Glaciation on Mercury: Accumulation and flow of ice in permanently shadowed circum-polar crater interiors

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A B S T R A C T

Radar-bright deposits that coincide with regions in permanent shadow typically found within impact craters at and near the poles of Mercury have been interpreted as being composed of water ice. We investigate the dynamic properties of these solid water-ice deposits (glaciers) with the goal of constraining their movement, flow rates, possible deformation, and related structures and deposits. As an end-member, we treat the extreme case of maximum ice accumulation where these deposits fill the crater to the shadow line, an event we would only expect following maximum accumulation conditions such as a large cometary impact or a comet shower. We find that, given the extremely cold conditions and the limited thickness of glaciers, even under the most favorable accumulation conditions, glaciation is cold-based, ice flow velocities are very low (4 × 10⁻⁸ m/yr) and, with the exception of a sublimation lag deposit, glaciation is unlikely to leave any significant impact on the terrain. An important accelerator of ice flow is found to be the contribution of lateral heat conduction from the surrounding extremely hot surface terrain (~ 225 K average with up to 450 K maximums), which for 55 km craters yields velocities of 10⁻³ m/yr while for smaller 10 km craters velocities may exceed 1 m/yr, enough to leave a detectable ice deformation imprint such as drop moraines. However given that the observed deposits are substantially below their maximum fill capacities these estimates are unlikely to be attained.

1. Introduction

Earth-based radar observations have revealed the presence of radar-bright materials near the poles of Mercury shown in Fig. 1 (Slade et al., 1992; Harmon and Slade, 1992; Butler et al., 1993; Harmon, 2007; Harmon et al., 1994; Harmon et al., 2011). Analysis of the radar data suggests that these highly reflective materials are composed of nearly pure water ice, with less than ~5% volume fraction of silicates (Butler et al., 1993). Furthermore, enhanced concentrations of hydrogen in the north polar region of Mercury are also consistent with a water-ice composition for the ice, as suggested from measurements by the Neutron Spectrometer on board the MErcury Surface, Space ENVIRONMENT, Geochemistry, and Ranging (MESSENGER) spacecraft (Lawrence et al., 2013). Extensive data from MESSENGER provides evidence that these water-ice deposits are distributed within the permanently shadowed terrains near the poles of Mercury (Chabot et al., 2012; Chabot 2013; Deutsch et al., 2016). Both reflectance (Neumann et al., 2013) and visible imaging (Chabot et al., 2014, 2016) of the permanently shadowed surfaces reveal that ice on Mercury today is present in two configurations: (1) as stable surface ice, where surface reflectance is anomalously high (Neumann et al., 2013) or (2) as stable subsurface ice, buried beneath insulating materials that have anomalously low reflectance (Neumann et al., 2013). This insulating layer is estimated to be relatively thin (~ 10–30 cm thick) on the basis of comparative variation of the flux of epithermal and fast neutrons with latitude (Lawrence et al., 2013). The low-reflectance materials are interpreted to have formed as a sublimation lag, and are suggested to be composed of volatiles species other than water, including organic-rich compounds (Paige et al., 2013).

Thermal models suggest that water ice is stable within the permanently shadowed terrains on Mercury on geologic timescales (Paige et al., 1992; Paige 2013). Mercury provides a unique environment for ice accumulation due to lack of an appreciable atmosphere to trap heat and the planet’s low obliquity of 0.034° (Margot et al., 2012). Insolation on Mercury varies not only with latitude, but also with longitude due to the planet’s eccentric orbit and 3:2 spin-orbit resonance: the average insolation is higher at longitudes 0°E and 180°E than the average insolation at longitudes 90°E and 270°E (Vasavada et al., 1999). A detailed energy-balance model of the surface temperatures around and within polar craters (Vasavada et al., 1999)...
suggests an annual mean temperature near 110 K in permanent shadow, and surface temperatures close to 400 K in sunlit regions, creating a substantial thermal gradient across permanently shadowed terrains in the polar regions.

Here, we describe possible dynamical properties of the ice accumulations in the unique thermal environment of Mercury. While thickness estimates for the current state of the ice deposits on Mercury range from several m (Harmon, 2007; Black et al., 2010) to ~50 m (Eke et al., 2017; Susorney et al., 2017; Deutsch et al., 2018), there is little understanding of how the state of the ice accumulations may have changed through time. Here, we explore dynamics of ice under mercurian conditions in order to delineate the effects of ice deformation and the potential for flow that could modify the observed ice deposits over geologic time. In order to evaluate the potential for flow, we treat the extreme end-member case in which ice accumulates in craters to the shadow line (permanent shadow terminator, or PST), and assess glacial movement styles (wet-based or cold-based) and glacial flow rates compared to glaciers on the Earth and Mars. We then analyze the role of heat conduction from surrounding terrain into the margins and base of the glacier and its influence on flow rates. We conclude with an assessment of the range of processes involved in glaciation on Mercury (from initial ice accumulation to the post-glacial period), comparing these to the Earth and Mars, and listing predictions for recognition of periods of earlier and more extensive glaciation on Mercury.

