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Global-scale water circulation in the Earth's mantle: Implications for the mantle water budget in the early Earth

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ABSTRACT

We investigate the influence of the mantle water content in the early Earth on that in the present mantle using numerical convection simulations that include three processes for redistribution of water: dehydration, partitioning of water into partially molten mantle, and regassing assuming an infinite water reservoir at the surface. These models suggest that the water content of the present mantle is insensitive to that of the early Earth. The initial water stored during planetary formation is regulated up to 1.2 OMs (OM = Ocean Mass; 1.4×10^{21} kg), which is reasonable for early Earth. However, the mantle water content is sensitive to the rheological dependence on the water content and can range from 1.2 to 3 OMs at the present day. To explain the evolution of mantle water content, we computed water fluxes due to subducting plates (regassing), degassing and dehydration. For weakly water dependent viscosity, the net water flux is almost balanced with those three fluxes but, for strongly water dependent viscosity, the regassing dominates the water cycle system because the surface plate activity is more vigorous. The increased convection is due to enhanced lubrication of the plates caused by a weak hydrous crust for strongly water dependent viscosity. The degassing history is insensitive to the initial water content of the early Earth as well as rheological strength. The degassing flux from Earth's surface is calculated to be approximately $O(10^{13})$ kg/yr, consistent with a coupled model of climate evolution and mantle thermal evolution.

1. Introduction

Geochemical observations and mineral physics experiments suggest that the water content of Earth's mantle should be around 1 to 2 OMs at the present time (OM = Ocean Mass; 1 Ocean Mass = 1.4×10^{21} kg) (e.g. Hirschmann, 2006). However, the origin and evolution of water in the Earth's mantle is not well-understood. Outstanding questions include: 1. Could the early Earth's mantle store some water during its formation, and if so, how much? 2. Could the water in the early Earth's formation still be stored in the present day mantle?

Regarding the first question, the water content of the early Earth's mantle is still debated in geo- and cosmo-chemical analyses as well as theoretical studies of planetary formation (Albarede, 2009; Tian and Ida, 2015; Genda and Ikoma, 2008; Elkins-Tanton, 2008; Hamano et al., 2013). Their estimates of water content in the early Earth's mantle range from Dry (0 OM) to ~10 OMs. Albarede (2009) predicted the mantle to be essentially dry right after the moon-forming impact and later supplied from icy asteroids inferred from geo- and cosmo-chemical analysis. Alternatively, inferences from observation and theoretical modeling on the formation of exosolar planets, a huge amount of water (~1.0 wt.%) was preserved in early planetary mantle rocks (Tian and Ida, 2015). In addition, during the solidification of a magma ocean, the water content of the early Earth's mantle could be up to 10 OMs (Hamano et al., 2013; Genda and Ikoma, 2008) or ~1 OM (Elkins-Tanton, 2008). Given such a wide range of estimates, the mantle water content of early Earth's mantle is still uncertain.

Regarding the second question, mantle thermal evolution calculations suggest that the residence time of water stored in the early Earth's mantle may be around 1 to 2 billion years and that the water content in the early Earth could be up to 4 OMs (Sandu et al., 2011). However, the water migration process in their model assumed only a degassing-regassing process, with the degassing process only occurring at Mid-Oceanic-Ridges (MORs). In mantle dynamics, the dehydration of volatiles during subduction is also important as well as the degassing-regassing process

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Table 1
Mantle model physical parameters. $Ra_0 = \rho_0 g \alpha_0 \Delta T_{sa} d^3 / \kappa_0 \eta_0$. Activation energy and volume for dry mantle are taken from Yamazaki and Karato (2001) and for wet mantle are from the wet rheology of olivine (Korenaga and Karato, 2008). The latent heat shown here is used for temperature feedback caused by partial melting (see Xie and Tackley, 2004).

Symbol	Meaning	Value
η_0	Reference viscosity	1.4×10^{21} Pa s
ρ_0	Surface density	3300 $\mathrm{kg\,m^{-3}}$
g	Surface gravity	9.8 $\mathrm{m\,s^{-2}}$
α_0	Surface thermal expansivity	5×10^{-5} $\mathrm{K^{-1}}$
κ_0	Surface thermal diffusivity	7×10^{-7} $\mathrm{m^2\,s^{-1}}$
ΔT_{sa}	Temperature scale	2500 K
C_p	Heat capacity	1250 $\mathrm{J\,kg^{-1}\,K^{-1}}$
L_m	Latent heat	6.25×10^5 $\mathrm{J\,kg^{-1}}$
H	Internal heating rate	3.7×10^{-12} $\mathrm{W\,kg^{-1}}$
E_d	Activation energy of dry mantle	290 $\mathrm{kJ\,mol^{-1}}$
V_d	Activation volume of dry mantle	2.4×10^{-6} $\mathrm{m^3\,mol^{-1}}$
E_w	Activation energy of wet mantle	380 $\mathrm{kJ\,mol^{-1}}$
V_w	Activation volumes of wet mantle	4×10^{-6} $\mathrm{m^3\,mol^{-1}}$

(e.g. Iwamori and Nakakuki, 2013; Wilson et al., 2014). A recent investigation, (Nakagawa et al., 2015) using global-scale mantle convection simulations that approximate water migration, suggests that the dehydration process might be more important than the degassing-regassing process because the mantle water content is strongly regulated by the dehydration process on a very short time-scale. In that study, the water content was regulated assuming only dehydration and did not include degassing-regassing and partitioning of water into partially molten material. Those processes should also be crucial for surface climate evolution as well as water content evolution in Earth's mantle (e.g. Tajika and Matsui, 1992; Sandu et al., 2011). In this study, we check the sensitivity of the water content of the current day mantle to the initial amount of water in the mantle after 4.6 billion years of planetary evolution. We present implications for possible water content in early Earth and its impact on the surface climate evolution and plate tectonics.

