Enhanced atmospheric loss on protoplanets at the giant impact phase in the presence of oceans

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The atmospheric compositions of Venus and Earth differ significantly, with the venusian atmosphere containing about 50 times as much ^{36}Ar as the atmosphere on Earth[1]. The different effects of the solar wind on planet-forming materials for Earth and Venus have been proposed to account for some of this difference in atmospheric composition[2,3], but the cause of the compositional difference has not yet been fully resolved. Here we propose that the absence or presence of an ocean at the surface of a protoplanet during the giant impact phase could have determined its subsequent atmospheric amount and composition. Using numerical simulations, we demonstrate that the presence of an ocean significantly enhances the loss of atmosphere during a giant impact owing to two effects: evaporation of the ocean, and lower shock impedance of the ocean compared to the ground. Protoplanets near Earth's orbit are expected to have had oceans, whereas those near Venus' orbit are not, and we therefore suggest that remnants of the noble-gas rich proto-atmosphere survived on Venus, but not on Earth. Our proposed mechanism explains differences in the atmospheric contents of argon, krypton and xenon on Venus and Earth, but most of the neon must have escaped from both planets' atmospheres later to yield the observed ratio of neon to argon.

According to recent planetary formation theory, the terrestrial planets were formed in two stages: the formation of several tens of Mars-sized protoplanets through accretion of planetesimals[4], which was followed by collisions among these protoplanets[5,6]—that is, giant impacts. We term the latter stage the stage of giant impacts.

The formation of an atmosphere by impact degassing of volatile-containing planetesimals is expected during the formation of protoplanets[7]. Simultaneously, a protoplanet gravitationally attracts the surrounding nebular gas with solar composition[8], because protoplanetary formation (timescale $\sim 10^5 - 10^6$ yr; ref. 4) is considered to be completed before the dissipation of the surrounding nebular gas (timescale typically $\sim 10^7$ yr; ref. 9). At the stage of giant impacts (timescale $10^7 - 10^8$ yr; ref. 5), surrounding nebular gas has probably already been lost. However, the atmosphere on the protoplanet is trapped by the gravity of that protoplanet. Thus, Mars-sized protoplanets have proto-atmospheres composed of a mixture of solar and planetesimal components[10]. Most of the noble gases in such atmospheres are derived from the nebular gas with solar abundance, whereas most of H_2O and CO_2 are derived from planetesimals[10].

When a giant impact occurs, the atmosphere near the impact site is expelled by the expansion of vapour plumes that are generated at the impact site. However, this type of direct stripping cannot remove atmosphere that lies far from the impact site—direct stripping affects only 25% of the entire atmosphere at the most[11]. Instead, a giant impact creates a strong shock wave that travels through the planetary interior, thereby inducing a global ground motion. Such motion may expel the entire atmosphere. According to direct three-dimensional smoothed particle hydrodynamic (3D SPH) simulations of the giant impact at the escape velocity, the velocity of the ground motion is estimated to be approximately 6 km s^{-1} at the antipode of the impact[12] and to be smaller elsewhere (4–5 km s^{-1} on average).

In a previous study[11], we performed simulations of the atmospheric motion induced by the global ground motion. We found that when a Mars-sized planet strikes an Earth-sized planet, the former loses 30% of its atmosphere, and the latter loses 10%. In other words, 90% of the atmosphere on the Earth-sized planet survives, and 70% of the atmosphere of the Mars-sized impactor is supplied to the Earth-sized planet. We also estimated that a mutual collision of planets of the same size results in a 30% atmospheric loss. Using these results, we calculated the changes in the atmospheric mass after every giant impact (Fig. 1). Figure 1 shows that the atmospheric mass after the stage of giant impacts is approximately 3–4 times the atmospheric mass of a protoplanet. This result indicates that the proto-atmospheres formed before the stage of giant impacts play an important role in the present terrestrial atmospheres[13].

