Exploring the Evolution and Habitability of Planets: Coupling of the Mantle-Atmosphere System. Matthew B. Weller[1], Alexander J. Evans[1], Daniel E. Ibarra[1,2], Alexandria V. Johnson[3], and Tyler Kukla[4], [1]Brown University, Providence RI. (corresponding email: mbweller@brown.edu), [2]University of California, Berkeley, Berkeley CA., [3]Purdue University, West Layfette, IN., [4]Stanford University, Stanford, CA.

Introduction: With numerous terrestrial-like exoplanets discovered from the Kepler mission [e.g., 1], ranging from ~ 0.5 to ~ 2 Earth radius, it is natural to consider how many of these bodies may have an atmosphere that allows for stable liquid water at the surface. Although many of these planets may fall within the classically defined habitable zone [e.g., 2] of their host stars, many do not and distance from the host star alone is likely an insufficient metric to assess the stability of liquid water (and habitability potential). The habitability of a planetary body is significantly influenced by both atmospheric and interior processes, such as mantle convection, the tectonic mode, geochemical evolution, core dynamos, melting and outgassing, atmospheric development, chemistry, and the development of a water cycle [e.g., 3-6].

As planetary atmospheric development is inherently linked with interior evolution, it is necessary to understand the thermal and chemical evolution of a rocky planetary body to understand how its atmosphere evolves. *Here, we couple planetary interior evolution models with equilibrium atmospheric models to understand the linked behavior and evolution of a planet, its atmosphere, and surface temperatures.* We explore differing tectonic states including Earth-like mobile lids, Mars-like stagnant lids, and heat-pipe planets. This allows us to obtain a more comprehensive understanding of processes that are likely to foster the presence of stable liquid water at the surface of a planet.

Numerical Methods: To understand the long-term thermal evolution of planetary bodies, we develop and apply a parameterized model of mantle convection, using as a first step a bulk Earth-like composition. Our model solves a 1D energy conservation equation [7]:

$$\left(R_p^{\,3}-R_c^{\,3}\right)\left[H(t)-\rho_m\cdot c_m\frac{dT_m}{dt}\right]=3\cdot\left(q_s\cdot R_p^{\,2}-q_c\cdot R_c^{\,2}\right)\quad(1)$$

where R_p and R_c are the planetary and core radii, respectively, ρ_m is the bulk mantle density, c_m is the specific heat, T_m is the bulk average temperature, q_s and q_c are the surface and core heat flux, respectively. The heat source $H(t)$ is the radiogenic heat production, which along with the core heat flux, are assumed to be the primary energy sources for the mantle.

For the mantle, the Nusselt number (Nu) is given by a Nusselt-Rayleigh (Ra) relationship [e.g. 8,9], and the temperature-dependent viscosity ($\eta(T)$), defined as:

$$Nu\sim(Ra)^{\beta}\quad(2)$$

$$Ra=g\rho\alpha\Delta Td^{3}/(\kappa\eta)\quad(3)$$

$$\eta(T)=\eta_0\exp\left(A/\left(RT_m\right)\right)\quad(4)$$

where α is the mantle thermal expansivity, g is surface gravity, κ is thermal diffusivity, d is mantle thickness, η_0 is the reference viscosity, A is the activation energy, and β is 0.33 (mobile lids) and 0.2 (stagnant lids) [8,9].

We use standard solidus and liquidus relationships [10], modified to allow for the effects of water on melt and viscosity [11,12] as well as melt reprocessing [13]. Using melt production to constrain greenhouse gas concentrations (e.g., CO_2) we solve, at steady state, a 1D zonal energy balance model (modifying [14,15]) for a range of distances from a Sun-like star. Assuming typical surface albedos (e.g., 0.3), we solve for the zonal-mean net heating of the atmosphere (Q_{net}), versus surface difference in the net downward energy flux, to solve for h, the near-surface moist static energy:

$$Q_{net}(x)=-\frac{p_s}{gR_p^{\,2}}\frac{d}{dx}\left[\left(1-x^2\right)\frac{dh}{dx}\right]\quad(5)$$

where p_s is surface air pressure, D is a diffusion coefficient, and x is the sine of latitude. The resulting zonal-mean moist static energy profile is then translated to a temperature profile. These profiles are averaged to calculate global mean temperatures and assess the possibility of stable liquid water.

Results: For simplicity, we only include results from cases that have planetary radii of 0.5 and 1.0 of Earth's radius (see Figure 1). Models yield surface temperatures that range from 30 K to 800 K. A nominal Earth-like mobile lid is illustrated in Figure 1a, which serves as the comparison point for all models. Despite early variations in melting rates, these systems quickly converge to a 'standard' background rate for all but the highest 800 K surface temperature case, indicating self regulation of the mantle.

Compared to the case of 1.0 Earth Radius, a 50% reduction in planetary radii (Figure 1b) shows an increase in melt production for the mobile lid cases, reaching maximum production rates (due to melting occurring at the core mantle boundary) early in the planets evolution. This increase in melt production for the smaller planetary radius case is likely due to the effects of g on the slopes of the adiabat, solidus, and liquidus. The adiabat scales linearly with g, whereas the solidus and liquidus scale quadratically. Decreasing g results in an adiabat closer to the solidus at lower temperatures, whereas increasing g results in and adiabat that is sufficiently removed from the solidus temperature to limit melt production.

