Table 3 and determined from standard deviation and error propagation for FTIR value and an assumed error of  $\pm 0.20$  wt.% for on  $\text{H}_2\text{O}^{\text{tot}}$  (2) determined by Raman is taken.

Figure 10: ppm  $\text{CO}_2$  versus wt.%  $\text{H}_2\text{O}$ .  $\Delta\text{FMQ}$  value is indicated next to each point. ● are 300 MPa data, ○ is 200 MPa 1-1-200 and ◇ is for 4-2-*us* nominally  $\text{CO}_2$ -free glass.

Solid lines are isobaric CO<sub>2</sub>-H<sub>2</sub>O solubility trends at 150, 200, 250 and 300 MPa obtained from thermochemical equilibrium of Dixon et al. (1995) and for relatively oxidized conditions.

Figure 11: ppm CO<sub>2</sub> versus  $f\text{CO}_2$ . Several dataset are represented in this plot: This study (*open circle*), rhyolite (Tamic et al., 2001; *filled triangle*); basalt (Pan et al., 1991; *inversed open triangle*; Pawley et al., 1992; *filled square*; Dixon et al., 1995; *filled diamond*; Botcharnikov et al., 2005; *filled hexagon*; Lesne et al., *subm.*; *stars*) and andesite (Botcharnikov et al., 2006; *open hexagon*). An insert is added to the plot with an increased scale showing only our data points with H<sub>2</sub>O<sup>tot</sup> indicated.

