



Planetary Evaporation

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Evaporation of magma oceans exposed to space may have played a role in the chemical and isotopic compositions of rocky planets in our Solar System (e.g., Earth, Moon, Mars) and their protoplanetary antecedents. Chemical depletion of moderately volatile elements and the enrichment of these elements' heavier isotopes in the Moon and Vesta relative to chondrites are clear examples. Evaporation is also thought to be an important process in some exoplanetary systems. Identification of evaporation signatures among the rock-forming elements could elucidate important reactions between melts and vapors during planet formation in general, but the process is more complicated than is often assumed.

KEYWORDS: evaporation; isotopic fractionation; magma ocean; atmospheric escape

INTRODUCTION

The relative volatility of different elements played a central role in determining the chemical and isotopic compositions of the rocky bodies in the Solar System and, presumably, in extrasolar rocks as well. A thorough understanding of evaporation and sublimation on the one hand, and condensation on the other, is, therefore, important for understanding the formation and evolution of rocky planets and the planetesimals that built them. Evaporation of molten rock in the Solar System is evidenced unequivocally by the millimeter-to-centimeter size calcium–aluminum-rich inclusions (CAIs) found in chondrite meteorites (e.g., Davis et al. 1990). Many of these objects display enrichments in heavy isotopes of their primary constituents, including the elements Mg and Si. Because lighter isotopes evaporate more readily than heavy isotopes, heavy isotope enrichment is a telltale sign that residual melts experienced evaporation. Conversely, chondrules, which are solidified molten spheres that, like their igneous CAI cousins, were also floating in space in a molten state, do not show elemental and isotopic evidence of evaporation. The contrasting behavior of these primitive melts in space is testament to the importance of the environments in which they formed. Igneous CAIs formed under conditions that permit evaporation; chondrules formed under conditions precluding evaporation.

Similarly, the primary evidence for evaporative losses during planet formation is heavy isotope enrichment in several major and minor rock-forming elements. Arguably, the most sensitive indicators of evaporation/condensation during the formation of rocky bodies are the isotope ratios of moderately volatile elements (elements with 50%

condensation temperatures of 800 K to 1,300 K under nominal solar nebula conditions). The moderately volatile elements of K, Zn, and Rb in the Moon, Mars, and Vesta are enriched in heavy isotopes compared to those of the Earth and in chondrites (FIG. 1A). One explanation for this heavy isotope enrichment observed for the Moon and Vesta could be evaporation from magma oceans. Higher $^{25}\text{Mg}/^{24}\text{Mg}$ relative to chondrites in Earth (FIG. 1B) is another important, though more subtle, example. Indeed, Earth, Mars, Vesta (i.e., the howardite–

eucreite–diogenite meteorites), and the angrite meteorite parent body are all high in $^{25}\text{Mg}/^{24}\text{Mg}$ relative to chondrites, suggesting that evaporative losses may have occurred. Silicon and iron isotope ratios of planets and meteorite parent bodies have also been interpreted as being affected by melt–vapor fractionation, but other high-temperature planetary-scale processes, such as partial melting and core formation, are competing mechanisms that lead to ambiguities in the interpretations (Sossi and Shahar 2021).

In this review, we discuss the process of evaporation and its potential consequences for the chemical and isotopic composition of the planetary bodies. Instead of treating evaporation as a convenient catch-all explanation for planetary evolution, quantitative treatments of evaporation from a magma ocean must account for the kinetics of the evaporation process itself, as well as the physics of rock–vapor atmospheres above the melts, and the interplay between the two.

THE HEAT SOURCES FOR MAGMA OCEANS

In our Solar System, early formation of planetesimals heated by the decay of the short-lived radionuclide ^{26}Al ($t_{1/2} = 0.73$ My), led to global melting of bodies larger than ~20 km in radius and that had formed within the first 1.5 million years (Castillo-Rogez and Young 2017). Decay of ^{26}Al in an early formed body with the initial ^{26}Al abundance in the early Solar System ($^{26}\text{Al}/^{27}\text{Al} = 5.2 \times 10^{-5}$) produced three orders of magnitude more heat per kilogram than the heat induced by tides in Jupiter's moon Io, one of the most volcanically active bodies in the Solar System today (e.g., Davies et al. 2018). This prodigious heat production from ^{26}Al decay should have triggered vigorous convection in magma oceans exposed to space for ~ 10^5 years (e.g., Young et al. 2019).