2. Modeling the glacial flow process

2.1. Crater morphology and bed topography

Modeling glacial flow requires topography on which to reconstruct the glacier or ice sheet in question. The depth of a crater can be approximated from the ratio of depth, d, to diameter, D, (d/D, (Nagel and Fechtig, 1980)), which Vasavada et al. (1999) modeled for mercurian craters from idealized crater geometries as 1:5 (depth is 1/5th the diameter) for craters less than 10 km in diameter and 1:25 for craters with diameters of 100 km. We found from looking at topographic profiles (transsects) across a mercurian crater that the d/D ratio could go as low as 1:40 and that crater-wall slopes were typically ~7.5° (see Barnouin et al., 2012, his Fig. 1, for a 120 km crater). More recently, various researchers used topographic data acquired by the Mercury Laser Altimeter (MLA) to derive various power-law fits for depth-diameter relationships for both simple craters (crater less than 10 km in diameter and generally bowl-shaped) and complex craters (those larger than 10 km and generally with a flat crater floor) (Barnouin et al., 2012; Talpe et al., 2012; Susorney et al., 2016). These power-law fits also provide estimates of crater-wall slope ranging from 15–45° for small 10 km craters to 5–30° for larger 100 km craters.

In our modeling, we utilize an idealized crater geometry that is defined by various power-law fits, is a function only of crater diameter, and was derived from topographic measurements of more than 300 mercurian craters (Susorney et al., 2016). Given only crater diameter, these power laws provide estimates of the depth, the rim-crest height, and the wall width from which we define our idealized crater geometry. Depth as a function of diameter is shown in Fig. 2(a). The red dot-dash line is depth from Talpe et al. (2012), and the blue dashed line is depth from Susorney et al. (2016), both with indicated uncertainties and good agreement. In calculating the effective shadowing depth, we include the rim-crest height, also from Susorney et al. (2016), indicated by the solid black line. Also shown in the top gray box are the power-law fits for the wall widths, defined as the lateral distance between the rim crest and the crater floor, (Susorney et al., 2016) that define the crater-wall slopes.

A profile generated from these power laws is shown in Fig. 2(b) for a 20 km crater (black solid line). The blue dashed line indicates the specified diameter, from which depth (red dashed line), rim-crest height (green dashed line) and wall width (magenta dashed line) are derived using the power-law fits in Susorney et al. (2016). Also shown are permanent shadow terminator elevations for various latitudes (85–89°, red, green, blue, magenta, and cyan solid lines, respectively). The PST elevations increase with latitude to the increasingly grazing angle of the Sun coupled with planet’s low obliquity. In the top gray box the profile is shown with a 1:1 scaling.
For input to the ice sheet model, a 2-dimension rectangular array of grid points representing the bed topography is generated by rotating the bed profile obtained from the power laws about the center of the crater. Performing an orthographic rotation of these grid coordinates such that the sun angle, defined by the specified latitude of the crater, is in a horizontal plane produces the shadow surface. Moving along each grid line we select the first highest point encountered, setting each subsequent point to be in shadow if it is below that first highest point. The original un-rotated height of that shadow surface defines the permanent shadow terminator elevation.

An example of this simplified geometry based on Crater C (a diameter of 50 km at latitude 87.7° from Vasavada et al. (1999) (their Table 1) is shown in Fig. 3. Fig. 3(a) shows bed topography, Fig. 3(b) the ice surface elevation, Fig. 3(c) the ice thickness, and Fig. 3(d) a 3D perspective view of the crater with the potential permanently shadowed volume shown as completely full of ice. While this end-member example greatly exceeds the realistic volumes for water-deposits on Mercury, given that ice thickness are estimated between several m (Harmon, 2007; Black et al., 2010) and ~50 m thick (Eke et al., 2017; Susorney et al., 2017; Deutsch et al., 2018), we model here the extreme case of maximum ice thickness allowable, the case in which these deposits fill the crater to the PST.

2.2. Ice loss and supply

This maximum ice configuration is then passed to our ice dynamics model (UMISM, the University of Maine Ice Sheet Model), a finite-element shallow-ice approximation model that has been used extensively both on Earth (Johnson and Fastook 2002; Kleman et al., 2002; Hooke and Fastook, 2007) and Mars (Fastook et al., 2008; Fastook et al., 2011; Fastook et al., 2012; Fastook and Head, 2014; Fastook and Head, 2015), here adapted for Mercury gravity and thermal environment. UMISM is a thermomechanically-coupled ice sheet model, where ice properties that control deformation are functions of temperature, which itself is derived from a time-dependent solution of the energy equation within the ice sheet volume, accounting for advection of heat by moving ice, internal shear heating, and geothermal heat flux.

An ablation rate of −1.0 m/yr is assumed in the sunlit areas, sufficient to quickly remove any ice that moves out of the PST volume. Minor accumulation (10⁻¹⁰ m/yr) is assumed for the ice surface within the PST, consistent with estimates of water-ice delivery via impacts to Mercury (Moses et al., 1999). With such a low accumulation rate the ice sheet surface will basically attempt to “relax” into a configuration where the driving stress (the product of density, gravitational acceleration, thickness, and surface slope) is minimized and any observed velocity is a result of internal deformation and not of mass-balance driven flux to accommodate mass input from surface accumulation.