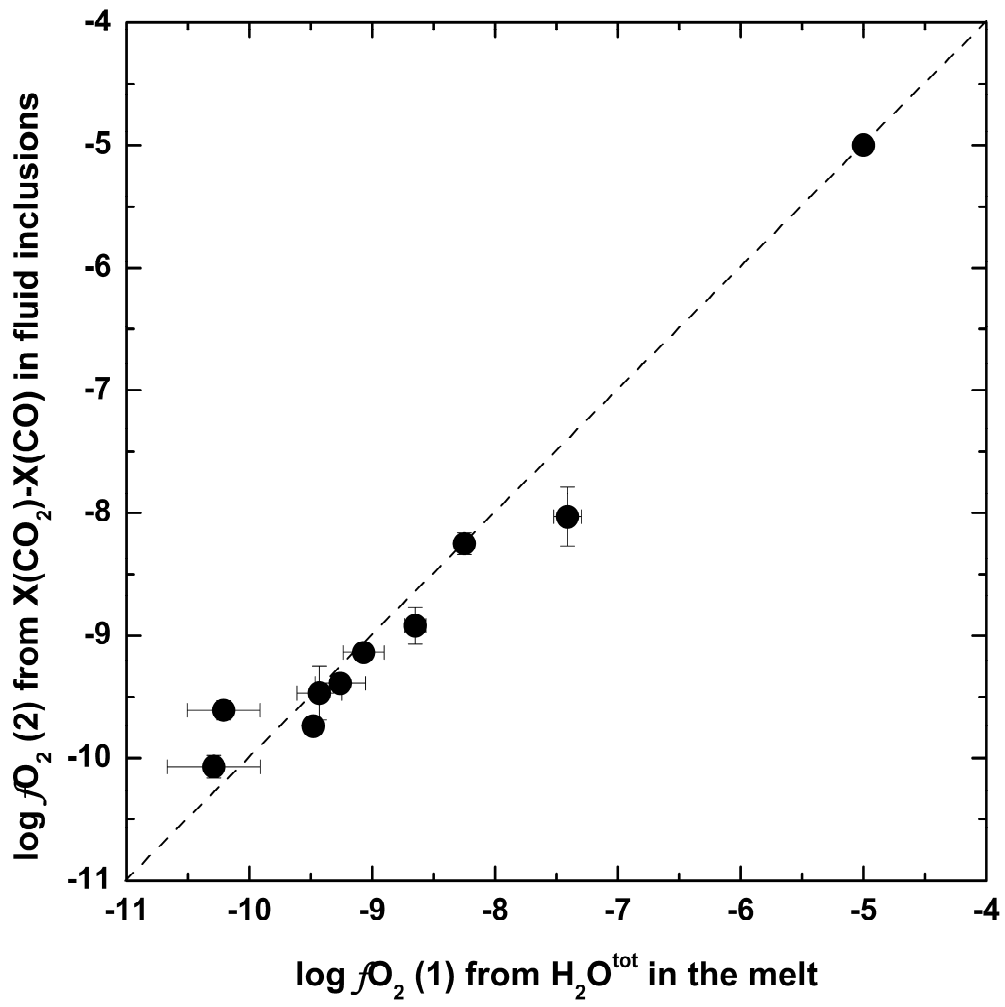


Figure 1

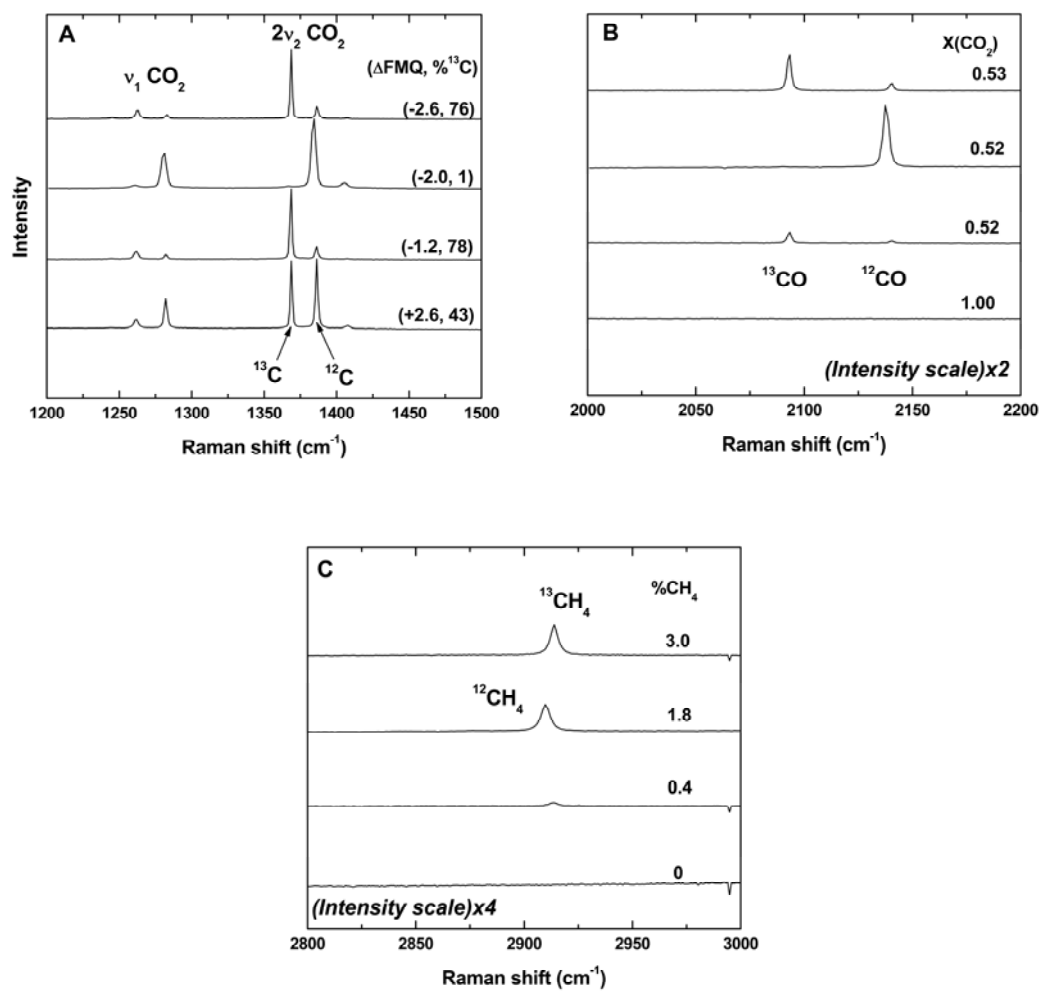