Like lava lakes on the present Earth today, melts exposed to space form thin, ductile, viscoelastic crusts. These crusts are likely to be dynamic and transient owing to the convec-

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tion beneath. Solomatov (1995) offered a scaling for the thickness of the crust in terms of the Rayleigh number that characterizes the convection in the melt. The implication is that the crusts that form on magma oceans exposed to space will be on the order of centimeters in thickness for planetesimal-sized bodies. Such a thin crust will tend to founder and provide plenty of opportunities for exposing melt to space on a semi-continuous basis as long as convection persists.

After several million years in the early Solar System, ^{26}Al became effectively extinct by virtue of radioactive decay. Magma oceans still formed, however, as a result of collisions. One well-known example is the giant impact that likely formed the Moon (e.g., Cameron and Ward 1976). The nascent Moon was composed of molten rock which formed a lunar magma ocean. Lunar geochronology reveals that the crystallization of the entire magma ocean took perhaps ~ 200 My, even as the surface must have cooled much more quickly, after only $\sim 1,000$ years (e.g., Tang and Young 2020). The magma ocean has been considered one possible vehicle for loss of moderately volatile elements and isotope fractionation for the Moon (e.g., Dhaliwal et al. 2018). Heat generated by tidal interactions is another perennial source of heat for the formation of magma oceans in general, as exemplified by Jupiter's effect on Io today.

THE KINETICS OF EVAPORATION

The classical kinetic theory for evaporation and condensation dates back to late 19th century and is based on a Maxwellian velocity distribution of molecules and atoms in the gas phase, such that the probability of a given velocity is proportional to $\exp(-(v_x - U_x)^2)$, where v_x is the velocity of molecules perpendicular to the evaporating surface, and U_x is the mean velocity (see FIG. 2A). Maxwellian velocities apply far from the surface of the condensed phase. However, the presence of a surface perturbs this velocity distribution in complex ways in a region called the Knudsen layer. There is a large literature devoted to describing the behavior of gas molecules in the Knudsen layer (e.g., Bird and Liang 2019). One consequence of a modified distribution of velocities near the melt surface is the reflection of molecules back to the surface by intermolecular collisions, resulting in a “near-field” return flux ($J = \text{flux}$) that reduces the net evaporation flux from the melt. The net molar flux of a molecular species i with molecular mass m_i , assuming a single temperature, T , for the melt and gas can be approximated using a simplified form of the Hertz–Knudsen equation:

$$J_{i,\text{net}} = \frac{\gamma_i (P_{i,\text{eq}} - P_i)}{\sqrt{2\pi m_i RT}} \quad (1)$$

where $P_{i,\text{eq}}$ is the equilibrium vapor pressure and P_i is the extant partial pressure. The evaporation coefficient is usually determined empirically. This equation underscores that the net evaporative flux is the difference between the evaporative flux into a vacuum (proportional to $P_{i,\text{eq}}$) and the condensation flux associated with the return of gas to the melt (proportional to P_i). The saturation index, $S_i = P_i/P_{i,\text{eq}}$, is a convenient measure of the ratio of evaporative to return flux: when $S_i = 1$, equilibrium is attained; when $S_i < 1$, the net flux leads to evaporation. Saturation can also be written in terms of the ratio of return flux to net flux explicitly, so that $S_i = 1 - 1/[1 + J_{i,\text{return}}/J_{i,\text{net}}]$, illustrating that when the return flux is much less than the net flux, $S_i \sim 0$, and when the return flux is much greater than the net flux, $S_i \sim 1$.

The concept of saturation becomes more complicated when evaporation from spherical bodies is considered. At steady state, $S_i \propto 1 - 1/[1 + (r_p/D_i)]$ where r_p is the radius of the body and D_i is the pressure-dependent diffusivity through the gas (Richter et al. 2002; Young et al. 2019). The value for S_i in the limit of $r_p/D_i = 0$ is 0; and the limit where r_p/D_i goes to infinity is 1. The approach to saturation will, thus, depend on the size of the body and the environment in which it evaporates.

For a small-sized evaporating droplet with a radius smaller than the mean free path of the surrounding gas molecules, evaporated molecules will tend to miss the evaporating surface when reflected back towards that surface by intermolecular collisions (FIG. 2B). In the case of millimeter-to-centimeter CAIs in the early Solar System, the mean free path of H_2 molecules comprising the gas of the protostellar disk was on the order of 0.1 meters (pressures of about 10^{-5} bar). The CAIs, thus, experienced “free evaporation” with little or no return flux. Chondrules must have experienced significant return flux because they exhibit no evidence for evaporation. This indicates near-equilibrium partial pressures of rock-forming elements where chondrules formed.

