Icarus 255 (2015) 78-87

Contents lists available at ScienceDirect

Icarus

journal homepage: www.elsevier.com/locate/icarus



Evolution of lunar polar ice stability

Matt Siegler^{a,*}, David Paige^b, Jean-Pierre Williams^b, Bruce Bills^a

^a Jet Propulsion Laboratory, California Institute of Technology, Pasadena, CA 91109, United States ^b University of California, Los Angeles, CA 90095, United States

ARTICLE INFO

Article history: Received 31 May 2014 Revised 17 September 2014 Accepted 20 September 2014 Available online 12 October 2014

Keywords: Moon, surface Ices Regoliths Planetary dynamics

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 obliguities 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. © 2014 Elsevier Inc. All rights reserved.

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 \sim 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





^{*} Corresponding author.

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 ~ 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 lavered temperature dependent thermal conductivity model assuming $k = k_c [1 + \gamma (T/350)^3]$ 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:

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

where P_t (for H₂O, 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 a	l. (2010a,b).

Depth range (cm)	$k_{\rm c}~({\rm W}~{ m m}^{-1}~{ m K}^{-1})$	X	$ ho~({ m kg~m^{-3}})$
0–2	0.000461	1.48	1300
>2	0.0093	0.073	1800



Fig. 1. Model results of lunar south polar temperatures for the current lunar orbit. (a) Yearly minimum surface temperatures (stretched 35–85 K), (b) yearly mean surface temperatures (stretched 50–200 K), and (c) yearly maximum surface temperatures (stretched 100–350 K).

diameter δ (75 µm here), the sublimation rate calculated by Eq. (3). Assuming *T* at depth *z*, and calculating *E*(*z*) (the sublimation rate if ice were exposed, Eq. (1)), the effective sublimation rate, *J*(*z*) is estimated by Schorghofer and Taylor (2007) as:

$$J(z) = \frac{\mu \delta E(z)}{2z} \tag{3}$$

which assumes diffusion to be well within the Knudsen regime (in which diffusion is controlled by collisions with pore walls rather than other molecules). J(z) also has units of kg m⁻² s⁻¹, which is roughly equal to mm s⁻¹. This simple model appears to explain the transition from surface to subsurface ice deposits on Mercury, which are credited to be remnants of a large comet impact or series of impacts (Paige et al., 2013).

A slightly more nuanced examination of where ice would survive would include a simple model of ice migration within the regolith. Ice that migrates into regolith pore space might survive longer than ice at the surface, even counteracting burial and mixing by impact gardening. Assuming a plentiful supply (such as delivery of a meters thick ice deposit after a large comet impact or a steady state supply from other sources, e.g. solar wind), ice would migrate from the surface into subsurface pores at a rate depending on the saturation vapor density gradient. Assuming no pore filling (which would inhibit diffusion) and that pore filling ice has no effect on thermal properties (which will alter thermal gradients), ice will build in soil at a rate controlled by the saturation vapor density, ρ_{sy} .

$$\frac{\partial \sigma(z)}{\partial t} = -\frac{\partial J(z)}{\partial z} = \frac{\partial}{\partial z} \left(D(z) \frac{\partial \rho_{s\nu}(z)}{\partial z} \right)$$
(4)

where D(z) is the diffusion coefficient (typically ~10⁻¹⁰ m² s⁻¹ for the temperatures and grain sizes assumed in our lunar models; Schorghofer and Taylor, 2007), *J* is the flux (kg m⁻² s⁻¹) at a given depth and σ is the density (kg m⁻³). The saturation vapor density ρ_{sv} can be calculated (assuming the ideal gas law) from the saturation vapor pressure (Eq. (2)). If we assume *D* is constant as a function of depth, we can quantify the rate of ice deposition as directly proportionate to the second derivative of the vapor density, $\frac{\partial^2 \rho_{sy}(z)}{\partial z^2}$, for the length of time that a surface ice cover survives.

Additional non-thermal migration and loss of ice will occur. Ice exposed at the surface needs to contend with photo-dissociation and sputtering from Lyman-alpha radiation and high energy cosmic rays. These processes would limit the lifetime of a surface deposit and are not accounted for here. Extensive work has been done modeling mixing of non-thermally mobile ice with regolith by impact gardening of the surface (Crider and Killen, 2005; Hurley et al., 2012). This process is likely reasonable for the lack of radar backscattering ice deposits on the Moon (Campbell et al., 2006) and sets a minimum age of any large deposition event, as radar coherent blocks are likely to disappear within 10–100s of Myr (Hurley et al., 2012). The relationship between neutron spectrometer measurements (e.g. Feldman et al., 1998) and gardening is more dependent on the initial vertical and lateral extent of the deposit, but a roughly 10 cm thick deposit should be detectable for ~1 Gyr as burial rates tend to be on the order of 1 mm Gyr⁻¹ (Hurley et al., 2012). Lacking a full model including both ice diffusion and impact processes, gardening limits our ability to directly quantify the present day concentration of ancient ice. Here, we instead aim to simply geographically identify regions of ice stability and enhanced deposition under past lunar climates and discuss their relation to present day observables.