Table 1

<table>
<thead>
<tr>
<th>Crater locations</th>
<th>Diameter (km)</th>
<th>Latitude (N)</th>
<th>Longitude (E)</th>
</tr>
</thead>
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<tr>
<td>C</td>
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<td>188.8</td>
</tr>
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<td>188.4</td>
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<td>85.2</td>
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</tr>
<tr>
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<td>84.9</td>
<td>353.5</td>
</tr>
<tr>
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<td>88.8</td>
<td>148.9</td>
</tr>
<tr>
<td>Y</td>
<td>18.2</td>
<td>87.6</td>
<td>210.2</td>
</tr>
</tbody>
</table>

Fig. 2. (a) Crater depths are estimated for 10–55 km diameter craters from the power-law fits of Susorney et al. (2016) in the blue dashed line, and of Talpe et al. (2012) in the red dot-dash line. The effective shadow depth (calculated as the sum of the crater depth and rim-crest height) is shown by the solid black line. The wall width, defined as the distance between the crater rim crest and the crater floor, is shown by the solid black line in the top gray box. (b) Transect across a 20 km idealized crater defined by diameter (blue dashed line) from which depth (red dashed line), rim-crest height (green dashed line), and wall width (magenta dashed line) are derived through Susorney et al. (2016) power-law fits. Also shown are the permanent shadow terminator elevations for various latitudes. The gray box at the top contains the same transect at a 1:1 scale. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article).
This is the first fundamental difference between glaciation on Mercury and on planetary bodies such as Earth and Mars: the lack of an atmosphere on Mercury drives relatively high ablation rates, and therefore precludes net ice accretion in the accumulation zone, resulting in mass-balance driven flux to accommodate mass input from seasonal, annual, or longer-term surface ice accumulation.

2.3. Mechanical properties of the ice

Boundary conditions for the thermodynamic component of UMISM, used to calculate the mechanical properties of the ice (i.e., its hardness), include (1) mean-annual surface temperature, taken to be 110 K for exposed water ice (Vasavada et al., 1999), and (2) basal geothermal heat flux. Because the initial modeled ice filling the permanently-shadowed volume is constrained in thickness (basically the crater depth) and the slope is shallow (basically the sun angle), as well as the fact that the ice is cold (∼ 110 K), modeled velocities are extremely low, on the order of $10^{-8}$ m/yr for a uniform geothermal flux of 50 mW/m². This is 7 to 8 orders of magnitude slower than ice flow velocities of meters to decameters per year for typical glaciers on Earth (Paterson 1994). Ice flow velocity, shown in Fig. 4, is highest at the base of the crater wall where the maximum case thickness explored here is greatest at just over 2400 m, which we again acknowledge is unlikely to have ever been attained on Mercury, even during an episode of heavy bombardment by water-delivering comets. Long-term deformation with this velocity distribution would result in a thinning upstream toward the crater rim and a thickening downstream toward the crater center as the profile relaxes to a minimum driving stress configuration. This downstream thickening would tend to drive the ice in the crater center past the PST and into the illuminated portion of the crater, causing rapid ablation there. However, with velocities this low, little deformation would take place even over billion-year time spans. Maximum forward velocities would result in less than a meter glacial advance into sunlight over a 100 million years.

Warmer ice (assumed to be ordinary ice $I_h$) is exponentially softer than colder ice due to the temperature dependence of the rate parameter, $A(T)$, in the non-linear Glen's Flow Law (Glen, 1958; Paterson and Budd, 1982) relating strain rate ($\dot{\varepsilon}$) to stress ($\tau$). The exponent, $n$, is usually taken to be 3.

$$\dot{\varepsilon} = A(T)\tau^n$$

Fig. 3. Idealized geometry for Crater C from Vasavada et al. (1999), Table 1, with diameter of 50 km at 87.7°N, 188.8°E. (a) Bed elevation (m), (b) surface elevation (m), (c) ice thickness (m) that in our extreme-case simulation fills the permanently-shadowed volume, and (d) a 3D view of the maximum end-member case treated here in which ice is filled to the shadow line (permanent shadow terminator PST). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article).
The Arrhenius relationship is widely used to describe the temperature dependence of many physical properties, including chemical reaction rates, diffusion coefficients, as well as creep phenomena such as is described by Glen's Flow Law.

\[ A(T) = A_0 e^{-\frac{Q}{RT}} \]  

Here \( R \) is the Boltzmann constant, \( T \) is temperature in Kelvins, and \( A_0 \) (the pre-exponent factor) and \( Q \) (the activation energy) are constants obtained by fitting to data. Since Arrhenius is considered to be reasonably accurate down to a homologous temperature (the temperature divided by the melting point) of 0.4 (Cadek, 1988), we expect this reasonably extends the measured temperature dependence of ice (Paterson, 1994) down to the cold temperatures in Mercury’s permanently shadowed craters (∼−110 K). At these temperatures, a one-degree increase in temperature doubles the rate parameter in Glen’s Flow Law, effectively doubling the velocity obtained by integrating vertical strain rates from the bed to the ice surface to find the column-averaged velocities necessary for the conservation of mass calculation in UMISM.