In contrast, asteroids, moons, and planets have radii that far exceed the mean free paths of the gases into which evaporation would have taken place. In these cases, the return flux back to the melt is influenced by the “far-field pressure.” The far-field pressure is the pressure far removed from the

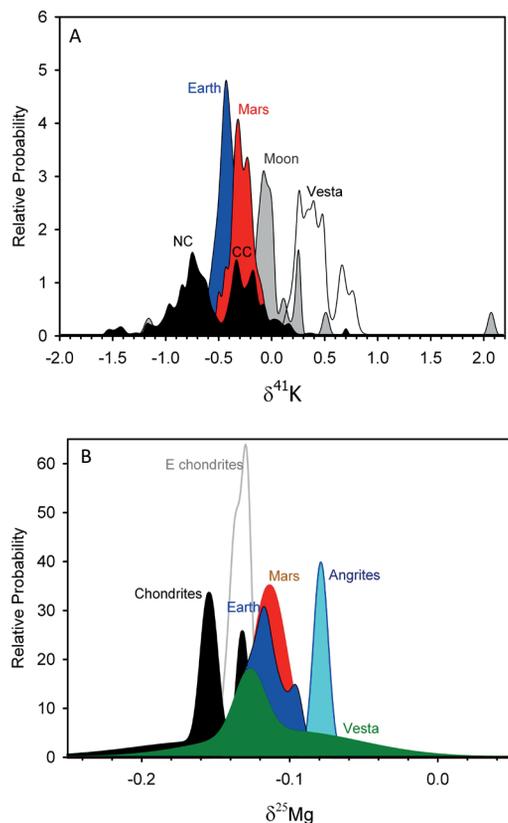


FIGURE 1 (A) Probability distributions for literature compilations of $^{41}\text{K}/^{39}\text{K}$ ($\delta^{41}\text{K}$) and their measurement uncertainties, representing a moderately volatile element (K) for various Solar System rocky bodies (Hin et al. 2017; Bloom et al. 2020; Ku and Jacobsen 2020; Tian et al. 2020a, b and references therein). (B) Probability distributions for literature compilations of $^{25}\text{Mg}/^{24}\text{Mg}$ ($\delta^{25}\text{Mg}$), Mg being a more refractory element. References as for 1A. Both isotope ratios are shown as per mil deviations from their respective standards. Abbreviations: NC = noncarbonaceous chondrites; CC = carbonaceous chondrites; E chondrites = enstatite chondrites.