Here we focus on the effect of an ocean on the surface during the stage of giant impacts, which has not been considered in the previous studies[11,14–16]. According to calculations made using the radiative–convective equilibrium model of an H_2O-CO_2 atmosphere, an ocean can form when the energy flux radiated from the planet into space (F_{pl}) is less than ~ 300 W m^{-2} (ref. 17). As the energy released by the accretion of planetesimals is negligible just after the formation of protoplanets, F_{pl} can be approximated by the net solar radiation flux, that is, $S(1 - A)/4$, where S and A are the solar radiation flux and the planetary albedo, respectively. Considering the S of the early Sun to be 70% of the present value[18], that is, 960 W m^{-2} at 1 AU, F_{pl} is estimated to be 168 and 321 W m^{-2} for $A = 0.3$ at the orbits of Earth and Venus, respectively. Therefore, all the protoplanets near the Earth's orbit should have had oceans during the stage of giant impacts.

We calculate the loss fractions of the atmosphere (X_{atm}) and ocean (X_{oce}) caused by the ground motion induced by a giant

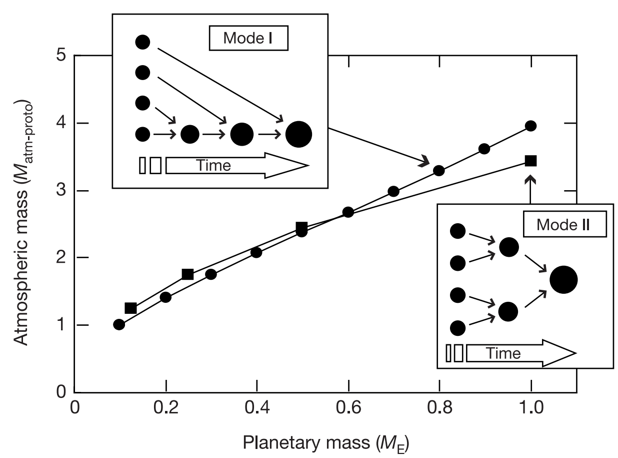

Figure 1 Changes in atmospheric mass during the stage of giant impacts. Two accretion patterns are considered. In mode I (top inset), a protoplanet grows to have the present Earth's mass (M_E) through nine giant impacts by protoplanets, each of mass 0.1 M_E. Each protoplanet has an atmospheric mass of $M_{atm-proto}$. In mode II (bottom inset), a protoplanet grows to mass M_E through three collisions with planets of the same size; in other words, a total of eight protoplanets (each of mass 0.125 M_E) collide with each other in the 'tournament'—each of these protoplanets has an atmospheric mass of 1.25 $M_{atm-proto}$. We consider giant impacts whose collision velocities are the escape velocity. The atmospheric mass of the Earth-sized planet finally formed does not entirely depend on these accretion patterns, and is approximately 3–4 times the mass of the proto-atmosphere of a Mars-sized protoplanet.

impact, assuming a spherically symmetric motion of the atmosphere and ocean (see Methods). Figure 2 shows the calculated X_{atm} and X_{oce} as a function of the initial ground velocity (u_g). We performed simulations for various initial conditions of the atmosphere (for example, various molecular weights and temperatures), ocean, and planetary mass. It is found that the relations between X_{atm} and u_g normalized by the escape velocity are sensitive only to the initial ratio of the atmospheric mass to the ocean mass (R_{mass}), and are insensitive to other initial conditions. For a given value of u_g, X_{atm} from an ocean-covered planet is always larger than that from a planet without an ocean. There are two enhancement mechanisms for atmospheric loss. One is the vaporization of the ocean; the ground motion induces complete vaporization of the ocean, and the vaporized ocean can efficiently push out the atmosphere. The other is impedance coupling. The shock impedance of the ocean is lower than that of silicate materials, and typically higher than that of the atmosphere (see Supplementary Information). Owing to such relations of shock impedances, the velocity at the ocean–atmosphere interface becomes larger than that of the ground motion.

In the following, we consider a proto-atmosphere composed of a mixture of solar and planetesimal components. We estimate the mass of the gravitationally attracted solar component on a Mars-sized protoplanet. Assuming the minimum mass disk model of a nebula[19], it is estimated to be $\sim 4 \times 10^{19}$ kg (see Supplementary Information). We assume an isothermal atmosphere, because the energy supply on the protoplanetary surface due to accretion of planetesimals has already finished during the stage of giant impacts.