Figure 2

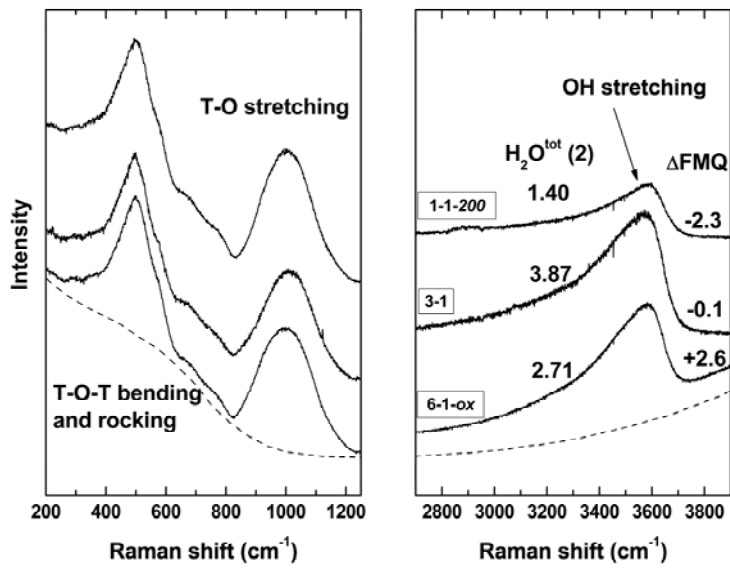


Figure 3

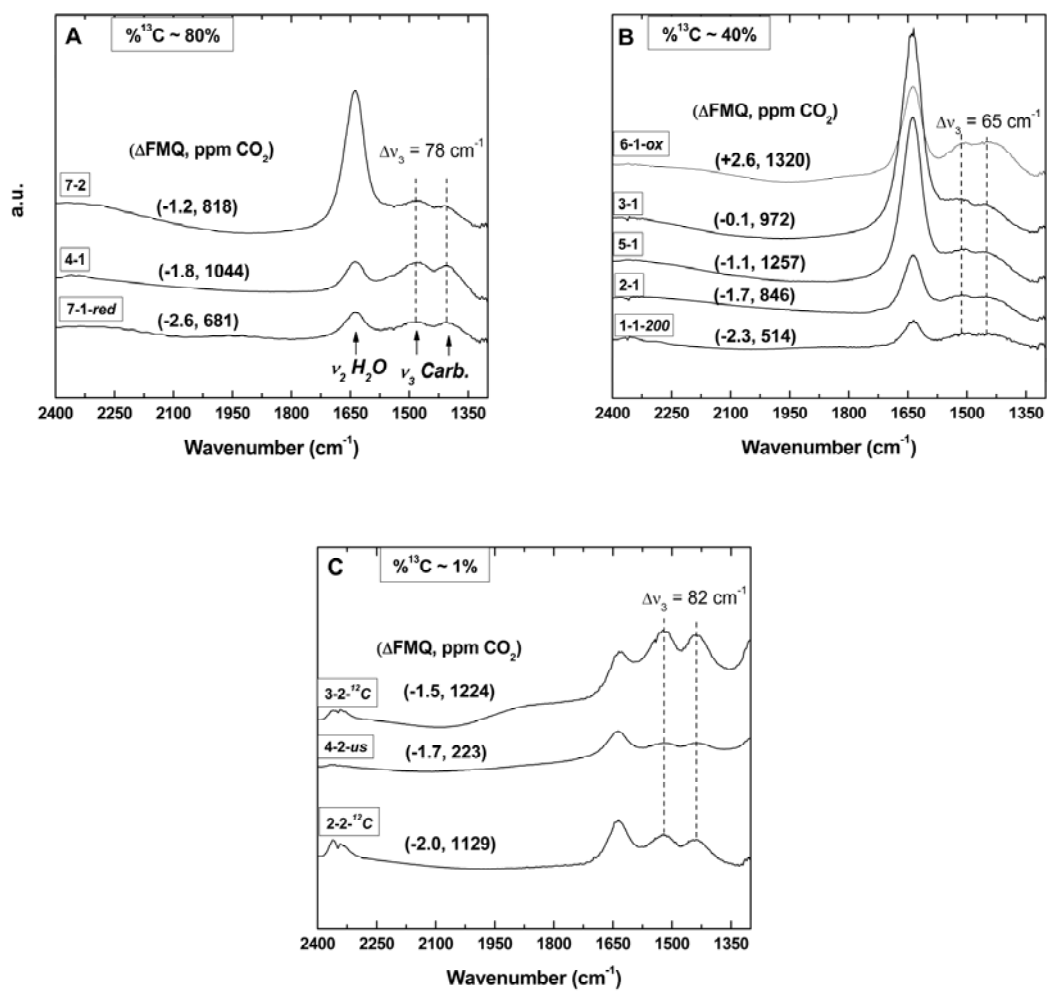


