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Hydrogen and helium ingassing during terrestrial planet accretion

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ABSTRACT

The amount of atmospheric gas absorbed into a planetary interior during formation depends on the temperature, pressure, composition, and dynamics of its atmosphere, along with the planet mass, its composition, rate of accretion, and the average age of its surface. We develop a boundary layer model for terrestrial planet ingassing during accretion that quantifies the amounts of hydrogen and helium dissolved into a global magma ocean from a hot, gravitationally captured nebular atmosphere in the first few million years of solar system history. Assuming that most of Earth's accretion occurred within the lifetime of its nebular atmosphere, one or more ocean masses of ingassed hydrogen are predicted if the surface pressure is high and the magma surface is young. In addition, Earth would have acquired hundreds of petagrams of helium-3 from the same nebular atmosphere. Our model predicts negligible hydrogen ingassing from a gravitationally captured nebular atmosphere on Mars, but shows that vast amounts of hydrogen ingassing – tens of ocean masses – are possible during the accretion of Super-Earth-sized terrestrial planets.

1. Introduction

The composition and abundance of volatiles offer major constraints on terrestrial planet evolution, starting from the earliest events in the solar system (Lammer et al., 2018). Processes that are widely recognized as being important for volatile loss from the interior of terrestrial planets include magmatism and volcanism, metamorphic devolatilization, and hydrothermal activity (Kerrick and Connolly, 2001; Wallace, 2005). Similarly, volatile loss is possible during planet formation through a variety of mechanisms, most notably giant impacts (Ahrens, 1993; Vickery and Melosh, 1990). Mechanisms that are recognized as being important for mantle volatile acquisition are harder to quantify; those that have been considered previously include solar wind implantation (Peron et al., 2017), incorporation of volatile ice phases, and accretion of hydrous bodies such as chondrites (Albarède, 2009).

In this paper we consider the efficacy of another mechanism, direct absorption of gas from a planet's atmosphere during accretion, a process hereafter referred to as atmospheric ingassing. Numerous studies have proposed that massive acquisition of hydrogen and helium occurred in the first few million years of Earth history (Mizuno et al., 1980; Harper and Jacobsen, 1996; Porcelli et al., 2001; Genda and Ikoma, 2008; Sharp et al., 2013; Sharp, 2017) when, it is assumed, accreting terrestrial planets were surrounded by atmospheres composed of nebular gas. The combination of high surface luminosity and high atmosphere pressure would lead to elevated surface temperatures and eventually a global magma ocean, especially if the nebular atmosphere persisted through the main phase of planetary growth. For example, calculations by Ikoma and Genda (2006) that consider wide ranges of planet luminosity, along with the mass, opacity, and surface pressure of its nebular atmosphere predict that once the mass of an accreting terrestrial planet exceeds 0.3 M_E (where M_E is Earth's mass), its surface temperature exceeds the melting points of mantle silicates. Impacts, atmospheric winds, and convection would then readily disturb the magma, promoting ingassing.

2. Atmosphere-ocean gas transfer

The transfer of gas between the atmosphere of an accreting terrestrial planet and an underlying liquid silicate magma ocean involves complex, multi-scale physics acting near the gas–liquid interface, including convection, diffusion, waves, and bubbles in the magma, plus convection, winds, evaporation, and condensation in the atmosphere. Fig. 1 shows the geometry in simplified form of the interfacial region between a silicate magma ocean and a hydrogen–helium rich nebular atmosphere, with the gas–liquid interface idealized as a planar surface. Ingassing from the atmosphere to the magma requires gas transfer across two adjacent boundary layers, one in the atmosphere, another in the magma. For magma volatile acquisition to occur, the gas concentration must decrease across the atmospheric boundary layer from its value in

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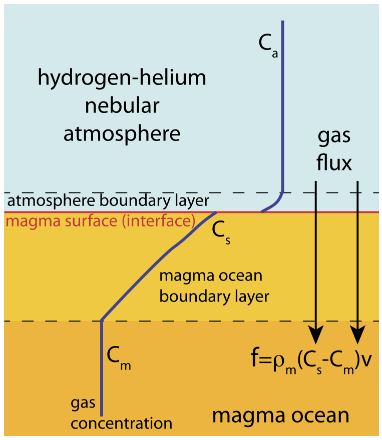

Fig. 1. Idealized geometry near the magma ocean–nebular atmosphere interface. Blue curve labeled C denotes the gas concentration in both fluids, f is the gas flux from atmosphere to magma ocean, v is the transfer velocity, ρ_m is magma density, the red curve marks the magma–atmosphere surface (here shown as planar) and the dashed lines mark the edges of the atmosphere and magma ocean boundary layers. Subscripts m, a, and s denote magma ocean, nebular atmosphere, and magma surface, respectively. (For interpretation of the colors in the figure(s), the reader is referred to the web version of this article.)

the open atmosphere (denoted by C_a) to a smaller value on the magma surface. The gas concentration decreases across the magma boundary layer from a value denoted by C_s, usually the equilibrium gas concentration in the silicate melt, to a smaller value in the magma ocean interior denoted by C_m. For weakly soluble gasses such as hydrogen or helium in silicate magma, the flux of gas from the open atmosphere (where its concentration is assumed to be nearly uniform because of turbulent mixing) to the magma interior (where its concentration is also assumed to be nearly uniform) is primarily limited by resistance within the magma boundary layer, across which the largest differences in gas concentration in the interfacial region occur.

