ISOTOPIC MIXING DUE TO INTERACTION BETWEEN THE PROTOLUNAR DISK AND THE EARTH'S ATMOSPHERE. **Steven J. Desch**[1] and **G. Jeff Taylor**[2]. [1]School of Earth and Space Exploration, Arizona State University, Tempe, AZ 85287. [2]Hawaii Institute for Geophysics and Planetology, Univ. of Hawaii, Honolulu, HI (steve.desch@asu.edu) .

Background: The isotopic compositions of the Earth and Moon are remarkably alike. The Moon is indistinguishable from the Earth in its isotopic ratios of O [1], Ti [2], Cr [3] and W [4], as well as K [5], typically matching to with experimental error $\ll 1\epsilon$. The O isotopes are indistinsuighable at the level of 0.05ϵ [1]. In contrast, the Moon differs significantly from the Earth in its isotopic ratios of Zn, with δ^{66}Zn $\approx 10-16\epsilon$ [6], and in its D/H ratio, with δD between about 200 permil and 1000 permil [7,8], or 2000 to 10,000ϵ. The overall trend, of identical isotopic composition to the Earth, is difficult to understand: in the context of the giant impact origin of the Moon, 80% of the Moon's mass comes from the impactor, Theia [9]. The fractions of Earth and Moon that came from Theia must have differed by tens of percent, yet even a 3% difference in composition would probably lead to measureable differences in O isotopes [10]. This argues strongly for substantial mixing between the protolunar disk and the Earth's mantle, and yet the Moon is considerably depleted in volatiles like H and Zn, relative to the Earth. **How could the protolunar disk be both well mixed with the Earth's mantle isotopically, and depleted in volatiles relative to the Earth?**

We suggest that the protolunar disk evolved in three distinct stages. **First**, the Earth's atmosphere extended to $r > 1.3\,R_\oplus$ and interacted strongly with the disk, with hydrodynamic instabilities mixing material across their interface, as suggested by [10]. **Second**, after ~ 1 yr the Earth's atmosphere gained enough angular momentum from the disk that its could stably match the angular velocity of the disk. Shear instabilities then ceased, cutting off dynamical and chemical communication between the atmosphere and disk. **Third**, after $\sim 10^2$ years, condensation and rainout of the silicate vapor allowed for hydrodynamic escape of species more volatile than silicates. The Moon would have retained heavier isotopes of species such as H that both isotopically fractionated as they partitioned between vapor and magma and were significantly depleted by hydrodynamic escape.

The Earth's Atmosphere: To calculate the structure of the Earth's atmosphere, we have adopted the formulism of [11] for the vertical structure of a silicate vapor atmosphere. The co-existence of silicates in liquid and vapor forms is assumed, so the constitutive relationship $P = P_0 \exp(-T_0/T)$, is applied [12]. Hydrostatic equilibrium is assumed, and the radial derivatives of T, P, density ρ, and gas mass fraction x all are proportional to the effective gravitational acceleration, which in the equatorial plane is $g_{\rm eff} = GM_\oplus/r^2 - \Omega^2 r$, where Ω is the angular velocity of the atmosphere. This contrasts with the *vertical* structure solutions of [11], which assumed $g_{\rm eff} = \Omega^2 z$, but otherwise the calculation is identical. For simplicity we assume the base of the atmosphere is at $\equiv 1\,R_\oplus$, on top of a planet of mass $\equiv 1\,M_\oplus$. At the base of the atmosphere we impose a gas mass fraction $x_{\rm c} = 0.5$ and a temperature $T_{\rm c} = 8000\,$K.

