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Atmospheric mass loss during planet formation: The importance of planetesimal impacts

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ABSTRACT

Quantifying the atmospheric mass loss during planet formation is crucial for understanding the origin and evolution of planetary atmospheres. We examine the contributions to atmospheric loss from both giant impacts and planetesimal accretion. Giant impacts cause global motion of the ground. Using analytic self-similar solutions and full numerical integrations we find (for isothermal atmospheres with adiabatic index $\gamma = 5/3$) that the local atmospheric mass loss fraction for ground velocities $v_g \lesssim 0.25 v_{esc}$ is given by $\chi_{loss} = (1.71 v_g/v_{esc})^{4.9}$, where v_{esc} is the escape velocity from the target. Yet, the global atmospheric mass loss is a weaker function of the impactor velocity v_{Imp} and mass m_{Imp} and given by $X_{loss} \simeq 0.4x + 1.4x^2 - 0.8x^3$ (isothermal atmosphere) and $X_{loss} \simeq 0.4x + 1.8x^2 - 1.2x^3$ (adiabatic atmosphere), where $x = (v_{Imp} m / v_{esc} M)$. Atmospheric mass loss due to planetesimal impacts proceeds in two different regimes: (1) large enough impactors $m \gtrsim \sqrt{2}\rho_0(\pi h R)^{3/2}$ (25 km for the current Earth), are able to eject all the atmosphere above the tangent plane of the impact site, which is $h/2R$ of the whole atmosphere, where $h,\ R$ and ρ_0 are the atmospheric scale height, radius of the target, and its atmospheric density at the ground. (2) Smaller impactors, but above $m > 4\pi\rho_0 h^3$ (1 km for the current Earth) are only able to eject a fraction of the atmospheric mass above the tangent plane. We find that the most efficient impactors (per unit impactor mass) for atmospheric loss are planetesimals just above that lower limit (2 km for the current Earth). For impactor flux size distributions parametrized by a single power law, $N(>r) \propto r^{-q+1}$, with differential power law index q, we find that for $1 < q < 3$ the atmospheric mass loss proceeds in regime (1) whereas for $q > 3$ the mass loss is dominated by regime (2). Impactors with $m \lesssim 4\pi\rho_0 h^3$ are not able to eject any atmosphere. Despite being bombarded by the same planetesimal population, we find that the current differences in Earth's and Venus' atmospheric masses can be explained by modest differences in their initial atmospheric masses and that the current atmosphere of the Earth could have resulted from an equilibrium between atmospheric erosion and volatile delivery to the atmosphere from planetesimal impacts. We conclude that planetesimal impacts are likely to have played a major role in atmospheric mass loss over the formation history of the terrestrial planets.

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1. Introduction

Terrestrial planet formation is generally thought to have proceeded in two main stages: The first consists of the accretion of planetesimals, which leads to the formation of several dozens of roughly Mars-sized planetary embryos (e.g. Ida and Makino, 1993; Weidenschilling et al., 1997), and the second stage consists of a series of giant impacts between these embryos that merge to form the Earth and other terrestrial planets (e.g. Agnor et al., 1999; Chambers, 2001). Understanding how much of the planets' primordial atmosphere is retained during the giant impact phase is crucial for understanding the origin and evolution of planetary atmospheres. In addition, a planet's or proptoplanet's atmosphere cannot only be lost due to a collision with a comparably sized body in a giant impact, but also due to much smaller impacts by planetesimals. During planet formation giant impacts begin when the planetesimals are no longer able to efficiently damp the eccentricities of the growing protoplanets. Order of magnitude estimates that balance the stirring rates of the protoplanets with the damping rates due to dynamical friction by the planetesimal population and numerical simulations find that giant impacts set in when the

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total mass in protoplanets is comparable to the mass in planetesimals (Goldreich et al., 2004; Kenyon and Bromley, 2006). Therefore about 50% of the total mass still resides in planetesimals when giant impacts begin and planetesimal accretion continues throughout the giant impact phase. Furthermore, geochemical evidence from highly siderophile element (HSE) abundance patterns inferred for the terrestrial planets and the Moon suggest that a total of about 0.01 $M_\oplus$ of chondritic material was delivered as 'late veneer' by planetesimals to the terrestrial planets after the end of giant impacts (Warren et al., 1999; Walker et al., 2004; Walker, 2009). This suggests that planetesimal accretion did not only proceed throughout the giant impacts stage by continued beyond. Therefore, in order to understand the origin and evolution of the terrestrial planets' atmospheres one needs to examine the contribution to atmospheric loss from both the giant impacts and from planetesimal accretion.