4. Orbit history

To examine the past lunar temperatures, we much first understand the forces that drive its present day orientation. The Moon currently resides at a constant angle with respect to the ecliptic (which we call θ_{max} , or declination, as it represents the maximum yearly solar declination angle) of 1.54° (a combination of its 6.69° obliquity and 5.15° orbital inclination). The current obliquity results from a spin–orbit coupling known as a Cassini State (Colombo, 1966; Peale, 1969). As dissipation within the Moon, which is likely lower now than it has ever been, is responsible for driving the Moon to such a state, it is a safe assumption the Moon has resided in a Cassini State for most of its history. This assumption, combined with a tidal evolution model (after Goldreich, 1966; Touma and Wisdom, 1994) allows for prediction of past orbital inclination and precession rates, provides a robust estimation of the evolution of the orientation of the Moon in space.

When the Moon was nearer to the Earth, tidal torques from the oblate figure of the Earth caused large variations in the Moon's orbital inclination on the timescale of the procession of the lunar orbit (varying from 15 to 80 years). These inclination variations lessened as the lunar semimajor axis grew and the oblate Earth acted more like a gravitational point source. This outward evolution drove the orbit to the nearly constant inclination to the ecliptic (by the time the Moon reached a roughly 35 RE semi major axis) (Goldreich, 1966; Touma and Wisdom, 1994).

This model assumes that early on, the Moon resided in what is known Cassini State 1. Though there are other stable Cassini States at this time, one requires a retrograde lunar orbit, the other one nearly perpendicular to the ecliptic (Peale, 1969). This argument has merit as all other known planetary bodies (including Mercury) locked in a Cassini State lie in State 1, having lacked the semimajor axis evolution of the Moon. Cassini State 1 begins with a roughly zero obliquity, increasing as semimajor axis increases. Midway through the Moon's outward orbital evolution, Cassini State 1 ceased to exist, giving rise to its current state, Cassini State 2. Obliquity has been estimated to have reached roughly 77° during this transition (Peale, 1969; Siegler et al., 2011).

Once these oscillations have damped, the spin axis will settle into a new stable state in the one remaining prograde Cassini State, State 2, with an obliquity of about 49° (and θ_{max} of 54.9°). Once in State 2, the θ_{max} will be simply derived as obliquity minus orbital inclination (e.g. the current $1.54^\circ = 6.69 - 5.15^\circ$). The resulting history of θ_{max} , or amplitude of the yearly oscillation of the subsolar point about the equator (also called *declination* here), in Fig. 2 (adapted from Siegler et al., 2011).

To calculate temperatures from the Paige et al. (2010a,b) model presented in Section 2, one needs simply to describe the position of the Sun in the sky as viewed from the Moon. This is done by identifying the subsolar longitude (ϕ_{ss}) and latitude (θ_{ss}) on the Moon.

The subsolar longitude cycles from 0 to 2π each draconic month (29.53 days currently), which is calculated as the inverse of the sum of the synchronous orbit frequency calculated from Kepler's 3rd law (currently $2\pi/27.3$ days) and for the motion of the Earth Moon system about the Sun (always assumed $2\pi/365.25$ days). Past orbital frequencies were provided from Touma and Wisdom (1994, data from J. Wisdom, per. comm.).

The subsolar latitude is modulated by the amplitude of θ_{max} . Most compactly (ignoring phase) subsolar latitude can be written as:

$$\theta_{ss} = (B + C\cos\omega_{pre}t)\sin\omega_{year}t \tag{5}$$

where ω_{pre} and ω_{year} are $2\pi/(\text{precession period})$ and $2\pi/(\text{draconic year})$ respectively, *B* is the mean value θ_{max} over a precession cycle and *C* is the amplitude of the variation of θ_{max} on the precessional time scale. For all times after roughly 35RE, *C* can be considered constant as inclination is roughly constant after this time.

The Sun is treated as a finite sized disc, which is important as the Sun is often partially set at the poles, composed of 128 model triangles, weighted in solar intensity to account for limb darkening (Negi et al., 1985). Insolation on a given lunar topographic model triangle is zero when the centroid of a solar triangle sets below the local horizon as identified by the topographic model.

The timeline associated with the outward evolution of the Moon is a matter of great debate. Approximations of outward lunar migration can be made (e.g. Bills and Ray, 1999), but the rate that the Moon moved outward since that time at a rate depending on the dissipation of tidal energy within the Earth and its oceans. The present day dissipation (leading to an outward evolution of 3.8 cm yr^{-1}) is anomalously high due to the current ocean geometry (Webb, 1982). This does not mean that current recession is the



Fig. 2. Modeled yearly maximum Sun angle, θ_{max} as a function of lunar semimajor axis. This angle represents the amplitude of subsolar point variations about the equator. The filled section represents rapid precessional time scale oscillation. This angle is currently about 1.54°.

fastest, as the effects of dissipation decrease with semimajor axis, *a* (proportionate to $a^{-11/2}$) (Lambeck, 1977). Following this rule, Table 2 provides upper and lower bounds of time at a given θ_{max} based on the fastest (assuming current recession rate) possible and slowest possible evolution models that would lead to a Moon forming at Earth's Hill radius 4.5 Gyr ago. Both of these models are consistent with robust tidal laminae measurements (as summarized in Bills and Ray, 1999).