We first calculate the structure assuming the atmosphere corotates with the Earth every 5 hours, at a rate $\Omega_\oplus = 3.5 \times 10^{-4}\,{\rm s}^{-1}$. Numerically integrating the equations we find a column density of rock vapor in the Earth's atmosphere that is equivalent to the expected $0.2\,M_\oplus$ if globally averaged, but which is considerably extended, with significant density at $r > R_{\rm edge} \approx 1.26 R_\oplus$. [We define $R_{\rm edge}$ to be where the atmospheric density provides sufficient drag force on the disk droplets $100\,\mu$m in radius that they inspiral in < 1 yr. $R_{\rm edge}$ varies only slightly with different assumed inspiral times or particle radii.] **The protolunar disk cannot extend inward of about $1.3\,R_\oplus$, because the Earth's atmosphere is too thick.**

Mixing with the Disk: As suggested by [10], KH (Kelvin-Helmholtz) instabilities will occur at the disk-atmosphere interface, provided the Richardson number ${\rm Ri} = (g_{\rm eff}/\rho)(\partial\rho/\partial r)\,(\partial U/\partial r)^{-2} < 1/4$ [17]. We rewrite this as ${\rm Ri} \sim (g/H)(\Delta U/\Delta R)^{-2}$, where $H \sim C^2/g_{\rm eff}$ is the scale height of the atmosphere, C being the sound speed. KH instabilities will result if the velocity mismatch $\Delta U \sim 4\,{\rm km\,s^{-1}}$ between the atmosphere and disk is in a zone with width $\Delta R < (\Delta U)C/(2g_{\rm eff}) \sim 10^3\,$Km. Notably, the gas mixed across the interface is mostly liquid droplets: $x \approx 0.2$ at the disk interface, meaning 80% of the mass of material is in liquid form (this result holds regardless of the value of x at the atmosphere's base, for $x_{\rm c} > 0.01$). **The Earth and disk can exchange refractory elements as well as volatiles**. As discussed by [10], likely levels of turbulent viscosity in the disk can then mix this Earth

material throughout the disk in $< 10^2$ years. Viscosity in the atmosphere can also mix disk material and angular momentum inward.

The atmosphere absorbs disk angular momentum and spins up, first near the interface and then further inward. It eventually matches the disk's angular velocity at $R_{\rm edge}$ (assumed to be 95% of the Keplerian velocity, due to pressure support), and transitions to $\Omega_\oplus$ at $R_{\rm edge} - \Delta R$. We ran a second calculation in which the angular velocity in the atmosphere is matched at these two points with the profile $\Omega(r) = A + B/r^2$, the only steady-state solution in the presence of viscosity [13]. The additional centrifugal support pushes $R_{\rm edge}$ out to $1.33 R_\oplus$. These profiles are shown in Figures 1-4, where it is seen that $\Omega = \Omega_\oplus$ out to about $1.13 R_\oplus$, matching the disk velocity at $R_{\rm edge} = 1.33\, R_\oplus$. With this velocity difference spread out over $0.2\, R_\oplus$, Ri $> 1/4$ and the mixing ceases. The mixing of angular momentum into the atmosphere probably takes a time $t_{\rm mix} \sim (\Delta R)^2/\nu$, where we estimate $\nu \sim \alpha (0.1r)^2 \Omega$, so that $t_{\rm mix} \sim 1/(\alpha\Omega)$. $t_{\rm mix} \sim 1\,$yr if $\alpha \sim 10^{-4}$.