Ice at depth is warmer than ice at the surface due to the insulating power of the ice thickness. As such, the heat flux delivered to the base of the ice is a key parameter controlling ice deformation, and hence, the ice velocity. Vasavada et al. (1999) uses an estimate for the heat flux of 20 mW/m² (Schubert et al., 1988). An estimate based on the height of a 40 mW/m² originating from the mantle (which they allow could be as much as 10 mW/m²) with the rest provided by the radiogenic contribution of the crust.

An additional consideration is the lateral transport of heat from the warm sunlit surroundings of the cold permanently shaded crater interior. This lateral transport into a crater is illustrated with a simplified geometry consisting of a rectangular region of water ice with a cold surface and a depth/diameter relationship appropriate for mercurian craters embedded in rock with a hot surface that is heated from below by a geothermal heat flux. A 2D solution of the steady-state heat flow equation (Becker, 1981) for such a rectangular domain with a “cold spot” on the surface results in a depression of the isotherms below the cold spot. This depression and warping of the isotherms tends to direct additional heat from the surroundings beneath the hot surface terrain into the center of the cooler region beneath the cold spot. This is demonstrated for a 50 km cold spot in Fig. 4 for a nominal heat flux of 50 mW/m² entering the base of the domain at a depth of 20 km. Fig. 5(a) illustrates the conductivity, showing higher conductivity rock (Eppelbaum et al., 2014) surrounding the low conductivity ice-filled region. The temperatures within the modeled region are shown in Fig. 5(b), and Fig. 5(c) shows the enhanced interior heat fluxes beneath the simplified crater exceeding 150 mW/m². Heat delivered to the bottom of the ice-filled region is shown in Fig. 5(d); most of the heat is concentrated near the edges of the crater, although even at the center the delivered flux exceeds the geothermal heat by a factor of 1.5 (75 mW/m²).

The effect of lateral heat flux is influenced by the diameter of such a simplified crater, with the effect being most pronounced near the edges at the surface-temperature discontinuity. For this reason smaller craters have much higher heat fluxes at their centers, decreasing as the crater diameter increases. This effect is illustrated in Fig. 6, which shows the heat flux delivered to the bottom of the simplified crater as a function of diameter for a nominal heat flux of 50 mW/m² at the base of the 20 km deep domain. This effect is further demonstrated in Fig. 7, which shows the temperature (top row) and heat flux fields (bottom row) beneath the cold spot for diameters 10, 20, 30, 40, and 50 km (a–e, respectively). Temperature isotherms are bowed down beneath the cold spot, and since heat flows perpendicularly to the isotherms, heat is preferentially focused onto the base of the cold spot. The region of enhanced lateral flux is most intense beneath the edge of the crater, where the temperature discontinuity is largest (400 to 110 K). With the smallest 10 km crater modeled here, the two contributions from the right and the left overlap in the center, resulting in the largest flux at the center of the crater. Larger craters result in a deeper penetration of the temperature anomaly, and hence steeper, temperature gradients carrying more heat flux, but since they are farther apart, the synergistic contribution is less at the center of the crater and less enhanced flux is delivered. Accounting for the lateral transport of heat conducted through the bedrock to small craters suggests the most potential for warming from enhanced geothermal heat flux, and the least likelihood that deposits would survive for significant periods of time.

3. Results and discussion

3.1. The effects of enhanced heat fluxes

Having derived the significantly increased heat flux that can be delivered to the base of the crater from lateral conduction (e.g., Figs. 6 and 7), we now can apply the ice sheet model with these enhanced fluxes instead of the uniform 50 mW/m², an example of which was shown in Figs. 3 and 4. To investigate the effect of the enhanced lateral heat flux, we run the ice sheet model with both uniform and enhanced heat fluxes for various size craters and various latitudes (10 to 55 km in 5 km steps and 85 to 89.5° in 0.5° steps, a total of 100 model runs). Fig. 8(a) shows the base-10 logarithm of the maximum velocity observed in each uniform flux model run (a value of −10 corresponds to 10⁻¹⁰ m/yr). Labeled black stars indicate the size and latitude of craters with known deposits from Vasavada et al. (1999) Table 1. Fig. 1 and Table 1 show locations and size of these craters. Crater C, which is as yet unnamed, was shown as an example in Figs. 3 and 4. Since depth is determined from the depth/diameter relationship, larger craters in general display higher velocities, primarily due to the thicker ice that can exist in the deeper craters. A secondary effect is the slope, which is steeper at lower latitudes due to the sun being higher above the horizon. This effect is negated by the thinner ice available with these steeper slopes, resulting in an optimal latitude at 87.0° where velocity is highest (4.1 × 10⁻⁸ m/yr). Lowest velocities (6.7 × 10⁻¹² m/yr) are observed in the smallest craters at the highest latitudes where the sun
We acknowledge that the model is not accounting for longitudinal changes in insolation, since temperature shows both latitudinal and longitudinal symmetries on the surface of Mercury due to the 3:2 spin-orbit resonance of the planet. The insolation at longitudes 90°E and 270° is less than the insolation at 0°E and 180°E (Vasavada et al., 1999) and at this point, we are accounting only for latitudinal variations. Future modeling should investigate the longitudinal effects, given that mapping of radar-bright materials within permanently shadowed craters suggest that cold pole longitudes provide a more favorable thermal environment for water ice (Chabot et al., 2013).