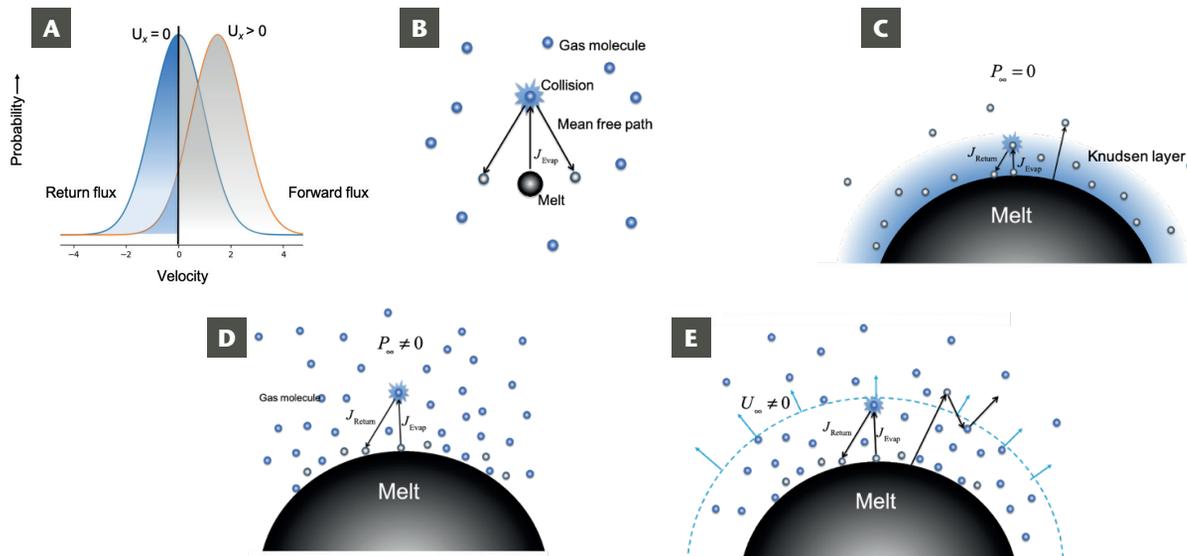


FIGURE 2 (A) Schematic illustration of the relative contributions of return flux (blue area) to a melt surface and forward evaporative flux (grey area) for two fictive molecular velocity distributions, one with no net velocity above the melt ($U_x = 0$, melt–vapor equilibrium) and the other with a significant net positive velocity ($U_x > 0$, evaporation). (B) Evaporating droplets are smaller than the mean free path of the enveloping gas, leading to free evaporation. This scenario applies to calcium–aluminum-rich inclusions (CAIs) in the early Solar System. Symbol: $J_{\text{Evap}} =$

evaporative flux. (C) Evaporation from the surface of a magma ocean exposed to space where there is a finite return flux (J_{Return}) due to the dispersion of velocities in the Knudsen layer above the melt. Symbol: $P_\infty =$ far-field pressure. (D) Evaporation from the surface of a magma ocean where a finite pressure of gas overlies the melt surface. Symbols as for 2B and 2C. (E) Similar to 2D but where there is a net advective velocity of gas, U_x , represented by the blue arrows, that decreases the return flux.

evaporating surface (e.g., several diffusion lengths scales). Where the far-field pressure P_∞ is zero, a finite return flux is still possible because of collisions in the Knudsen layer immediately above the evaporation surface (FIG. 2C). In the case where evaporation takes place in the presence of an ambient gas with $P_\infty \neq 0$, a return flux is expected regardless of the details of the Knudsen layer (FIG. 2D). The scenario in FIGURE 2C is perhaps most applicable for sublimation of ices from comets (Davidsson et al. 2021) whereas the scenario in FIGURE 2D is most applicable for evaporation of magma oceans where a steady-state atmosphere builds up (Tang and Young 2020). Lastly, the return flux will be less if there is an acceleration of evaporated molecules immediately above the evaporating surface (FIG. 2E).

Because of the prospects for significant return fluxes under almost all circumstances that are relevant for evaporation from a magma ocean, vapor that is emitted from magma on a planetary body will tend to approach saturation to varying degrees ($P_i \rightarrow P_{i,\text{eq}}$ and $S_i \rightarrow 1$). Planetesimals may evaporate into the H_2 gas of the protostellar disk, or a body with sufficient mass might accrete a higher-pressure primary atmosphere of hydrogen from the surrounding disk gas, hindering evaporation. Even in a pure vacuum, the existence of a Knudsen layer near the melt surface can produce saturation values as high as 70% ($S \sim 0.7$) (Anisimov 1968). Saturation values due to a return flux in the Knudsen layer of order 0.2 are often cited when considering sublimation of water-ice from comets (Davidsson et al. 2021).

ISOTOPE FRACTIONATION

Isotope fractionation associated with evaporation of a magma ocean will be sensitive to the ratio of evaporative flux to return flux (i.e., on the net flux), or, put another way, the degree of saturation. The isotope fractionation factor is $\alpha = R_{\text{vapor}}/R_{\text{melt}}$ where R is the isotope ratio of interest (e.g., $^{41}\text{K}/^{39}\text{K}$ or $^{25}\text{Mg}/^{24}\text{Mg}$). It is customary to express fractionation as per mil deviations from a reference value using the delta notation, and the difference between delta values

for two phases is commonly assigned the capital delta symbol (Ibañez-Mejía and Tissot 2021). Absolute values for equilibrium fractionations between vapor and melt at high temperatures are generally $\leq 0.1\text{‰}$ for the rock-forming elements. However, purely kinetic fractionation, due to differences in the velocities of the isotopic species for a given element, are generally tens of per mil (Watkins and Antonelli 2021). In the simplest case, the kinetic fractionation factor is the square root of the inverse of the isotopic masses, $\alpha_{\text{kinetic}} = \sqrt{m/m'}$, where m is the mass of the light isotopic species and m' is the mass of the heavy isotopic species (e.g., Graham’s law). The saturation index S_i can be used to estimate the relative contributions of equilibrium and kinetic fractionation to the isotope fractionation attending a net evaporative flux using

