A MODEL OF THE MOON'S VOLATILE DEPLETION. Steven J. Desch[1] and **G. Jeff Taylor**[2].
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The Moon is widely accepted to have formed from a proto-lunar disk, formed after the impact of a Mars-sized planetary embryo with the Earth [1]. The disk material derived 20% from the Earth's mantle, and 80% from the impactor's mantle [2,3], which probably was similar to Earth's. During the disk's existence, material is hypothesized to have exchanged between the Earth and disk, to explain the remarkably similar oxygen isotopic composition of the Earth and Moon [4]. In this picture, volatiles should have exchanged as well, so the Moon may be expected to have formed with the terrestrial abundances of H_2O (≈ 500 ppm; [5]) and Na (≈ 2500 ppm; [6]). In fact, the Moon is depleted in these and other volatiles. Recent analyses of lunar samples suggest a bulk lunar abundance of H_2O in the range 0.1 - 30 ppm [7-10]; to fix numbers we consider an abundance of 10 ppm. Our aim is to quantify the loss of volatiles during the protolunar disk stage, to explain why the Moon is depleted in volatiles.

All volatile loss must be by hydrodynamic escape. Loss of 500 ppm of H_2O from an area $\sim 50\pi R_\oplus^2$, in $< 10^2$ yr, implies fluxes $> 3 \times 10^{16}$ cm^{-2} s^{-1}, for which the crossover mass [11] is $\sim 10^7$ amu. Therefore gases are lost according to their proportions in a disk atmosphere above and below the magma disk. Our model calculates how much of each volatile (such as H_2O) is dissolved in the magma (with mass fraction x_s) in each disk annulus, and how much is in the disk atmosphere (with partial pressure P_{H2O}). For example, $x_s = (6 \times 10^{-7})(P_{H2O}/1\,\mathrm{dyn\,cm^{-2}})^{0.54}$ [12], and typically $1 - 15\%$ of the H_2O is dissolved in the magma. We calculate the rate of hydrodyanmic escape of these (orbiting) gases from the disk-Earth system, accounting for the fact that the gas has high temperature T and has varying mean molecular weight. An important result is that the mean molecular weight is typically ~ 20 amu unless $T > 2800$ K, when rock vapor dominates (assuming $X_{Fa} = 0.09$ and the vapor pressure of [13]). At low densities, we find, thermal dissociation of H_2O and thermal ionization of Na and K, can lower the mean molecular weight. We account for hydrodynamic loss deeper than the exosphere by using the variation of mass loss with Knudsen number derived by [14] based on molecular Monte Carlo simulations. Gases in the atmosphere are not bound to the Moon after it forms, which will thus accrete only those volatiles dissolved in the magma.

We consider a $\sim 2\,M_L$ disk from $1\,R_\oplus$ to $5\,R_\oplus$ that then viscously spreads as described by [1]. Viscous heating maintains high temperatures iuntil the surface density drops because of the disk spreading. Temperatures ~ 2100 K, consistent with solidifcation of rock and possibly the triggering of the Moon's formation, are reached at the Roche limit $2.9\,R_\oplus$ after about 20 years, and the Moon may form quickly after that. In Figure 1 we show the total disk mass beyond $2.9\,R_\oplus$, as well as the total mass of H_2O dissolved in the magma beyond $2.9\,R_\oplus$, as functions of time. At 20 years, the total mass to be swept up is $\approx 1.0\,M_L$, and the bulk water content is ≈ 10 ppm, thus potentially explaining the depletion of water in the Moon.
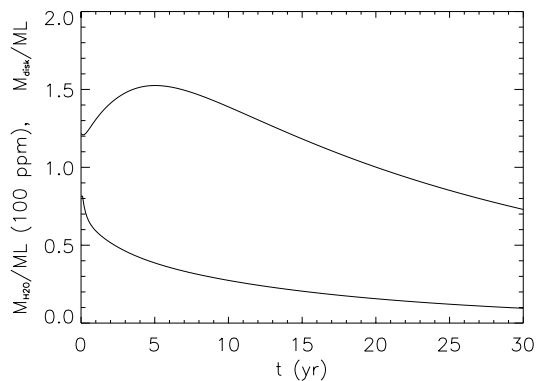
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