5. Results: past temperatures, ice stability, deposition, and mobility

Combining past orbital conditions with our topographic thermal model, we are able to robustly calculate surface and subsurface temperatures for past lunar epochs. As the Cassini State transition clearly marks a period when no lunar ice would have been stable, we are interested only in examining the later half of the Moon's outward evolution (while the Moon has resided in Cassini State 2). Given reasonable lunar thermal properties, the several 100 Myr heat wave during the transition would have made ice unstable to great depth. This model assumes that in the last 2-3 Gyr craters have not changed topography (true for most large south polar craters, Spudis et al., 2008) and the spin pole of the Moon has not migrated due to true polar wander or giant impacts. It is worth noting that polar shadowed ice deposits should have been stable prior to the Cassini State transition and might have left regions of hydrated minerals. If such mineralogic signatures were found, they could be used to identify a once ice-rich paleopole, making a case for or against true polar wander or dramatic latitudinal reorientation (e.g. Wieczorek and Le Feuvre, 2009). Additionally, small polar reorientation could explain the general enhancement of hydrogen in the greater polar region observed in neutron spectrometer data.

The present cold traps are as spatially extensive now as they have ever been (Fig. 1). Fig. 3a-l shows how temperatures in the South Polar region evolved as a function of time. The first row (a-c) represent annual minimum, average, and maximum temperatures at 4° tilt (θ_{max}), the second row (d–f) at 8° θ_{max} , the third row (g-i) at 12° θ_{max} , and the final row (j-l) at 16° θ_{max} . Minimum temperature maps share a common stretch of 35-85 K, average 50-200 K, and maximum 100-350 K (as in Fig. 1). Surface ice will only be stable when maximum temperatures remain below $\sim 100 \text{ K}$ (loss rate < 1 kg m⁻² Gyr⁻¹). Such regions are absent in the 12° and 16° declination models (Fig. 3i and 1). Subsurface ice and ice mobility require more detailed calculations using temperatures as a function of depth (Eqs. (3) and (4)). If thermal properties were constant as a function of temperature or temperatures constant with time, temperatures at depth (below the penetration of the yearly thermal wave, $\sim 1 \text{ m}$) can be roughly approximated by the yearly mean temperature (Fig. 3b, e, h, and k). If buried, ground ice can be stable up to 145 K (Schorghofer, 2008). Therefore, areas with average temperatures below 145 K could potentially harbor ground ice.

In Fig. 4a we see full model calculated results for ice stability in the current lunar thermal environment. The white areas are those where Eq. (1) (assuming *T* equals the yearly maximum surface temperature, Fig. 1c) would result in a water ice loss rate below 1 kg m⁻² Gyr⁻¹, totaling ~12,540 km² in area (reproduced from Paige et al., 2010a,b). This assumes no adsorption onto grain surfaces. Colored areas represent locations and depths where water ice loss rates would fall below 1 kg m⁻² Gyr⁻¹ within the top meter (according to Eq. (3)), totaling 128,380 km² in area (including locations where ice is stable on the surface). These are areas where ice could be stable if the Moon initially had a regolith entirely filled with water ice.



Fig. 3. (a–l) Model results of lunar south polar temperatures for past lunar *declination*, θ_{max} (a, d, g, j) yearly minimum surface temperatures (stretched 35–85 K), (b, e, h, k) yearly mean surface temperatures (stretched 50–200 K), and (c, f, i, l) yearly maximum surface temperatures (stretched 100–350 K).

Table 2

A summary of model results and input orbital parameters. This table quantifies the area minimum of stable (loss rate < 1 kg m⁻² Gyr⁻¹) ice in the top meter of the lunar south polar for past lunar *declination*, θ_{max} . Surface ice is essentially non-existent after 8° declination. Subsurface ice survives in only very few locations (such as Shackleton crater) at 16°.

θ_{\max} (°)	1.54	4.0	8.0	12.0	16.0
Semimajor axis (Earth radii)	60.2	41.4 (0.68 current)	36.7 (0.61 current)	34.9 (0.58 current)	33.9 (0.56 current)
Length of sidereal/synodic month (days)	27.3/29.53	15.6/16.3	13.0/13.5	12.1/12.5	11.6/12.0
Approximate time before present	Present day	0.9-3.2 Gyr	1.0-3.6 Gyr	2.1-4.0 Gyr	2.8-4.2 Gyr
Area available for surface ice	12,540 km ²	2250 km ²	\sim 1.1 km ²	0 km ²	0 km ²
Area available for subsurface ice (in top meter)	128,380 km ²	112,810 km ²	34,836 km ²	3060 km ²	337 km ²



Fig. 4. Model results of minimum depth of stable (loss rate < 1 kg m⁻² Gyr⁻¹) ice for present and past lunar *declination*, θ_{max} . Surface ice is essentially non-existent after 8° declination. Subsurface ice survives in only very few locations (such as Shackleton crater) at 16°.