Fig. 8(b) shows the maximum temperature (with the surface uniformly at 110 K) observed at the bed, generally at the thickest point at the base of the crater wall on the sunward side. As would be expected, the thickest ice in large, high-latitude craters with low sun angles experience the greatest warming of the bed (34.5 K).

Re-running the same set of crater sizes and latitudes, but with the enhanced heat fluxes that includes lateral heat transport of Fig. 6, we obtain a distinctly different appearing pattern of temperatures at the bed, shown in Fig. 8(d). Immediately obvious is that in all cases the bed of the ice is much warmer. The warmest ice (253 K) is now in the small craters that experience the greatest enhancement of the heat flux at the base of the ice (Fig. 6). Since higher latitude leads to lower sun angles, and hence thicker ice, basal ice is still warmest at high latitudes.

Fig. 8(f) shows the difference between Fig. 8(b) and (d), which is the amount of warming due solely to the use of the enhanced heat fluxes. The greatest warming (∼121 K) occurs for the small 10 km, highest latitude craters, and the least warming (34 K) occurs for the largest 55 km, lowest latitude craters. This warmer ice, being softer and easier to deform, yields higher velocities, as shown in Fig. 8(c).

With a specified surface temperature and a specified heat flux at the base of the ice, the ice column is warmer toward the bed, resulting in softer, more easily deformable ice. The overall pattern is very different from the uniform 50 mW/m² heat flux case of Fig. 8(a), with the highest velocities now in the smallest craters (as high as 5 m/yr at 87.0°). The order of magnitude speedup due to this warmer softer ice is shown in Fig. 8(c). Even with large craters at low latitudes, where the warming is least (34.5 K), a 4.3 orders of magnitude speedup is observed (from $2.6 \times 10^{-8}$ m/yr to $5.3 \times 10^{-4}$ m/yr). Large craters where uniform-flux velocities are a maximum, experience a 4.6 orders of magnitude increase in velocity with the enhanced fluxes (from $4.1 \times 10^{-8}$ m/yr to $1.6 \times 10^{-3}$ m/yr). Small craters at middle latitudes (87.5°) see the largest speedup (10.9 orders of magnitude faster, albeit with very low velocities to begin with; from $6.7 \times 10^{-12}$ m/yr to $5.2 \times 10^{-1}$ m/yr).

Given the sensitivity to heat flux observed with the uniform versus

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Fig. 5. Lateral heat transport from hot (∼400 K) sunlit terrain to cold (∼110 K) crater interior. (a) Conductivity, (b) temperature, and (c) heat flux are shown for a 50 km diameter cold spot, and (d) the heat delivered to the bottom of a completely ice-filled crater is modeled. Boundary conditions consist of specified temperatures across the top (grading from a ∼400 K sunlit terrain to a ∼110 K permanently shadowed cold spot) and specified flux along the bottom at 20 km depth (50 mW/m²). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article).
enhanced heat flux, an experiment was run for various base heat fluxes, both with (solid line) and without (dashed line) lateral heat transport. Using as an example, Crater C (displayed in Fig. 3 and 4), with a diameter of 50 km and at a latitude of 87.7°, we obtained the results shown in Fig. 9 for basal heat flux ranging from 0 to 100 mW/m². While anything over 60 mW/m² is probably unreasonable for anything except a landscape experiencing volcanism or a recent projectile impact, we show results to emphasize how strongly dependent velocity is on heat flux through the Arrhenius-activated thermal softening of ice. Over the range from 10 to 100 mW/m² we see more than eight orders of magnitude increase in velocity from $1.4 \times 10^{-12}$ m/yr to $4.0 \times 10^{-4}$ m/yr for the crater with uniform heat flux, and more than five orders of magnitude increase in velocity from $3.6 \times 10^{-8}$ m/yr to $1.09$ m/yr when lateral heat transport is included. Note that for the unreasonable 100 mW/m² heat flux value, the maximum observed velocity of $4 \times 10^{-4}$ m/yr for the uniform heat flux case is within an order of magnitude of velocities observed on slow-moving terrestrial glaciers such as the debris-covered Mullins glacier in the Antarctic Dry Valleys (Marchant et al., 2007), whereas in the lateral heat transport case as little as 50 mW/m² is sufficient to match the velocity of the slow-moving terrestrial glaciers.