$$\Delta_{\text{Net}} = \Delta_{\text{Eq}} + (1 - S_i)\Delta_{\text{Kin}} \quad (2)$$

where Δ_{Net} , Δ_{Eq} , and Δ_{Kin} refer to the net, equilibrium, and kinetic fractionation factors expressed as Δ values (vapor–melt). More complicated formulae are required to relate the saturation, and, thus, the net fractionation factor, to pressure (Young et al. 2019). Two relevant examples are shown in FIGURE 3. They illustrate that the far-field, or background, pressure and the size of the evaporating body control the extent to which isotopic equilibrium is achieved at the melt–vapor interface. The background pressure is controlled by the structure and dynamics of the atmosphere overlying the magma ocean.

The far-field pressures of gas above a molten planetary body, and, thus, the isotopic effects of evaporation, will be controlled by a balance between the rate of evaporation, i.e., the source of the atmosphere, and the rate of vapor escape (the sink). Both of these rates will be affected by the structure and behavior of the atmosphere.

PREDICTING EVAPORATIVE FLUXES

Substitution of the thermodynamically defined equilibrium vapor pressures into Equation (1) leads to a general expression for the free evaporation flux of metal M from the

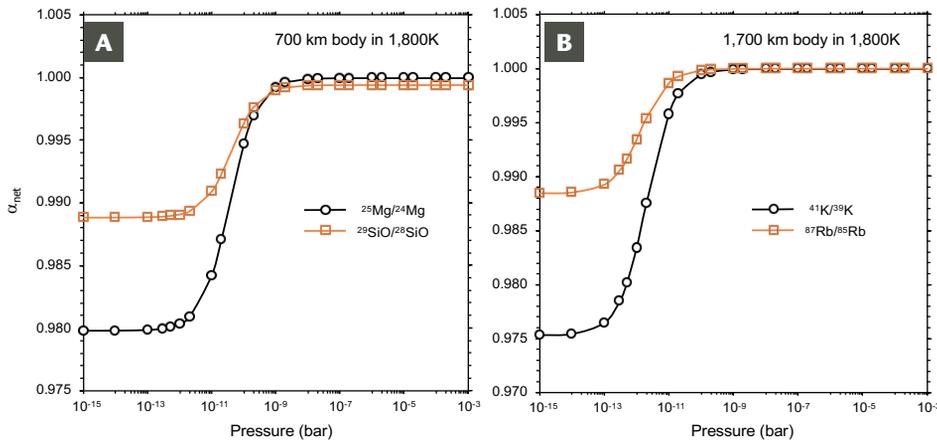


FIGURE 3 Effect of far-field pressure and body size on isotope fractionation. **(A)** Net isotope fractionation factors for Mg and Si for evaporation from the surface of a 700 km body. Symbol: α_{net} = net value for $R_{\text{vapor}}/R_{\text{melt}}$ where R is the isotope ratio of interest. **(B)** Net isotope fractionation factors for K and Rb for evaporation from a Moon-sized body. Temperature is 1,800 K in both cases.

simple oxide melt component MO_x (x is the stoichiometric coefficient for the oxygen):

$$J_M = a_{\text{MO}_x}^{\text{melt}} \frac{\exp\left(-\frac{\Delta\hat{G}_{\text{MO}_x \rightarrow M + 1/2x\text{O}_2}}{RT}\right)}{\sqrt{2\pi RT m_M}} P_{\text{O}_2}^{-(1/2)x} \quad (3)$$

where $\Delta\hat{G}_i$ is the molar free energy of evaporation reaction $\text{MO}_x \rightarrow M + (1/2)x\text{O}_2$ at standard state; $a_{\text{MO}_x}^{\text{melt}}$ is the activity for oxide MO_x in the evaporating melt; and m_M is the molar mass of metal species M . This expression illustrates that the activities of the oxides in the melt, at least in part, control evaporation rates. If the actual speciation in the liquid from intramelt reactions such as $\text{MgO} + \text{SiO}_2 \rightarrow \text{MgSiO}_3$ can be characterized, the activities can be modeled. This is the approach taken in the MAGMA code (e.g., Fegley and Cameron 1987). The alternative is to measure the activities experimentally. In practice, the evaporation coefficient is difficult, if not impossible, to deconvolve from a nonideal activity–composition relationship, and for this reason is omitted in this purely thermodynamic expression.

