THE TRANSITION FROM PRIMARY TO SECONDARY ATMOSPHERES ON ROCKY EXOPLANETS.

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Introduction: How massive are rocky-exoplanet atmospheres? For how long do they persist? These questions are compelling in part because an atmosphere is necessary for surface life. Magma oceans on rocky exoplanets are significant reservoirs of volatiles, and could potentially assist a planet in maintaining its secondary atmosphere [1,2].

We are modeling the atmospheric evolution of R ≲ 2 R_{Earth} exoplanets by combining a magma ocean source model with hydrodynamic escape. This work will go beyond [2] as we consider generalized volatile outgassing, various starting planetary models (varying distance from the star, and the mass of initial "primary" atmosphere accreted from the nebula), atmospheric conditions, and magma ocean conditions, as well as incorporating solid rock outgassing after magma ocean solidification.

Through this, we aim to predict the mass and longevity of secondary atmospheres for various sized rocky exoplanets around different stellar type stars and a range of orbital periods. We also aim to identify which planet sizes, orbital separations, and stellar host star types are most conducive to maintaining a planet's secondary atmosphere. Such a model could be tested with the James Webb Space Telescope.

Multiple studies have investigated the importance of maintaining an early water based atmosphere to the extension of cooling time of the magma ocean, as well as to the eventual formation of a liquid water ocean on the planetary surface [1,2,3,4,5].

These secondary atmospheres are susceptible to depletion by hydrodynamic escape [6,7]. In order to accurately determine the magnitude of the secondary atmosphere remaining after magma ocean degassing, one must track atmospheric loss. Various studies have created and analyzed models which study the effect of hydrodynamic escape on long-term atmospheric evolution, both for exoplanets [8,9,10] and for solar system protoplanets [11]. These studies do not couple this atmospheric loss with magma ocean or solid rock outgassing, both of which would extend the longevity of a planet's secondary atmosphere.

Methods: We combine basic models of magma ocean degassing [1,5] and hydrodynamic escape of atmospheres [7,8,9] to study the transition from primary to secondary atmospheres on rocky exoplanets. The model flow through of volatile outgassing and hydrodynamic escape is illustrated below in Figure 1.

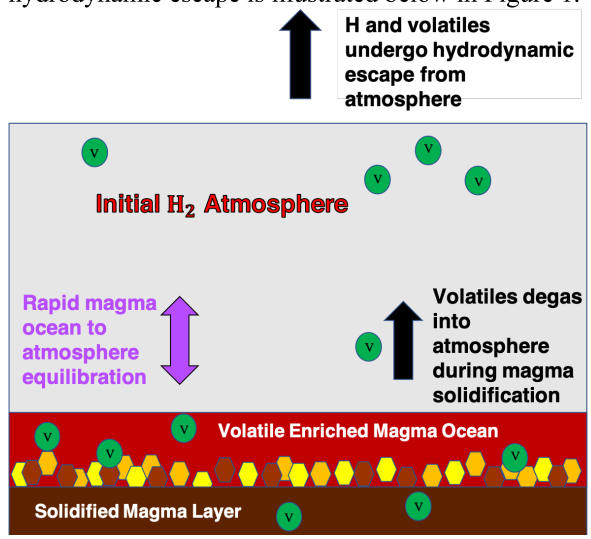

Figure 1: Key processes of our magma ocean and hydrodynamic escape model are shown above. The green circles marked with a V indicate volatiles that are outgassed from the solidifying magma and accumulate in the atmosphere. These volatiles are lost from the exoplanet's atmosphere through hydrodynamic escape aided by outflow of hydrogen, which is derived from the nebula. The purple arrow denotes our model assumption of rapid magma ocean to atmosphere equilibration during degassing. Results from this model flow-through are currently being gathered.

For the magma ocean outgassing component, we calculate the rate of solidification and emissivity of the atmosphere, following Ref. [5]. Magma oceans overlain by massive atmospheres cool slowly [12].

In our model, we set an initial depth of the magma ocean and an initial volatile inventory that is contained within the magma. Each time step, the rate of solidification is calculated (taking into account the emissivity of the atmosphere as well as the luminosity of the host star), and the mass of the remaining liquid magma is determined. Using the mass of the remaining liquid magma, we interpolate over two corresponding pressure and total volatile mass arrays to determine which atmospheric pressure fits the total volatile mass in our model. We then convert that pressure into atmospheric mass of the volatile. From that, we use $wt_{volatile}[\text{in wt\%}] = 100(m_{volatile\ magma})/(m_{volatile\ magma} + m_{magma})$ where $m_{volatile\ magma} = (m_{total\ volatile} - m_{atmospheric\ volatile})$.

For the hydrodynamic escape component, we calculate the escape fluxes of hydrogen and various heavier molecular species (for the results shown below, the heavier molecular species in question is O) following [9]. In this hydrodynamic escape component, we consider the atmospheric column to be well mixed, full molecular dissociation to occur at the homopause, and for hydrodynamic escape to occur from the homopause.