Compared to planets with a mobile lid, planets with thick high-viscosity (stagnant) lids have much higher overall internal temperatures, as much as ~45% greater. For planets with a stagnant lid, melting rates remain only slightly elevated above those of the mobile lid, with a small percentage of that melt ($\mathcal{O}$(1-10%)) reaching the surface. The thick high viscosity lid, which is the key difference from a mobile lid state, reduces internal convective velocities, limits extraction of melt to the surface, and insulates the interior from the effects of surface temperatures. Reducing the planetary radius, however, allows the surface temperature to have a more pronounced effect on the considered time scales. For example, surface temperatures below 100 K result in a purely conductive planet at ~4 Gyr if the planetary radius is 40% that of the Earth's or less. This suggests that atmospheric insolation may allow for planets to remain thermally active for longer time scales.

We illustrate the effects of our modified 1D zonal energy balance model [14,15] for the 1.0 Earth radius stagnant lid results (Figure 1a), for a range of stellar distances (Figure 2). The Sun-Earth distance (1.0 AU) stagnant lid case starts with surface temperatures that disallow liquid water (so-called snowball state). However, as CO_2 builds and luminosity increases in our model, this case emerges from a snowball state at ~ 1.5 Ga, with surface temperatures allowing for liquid water to the present. At 0.86 AU, this transition occurs at ~ 3.4 Ga, whereas the 0.72 (Venus distance) may allow for lower temperatures in the first ~100 Myr, but warms soon thereafter. This suggests that both an Earth-like stagnant lid in an Earth like orbit, and a Venus-like orbit could have the potential for stable liquid water at their surfaces over geologic time.

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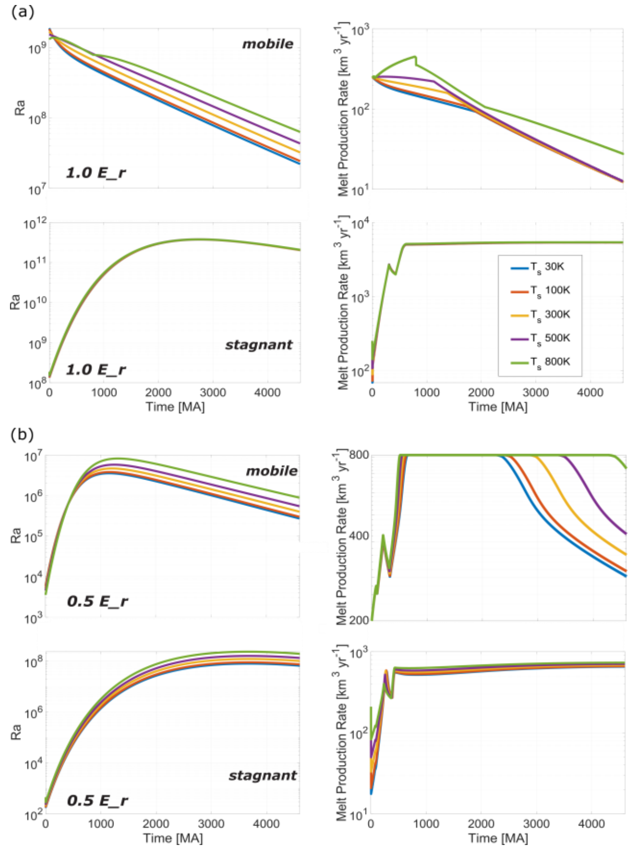

Figure 1: Mantle thermal evolution results for a 1.0 (a) and 0.5 (b) Earth Radius (E_r) models. Gravity, mass, and radius relationships follow from [16] and are Earth-like, g = 9.8 (1 E_r) and 3.1 m s^{-2} (0.5 E_r). Initial mantle temperatures are 2550 K and 1900 K, respectively.

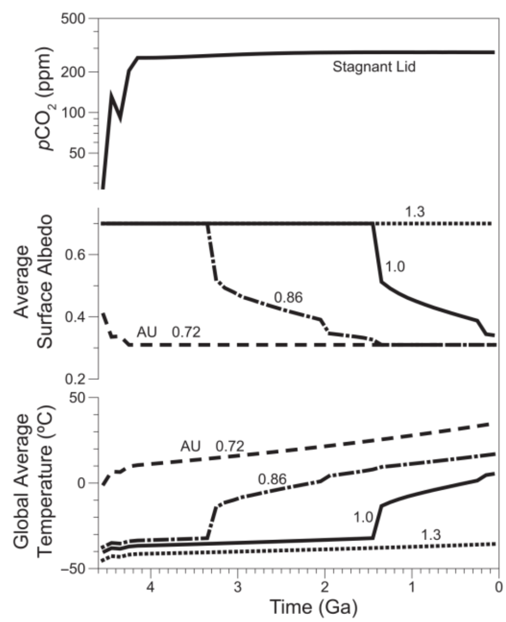

Figure 2: Results from the 1D zonal energy balance model with an Earth-like stagnant lid. pCO2 results from mantle melt production. (Fig. 1a). This climate response is from CO_2 forcing [15] from Venus-like orbit (0.72 AU) to inside Mars' orbit (1.3 AU). Here, global average temperatures above 0°C indicate the ability for liquid water to be present.