Equation (3) shows that the partial pressure of O_2 (P_{O_2}) may affect the rate of evaporation of metal species M through different values for the stoichiometry of the reactant oxides. This prompted O'Neill and Palme (2008) to suggest that the ratio of fractional losses of Na to those in Mn, two moderately volatile elements but with different oxide stoichiometries, $\text{NaO}_{0.5}$ and MnO , with x values of $1/2$ and 1 in Equation (3), respectively, could be used to monitor the oxygen fugacity (nonideal P_{O_2}) during volatilization; high oxygen fugacity should lead to greater fractional losses of Na relative to Mn.

The ratio of the fractional change in elemental abundance i ratioed to that in j in the evaporating melt is

$$\frac{\left(\frac{\Delta n_i}{n_i^0}\right)}{\left(\frac{\Delta n_j}{n_j^0}\right)} = \frac{J_i x_j^0}{J_j x_i^0} \quad (4)$$

where x_i are the mole fractions of the elements in the melt; superscript 0 refers to the initial concentrations; and Δn_i and Δn_j are the changes in molar abundances. This handy relationship permits an estimation of the consequences of evaporation based on Equation (3) and tabulations of thermodynamic data. Application of Equation (4) shows that if bodies with evidence for heavy Mg isotope enrich-

ment relative to chondrites, such as Earth (FIG. 1), inherited this signature from evaporation of planetesimal magma oceans, little if any of the moderately volatile elements remained in those planetesimals, requiring later addition of those elements by other bodies that did not experience the same level of evaporative losses (Hin et al. 2017; Young et al. 2019).

STRUCTURES OF ROCK-VAPOR ATMOSPHERES

The structures of rock-vapor atmospheres are becoming central to studies of bodies in the early Solar

System and for the evolution of rocky exoplanets, especially short-period planets tidally locked to their host star (e.g., Kite et al. 2016). In the latter case, winds driven from the day side to the night side are a critical facet of the prospects for volatilization from the surface. Bodies in the Solar System most likely did not have such stark temperature contrasts due to external sources of irradiation. The Moon immediately after its formation is a potential exception due to Earth shine, although this would require that outgassing from the Earth failed to form an optically thick atmosphere by the time the Moon coalesced.

Did the atmosphere overlying the magma ocean of a planetary body have a homosphere (the lower atmospheric layer where the bulk gases are well mixed) dominated by advective mixing and an overlying heterosphere (the overlying layer where the gases are sorted by molecular weight) in hydrostatic equilibrium, or did hydrodynamic escape modify or disrupt this equilibrium structure? Because the timescales to build up an atmosphere from evaporation of the global magma ocean are relatively short—on the order of days—a steady state between the surface-integrated evaporative flux (i.e., moles or kg s^{-1}) and the surface-integrated escape flux is expected. Low escape fluxes relative to evaporative fluxes will yield relatively high steady-state pressures in the vicinity of the molten planetary body, and an evaporative flux balanced by a higher escape flux will produce low steady-state pressures. Rock-vapor rain, or clouds, are expected if the total entropy of the vapor is less than that of liquid and vapor together (Tang and Young 2020). Isothermal atmospheres inhibit cloud formation, while adiabatic expansion and attendant cooling favors cloud formation.

ATMOSPHERIC ESCAPE

Regardless of the details of the evaporation process itself, a zero-order constraint on the viability of evaporation of magma at the surface is the ability of the vapor to escape by overcoming the gravity of the body. Equating the surface escape velocity with the mean thermal velocity of the gas molecules indicates that planetoids more massive than about 0.1% that of Earth may retain gases emitted by evaporation. At face value, this suggests that asteroids the size of Vesta, which experienced melting, could have lost volatile elements by evaporation, but larger bodies such as the Moon could not.

