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Evolution of lunar polar ice stability

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ABSTRACT

The polar regions of the Moon and Mercury both have permanently shadowed environments, potentially capable of harboring ice (cold traps). While cold traps are likely to have been stable for nearly 4 Gyr on Mercury, this has not been the case for the Moon. Roughly 3 ± 1 Gya, when the Moon is believed to have resided at approximately half of its current semimajor axis, lunar obliquities have been calculated to have reached as high as 77°. At this time, lunar polar temperatures were much warmer and cold traps did not exist. Since that era, lunar obliquity has secularly decreased, creating environments over approximately the last 1–2 Gyr where ice could be stable (assuming near current recession rates). We argue that the paucity of ice in the present lunar cold traps is evidence that no cometary impact has occurred in the past billion years that is similar to the one(s) which are thought to have delivered volatiles to Mercury's poles. However, the present ice distribution may be compatible with a cometary impact if it occurred not in today's lunar thermal environment, but in a past one. If ice were delivered during a past epoch, the distribution of ground ice would be dictated not by present day temperatures, but rather by these ancient, warmer, temperatures. In this paper, we attempt to recreate the thermal environments for past lunar orbital configurations to characterize the history of lunar environments capable of harboring ice. We will develop models of ice stability and mobility to examine likely fossil remains of past ice delivery (e.g. a comet impact) that could be observed on the present Moon. We attempt to quantify when in the Moon's outward evolution areas first became stable for ice deposition and when ice mobility would have ceased.

1. Introduction

The polar regions of the Moon and Mercury both have permanently shadowed environments, potentially capable of harboring ice (cold traps). While the distribution and temperatures of Mercury's cold traps have likely been stable for nearly 4 Gyr (Siegler et al., 2013), this has not been the case for the Moon. Roughly 3 ± 1 Gya, when the Moon is believed to have resided at approximately half of its current semimajor axis, lunar obliquities have been calculated to have reached as high as 77° (Goldreich, 1966; Ward, 1975; Arnold, 1979; Wisdom, 2006; Siegler et al., 2011). This is due to a dissipation-driven spin–orbit coupling known as a Cassini State. Combined with the modeled orbital inclination for this time period, this left the lunar poles with a maximum solar illumination angle (here termed θ_{max}, or *declination*) of approximately 83° (Siegler et al., 2011). At this time, lunar polar temperatures were much warmer and cold traps did not exist. Since that era, lunar obliquity has secularly decreased, creating environments over approximately the last 1–2 Gyr where ice could be stable (assuming near current recession rates).

On Mercury evidence points to nearly pure ice deposits resulting from a large cometary impact within the last several 10s of Mys (Crider and Killen, 2005; Lawrence et al., 2013; Neumann et al., 2013; Paige et al., 2013). A geologically recent comet impact is favored here, as it would explain the thickness and purity of the ice (to be consistent with radar data) and provide a mechanism to bury it to depths of 10s of centimeters (consistent with neutron spectrometer and radar loss data). The generally similar thermal environments on the Moon also would be expected to retain relatively pure water ice for 10–100s of Mys. However, there is no evidence for Mercury-like nearly pure ice deposits at least 10s of cm thick on the Moon, with ice concentrations less than a few percent in the top meter of regolith (Feldman et al., 1998, 2001; Campbell et al., 2006; Coleprete et al., 2010). It is difficult to explain how nearly all ice from a large impact over the past ~1.5 Gyr could be lost. Though impact gardening will bury ice and remove radar scattering blocks, even a 10 cm thick ice layer should be visible by neutron spectrometer measurements for 1 Gyr (Hurley et al., 2012). The Mercury deposits need to be much thicker than the 12.6 cm S-band Arecibo wavelength to return the observed coherent backscatter signal (Harmon et al., 2011). Essentially, one cannot explain the paucity of lunar ice in locations where it would be stable in the

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current thermal environment, unless no comet similar to the one(s) which struck Mercury has struck the Moon in the past billion years or more.

Cometary impacts may be consistent with the present lunar volatile distribution if they occurred not in today's lunar thermal environment, but a past one. If ice were delivered during a past epoch, the distribution of ground ice would be dictated not by present day temperatures, but rather by these ancient temperatures. This ancient ice, buried and mixed into the regolith by impact gardening could still be observable, given a large initial deposit (Hurley et al., 2012). Additionally, if thermal environments are favorable to ice mobility, ice may re-equilibrate to a stable depth, countering burial by gardening. This may be the case on Mercury (and Mars), as all observed ice deposits could be interpreted as consistent with depths predicted by thermal equilibrium but inconsistent with a steady burial (Paige et al., 2013). On the Moon, ice may have remained at a steady equilibrium depth for a substantial time before the current "deep freeze" led to conditions where burial by gardening outpaced thermal mobility (Siegler et al., 2011). If so, the important age for determining the beginning of substantial ice burial and loss to gardening might not be the time of ice delivery, but the secession of ice mobility (when the deposit cooled below $\sim$100 K).

In this paper, we attempt to recreate the thermal environments for past lunar orbital configurations to characterize the history of lunar environments capable of harboring ice. We will develop models of ice stability and mobility to examine likely fossil remains of past ice delivery (e.g. a comet impact) that could be observed on the present Moon. We attempt to quantify when in the Moon's outward evolution areas first became stable for ice deposition and when ice mobility would have ceased. These models are qualitatively compared to current evidence for ice enhancement (Feldman et al., 1998, 2001; Mitrofanov et al., 2010; Gladstone et al., 2010; Lucey et al., 2014) but a model quantitatively comparable to data will require future work incorporating models of ice supply, impact gardening, and assumptions of the timeline of lunar orbital evolution.