Figure 4

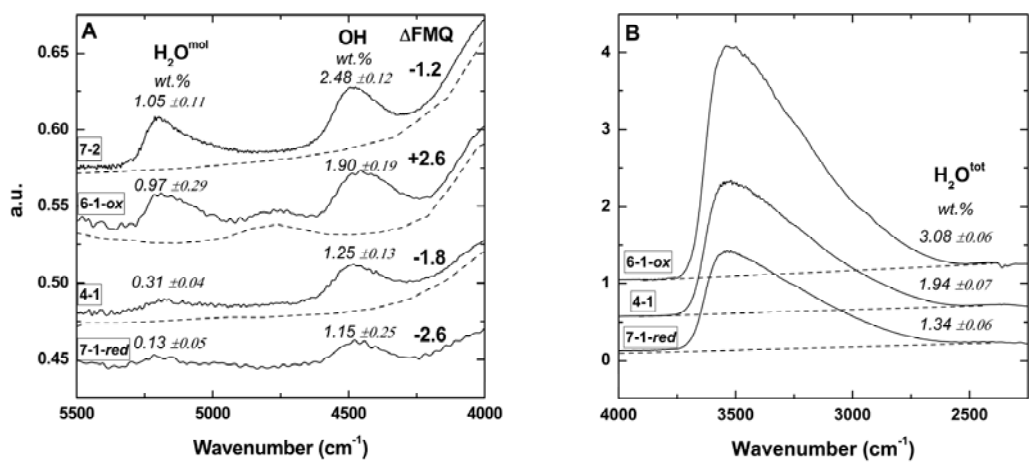


Figure 5



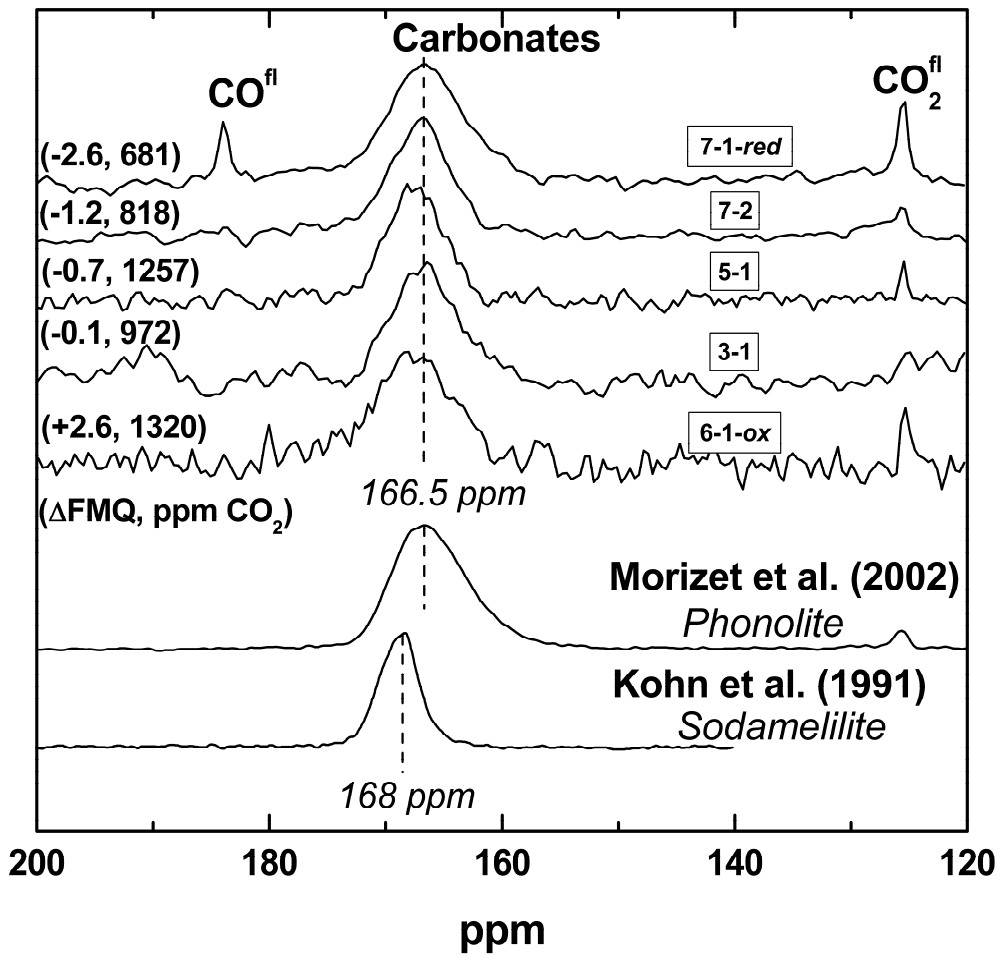