2. Numerical model

The governing equations and numerical procedure of mantle convection simulations have been described in detail in Nakagawa et al. (2015). All physical parameters for mantle convection are listed in Table 1. Detailed information on the governing equations and numerical procedures for water migration is referred to in the supplemental material. Briefly, we use the numerical code StagYY (Tackley, 2008), which includes solution of compressible and truncated anelastic mantle convection with simplified partial melting to form the oceanic crust and free-slip velocity boundary conditions at both top and bottom boundaries.

The solid rheology is temperature-, depth-, yield stress- and water-dependent. The viscosity formulation is given as

$$\eta_d = A_{d0} \sum_{i,j=1}^{nphase} \Delta\eta_{ij}^{\Gamma_{ij} f_j} \exp\left[\frac{E_d + pV_d}{RT}\right] \tag{1}$$

$$\eta_w = A_{w0} \left(\frac{C_w}{C_{w0}}\right)^{-r} \sum_{i,j=1}^{nphase} \Delta\eta_{ij}^{\Gamma_{ij} f_j} \exp\left[\frac{E_w + pV_w}{RT}\right] \tag{2}$$

$$\eta_Y = \frac{\sigma_Y}{2\dot{e}} \tag{3}$$

$$\sigma_Y = \min\left(\sigma_b + \frac{d\sigma_b}{dP} P, C_Y + \mu P\right) \tag{4}$$

$$\eta = \left(\frac{1}{\eta_d} + \frac{1}{\eta_w} + \frac{1}{\eta_Y}\right)^{-1} \tag{5}$$

where η_d and η_w are viscosities for dry and wet materials respectively. A_{d0}, A_{w0} are prefactors for dry and hydrous materials calculated at $d = 0$ km and $T = 1600$ K, C_w is the water content, C_{w0} is the reference water content set at 620 ppm (Arcay et al., 2005), r is the exponent of viscosity dependence of water content, $\Delta\eta_{ij}$ is the viscosity jump at the phase boundary between spinel and perovskite, Γ_{ij} is the phase function (indexes of i and j in both viscosity jump and phase function indicate mineral systems: i as the olivine system and j as the pyroxene system), f_j is the fraction of basaltic composition (e.g. Keller and Tackley, 2009), $E_{d,w}$ is the activation energy, $V_{d,w}$ is the activation volume (the subscripts d and w indicate dry and hydrous materials), R is the gas constant, and σ_b and $d\sigma_b/dP$ are the yield stress at the surface (50 MPa), and the yield stress gradient (0.009863) for the brittle-ductile transition layer, respectively, P is the lithostatic pressure, C_Y and μ are the cohesion (0 MPa) and friction coefficient (0.1) of the brittle layer, respectively, and $\dot{e}$ is the second invariant of the strain tensor. Regarding the pseudo-plastic yielding, again, we use the similar formation on the brittle-ductile transition (see Tackley (2000)) and the numerical code using here was benchmarked with 'viscoplastic mantle convection benchmark' (Tosi et al., 2015).

The solution domain is a 2-D spherical annulus, which is equivalent to the equatorial section of 3-D spherical shell (Hernlund and Tackley, 2008). Since this geometry is nearly equivalent to a 3-D spherical shell in terms of heat transfer contribution (Nakagawa and Tackley, 2010), we assume that the ratio of inner to outer radii is the same as that in Earth's mantle. The numerical resolution is assumed as 1024×128 with 4 million tracer particles to track chemical composition, melt fraction and water content of the solid phases. These particles only travel with the solid but track both basaltic and ambient mantle compositions. The initial conditions for temperature and compositional fields are assumed to be adiabatic temperature profiles at 2000 K of surface plus thin thermal boundary layers (30 km) with small random perturbations (20 K) and uniform composition (20% of basaltic material). The temperature at the surface and core-mantle boundary (CMB) is fixed as 300 K and 4000 K respectively. Here the core cooling effect is not assumed because we focus on investigating influences of initial water content, rheological properties to water content evolution as well as its water flux contributions. However, effects of core-mantle cooling would be important for understanding the evolution of water content of Earth but this would be the next step.

To calculate water content and excess water content in the mantle, water is transported as hydrous minerals in particles with either basaltic, or peridotitic composition. The concentration of water bound in each particle C_w is governed by

$$\frac{\partial C_w}{\partial t} + \underline{u} \cdot \nabla C = \frac{dC_w}{dt} = S_w \tag{6}$$

where $\underline{u}$ is the convective solid velocity and S_w is the net sources and sinks that a particle experiences. These sources/sinks include

1) Rehydration: interaction with the surface ocean to a maximum saturation of 6%, and any re-equilibration with available "excess water".
2) Melting: fractionation of water into a melt-phase for partially molten particles that are then instantaneously transported to the surface and "erupted".
3) Dehydration: production of "excess water" for particles that exceed their local saturation as determined by look-up tables of water solubility in upper mantle minerals (Iwamori (2004, 2007) and Fig. S2). In the lower mantle, water solubility is set at 0.01 wt.% (Karato, 2011). (See supplemental information for more details.)