Depending on impactor sizes, impact velocities and impact angles, volatiles may be added to or removed from growing planetary embryos by impacts of other planetary embryos and smaller planetesimals. The survival of primordial atmospheres through the stage of giant impacts during terrestrial planet formation has been examined by Genda and Abe (2003, 2005). These works numerically integrate the hydrodynamic equations of motion of the planetary atmosphere to determine the amount of atmospheric loss for various ground velocities. In contrast to giant impacts, smaller impactors cannot eject the planet's atmosphere globally but are limited to, at best, ejecting all the atmosphere above the tangent plane of the impact site. Some of the first calculations of impact induced atmospheric erosion were performed using the Zel'dovich and Raizer (1967) solution for the expansion of a vapor plume and momentum balance between the expanding gas and the mass of the overlying atmosphere (e.g. Melosh and Vickery, 1989; Vickery and Melosh, 1990; Ahrens, 1993). The results of these calculations were used to investigate the evolution of planetary atmospheres as a result of planetesimal impacts (e.g. Zahnle et al., 1990, 1992). Newman et al. (1999) investigated by analytical and computational means the effect of $\sim$10 km impactors on terrestrial atmospheres using an analytical model based on the solutions of Kompaneets (1960). Atmospheric erosion calculations were extended further by, for example, Svetsov (2007) and Shuvalov (2009) who investigated numerically atmospheric loss and replenishment and the role of oblique impacts, respectively. In the work presented here, we use order of magnitude estimates and numerical simulations to calculate the atmospheric mass loss over the entire range of impactor sizes, spanning impacts too small to eject significant amounts of atmosphere to planetary-embryo scale giant impacts. Our results demonstrate that the most efficient impactors (per impactor mass) for atmospheric loss are small planetesimals which, for the current atmosphere of the Earth, are only about 2 km in radius. We show that these small planetesimal impacts could have potentially totally dominated the atmospheric mass loss over Earth's history and during planet formation in general.

Our paper is structured as follows: in Section 2 we use analytic self-similar solutions and full numerical integrations to calculate the amount of atmosphere lost during giant impacts for an isothermal and adiabatic atmosphere. We analytically calculate the atmospheric mass loss due to planetesimal impacts in Section 3. In Section 4, we compare and contrast the atmospheric mass loss due to giant impacts and planetesimal accretion and show that planetesimal impacts likely played a more important role for atmospheric loss of terrestrial planets than giant impacts. We discuss the implications of our results for terrestrial planet formation and compare our findings with recent geochemical constraints on atmospheric loss and the origin of Earth's atmosphere in Section 5. Discussion and conclusions follow in Section 6.

2. Atmospheric mass loss due to giant impacts

When an impact occurs the planet's atmosphere can be lost in two distinct ways: first, the expansion of plumes generated at the impact site can expel the atmosphere locally but not globally. Atmospheric loss is therefore limited to at best $h/(2R)$ of the total atmosphere, where h is the atmospheric scale height and R the planetary radius (see Section 3 for details). Second, giant impacts create a strong shock that propagates through the planetary interior causing a global ground motion of the protoplanet. This ground motion in turn launches a strong shock into the planetary atmosphere, which can lead to loss of a significant fraction of or even the entire atmosphere Fig. 1.

It was realized several decades ago that self-similar solutions provide an excellent description for a shock propagating in adiabatic and isothermal atmospheres (e.g. Raizer, 1964; Grover and Hardy, 1966). Here, we take advantage of these self-similar solutions and use them together with full numerical integrations to calculate the atmospheric mass loss due to giant impacts.

2.1. Self-similar solutions to the hydrodynamic equations for an isothermal atmosphere

Terrestrial planet's atmospheres, like the Earth's, are to first order isothermal, giving rise to an exponential density profile. We therefore solve the hydrodynamic equations for a shock propagating in an atmosphere with an exponential density profile given by

$$\rho = \rho_0 \exp[-z/h], \tag{1}$$

where ρ_0 is the density on the ground, z the height in the atmosphere measured from the ground and h the atmospheric scale height. The atmosphere is assumed to be planar, which is valid for the terrestrial planets since their atmospheric scale heights are small compared to their radii. We further assume that radiative losses can be neglected such that the flow is adiabatic. The adiabatic hydrodynamic equations are given by

$$\frac{1}{\rho}\frac{D\rho}{Dt} + \frac{\partial u}{\partial z} = 0 \tag{2}$$

$$\frac{Du}{Dt} + \frac{1}{\rho}\frac{\partial p}{\partial z} = 0 \tag{3}$$

$$\frac{1}{p}\frac{Dp}{Dt} - \frac{\gamma}{\rho}\frac{D\rho}{Dt} = 0, \tag{4}$$

where γ is the adiabatic index and D/Dt the ordinary Stokes time derivative.

Thanks to the self-similar behavior of the flow, the solutions to hydrodynamic equations above can be separated into their time-dependent and spatial parts and can be written as

$$\rho(z,t) = \rho_0 \exp[-Z(t)/h]G(\zeta), \quad u(z,t) = \dot{Z}U(\zeta),$$
$$p(z,t) = \rho_0 \exp[-Z(t)/h]\dot{Z}^2 P(\zeta) \tag{5}$$

where $Z(t)$ is the position of the shock front and $\zeta = (z - Z(t))/h$. The similarity variables for the density, velocity and pressure are given by $G(\zeta)$, $U(\zeta)$ and $P(\zeta)$, respectively. Using the expressions in Eq. (5) and substituting them into the hydrodynamic Eqs. (2)–(4) yields for the spatial parts

$$\frac{1}{G}\frac{dG}{d\zeta}(U-1) + \frac{dU}{d\zeta} = 1 \tag{6}$$

$$(U-1)\frac{dU}{d\zeta} + \frac{1}{G}\frac{dP}{d\zeta} = -\frac{U}{\alpha} \tag{7}$$

$$(U-1)\left(\frac{1}{P}\frac{dP}{d\zeta} - \frac{\gamma}{G}\frac{dG}{d\zeta}\right) = -\frac{2}{\alpha} - \gamma + 1, \tag{8}$$

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