The fact that we do not find ice in all these regions is proof that the Moon is supply limited. Lunar nitches for water ice stability have either lost water by non-thermal processes (e.g. impact gardening) at a rate faster than water has been resupplied, or there was never a large enough supply to fill these coldtraps to begin with. Recent data from Mercury (Harmon et al., 2011; Lawrence et al., 2013; Neumann et al., 2013) also imply water ice exists in nearly all available niches (Paige et al. (2013) uses an identical calculation to that presented here), implying a recent, or even continuous, source of water there. As noted in Section 1, such a continuous source could be thermal diffusion of subsurface ice resurfacing after burial by impact gardening.

Fig. 4b-e show results for past lunar temperature conditions. At 4° declination, most locations where ice would currently be stable could still harbor ice, but generally at deeper depths (87% of areas still could have ice in the top meter - see Table 1). Many large areas of surface ice stability, such as Amundsen and Cabeus craters, have vanished. Results such as this may help explain data from the LCROSS impact, which saw an apparent increase in ice concentration with depth in Cabeus crater (Coleprete et al., 2010). If ice were supplied to the Moon during this epoch (e.g. by comet impact, etc.), ice would potentially have migrated to depth (or formed a lag like the deposits on Mercury) in Cabeus, leading to higher ice concentrations below the surface. Meanwhile, the three large polar craters Havworth, Shoemaker, and Faustini, would have ice stable at the surface (if supplied there in large quantities). In these regions, delivered water would concentrate at the surface, rather than migrating to depth, leaving it exposed to non-thermal loss from Lyman-alpha light, cosmic rays, and sputtering. A large comet impact during this period could explain the relative enhancement in hydrogen observed in Cabeus over these other large polar craters.

As we go further back in time (Fig. 4c), surface cold traps completely disappear, with only 9 model triangles retaining surface ice at 8° *declination*. Water molecules delivered to Cabeus and Amundsen would have had relatively less area for deposition than Hayworth, Shoemaker, and Faustini. Going back to periods at 12° (Fig. 4d) and 16° (Fig. 4e) *declination*, ice could only have been present in a few small, very deep craters, and as predicted from the maximum temperature models, only at depth. These results are compiled in Table 2.

Most of the small craters that could have harbored ground ice at 16° declination are likely to have formed more recent than this epoch, which can be roughly approximated to be at least 2 Gyr ago. However, Shackleton crater, which has been dated as roughly 3.2–3.8 Gyr old (Zuber et al., 2012), likely predates the Cassini State transition, and is therefore one of the first lunar cold traps to come into existence. Shackleton should potentially hold a record of the oldest ice on the Moon and any volatile delivery since. This long term stability lends credence to the apparent enhancement of hydrogen in near polar neutron spectrometer studies (Miller et al., 2014) and enhanced laser altimeter reflectance (Lucey et al., 2014), making Shackleton an especially exciting target for future study.

These past epochs where ice was only stable if buried lead to a question of how ground ice might form. Survival of delivered comet ice to the Moon during these epochs may depend highly on the purity of the comet itself. Paige et al. (2013 and described Lucey, 2013) proposed that the buried ice on Mercury survived by forming a lag of organic molecules that had been imbedded in the cometary ice as it sublimed. If a comet were especially pure ice, a lag might not form in time to protect ice from surface loss processes. It may be that only ice from the dirtiest comets survive and that the Moon has simply been hit with relatively clean ones. Additionally, it has been suggested that the lag material will form from the interaction of high-energy particles trapped in Mercury's magnetic field with simple non-lag-forming molecules (Paige et al., 2013: Delitsky et al., in press). Most models and observations of a lunar dynamo (e.g. Dwyer et al., 2011; Garrick-Bethell et al., 2009) predict the lunar magnetic field to have ceased well before the Cassini transition, potentially limiting the ability of comet delivered ice to form a protective lag on the Moon.

Another aspect of ground ice survival that needs to be examined is that of the relative strength of thermal *pumping* (Schorghofer and Taylor, 2007). Pumping provides fast migration of ice through the regolith due to thermal oscillations and could provide rapid burial (and therefore protection) of ice from loss processes without a lag or burial by neighboring impacts. Water molecules (if the environment is saturated) will migrate toward areas of lower vapor density, ρ_{SV} , at a rate determined by the saturation vapor density gradient $\frac{\partial \rho_{SV}(z)}{\partial z}$. Thermal *pumping* occurs because this gradient will be large and downward during the warm day (and summer), and small and upward during the cold night (and winter).

If there is no surface supply, molecules will be driven upward (due to the vapor pressure difference with vacuum of space), but at a rate controlled by this same temperature dependent migration. If deep ice is present and it is warm enough for ice to be mobile, it will migrate upward toward the ice table depth (shown in Fig. 4). Therefore, this pumping strength can also be used to estimate the ability of ice buried by impact gardening to re-equilibrate to the ice table depth.