2. Current lunar temperatures

Few will deny the statement that the present day lunar poles are cold, but thermal environments vary dramatically over short geographic distances. The current low maximum solar declination, θ_{max}, of 1.54° leads to regions that are permanently topographically shadowed from the Sun down to roughly 60° latitude (McGovern et al., 2013; Hayne et al., 2013). In doubly shadowed craters (those shadowed from the first "bounce" of reflected or reradiated illumination) temperatures have been found to dip as low as 20 K (Paige et al., 2010a,b; Siegler et al., 2012b; Aye et al., 2013). However, yearly maximum temperatures in excess of 330 K can be observed on the rim of near-polar Shackleton crater (89.7°S, 111°E) (Paige et al., 2010a,b). Topography is the dominant control of polar temperatures on the Moon.

As temperatures are so dominated by topography, detailed topographic models are required to accurately predict where water ice might be stable on the lunar surface. Such a topographic model was developed to match and extend temperature measurements from the Diviner Lunar Radiometer (Paige et al., 2010a,b). Detailed work is in progress refining these models to identify variations in near surface thermal properties, surface albedo, and emissivity, which will lead to an improved data-model match. However, despite nearly 5 years of mapping, due to the exact orbit phasing required to map a location at local noon on summer solstice or local midnight midwinter, models are required to interpolate between Diviner data points in order to compute maximum, minimum and average surface temperatures of the lunar polar regions than Diviner itself. Additionally, these models allow for extrapolation of temperatures below the surface, which represent a far larger region for ice stability than the surface alone and robust calculations of temperatures in the Moon's distant past (Paige et al., 2010a,b).

The Diviner south polar thermal model (Paige et al., 2010a,b) uses a triangular mesh with vertices based on Kaguya Laser Altimeter (Araki et al., 2008) and LOLA data (Smith et al., 2010). Each of the 2,880,000 isosceles triangles measures 500 m on the two shortest sides. Surface reflectance properties were assigned to be a highlands average from Clementine albedo measurements or about 0.2 (Isbell et al., 1999). Infrared emissivity was assigned as 0.95. For this simple model, visible and infrared scattering is assumed isotropic. The models published in Paige et al. (2010a,b) assume a layered temperature dependent thermal conductivity model assuming $k = k_c\ [1 + \chi\ (T/350)^3]$ with parameters in Table 1. Heat capacity was assumed temperature dependent, as measured from Apollo samples (Robie et al., 1970). The model has 114 layers (the top four are 5 mm thick, all others 25 mm) and reaches to 2.8 m depth. The bottom boundary assumes a fixed 16 mW m^{-2} heat flux. Model timesteps were 1/52nd of an Earth day.

Fig. 1 illustrates results of the Paige et al. (2010a,b) model of yearly minimum, average, and maximum surface temperature of the lunar South Pole. Temperatures are scaled 35–85 K, 50–200 K, and 100–350 K respectively (for direct comparison with Fig. 3). Our paper will focus primarily on the South Pole as there is greater evidence for subsurface ice deposits within shadowed regions (present and past) than in the North (Feldman et al., 1998, 2001).

3. Ice deposition/migration/destruction concepts

In the simplest concept, ice will be most stable where it is coldest. In the case of a block of ice sitting on the surface, this is true. Sublimation of an exposed volatile will slow with decreasing temperature. 100 K is often used as an estimate for ice stability on geologic time scales as the sublimation rate of exposed water ice will slow to roughly 1 kg m^{-2} Gyr^{-1}, or about 1 mm Gyr^{-1}. This loss rate can be calculated (Schorghofer and Taylor, 2007; Siegler et al., 2011):

$$E = \frac{P_{sv}}{\sqrt{2\pi RT/\mu}} \tag{1}$$

where E is the sublimation rate (kg m^{-2} s^{-1}) (Langmuir, 1913; Watson et al., 1961), R the Boltzmann constant (8.314 J K^{-1} mol^{-1}), T temperature, and μ the molecular weight of water. This formulation represents the maximum possible sublimation rate as it assumes a condensation coefficient of unity (actual values may fall between 0.7 and 1; Schorghofer and Taylor, 2007). P_{sv}, the saturation vapor pressure, can be calculated:

$$P_{sv} = P_t \exp\left[\frac{-Q}{R}\left(\frac{1}{T} - \frac{1}{T_t}\right)\right] \tag{2}$$

where P_t (for H_2O, 611.7 Pa) and T_t (237.16 K) are the triple point pressure and temperature, Q is the sublimation enthalpy (51.058 kJ/mol), and R is the universal gas constant (8.314 J K^{-1} mol^{-1}). These derivations can be used for any volatile with by changing P_t, T_t, Q, and μ. If ice is buried, either by thermal migration or gardening, beneath a regolith layer (z m thick) of particles

Table 1
Thermal properties used in Paige et al. (2010a,b).

Depth range (cm)	k_c (W m^{-1} K^{-1})	X	ρ (kg m^{-3})
0–2	0.000461	1.48	1300
>2	0.0093	0.073	1800