Figs. 10 and 11 show results analogous to Fig. 8 for base geothermal heat fluxes of 20 and 30 mW/m². Patterns of maximum velocity and basal temperatures with and without lateral heat transport are similar to those shown in Fig. 8(a–d) for the 50 mW/m² case, except of course that both temperatures and velocities are lower. Acceleration factors and warming due to lateral heat transport show the same patterns as Fig. 8(e) and (f).

Again, for the uniform flux cases the thickest ice in large, high-latitude craters with low sun angles experience the greatest warming of the bed (13.8 K for 20 mW/m² and 20.5 K for 30 mW/m²). The greatest warming due to lateral heat transport (115 K for 20 mW/m² and 117 K for 30 mW/m²) is still in the smallest 10 km, highest latitude craters, and the least warming (33.6 K for 20 mW/m² and 33.8 K for 30 mW/m²) is still found in the largest 55 km, lowest latitude craters. For the uniform flux cases we still see an optimal latitude for 55 km craters (85.5°, $4.0 \times 10^{-11}$ m/yr for 20 mW/m² and 86.0°, $4.4 \times 10^{-10}$ m/yr for 30 mW/m²) where velocity is highest. Lowest velocities ($2.1 \times 10^{-14}$ m/yr for 20 mW/m² and $1.6 \times 10^{-13}$ m/yr for 30 mW/m²) are still observed in the smallest craters at the highest latitudes where the sun angle is lowest.

The highest velocities for the lateral heat transport case are still in the smallest craters (0.88 m/yr for 20 mW/m² and 1.6 m/yr for 30 mW/m², both at 87.0°). Speedup due to the warmer softer ice for large low-latitude craters where the warming speedup spans 5.2 orders of magnitude increase (from $4.0 \times 10^{-11}$ m/yr to $6.5 \times 10^{-6}$ m/yr) for 20 mW/m² and 4.9 orders of magnitude (from $3.9 \times 10^{-10}$ m/yr to $2.9 \times 10^{-5}$ m/yr) for 30 mW/m². Large craters where uniform-heat velocities are a maximum experience a 5.3 orders of magnitude increase in velocity with the enhanced fluxes (from $4.1 \times 10^{-11}$ m/yr to $8.5 \times 10^{-6}$ m/yr) for 20 mW/m² and a 5.1 orders of magnitude increase in velocity (from $4.3 \times 10^{-10}$ m/yr to $5.0 \times 10^{-5}$ m/yr) for 30 mW/m². Small craters at high latitudes (89.5°) experience the

**Fig. 6.** Heat flux delivered to the bottom of an idealized crater for nominal heat flux of 50 mW/m² at the base of the 20 km domain.

**Fig. 7.** Temperature fields (top row) and heat flux fields (bottom row) for (a) 10, (b) 20, (c) 30, (d) 40, and (e) 50 km-diameter (left to right) cold spots. In all of the figures, the black and white outlined rectangle indicates the low conductivity ice. In the flux figures (bottom row), a green line outlines the region where the flux exceeds the 50 mW/m² flux specified along the bottom of the domain at a depth of 20 km delineating regions of enhanced heat flux. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article).
largest speedup (12.4 orders of magnitude faster from $2.1 \times 10^{-14}$ m/yr to $5.6 \times 10^{-2}$ m/yr for 20 mW/m² and 11.9 orders of magnitude faster from $1.6 \times 10^{-13}$ m/yr to $1.2 \times 10^{-1}$ m/yr for 30 mW/m²).

All of these results from Figs. 8, 10, and 11 are summarized in Table 2.

4. Summary and conclusions

On the basis of our analysis of the factors that are important in ice accumulation and flow processes in permanently shadowed areas on Mercury, we characterize Mercury glaciation as follows (Fig. 12):

1. Geometry of ice accumulation: In order for water ice to accumulate on Mercury, permanently shadowed regions are required and thus ice accumulation depends on location and on the size of the depression (Fig. 12(a)); impact crater interiors are clearly the most likely geologic environment for such accumulation, although volcanic calderas and pit craters could create a similar environment, as could some shallower topographic depressions and rough terrain at higher latitudes.

2. Ice accumulation environment: The ice accumulation environment on Mercury is fundamentally different than on planetary bodies such as Earth and Mars. The lack of an atmosphere on Mercury precludes atmospheric deposition as snow and net ice accretion in the accumulation zone, resulting in mass-balance driven flux to accommodate mass input from seasonal, annual, or longer-term surface ice accumulation. Instead, water molecules are delivered to the permanently shadowed cold traps by thermal migration processes, and are derived from some combination of internal sources (magmatic volatiles) and external sources (cometary or other water-rich impact events).
3. Ice accumulation rates: Unknown are the actual ice accumulation rates and time of emplacement. Moses et al. (1999; their Table 1) assess a wide range of volatile sources (Interplanetary dust particles, asteroids, Jupiter family comets, and Halley-type comets) and calculate that they would deliver from 1 to 95 m of ice to the cold traps over $3.5 \times 10^9$ yr, an accumulation rate of $2.8 \times 10^{-10}$ to $2.7 \times 10^{-9}$ m/yr. They point out that delivery in their Monte Carlo simulation is dominated by a few large impactors, so the uniform

Fig. 10. Analogous to Fig. 8 but for a heat flux of 20 mW/m². (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article).