Figure 6

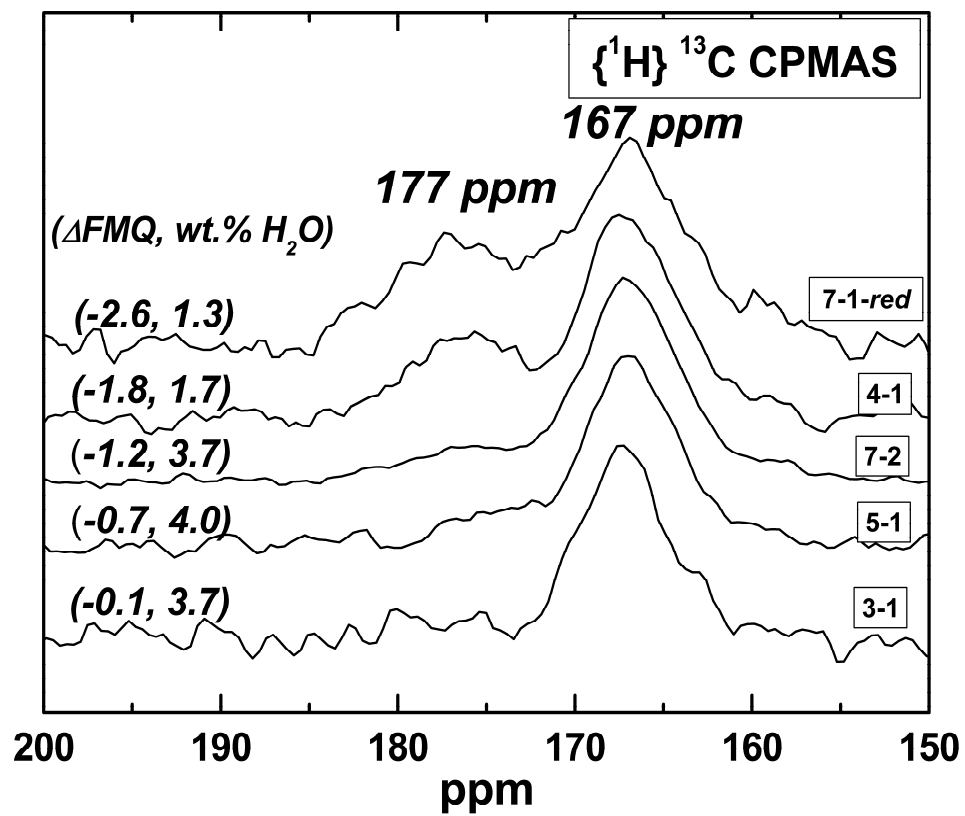


Figure 7