After the Mixing: Isotopic exchange of all elements is presumably rapid early on, but then ceases altogether once the angular velocity profile stabilizes. Subsequent volatile loss during the $\sim$ 250 years it takes for the Moon to form [15-16] lets the disk evolve relative to the Earth. As discussed by [14], significant loss in $\sim 10^2$ yr is possible only by it hydrodynamic escape, in which the atmosphere leaves in a thermal wind from the disk. The criterion for hydrodynamic escape is $\lambda \approx (GM_\oplus \bar{m}/2kT) < 2$. For likely conditions ($T \approx 2000\,$K, $r = 5\, R_\oplus$), escape requires the mean molecular weight of the gas to be $\bar{m} < 6\, m_{\rm p}$. This is not possible while the gas contains significant silicate vapor, but may be possible if the gas is predominantly water vapor that dissociates (in which case $\bar{m} \approx 6 m_{\rm p}$). Because condensation and rainout of the silicate vapor is necessary before hydrodynamic escape of the atmosphere can take place, we excpect little loss or isotopic fractionation of W, Ti, Cr, or even O, since these will sequestered into the silicate magma. Only species in the vapor phase can be depleted along with H_2O, meaning moderate volatiles like Zn, Cd, Hg and S *may* be depleted. There is no significant isotopic fractionation during hydrodynamic escape itself, but H fractionates as it partitions between the liquid magma and the vapor phase, so the light isotope in the vapor phase is preferentially lost [15]. Depending on the fraction of H that is lost, significant enhancements in δD are possible, as shown in Figure 5. Zn may also be volatile enough to partition sufficiently to cause enrichment of heavy Zn isotopes.

References: [1] Wiechert, U., et al. 2001, *Science* 294, 345. [2] Zhang, J., et al. 2012, *Nat.Geo.* 5, 251. [3] Lugmair, G. & Shukolyukov, A. 1998, *GCA* 62, 2863. [4] Touboul, M. et al. 2007, *Nature* 450, 1206. [5] Humayun, M. & Clayton, R. 1995, *GCA* 59, 2131. [6] Paniello, R. et al. 2012, *Nature* 490, 376. [7] Robinson, K. & Taylor, G. J. 2012, *LPSC* xxxx. [8] Greenwood et al. 2011, *Nat.Geo.* 4, 79. [9] Canup, R. 2004, *Icarus* 168, 433. [10] Pahlevan, K. & Stevenson, D. J. 2007, *EPSL* 262, 438. [11] Ward, W. 2012, *Ap.J.* 744, 140. [12] Thompson, C. & Stevenson, D. J. 1988, *Ap.J.* 333, 452. [13] Chandrasekhar, S. 1961, *Hydrodynamic and Hydromagnetic Stability* [14] Desch, S. J. & Taylor, G. J. 2011, *LPSC* 42, 2005. [15] Hoog, J. C. M., Taylor, B. E. & Van Bergen, M. 2009, *Chem.Geo.* 266, 256.

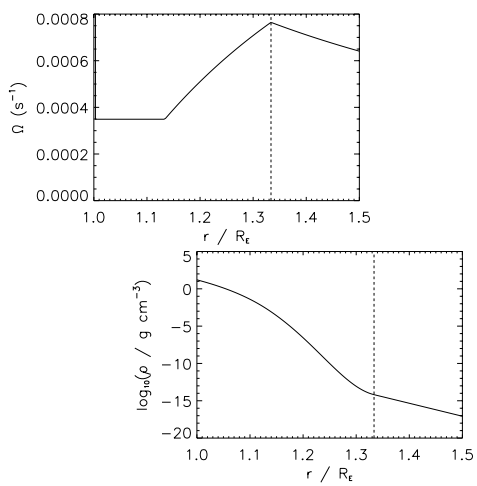

Figure 1: $\Omega(r)$ and $\rho(r)$ in Earth's atmosphere after absorbing angular momentum.

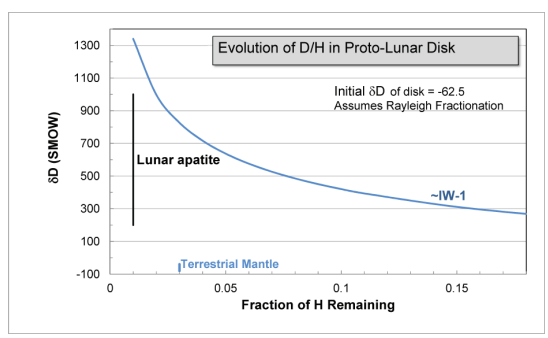

Figure 2: Change in δD as a function of H remaining in the protolunar disk, compared to the range observed in lunar apatite samples, and the Earth initial value.