Fig. 11. Analogous to Fig. 8 but for a heat flux of 30 mW/m². (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article).
rate is a minimum and could have been larger episodically. The overall impact flux has declined through time, but individual cometary impacts or comet showers are likely to be the most significant water source in recent geologic history. Water can also be delivered to planetary surface through volcanic outgassing (Wilson and Head, 2008; Head and Wilson, 2015), however the low-FeO content in surface silicates on Mercury suggests highly reducing conditions (Robinson and Taylor, 2010). Even if water is outgassed in mercurian eruptions, declining rates of volatile input from effusive and explosive eruptions (Wilson and Head, 2008; Denevi et al., 2013; Ostrach et al., 2015; Frockter et al., 2010; Head and Wilson, 2015; Head et al., 2009) suggest that volcanism has not been a major source of water in the last third of the history of Mercury.

Moses et al. (1999) point out that the observed purity of the radar-bright materials supports episodic large accumulations, because the ice would be much dustier with slow continuous deposition. Episodic deposition is also consistent with the observed sharp albedo boundaries of the deposits, which also suggests that the ice was delivered relatively recently (Chabot et al., 2014). Crider and Killen (2005) estimates the rate of burial of ice deposits in polar craters on Mercury to be 0.43 cm/Myr. If the ice deposits in the polar craters on Mercury consist of clean ice and the deposits are buried by 20 cm of regolith, they must be relatively recently emplaced, less that ~50 Myrs ago (Crider and Killen, 2005).

4. Predicted ice stratigraphy: On the basis of stochastic episodic supply events (individual impactors that vary in water content, velocity, impact angle, impact location and frequency), individual supply events are predicted to produce a layered accumulation sequence representing the cumulative record of individual events. Current lack of knowledge of the specific role of scattered light on favorable water molecule destinations within the cold trap, and an incomplete understanding of thermal migration dynamics in general, precludes a more detailed picture, but water molecules are unlikely to be emplaced in a layer of even thickness across the cold trap footprint. Also uncertain is the time between events; the presence of an organic-rich lag deposit on some current ice occurrences (Neumann et al., 2013; Chabot et al., 2014) suggests that ablation between events could produce a distinctive stratigraphy of paired ice layers overlain by a darker lag.

5. The nature of glacial flow and ice flow velocities in the maximum ice accumulation case (crater filled to the permanently shadowed terminator, as modeled here): As an end-member for the maximum possible extent of glaciation in permanently shadowed regions, we treated the extreme case where ice deposits fill the crater to the shadow line (PST) (Fig. 12(b)), a situation we would only expect following maximum accumulation conditions such as a large cometary impact or a comet shower. We found that temperatures are sufficiently cold that even under these maximum accumulation conditions, flow velocities were vanishingly small (10−8 to 10−11 m/yr without lateral conduction), compared to typical flow velocities for slow, cold debris-cover glaciers on Earth (~10−5 m/yr (Marchant et al., 2007)), and closer to, but still considerably below, typical flow velocities on Mars (~10−5 m/yr (Fastook et al., 2011)). Including lateral conduction increases potential velocities for craters filled to the PST to 10−6 m/yr for the 55 km craters and to potentially 1 m/yr for 10 km craters, certainly enough to produce observable deformation. However, these are for the extreme case of thick ice (>1000 m even in the 10 km craters). Given current estimates of the timing of ice deposits (<50 million years old (Crider and Killen, 2005)) and thickness of the ice (between several m (Harmon, 2007; Black et al., 2010)) and ~50 m (Eke et al., 2017; Susorney et al., 2017; Deutsch et al., 2018), this suggests that the ice on Mercury today was deposited and likely has not advanced at all.

6. The effects of lateral conduction on glacier thermal structure and flow rates: The extreme temperature difference (Paige et al., 2013) between permanently shadowed regions (75–100K) and adjacent regions illuminated by the Sun (200–400 K) means that heat will be conducted through the glacial substrate toward the deposit base (Fig. 12(a) and (b)). We explored a range of parameters and found that although maximum effects on ice thermal properties varied depending on crater geometry and illumination, the net effect was insufficient to bring flow style into the realm of wet-based conditions, even for a maximum end-member case where ice is filled to the PST (Fig. 12(c)). When assisted by lateral conduction the maximum ice flow rates range from 10−3 to 1 m/yr on Mercury, a rate that would certainly produce observable deformation over the lifetime of these deposits. For small ice patches (Deutsch et al., 2017), lateral conduction is likely to be a limiting factor in ice accumulation and retention. In the smallest crater for which we modeled the lateral heat flow (10 km, Fig. 6), the heat flux delivered to the base of the crater was five times the basic geothermal heat flux (50 mW/m²), and was increasing exponentially as crater size decreases. For a feature ~1 km in size the flux would be over ten times the basic flux delivering considerable heat to the base of the deposit making its long-term existence unlikely.