In-situ measurements of atmospheric CO_2 transfer across the air–sea interface (Wanninkhof et al., 2009) have demonstrated that, under conditions similar to Fig. 1, the combined effects of all the dynamical processes that contribute to atmospheric ingassing can be parameterized in terms of a gas transfer velocity v, defined here so that the local mass flux f of gas from the atmosphere to the underlying liquid can be expressed as

$$f = \rho_m(\Delta C)v \tag{1}$$

where ρ_m is liquid density and $\Delta C = C_s - C_m$. In (1) and hereafter, the subscripts m and s denote magma ocean interior and magma surface, respectively, and subscript a denotes atmosphere, as shown in Fig. 1. In (1), f has units of mass per surface area per time and is positive for ingassing, while C_s and C_m are concentrations measured in terms of mass of gas per magma mass. For highly intermittent processes, it has been shown (Garbe et al., 2004) that atmospheric gas transfer is well-represented as a surface renewal process in which gas diffusion in the liquid and the local age of the liquid surface define the transfer velocity in (1). For intermittent surface renewal, as would result from a sequence of impacts or discrete atmospheric disturbances, the duration of the resurfacing events is very short compared to the average time between events. In that situation, the transfer velocity is just

$$v = (\kappa/\pi\tau)^{1/2} \tag{2}$$

where κ is the diffusivity of the gas species in the liquid and τ is the local age of the surface, the amount of time that particular liquid parcel has resided on the magma surface since the last resurfacing event.

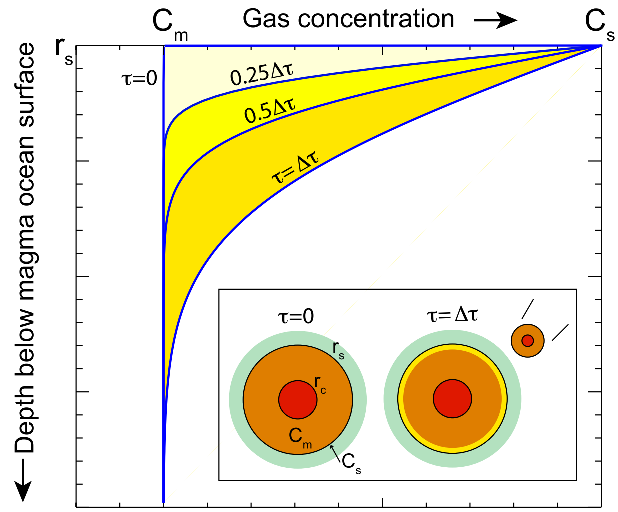

Fig. 2. Illustration of atmospheric ingassing through global resurfacing by intermittent impacts. Inset diagram shows beginning and ending stages of once cycle of surface renewal. C_m and C_s denote magma and magma surface gas concentrations, r_s and r_c denote radii of magma ocean and core, respectively. Light green represents the nebular atmosphere. At time $\tau = 0$ the surface has just been renewed by an impact; at $\tau = \Delta\tau$, ingassing has generated a mixed layer (light colored), immediately before the next impact and resurfacing event. Blue curves denote the depth variation of the gas concentration in the mixed layer of the magma at various surface ages τ during the cycle.

Fig. 2 illustrates the process of atmospheric ingassing by impacts that produce intermittent global resurfacing. The inset diagrams show the beginning and ending stages of one resurfacing cycle, immediately after an impact on the left and immediately before the next impact on the right. Light green shading represents the nebular atmosphere, yellow and brown represent the growing magma ocean (radius r_s) and red represents the growing core (radius r_c). At $\tau = 0$ in Fig. 2 the magma surface has just been renewed by an impact, its surface age is zero, and the magma interior has a uniform gas concentration throughout. Just before the next impact and resurfacing event the magma surface age is $\Delta\tau$ and the atmosphere ingassing has generated a gas-enriched boundary layer, depicted by the yellow-shaded region. The blue curves in Fig. 2 show the gas concentration versus depth in the magma ocean boundary layer at the beginning, ending, and at two intermediate stages of this cycle.

Using $\Delta\tau$ to denote the intermittency, i.e., the characteristic time between events that produce new surface, the time average of the local gas flux across the magma ocean surface produced by intermittent resurfacing is, from (1) and (2)

$$\bar{f} = 2\rho_m \Delta C(\kappa/\pi\Delta\tau)^{1/2}. \tag{3}$$

Here we have assumed negligible changes in ΔC between successive resurfacing events. Then according to (3), the average rate of gas transfer across the entire planet's atmosphere–magma surface (in units of mass per unit time) is given by

$$F_s = \bar{f} A_s = 2\rho_m \Delta C(\kappa/\pi\Delta\tau)^{1/2} A_s \tag{4}$$

where $A_s = 4\pi r_s^2$ is the surface area of the magma ocean, assumed spherical. For highly intermittent resurfacing, the intermittency is just twice the mean surface age $\bar{\tau}$, so that the total surface gas transfer (4) can alternatively be written

$$F_s = \rho_m \Delta C \bar{v} A_s \tag{5}$$