

## COMPRESSIBILITY OF WATER IN MAGMA: EFFECT ON PLANETARY DIFFERENTIATION

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**Introduction:** “Density crossovers” occur between silicate liquids and their coexisting equilibrium crystalline phases when compression curves of the liquids and the crystals intersect at high pressure. This can occur between certain crystal-liquid pairs because the compressibilities of silicate liquids are higher than the compressibilities of silicate crystals, and at high enough pressures the compression curves “crossover”. This phenomenon has been confirmed experimentally in anhydrous systems by numerous studies over the past 25 years [1-7]. Density crossovers may have important implications for planetary differentiation scenarios especially within a magma ocean environment, where temperatures are assumed to be high enough for wholesale melting of anhydrous mantle to occur at great depths and very high pressures. For recent planetary differentiation, with more modest mantle geotherms, deep melting is probably restricted to heterogeneous regions with thermal boundary layers such as the core-mantle-boundary (CMB) or regions where the mantle solidus has been lowered by the presence of water. Hydrous magma density crossovers for such deep mantle environments are not predictable from experimental data on anhydrous melts. Knowledge of the compressibility of water bearing magmas at high pressure is required. To this end we examine data from studies on hydrous melt density and compressibility at high pressure, and derive the effect of pressure on the partial molar volume of H<sub>2</sub>O in silicate melt ( $\bar{V}_{\text{H}_2\text{O}}$ ).

**Partial Molar Volume of Water in Silicate Melt at High Pressure:** Figure 1 shows data from various studies [8-12] that measured the density of hydrous silicate liquids or glasses at low pressure and high pressure, and from which we derive values for  $\bar{V}_{\text{H}_2\text{O}}$  normalized to 1800°C. The experimental densities were measured on hydrous silicate glasses near the glass transition temperature at 1-bar with dilatometry, on hydrous albite melt up to 0.85 GPa in an internally heated, gas pressurized, P-V-T vessel, on hydrous Fe-rich komatiite melt at 1.5-9 GPa in a piston-cylinder and multi-anvil apparatus using sink/float technique, on hydrous Fe-rich ultrabasic melt 10-16 GPa, and on molten hydrous MORB in multi-anvil apparatus using sink/float technique.

It can be seen from the best fit curve in figure 2 that  $\bar{V}_{\text{H}_2\text{O}}$  decreases from ~30 cm<sup>3</sup>/mol to ~10 cm<sup>3</sup>/mol in the range 0-8 GPa and approaches 3.918 cm<sup>3</sup>/mol as  $P \rightarrow \infty$ . This suggests that the compressibility of H<sub>2</sub>O in

magma has distinctly different character at low pressure versus high pressure. At low pressure (0-2 GPa) H<sub>2</sub>O is highly compressible compared to other silicate melt components, however at high pressure >3 GPa H<sub>2</sub>O compressibility is comparable to the other “liquid oxides” [13] that make up silicate melts.

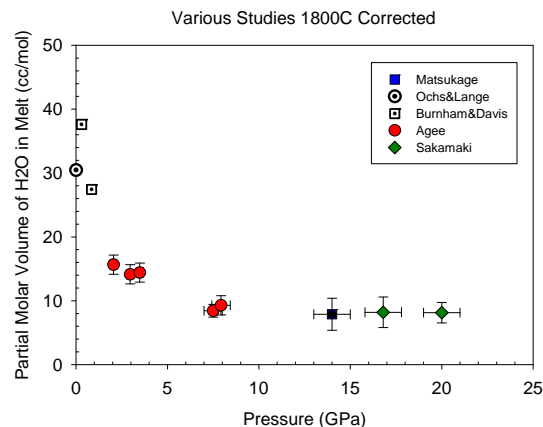


Figure 1. Pressure versus partial molar volume of H<sub>2</sub>O in silicate melt normalized to 1800°C. Data from Burnham and Davis (1971), Ochs and Lange (1999), Matsukage et al. (2005), Sakamaki et al. (2006), Agee (2007).

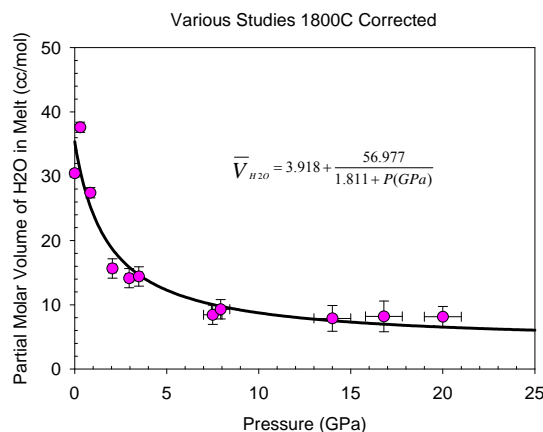


Figure 2. Pressure versus partial molar volume of H<sub>2</sub>O in silicate melt with data fit to a three parameter hyperbolic decay curve of the form  $f=y_0+(a*b)/(b+x)$ .

**Density Crossovers in Planetary Differentiation:** The best fit curve to the data in figure 2 allows estimation of  $\bar{V}_{\text{H}_2\text{O}}$  and hence calculation of hydrous magma density at high pressures within the mantles of the terrestrial planets. In figure 3 we apply these data to the mantle of Mars and illustrate the density-pressure relationship between anhydrous FeO-rich

komatiite, FeO-rich komatiite with 5 wt% H<sub>2</sub>O and possible liquidus crystalline phases olivine, majorite garnet, and perovskite.