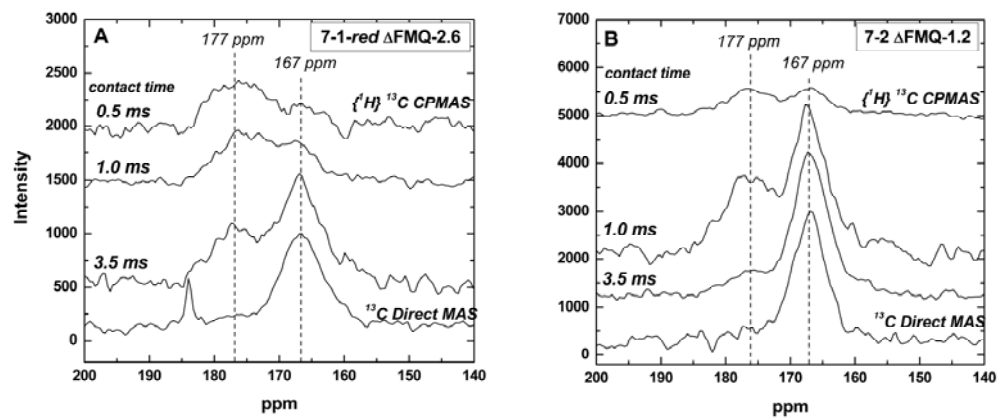


Figure 8

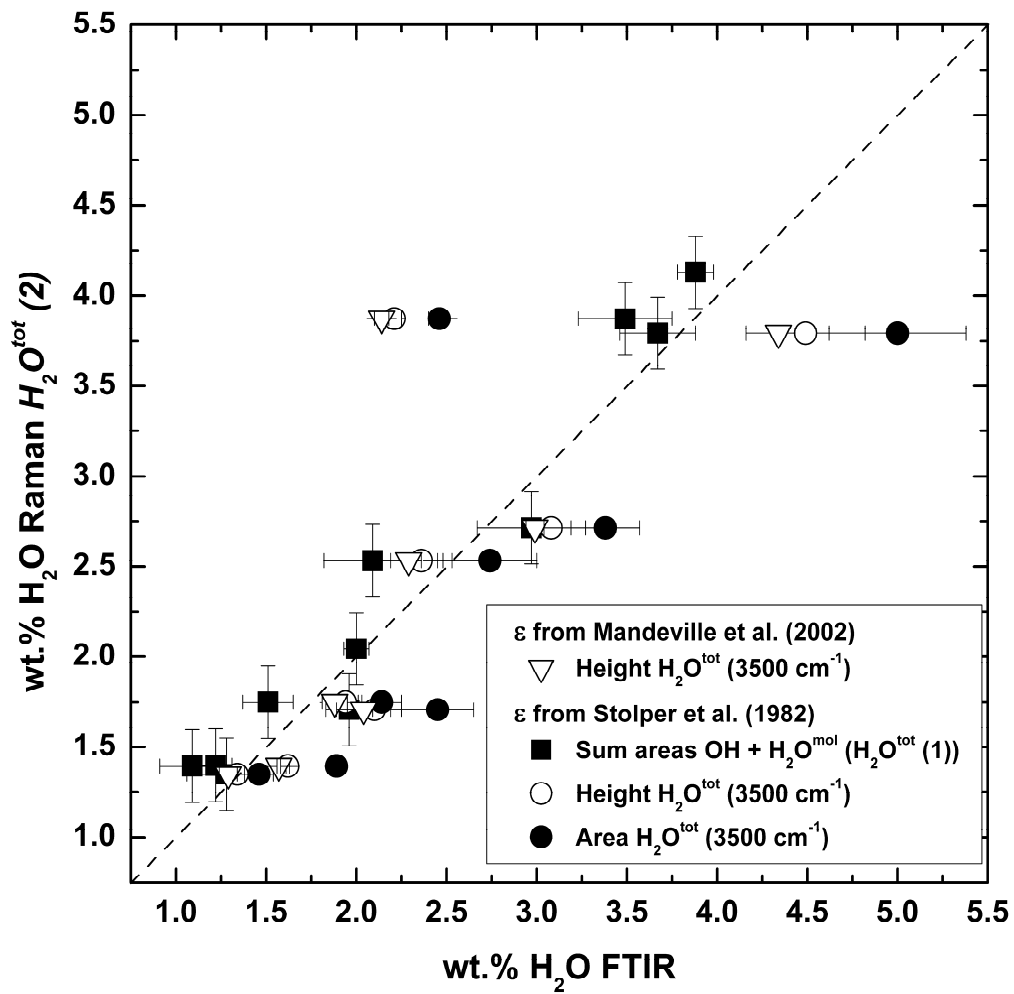


Figure 9