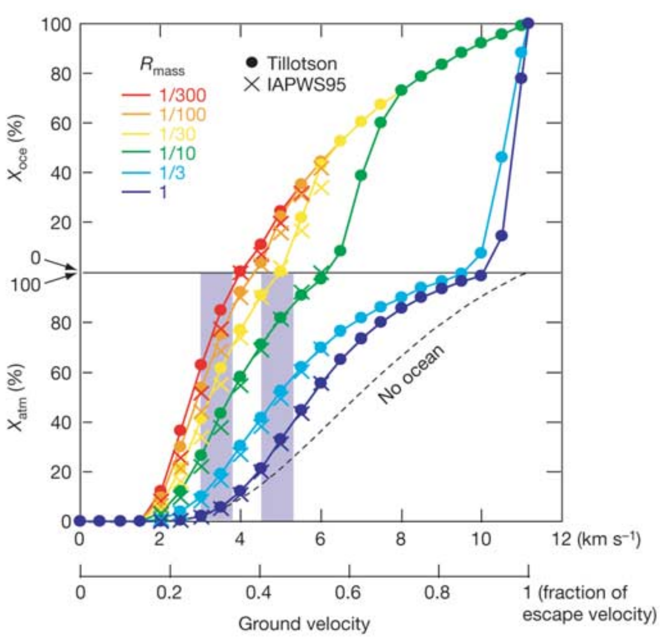

Figure 2 The loss fractions of atmosphere (X_{atm}) and ocean (X_{oce}) induced by the global ground motion with various initial ground velocities (u_g). The initial conditions of the atmosphere and ocean are described in the Methods. The results below $X_{atm} = 100\%$ imply some atmospheric loss and no oceanic loss. The results above $X_{oce} = 0\%$ imply complete atmospheric loss and some oceanic loss. In an actual giant impact, some fraction of the ocean is probably lost before the complete loss of the atmosphere. However, it is unlikely that X_{oce} is larger than X_{atm}, because the ocean initially exists below the atmosphere. Dashed curve, results without an ocean (case 1 atmosphere in figure 6 of ref. 11). Left- and right-hand vertical blue bars, range of u_g for an ocean-covered Earth-sized planet and an ocean-covered Mars-sized impactor, respectively, when the collision velocity of a giant impact is the escape velocity. u_g changes roughly linearly with the collision velocity.

As the solar abundance of ^{36}Ar is $7.6 \times 10^{-5}\,\mathrm{g\,g^{-1}}$ (ref. 20), this proto-atmosphere contains $\sim 3 \times 10^{15}$ kg of ^{36}Ar, which is approximately 15 times as much as Earth's ^{36}Ar content (2.06×10^{14} kg), and approximately one-third of the venusian ^{36}Ar content (1.0×10^{16} kg)[1]. The proto-atmosphere, containing a large amount of noble gases with the solar abundance, is quite different from the present Earth's atmosphere. However, such an atmosphere bears some resemblance to the present venusian atmosphere, as the venusian Ar/Kr and Kr/Xe ratios appear to be similar to the solar ones[21,22].

When the protoplanets are composed of planetesimals containing 1 wt% of H_2O on average[10], an ocean with a maximum mass of 6×10^{21} kg is formed, which corresponds to $R_{mass} \approx 1/100$. Because, in reality, the ocean mass depends on the partitioning of H_2O between the proto-atmosphere and the planetary interior, here, as a reference case, we consider that $R_{mass} = 1/10$. The u_g value estimated by the 3D SPH simulations is 4–5 km s^{-1} averaged over the entire surface of an Earth-sized planet. However, this value should not be directly applied to an ocean-covered planet, because the ground motion is suppressed by impedance coupling between water and rock, as compared with the free-surface case. We estimate that the decrease in u_g is approximately 25% (see Methods). Hence, we should adopt $u_g = 3$–$3.8\,\mathrm{km\,s^{-1}}$ instead of $u_g = 4$–$5\,\mathrm{km\,s^{-1}}$ for an Earth-sized planet covered with an ocean. From Fig. 2, X_{atm} is 30% for an Earth-sized planet with an ocean of $R_{mass} = 1/10$, while it is 10% for the case without an ocean. For a Mars-sized impactor and mutual collision of planets of the same size, X_{atm} is estimated to be 70% in the case of $R_{mass} = 1/10$, while it is 30% for the case without an ocean. In any case, no ocean escapes.