However, this simple criterion does not consider the details of the physics of escape. The capacity of a heated gas to escape the host body is often evaluated using the escape parameter, $\lambda = (GMm_{\text{gas}}/r)/(k_b T(r))$, where M = the mass of the body; $T(r)$ = the temperature at radius r from the center of the body; G = the gravitational constant;

m_{gas} = the mass of the vapor; k_b = the Boltzmann constant. The escape parameter is the dimensionless ratio of gravitational energy to thermal energy. The escape parameter at the surface of the body, λ_o , is a convenient predictor for the mechanism of escape. Values larger than ~ 3 suggest that the gravity of the body wins over the thermal energy of the atmospheric molecules (Volkov et al. 2011), and the primary escape mechanism is Jeans' escape, which occurs at the top of the atmosphere and for which individual molecules are virtually "collisionless" with clear paths for unhindered escape so that a small fraction escape purely by virtue of their thermal velocities exceeding the escape velocity of the body. Values of λ_o less than ~ 3 signify that the thermal energy of the gas is sufficient for the formation of a planetary "wind" resulting from hydrodynamic escape (also known as a Parker wind). Hydrodynamic escape is a thermally driven motion of colliding gas molecules that accelerates them outward en masse. In general, Jeans' escape is expected for a colder, thinner atmosphere on a planet-sized body, whereas hydrodynamic escape is expected for a hot atmosphere with relatively higher pressures on a smaller planetary body. The transition corresponds to about 10% of the mass of the Moon. In the special case of the Moon, the presence of the nearby Earth may have facilitated hydrodynamic escape. The closer the Moon was to the Earth, the lower the critical radius for hydrodynamic escape (Tang and Young 2020).

Isotope fractionation attending Jeans' escape is notoriously large, on the order of tens of per mil. This is because the velocity of the escaping molecules is proportional to the inverse of mass to the $\frac{1}{2}$ power, in keeping with Graham's law. Whereas the fractionation is large, the fraction of rock-forming elements in the atmosphere at any given time is orders of magnitude smaller than that in the melt, and exchange between melt and the upper atmosphere will be transport limited. Mass balance, therefore, works against Jeans' escape as an effective mechanism for fractionating the isotopic compositions of magma oceans.

Isotope fractionation attending hydrodynamic escape is notoriously small. This is because the upward acceleration of gas molecules experiencing mutual collisions overwhelms gravitational settling effects that depend on mass. For a rock-vapor atmosphere, isotopic fractionation is negligible for elements such as K, Mg, and Si.

EVAPORATION OF PLANETESIMALS AND THE MOON

Based on the above considerations, isotope fractionation attending evaporation of a magma ocean exposed to space is not as efficient as is sometimes supposed. The fractionation must occur at the melt surface, and the buildup of rock vapor above the melt must not be so great as to cause a return flux to the melt, so rivaling the evaporative flux from the melt. Two cases relevant to the Solar System illustrate these effects. The first is the potential for evaporation of planetesimals in the early Solar System. Hin et al. (2017) and Norris and Wood (2017) proposed that escape of vapor that is in near-equilibrium with a melt explains the Mg isotopic compositions of Solar System rocky bodies (FIG. 1) and the relative abundances of volatile chalcophile elements. In the latter case, near-equilibrium evaporation at oxygen fugacities (nonideal partial pressures of O_2) orders of magnitude greater than that defined by a hydrogen-rich protoplanetary gas is indicated, similar to the intrinsic oxygen fugacities of many rocky bodies in the Solar System. Both studies argued that perhaps Earth and other large bodies inherited the chemical signatures of evaporation from their antecedent planetesimals.

Evaporating Planetesimals

Some planetesimals in the early Solar System were heated from within by the decay of ^{26}Al to ^{26}Mg . The presence of an internal heat source would have produced a magma ocean at, or very near to, the surface of the body for timescales of 10^5 years (Young et al. 2019). The presence of the hydrogen-rich protostellar disk gas had two opposing effects on the evaporation process. Firstly, hydrogen accelerates the rate of evaporation, as can be seen in laboratory experiments (e.g., Young et al. 1998) and is understood in the context of thermodynamics as resulting from the decrease in the partial pressure of O_2 liberated by evaporation due to reaction with hydrogen. Secondly, the ambient hydrogen gas enhances the opportunities for a return flux, slowing the net rate of evaporation. Young et al. (2019) found that the best fit for the Mg and Si isotopic compositions of the bulk Earth relative to chondrites is obtained by the evaporation of the planetesimals that built the Earth at near-equilibrium conditions, as imposed by the surrounding H_2 gas of the protostellar disk on timescales of 10^4 years (FIG. 4). This result suggests that evaporation from the magma oceans of planetesimals that formed within the first ~ 1 My of the formation of the Solar System could have imparted small, but measurable, enrichments in the heavy isotopes of major rock-forming elements (e.g., Mg) to the larger bodies formed by accretion. However, this scenario does not explain the higher oxygen fugacities implied by the volatile chalcophile element abundances.