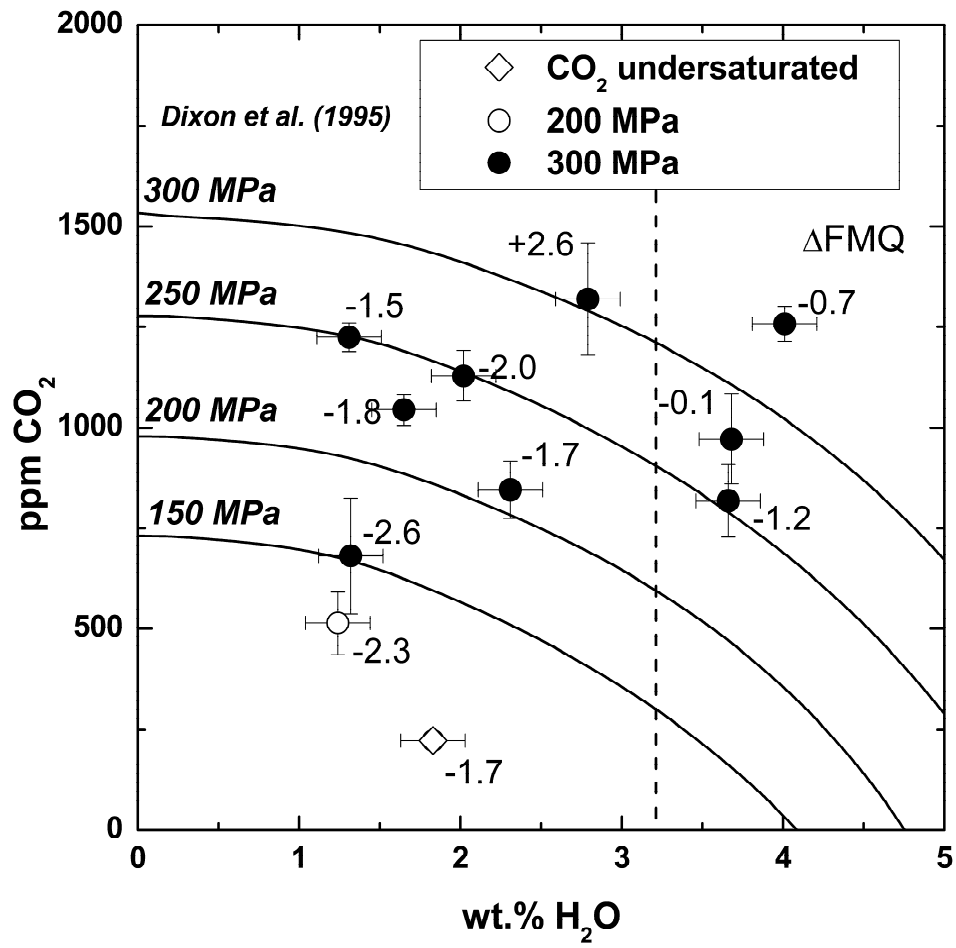


Figure 10

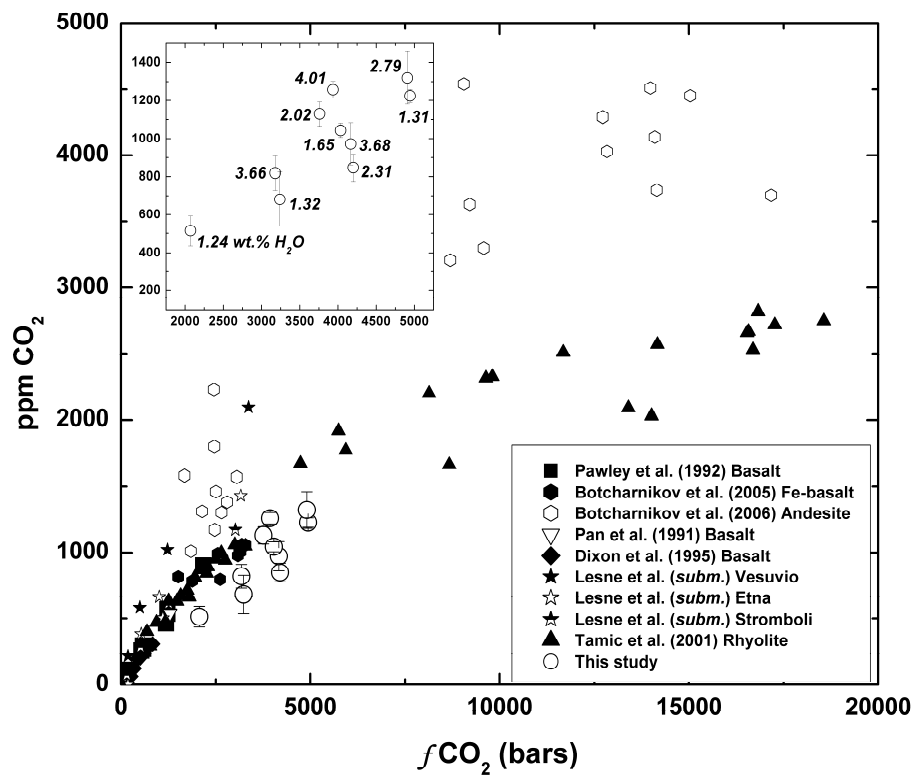


Figure 11