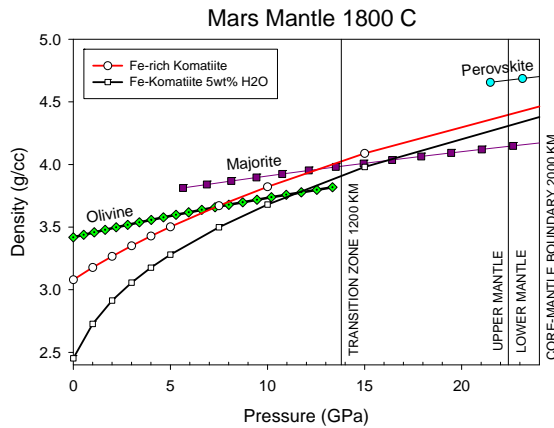


Figure 3. Pressure versus density diagram showing compression curves for anhydrous and hydrous Fe-rich komatiite melt and likely crystalline liquidus phases olivine, majorite garnet, and perovskite.

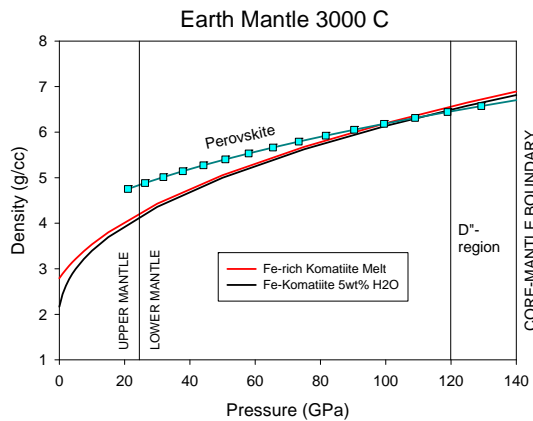


Figure 4. Pressure versus density diagram showing compression curves for anhydrous and hydrous Fe-rich komatiite melt and crystalline perovskite.

The anhydrous melt compression curve shown in red with open dots in figure 3 is calculated from the Birch-Murnaghan equation with an isothermal bulk modulus ( $K_T$ ) of 30 GPa and pressure derivative of the bulk modulus ( $K'$ ) of 4 [12]. The hydrous melt compression curve shown in black with open squares, was calculated from the best fit  $P$  versus  $\bar{V}_{H_2O}$  curve in figure 2, and “by difference” from the anhydrous melt compression curve. A density crossover exists for hydrous silicate melt and olivine atop the mantle transition zone which in the martian mantle occurs at approximately 1200 km depth. This could mean that a region of low seismic velocity exists at 1200 km in Mars, analogous to the low S-wave velocities observed atop the 410 km discontinuity in the Earth [14], associated with the

hypothesized presence of neutrally buoyant hydrous melts [15].

Figure 4 shows the density-pressure relationships for anhydrous Fe-rich komatiite melt, hydrous Fe-rich komatiite melt, and crystalline perovskite in the Earth’s lower mantle. At very high pressures (>20 GPa) the compression curve slope and the density differences between anhydrous and hydrous melt become small. This is owing to the relatively small value for  $\bar{V}_{H_2O}$  and lower value for compressibility of water in magma at these pressures. As a result, the density crossover for both anhydrous and hydrous melt and equilibrium perovskite occur at similar pressures at the base of the low mantle in the D’’-layer. This consistent with the observed ultralow seismic velocity zone (ULVZ) [16] thought to be associated with partial melts trapped by negative buoyancy at these depths. Interestingly, figure 3 shows that Mars is unlikely to possess a similar ULVZ since partial melts atop the core-mantle-boundary will be positively buoyant relative to coexisting crystalline perovskite. On the other hand, if thermal anomalies extend into the martian upper mantle transition zone, then hydrous melts there will be negatively buoyant relative to coexisting crystalline majorite garnet. Thus a mid-transition zone ULVZ could exist at 1300-1700 km depth in Mars.

#### References:

- [1] Rigden, S. M., et al. (1984) *Science*, 226, 1071
- [2] Agee, C. B. and Walker, D. (1988) *Journal of Geophysical Research*, 93, 3437
- [3] Miller, G. H., et al. (1991) *Journal of Geophysical Research*, 96, 11
- [4] Agee, C. B. and Walker, D. (1993) *Earth and Planetary Science Letters*, 114, 315
- [5] Suzuki, A., et al. (1995) *Science*, 269, 216
- [6] Circone, S. and Agee, C. B. (1996) *Geochimica et Cosmochimica Acta*, 60, 2709
- [7] Ohtani, E. and Maeda, M. (2001) *Earth and Planetary Science Letters*, 193, 69
- [8] Burnham, C. W. and Davis, N. F. (1971) *American Journal of Science*, 270, 54
- [9] Ochs, F. A. and Lange, R. A. (1999) *Science*, 283, 1315
- [10] Matsukage, K. N., et al. (2005) *Nature*, 438, 488
- [11] Sakamaki, T., et al. (2006) *Nature*, 439, 192
- [12] Agee, C. B. (2007) *Earth and Planetary Science Letters*, accepted for publication 4-nov-2007.,
- [13] Bottinga, Y. and Weill, D. F. (1970) *American Journal of Science*, 269, 169
- [14] Revenaugh, J. and Sipkin, S. A. (1994) *Nature*, 369, 474
- [15] Bercovici, D. and Karato, S. (2003) *Nature*, 425, 39
- [16] Williams, Q. and Garnero, E. (1998) *Science*, 273, 1528