Pumping strength can be estimated by its effect on ice deposition rates as described in Eq. (4). Since temperature oscillations at a given location will repeat (each year, or more accurately each orbit precession cycle), we can estimate the net effect of thermal oscillations by taking a time average of the saturation vapor density, ρ_{sv} . One first calculates the $\rho_{sv}(z,t)$ for all depths, then takes a time average to get $\langle \rho_{sv}(z) \rangle$. Assuming ice above the local ice table depth is rapidly lost, we can then rewrite Eq. (4) to solve for the total mass of ground ice that would build up due to pumping from an ice source at the ice table as:

$$\sigma(t) = \int_{-1m}^{z_{\text{icetable}}} \frac{\partial}{\partial z} \left(D(z) \frac{\partial \langle \rho_{sv}(z) \rangle}{\partial z} \right) t \partial z \tag{6}$$

With units of kg m⁻². The diffusion coefficient, D(z), also should vary with temperature and ice content (Schorghofer, 2010), but can be taken out of the integral if it is assumed constant. Assuming a D(z) is a constant 10^{-10} m² s⁻¹ (which is expected for temperatures ~145 K; Schorghofer and Taylor, (2007)) and time, *t*, of 1 Myr, Fig. 5a (and S1) shows the log (base 10) of the net deposition of ice in the subsurface after 1 Myr for current conditions (or roughly interchangeable, because ice is only allowed to build up until pore space is full (assuming 43% porosity).

In this length of time, mobile area (Fig. 5, areas in blue) tend to build a few % by mass ice, with roughly $2.5-5 \text{ kg m}^{-2}$ of ice over the available column for ice deposition (with higher concentrations near the ice table). This column available for ice varies as geothermal heat will prevent ice from filling below the penetration depth of the yearly thermal wave, which is generally less than 1 m and changes with location. Greater quantities of ice could form in these areas if thermal conductivity was allowed to increase with ice content, which would allow deeper penetration of surface thermal forcing. The direct effect of ice content on thermal conductivity depends highly on the ice formation within the pores (Siegler et al., 2012a,b), which is unknown for lunar conditions.

Additionally, ice buried by impact gardening will migrate relatively quickly (to re-equilibrate at the ice table depth predicted in Fig. 4) in these areas. Though this "fast" rate of ice build-up is still fairly slow, an ice deposit lasting the 10s of Myr (as is believed to be observed on Mercury) could substantially fill available pore space. Ice from above (i.e. during periods of after a comet impact) or from buried ice from below (during periods of net surface loss) would outpace thermal loss by 3–4 orders of magnitude at the ice table (2.5–5 kg m⁻² Myr⁻¹ gain vs 1 kg m⁻² Gyr loss).

Areas not experiencing maximum temperatures above ~100 K will have orders of magnitude slower ice deposition. For instance, at current conditions, the center of Shoemaker crater builds up only 10^{-28} kg m⁻² of ice in 1 Myr from thermal migration of ice. This dramatic change in ice mobility between cold traps and



(e)

Fig. 5. Model results of net deposition (kg m⁻² Myr⁻¹) of near surface ice (assuming a continuous supply of ice at the depths given in Fig. 4) for past lunar *declination*, θ_{max} . Currently, most of the large cold traps have weak "pumping", but many, such as Cabeus crater, had stronger pumping in the relatively recent past. Due to the extreme difference in deposition in and out of the cold traps, information was lost in (a and b). Supplemental Figs. S1 and S2 show a rescaled map of information lost in Fig. 5a and b.

neighboring regions is similar to the "ice pump" described in Schorghofer and Aharonson (2014), and leads to large areas that could pump and store ice within the regolith, not in the coldtraps, but around them. In Fig. 5b–e, we see how this area of fast ice

deposition changes with *declination*. Fig. 5b (and S2) shows that in a recent epoch (when $\theta_{max} \approx 4^{\circ}$), current cold traps like Cabeus and Admundsen craters would have had relatively fast deposition of ice supplied in a cometary impact compared to Hayworth,

Shoemaker, and Faustini. The center of these craters did not act as an efficient *ice pump* until further back in history when *declination* was roughly 8° (Fig. 5c). Shackleton crater currently would build up ice at a rate of $\sim 10^{-16}$ kg m⁻² Myr⁻¹ under current conditions, 10^{-7} at 4°, $10^{-3.5}$ at 8°, $10^{-1.5}$ at 12° (Fig. 5d) and ~ 1 kg m⁻² Myr⁻¹ at 16° (Fig. 5e). This is a smaller jump in deposition rate than a location like the LCROSS impact site, which currently builds ice at a rate of 10^{-35} kg m⁻² Myr⁻¹, but increases dramatically to 2.7 kg m⁻² Myr⁻¹ at 4° and 3.7 kg m⁻² Myr⁻¹ at 8° (Cabeus ceases to be a cold trap at higher *declination*). If a comet had impacted the Moon while at 4° *declination*, Cabeus crater would have been a much better ice pump than Shoemaker (which had a rate deposition of 10^{-21} kg m⁻² Myr⁻¹ at its center).