Evaporation of the Lunar Magma Ocean

The source of heavy isotope enrichment in the lunar moderately volatile elements is the target of intense study, and a number of scenarios related to the existence of the lunar magma ocean have been put forward. Tang and Young (2020) concluded that a significant return flux from the rock-vapor atmosphere above the surface of the magma ocean would have prevented anything but negligible isotope fractionation in elements such as K and Rb. In this work, the rock-vapor atmosphere of the lunar magma ocean is considered to be composed of two layers,

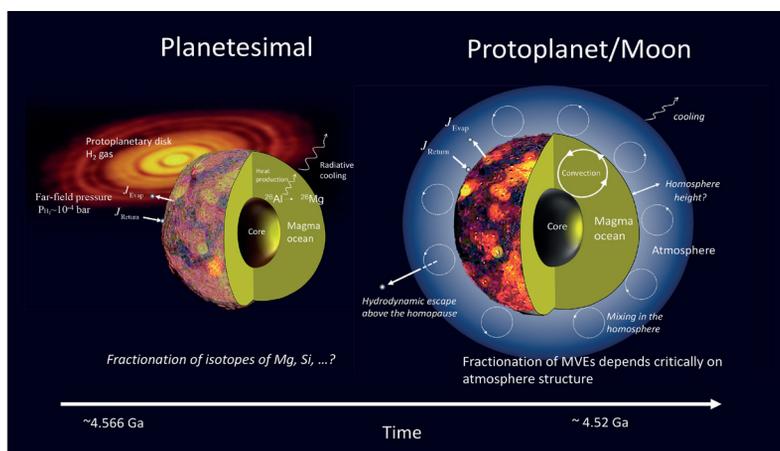


FIGURE 4 Schematic illustration of two scenarios for evaporation from magma oceans. (LEFT) The evaporation of a planetesimal into a background gas of hydrogen in the protostellar disk results in isotope fractionation of rock-forming elements (Young et al. 2019). (RIGHT) Evaporation of a magma ocean on a proto-planet or moon. Depending upon the structure of the atmosphere, fractionation of moderately volatile elements, such as K, would be limited by the presence of a homosphere, hydrodynamic escape occurring above the homopause (Tang and Young 2020). This scenario may apply to the Moon. Conversely, if hydrodynamic escape begins at the undersaturated melt surface, fractionation is possible (Charnoz et al. 2021). Symbols: J_{Evap} = evaporative flux; J_{Return} = return flux; MVEs = moderately volatile elements.

a homosphere that was well-mixed by convection and a heterosphere above the convective layer. Acceleration of the hot atmosphere resulted in hydrodynamic escape, augmented by the presence of the nearby Earth (FIG. 4). Neither the near-equilibrium evaporation process nor the atmospheric escape process would have been effective in causing isotope fractionation of moderately volatile elements. If there was no homosphere, the only source of a return flux would be the dispersion in molecular velocities directly above the surface of the lunar magma ocean modified by upward acceleration of the molecules, again facilitated by close proximity of the Earth. Charnoz et al. (2021) considered this alternative circumstance as a mechanism for depleting the Moon in moderately volatile elements. In this case, isotope fractionation attending evaporation would be mainly kinetic, and a short duration of evaporation of <100 years must be invoked to limit the degree of heavy isotope enrichments in K, Rb, and other moderately volatile elements.

If as some studies suggest, evaporation from the magma ocean was not the source of moderately volatile element depletions and isotope fractionation, focus must turn to the protolunar disk of melt and gas from which the Moon accreted. Incomplete condensation as the Moon formed is an important mechanism for separating elements and isotopes by volatility (Lock and Stewart 2017; Pahlevan

and Stevenson 2007), as is gas transport, which can, under the right conditions, reduce the saturation of moderately volatile elements in the disk gas (Nie and Dauphas 2019).

SUMMARY

Evaporation of magma oceans is a complicated process that entails feedbacks between the evaporation process itself and the dynamics of the atmosphere above the melt. In principle, evaporation can modify planetary isotopic compositions. Answering the specific question of whether evaporation of the lunar magma ocean could have shifted the Moon's moderately volatile element isotopic compositions will require a further understanding of rarefied atmospheres above magma oceans. Gone are the days where one can simply appeal to evaporation of melt into space as a convenient explanation for depletions in moderately volatile elements and enrichments in their heavy isotopes. The processes proposed must be evaluated using the kinetics of the evaporation process and the physics of rock-vapor atmospheres.

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