We calculate the F_{XUV} value at each timestep by using an equation that was interpolated from the graphical results from research surveying the UV and X-ray activity of M dwarfs within 10 pc of our Sun [13]. The resulting interpolated equation is as follows:

$$\log_{10}(L_{XUV}[ergs]) = -1.1[\log_{10}(ergs\ Gyr^{-1})](\log_{10}(t\ [Gyr])) + 27.6\ [\log_{10}(ergs)]).$$

A net build up rate of O_2 for models containing H_2O is also considered at each time step, and is represented by the equation $m_{O2\ buildup}[g\ cm^{-2}s^{-1}] = ((F_H[cm^{-2}s^{-1}]-2F_O[cm^{-2}s^{-1}])m_{O2}[g])/4$. After these values are determined, the resulting net mass loss values are subtracted from the atmospheric pressure generated from the magma ocean degassing to calculate the atmospheric pressure values for the next time-step.

Anticipated Results: For our rocky exoplanet models, we will consider a range of planet masses, initial magma ocean depths, initial magma volatile inventories, and initial atmospheric pressures. We will also consider a range of host star stellar types and orbital separations, which strongly effect the strength of hydrodynamic escape. The magma ocean solidification portion of the model will run until solidification is complete, and atmospheric evolution will continue until all atmospheric mass is depleted (or 10 Gyr – whichever is sooner).

Through our two component model, we will use the results of variety of planet starting conditions to determine which range of starting parameters produce the longest maintained secondary atmosphere. Once this correct range of parameters is established, further modeling will be done to find which parameters have the largest effect on atmospheric retention.

Future Work: To follow up on our initial results, we will expand our model to include volatile storage in solid rock and long term outgassing (Figure 2), using a parameterized approach to mantle convection (e.g. [14]). In the results above, we assume all volatiles are outgassed from the solid rock upon full solidification. This approximation captures the behavior of most of the volatiles, but outgassing of even a small fraction of the volatiles late in a planet's history could build up an atmosphere that would be detectable by JWST. In order to correctly model the planets' atmospheric evolution, we must consider the extension in longevity of the planet's atmosphere that can occur due to this residual outgassing.

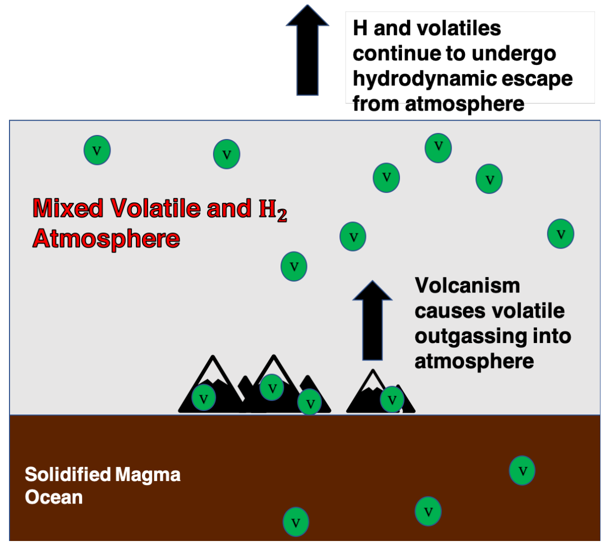

Figure 2: The intended model flow-through for volcanic degassing is shown above. When the magma ocean is solidified, volcanic outgassing can continue (e.g., [14]). We intend to use a parameterized approach to mantle convection [14] to model this continued outgassing. During this process, hydrodynamic escape may continue, and we model the atmospheric evolution until the atmosphere is depleted or 10 Gyr is reached.

Additionally, we seek to include the significant effect of the outgassed atmosphere's greenhouse effect on the surface of the planet during magma ocean cooling. This feedback from the gases released from the solidifying magma ocean in turn delays solidification, which extends the duration of volatile outgassing [3,15,16].

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References: [1] Elkins-Tanton, L.T. Astrophys Space Sci. 332, 359 (2011). [2] Schaefer, L. 2016. ApJ. 89, 63. [3] Abe, Y. and Matsui, T. J. Geophys. Res. 91, E291 (1986). [4] Abe, Y., Matsui, T. J. Geophys. Res. 90, C545 (1985). [5] Elkins-Tanton, L.T. EPSL. 271, 181 (2008). [6] Zahnle, K.J., Kasting, J.F., 1986. Icarus 68, 462-480. [7] Hunten, D.M., 1973. J. Atmos. Sci. 30, 1481-1494. [8] Luger, R., Barnes, R. 2015. Astrobiology 15, 119-143. [9] Tian, F. 2015. EPSL. 432, 126-132. [10] Tian, F. 2009. ApJ. 703, 905-909. [11] Odert, P., et al. 2018. Icarus 307, 327-346. [12] Salvador, A., et al. 2017. JGR-Planets 122, 1458-1486. [13] Stelzer, B., et al., 2013. Mon. Not. R. Astron. Soc. 431, 2063-2079. [14] Kite, E. S., et al. 2009. ApJ. 700, 1732-1749. [15] Zahnle, K.J. 1988. Icarus 74, 62-97. [16] Hamano. K., et al. 2013. Nature 497 607-610.