Fig. 6 summarizes these results (plus models run for every 1° of declination from 3° to 20°). The 3 lines illustrate deposition rates for areas at the LCROSS impact site and the centers of Shackleton and Shoemaker craters. The vertical dotted line marks the present day. The horizontal line approximate the burial rate due to impact gardening, roughly 1 mm Gyr⁻¹ (Crider and Killen, 2005). Once a location cools and drops below the horizontal dotted line, burial by impact gardening should outpace thermal ice mobility. In the present day, all 3 locations are poor "ice pumps", though Shackleton exceeds the deposition rate of the others by >10 orders of magnitude. However, thermal migration is still likely outpaced by burial from impact gardening. Here one might expect an occasional resurfacing of a water molecule buried by gardening, whereas ice mobility in the other craters will be entirely driven by impact processes (especially if our assumption of constant D(z) underestimates mobility). In the distant past, each crater had periods when it acted as a strong ice pump, aiding both deposition of delivered ice and retention of existing ground ice. While LCROSS and Shoemaker cease to be stable for ice (to 1 m depth) at about 11° declination, Shackleton has been shadowed for most of its history, harboring potential ground ice to roughly 20° declination.

Perhaps the most intriguing observation is that the LCROSS impact site was a strong ice pump very recently compared to these other locations. This means that this location was one of the most likely places to pump and store ice in the regolith from impacts over the last \sim 1 Gyr. This enhanced mobility also means that ice buried by impact gardening at this location would have been mobile enough to return to its stable equilibrium depth (about 1 cm below the surface at 4° declination). Therefore, it seems plausible that a substantial amount of ice was very near the surface at



Fig. 6. Model results for the net downward flux of water (kg m⁻² Myr⁻¹) of near surface ice (assuming a continuous supply of ice at the depths given in Fig. 4) for past lunar *declination*, θ_{max} . The vertical dotted line marks the present day *declination*, the horizontal line marks a steady burial rate of 1 mm Gyr⁻¹. Below the horizontal line, ice will migrate faster through impact gardening burial. Note that Shackleton has slow continuous deposition and the LCROSS impact site has relatively recent fast deposition.

this location when the Moon was at \sim 3–4° declination. Since that time, this area became one of the coldest environments of permanent shadow on the Moon (Paige et al., 2010a,b), leading ice to be immobile and subject to burial and mixing rates controlled by impact gardening. It may be this combination of ice mobility during a time of ice delivery (e.g. comet impact), followed by a period of "deep freeze" has made areas such as this especially good at retaining ancient lunar ice. This same ice pump to gardening dominated cold trap transition has occurred in many permanently shadowed craters, but occurred relatively recently (and in a large areal extent) near the LCROSS impact site. Therefore, there has been less time for impact gardening to "erase" ground ice in this region. This thermally-driven history could account for both the observed neutron enhancements at Cabeus (Feldman et al., 2001; Mitrofanov et al., 2010) and near surface ice detected by LCROSS.

6. Conclusions

Temperature conditions in the lunar polar regions have changed dramatically over the past 2-3 Gyr. This was not the case for Mercury (Siegler et al., 2013), and presents a plausible difference in the quantity of water ice present in the polar regions of these two otherwise similar bodies. As the lunar orbit evolved outward and obliquity decreased, polar cold traps began to form (Ward, 1975; Arnold, 1979). As each of these coldtraps formed, it first went through a period during which ice would be stable only in the subsurface and thermal pumping of ice into subsurface pore space would have been efficient (Siegler et al., 2011). Many of these coldtraps have since evolved to have yearly maximum temperatures below 100 K, causing thermal ice mobility to essentially stop. Though these areas can collect surface ice, they would not provide a mechanism to protect the ice from non-thermal erosion beyond slow burial from neighboring impacts and therefore risk not preserving ice long term.

Past temperature conditions may provide a way to understand current lunar ice deposits. Current putative ice concentrations hint that if the Moon has been hit by a large comet, as is hypothesized for Mercury, it was likely several 100 Myr ago, but not earlier, when the Moon was at roughly 12–16° declination (2–3 Gyr ago). Ice delivered to the Moon from an ancient comet impact likely saw different temperatures that on the present Moon. These conditions should have lead to preferential deposition in geographic regions favorable to fast burial and protection of ice at the time. In the epochs the followed, ground ice areas where it remained thermally mobile (blue in Fig. 5) would have been buried by nearby impacts (Hurley et al., 2012), but could potentially re-equilibrate to stable depths (as predicted in Fig. 4). Once maximum temperatures dropped below ~100 K, ice would become thermally immobile and subject to continuous burial by impact gardening. These cold areas are extremely inefficient at burying ice below a protective layer, which may explain the apparent lack of evidence for near-surface ice in many of these areas. Shackleton crater is especially interesting in that it is also likely the longest lived cold trap on the Moon, however, it has been an "ice pump" for most of its history, potentially explaining recently observed hydrogen enhancements (Miller et al., 2014). Potentially even more intriguing, hydrogenrich Cabeus crater (Feldman et al., 2001; Mitrofanov et al., 2010) has had much more recent ice mobility than many hydrogen-poor areas (such as Hayworth, Shoemaker and Faustini), potentially explaining those enhancements and the detection of near surface ice in the LCROSS impact. This preliminary study hints that past thermal environments and the effects they have on ice deposition may play an important factor in today's volatile distribution. This could be used to predict expected locations of ice concentration or, given age dated volatile samples (from isotopes or impact gardening burial depth), constrain the rate of evolution of the lunar orbit.