Figure 3 shows the changes in the atmospheric mass after every giant impact for an ocean-covered planet. Owing to extensive loss of the proto-atmosphere, the final atmospheric mass of an Earth-sized planet is much smaller than that without an ocean. For example, when $R_{mass} = 1/30$, it is approximately two orders of magnitude less than that on a planet without an ocean. In Fig. 3, we assume R_{mass} to be constant throughout the stage of giant impacts. However, in

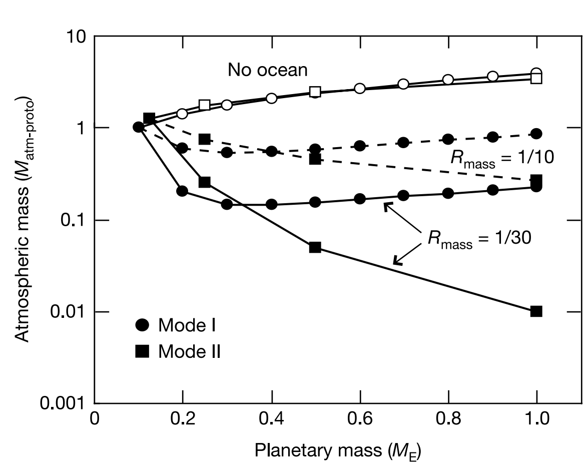

Figure 3 Changes in atmospheric mass for the cases with oceans. Filled circles and squares, results for accretion in mode I and mode II, respectively (see Fig. 1 legend). Open data points, results for the case without an ocean, which are the same as those of Fig. 1, but plotted on a logarithmic scale. We consider the cases for giant impacts whose collision velocities are the escape velocity. The amount of atmospheric loss from ocean-covered planets by each giant impact is estimated from the results shown in Fig. 2. The amount of atmosphere in the cases with an ocean becomes much smaller than that in the cases without an ocean. If impacts of ocean-covered and ocean-free planets occurred alternately, an atmospheric mass of approximately 1–2 $M_{atm\text{-}proto}$ finally survives the stage of giant impacts.

reality, R_{mass} is expected to decrease after each giant impact, because a large amount of the atmosphere is lost, while the entire ocean survives. Therefore, the final amount of atmosphere on an ocean-covered planet would be smaller than that shown in Fig. 3.

As discussed before, all planets near Earth's orbit should have oceans during the stage of giant impacts. Thus, the early Earth has experienced large-scale atmospheric losses after every giant impact, but almost the entire ocean would have survived. Although ocean formation on the planets near the orbit of Venus depends on the planetary albedo, the planets inside the orbit of Venus cannot have oceans. These atmospheres are in the runaway greenhouse state. Thus, a large amount of proto-atmosphere (approximately 3–4 times the mass of a proto-atmosphere of a protoplanet) survives the giant impacts on Venus.

Just after the stage of giant impacts, a tiny amount of the proto-atmosphere (except for H_2O) remains on the Earth, while a large amount of the proto-atmosphere with a solar-like noble gas pattern remains on Venus. This can explain the large difference in Ar abundance between these planets. However, proto-atmospheres with gravitationally attracted solar components also have a large abundance of Ne. Thus, they have a Ne/Ar ratio that is about 100 times higher than those of the present atmospheres. The observed Ne/Ar ratio may be created by subsequent evolutions, such as the supply and/or erosion of the volatile components during heavy bombardment[23–25], and the hydrodynamic escape of hydrogen (possibly also H_2O on Venus) due to solar ultraviolet radiation[21,22,26].