Acknowledgments

Thank you to the Diviner Lunar Radiometer and the Jet Propulsion Laboratory Early Career Fellowship for partial funding of this work. Thanks to Dr. Norbert Schorghofer for help digesting his collection of papers on this topic over the course of several years. The research was carried out at the Jet Propulsion Laboratory, California Institute of Technology, under a contract with the National Aeronautics and Space Administration. © 2014. All rights reserved.

Appendix A. Supplementary material

Supplementary data associated with this article can be found, in the online version, at http://dx.doi.org/10.1016/j.icarus.2014.09. 037.

References

- Araki, H., Tazawa, S., Noda, H., Tsubokawa, T., Kawano, N., Sasaki, S., 2008. Observation of the lunar topography by the laser altimeter LALT on board Japanese lunar explorer SELENE. Adv. Space Res. 42 (2), 317–322.
- Arnold, J.R., 1979. Ice in the lunar polar regions. J. Geophys. Res.: Solid Earth (1978– 2012) 84 (B10), 5659–5668.
- Aye, K.M., Paige, D.A., Foote, M.C., Greenhagen, B.T., Siegler, M.A., 2013. The Coldest Place on the Moon. LPI Contributions, 1719, p. 3016.
- Bills, B.G., Ray, R.D., 1999. Lunar orbital evolution: A synthesis of recent results. Geophys. Res. Lett. 26 (19), 3045–3048.
- Campbell, D.B., Campbell, B.A., Carter, L.M., Margot, J.L., Stacy, N.J., 2006. No evidence for thick deposits of ice at the lunar South Pole. Nature 443 (7113), 835–837.
- Coleprete, A. et al., 2010. The detection of water within the LCROSS ejecta plume. Science 330 (6003), 463–468.
- Colombo, G., 1966. Cassini's Second and Third Laws. SAO Special Report, 203.
- Crider, D., Killen, R.M., 2005. Burial rate of Mercury's polar volatile deposits. Geophys. Res. Lett. 32 (12).
- Delitsky, M.L., Paige, D.A., Siegler, M.A., Harju, E.R., Schriver, D., Johnson, R.E., Travnicek, P., 2014. Ices on Mercury: Chemistry of volatiles in permanently cold areas of Mercury's north polar region. Icarus (in press).
- Dwyer, C.A., Stevenson, D.J., Nimmo, F., 2011. A long-lived lunar dynamo driven by continuous mechanical stirring. Nature 479 (7372), 212–214.
- Feldman, W.C., Maurice, S., Binder, A.B., Barraclough, B.L., Elphic, R.C., Lawrence, D.J., 1998. Fluxes of fast and epithermal neutrons from Lunar Prospector: Evidence for water ice at the lunar poles. Science 281 (5382), 1496–1500.
- Feldman, W.C. et al., 2001. Evidence for water ice near the lunar poles. J. Geophys. Res.: Planets (1991–2012) 106 (E10), 23231–23251.
- Garrick-Bethell, I., Weiss, B.P., Shuster, D.L., Buz, J., 2009. Early lunar magnetism. Science 323 (5912), 356–359.
- Gladstone, G.R. et al., 2010. LRO-LAMP observations of the LCROSS impact plume. Science 330 (6003), 472–476.
- Goldreich, P., 1966. History of the lunar orbit. Rev. Geophys. 4 (4), 411-439.
- Harmon, J.K., Slade, M.A., Rice, M.S., 2011. Radar imagery of Mercury's putative polar ice: 1999–2005 Arecibo results. Icarus 211 (1), 37–50.
- Hayne, P.O. et al., 2013. American Astronomical Society. DPS Meeting #45, #107.03. Hurley, D.M., Lawrence, D.J., Bussey, D.B.J., Vondrak, R.R., Elphic, R.C., Gladstone,
- G.R., 2012. Two-dimensional distribution of volatiles in the lunar regolith from space weathering simulations. Geophys. Res. Lett. 39 (9).
- Isbell, C.E. et al., 1999, March. Clementine: A multi-spectral digital image model archive of the Moon. Lunar Planet. Sci. 30, 1812.