Although we have considered a mixed proto-atmosphere of solar and planetesimal components as an example, our main result (that is, the enhancement of the atmospheric loss due to an ocean) can apply irrespective of the type of pre-existing proto-atmosphere. In any case, the presence of an ocean during planetary formation is an important factor that caused the differences between the atmosphere of Earth and of Venus. □

Methods

Motion of an atmosphere and ocean

We consider a spherically symmetric motion of an atmosphere and ocean induced by the ground motion. We ignore radiative cooling and consider no ambient nebular gas, as in the previous study[11]. We use an ideal gas law for the EOS (equation of state) of the atmosphere. For the ocean, which is newly introduced here, we use two kinds of EOS; the Tillotson EOS[27] and the IAPWS95 (International Association for the Properties of Water and Steam Formulation 1995) EOS[28]. The Tillotson EOS is widely used in the simulation of shock waves; we use the parameter sets for water[29]. The IAPWS95 EOS is a high precision EOS specialized for H_2O, and has been used in the industrial field. We can exactly treat the vaporization of water by using the IAPWS95 EOS. Although the Tillotson EOS does not directly provide the fraction of vaporization, we can empirically estimate it from the internal energy.

As the initial conditions for an atmosphere, we consider a hydrostatically equilibrated polytropic atmosphere with given polytropic exponent (γ_a), atmospheric pressure (p_0) and temperature (T_0) at sea level, molecular weight (m_a) and specific heat ratio (γ). In Fig. 2, various values of p_0 (300, 100, 30, 10, 3 and 1 bar), $\gamma_a = 1.4$, $T_0 = 300$ K, $m_a = 2\,\mathrm{g\,mol^{-1}}$ and $\gamma = 1.4$ are adopted. ($p_0 = 300$, 100, 30, 10, 3 and 1 bar correspond to $R_{mass} = 1$, 1/3, 1/10, 1/30, 1/100 and 1/300, respectively, where R_{mass} is the initial ratio of the atmospheric mass to the ocean mass.) As the initial conditions for an ocean, we consider a hydrostatically equilibrated ocean with a depth of 3 km on an Earth-sized planet. We consider a constant internal energy distribution for the Tillotson EOS with $120\,\mathrm{J\,kg^{-1}}$, and an isentropic distribution for the IAPWS95 EOS with 300 K at sea level.

We do not solve the motion of the planetary interior induced by a giant impact. Instead, the ground motion is treated as the boundary condition of the bottom of the ocean. As in previous studies[11,16], we give the initial ground velocity u_g, and consider the subsequent ballistic motion of the ground—that is, the slow-down of the ground motion by gravity. We also assume that the motion ceases once the ground returns to the initial position.

Using a standard, one-dimensional, lagrangian, finite-differencing scheme[30], we integrate the conservational equations of the mass, momentum and energy for the atmosphere and ocean, and solve the motions of the atmosphere and ocean. We take 500 mass grids in the atmosphere, and 500 or 200 mass grids in the ocean for the Tillotson or IAPWS95 EOS, respectively. Although mixing of an ocean and atmosphere may occur due to Rayleigh–Taylor instability at the ocean–atmosphere interface, it could not be treated in the one-dimensional calculation. Nevertheless, we can estimate the upper bound of the mixing effect from the time during which the flow is subject to Rayleigh–Taylor instability. Since this duration time is less than 50 s, even if mixing occurs at the sound velocity, the mixed region is less than 20% of the entire atmosphere and ocean. Thus, the loss fraction of the atmosphere and ocean in Fig. 2 would not be significantly affected by the instability.

Ground velocity for an ocean-covered planet

When the shock wave travelling in the planetary interior (induced by a giant impact) arrives at the ground surface, the ground surface expands and its velocity (u_g) becomes faster than the particle velocity (u_p) in the planetary interior. When the ground is covered only by the atmosphere, the ground surface can be regarded as a free surface, and u_g accelerates up to $\sim 2u_p$. When the ground is covered by an ocean, the velocity at the ground surface u_g accelerates up to $\sim 1.5u_p$ (see Supplementary Information). This difference of the acceleration of the ground surface is due to the difference in the Hugoniot curves (the relation between the particle velocity and the shock pressure) of gas and water. Therefore, the ground velocity with an ocean is slower by ~25% than that without an ocean, when the impact conditions are the same, and thus the particle velocities (u_p) in the planetary interior are the same in both cases.

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