- Lambeck, K., 1977. Tidal dissipation in the oceans: Astronomical, geophysical, and oceanographic consequences. Philos. Trans. R. Soc. London A287, 545–594. Langmuir, I., 1913. The vapor pressure of metallic tungsten. Phys. Rev. 2 (5), 329.
- Lawrence, D.J. et al., 2013. Evidence for water ice near Mercury's north pole from MESSENGER neutron spectrometer measurements. Science 339 (6117), 292–
- 296. Lucey, P.G., 2013. A wet and volatile Mercury. Science 339 (6117), 282-283.
- Lucey, P.G. et al., 2014, March. Evidence for water ice and temperature dependent space weathering at the lunar poles from Lola and Diviner. Lunar Planet. Sci. 45, 2325.
- McGovern, J.A., Bussey, D.B., Greenhagen, B.T., Paige, D.A., Cahill, J.T., Spudis, P.D., 2013. Mapping and characterization of non-polar permanent shadows on the lunar surface. Icarus 223 (1), 566–581.
- Miller, R.S., Lawrence, D.J., Hurley, D.M., 2014. Identification of surface hydrogen enhancements within the Moon's Shackleton crater. Icarus 233, 229–232.
- Mitrofanov, I.G. et al., 2010. Hydrogen mapping of the lunar South Pole using the LRO neutron detector experiment LEND. Science 330 (6003), 483–486.
- Negi, B.S., Bhowmik, N.C., Mathur, S.S., Kandpal, T.C., 1985. Solar limb darkening and ray trace evaluation of solar concentrators. Appl. Opt. 24 (2), 296–298.
- Neumann, G.A. et al., 2013. Bright and dark polar deposits on Mercury: Evidence for surface volatiles. Science 339 (6117), 296–300.
- Paige, D.A. et al., 2010a. Diviner observations of cold traps in the lunar south polar region: Spatial distribution and temperature. Science 330 (6003), 479–482.
- Paige, D.A., Siegler, M.A., Vasavada, A.R., 2010b. Constraints on lunar heat flow rates from Diviner Lunar Radiometer polar observations. American Geophysical Union (Fall). Abstract #P31E-04.
- Paige, D.A. et al., 2013. Thermal stability of volatiles in the north polar region of Mercury. Science 339 (6117), 300–303.
- Peale, S.J., 1969. Generalized Cassini's laws. Astron. J. 74, 483-489.
- Robie, R.A., Hemingway, B.S., Wilson, W.H., 1970. Specific heats of lunar surface materials from 90 to 350 K. Science 167, 749–750.
- Schorghofer, N., 2008. The lifetime of ice on main belt asteroids. Astrophys. J. 682 (1), 697.
- Schorghofer, N., 2010. Fast numerical method for growth and retreat of subsurface ice on Mars. Icarus 208 (2), 598–607.
- Schorghofer, N., Aharonson, O., 2014. The lunar thermal ice pump. Astrophys. J. 788 (2), 169.
- Schorghofer, N., Taylor, G.J., 2007. Subsurface migration of H₂O at the lunar poles. J. Geophys. Res. 112, E02010.
- Siegler, M.A., Bills, B.G., Paige, D.A., 2011. Effects of orbital evolution on lunar ice stability. J. Geophys. Res.: Planets (1991–2012) 116 (E3).
- Siegler, M.A., Aharonson, O., Choukroun, M., Hudson, T., Schorghofer, N., Xu, S., 2012a. Measurements of thermal properties of icy Mars regolith analogs. J. Geophys. Res.: Planets 117. http://dx.doi.org/10.1029/2011JE003938.o.
- Siegler, M.A., Paige, D.A., Williams, J.-P., Smrekar, S., 2012b. The Lowest Temperatures on the Moon: What can we learn? NASA Lunar Science Institute Forum, NASA Ames Research Center.
- Siegler, M.A., Bills, B.G., Paige, D.A., 2013. Orbital eccentricity driven temperature variation at Mercury's poles. J. Geophys. Res.: Planets 118 (5), 930–937.
- Smith, D.E. et al., 2010. Initial observations from the lunar orbiter laser altimeter (LOLA). Geophys. Res. Lett. 37 (18).
- Spudis, P.D., Bussey, B., Plescia, J., Josset, J., Beauvivre, S., 2008. Geology of Shackleton crater and the south pole of the Moon. Geophys. Res. Lett. 35, L14201.
- Touma, J., Wisdom, J., 1994. Evolution of the Earth–Moon system. Astron. J. 108, 1943–1961.
- Ward, W.R., 1975. Past orientation of the lunar spin axis. Science 189 (4200), 377-379.
- Watson, K., Murray, B.C., Brown, H., 1961. The behavior of volatiles on the lunar surface. J. Geophys. Res. 66 (9), 3033–3045.
- Webb, D.J., 1982. Tides and the evolution of the Earth–Moon system. Geophys. J. Int. 70 (1), 261–271.
- Wieczorek, M.A., Le Feuvre, M., 2009. Did a large impact reorient the Moon? Icarus 200 (2), 358–366.
- Wisdom, J., 2006. Dynamics of the lunar spin axis. Astron. J. 131, 1864-1871.
- Zuber, M.T. et al., 2012. Constraints on the volatile distribution within Shackleton crater at the lunar South Pole. Nature 486 (7403), 378–381.