7. Style of glaciation: wet-based or cold-based?: On the basis of our analysis, all glaciation on Mercury in the range of maximum accumulation conditions that we treated would be frozen to the bed, and thus cold-based (Fig. 12(C)). Cold-based glaciation would severely limit the production of geomorphic features (such as drumlins, eskers and other features typical of wet-based glaciation)
that might provide evidence of previous thick ice deposits in permanently or formerly permanently shadowed regions (Fig. 12(d)). Wet-based conditions are predicted to occur only in special situations such as: (1) a magmatic hot spot (likely to occur only earlier in the history of Mercury), (2) a magmatic intrusion into ice deposits, as seen on Earth and Mars (Head and Wilson, 2002; Head and Wilson, 2007; Wilson and Head, 2002, Wilson and Head, 2007; Kadish et al., 2008; Scanlon et al., 2014; Scanlon, 2015), or (3) an area of elevated geothermal flux remaining from the waning stages of a newly formed impact at high enough latitudes to produce permanent shadow (the initial thermal flux is likely to be sufficiently high that water ice would not accumulate).

8. Detecting the presence of previous ice deposits: Under current obliquity conditions (0.034° (Margot et al., 2012)), ice deposits are stable in the present permanently shadowed regions and ablate with time from scattered light. On the basis of our findings, following complete ablation of the ice, little positive evidence should remain indicating that ice previously existed in these regions due to the cold-based nature of the glaciation and the extremely low flow velocities predicted for the ice deposits (Fig. 12(d)). Development of a low-albedo layer (Neumann et al., 2013; Chabot et al., 2014) of possible organic composition (Paige et al., 2013) could result in a lag deposit remaining on the crater floor following complete loss of ice. However, low-albedo lag deposits on Mercury are found exclusively in permanently shadowed regions, suggesting that they, too, are volatile-rich (Paige et al., 2013), thus even this glaciation-associated remnant proxy for previous ice deposits would not survive for an extended period of time beyond water ice loss. However, their existence in permanently shadowed terrains that are of higher temperatures than terrains where water ice is exposed suggests that these volatile species are stable at higher temperatures than water ice (Paige et al. 2013; Neumann et al., 2013; Chabot et al. 2016). Cold-based glacial deposits are often accompanied by drop moraines, which are ridges of supraglacial debris deposited at the margins of the ice deposit (Head and Marchant, 2003; Kadish et al., 2014; Mackay et al., 2014). Drop moraines form at a stationary ice front where the forward advance rate is equal to the marginal ablation rate: ice continues to move forward in the glacier but any supraglacial debris is deposited at the stationary front as a moraine (Paterson, 1994). On the basis of the extremely slow moving ice in our simulations even for a maximum end-member case where ice is filled to the PST, and the low likelihood of significant waxing and waning of the ice deposit front, we would not expect abundant drop moraines in the geologic record. Furthermore, the paucity of potential debris observed on current Mercury ice deposits, and the extremely slow flow rates that might serve to redistribute debris, also reduce the possibility of drop moraines. Future mission imaging observations (Benkhoff et al., 2010) into the permanently shadowed regions illuminated by scattered light might search for (1) marginal deposits of low-albedo sublimation lags (a type of drop moraine), (2) boulder piles or ridge segments formed from debris excavated through the glacier by impacts, or (3) rockfalls from the crater wall, transported laterally to the glacier margin, and subsequently deposited (Fig. 12(d)). One of the strongest tell-tale signs of the previous presence of cold-based ice is the paucity of small craters in permanently shadowed portions of the crater floor (caused by “filtering” of craters formed solely in the ice), relative to the number of craters in the sunlight portions of the floor. This, too, can be investigated with high-resolution images of permanently shadowed surface.

9. Rates of horizontal ice movement versus vertical ablation: Terrestrial and martian cold-based glaciers advance relatively slowly, but still rapidly (six to ten orders of magnitude) compared to cold-based glaciers on Mercury. Because of this, if ice on Mercury ever accumulated to a significant thickness, then downward areal ablation rates may exceed flow velocity rates for forward movement ice, resulting in a different geometry of evolution than that typically seen on terrestrial glaciers (Fig. 12(c)). If the ice deposits ablate more quickly than they flow one would expect the low-albedo material would cover the entirety of the ice deposits. This may account for the fact that in almost every example of ice-bearing craters where there is a lag deposit, the low-reflectance material
covers the entirety of the ice deposit. Additional modeling of ice ablation rates in permanently shadowed areas under various conditions on Mercury, as well as knowledge of the effects of a low-albedo lag, are needed to assess this further.

Clearly, in the cold environment of the Mercury polar and circum-polar craters, flow velocities are extremely low (10⁻⁶ m/s/yr), and hence the potential for significant deformation of the ice deposits is small. It is clear that accounting for the enhanced flux of heat from the surrounding hot sun-lit terrain is important, offering orders of magnitude speedup over a uniform flux case. As is expected, larger, thicker deposits with the greatest surface slope would flow the fastest, leading to an optimal thermal environment conducive to the accumulation of thick ice due to the lower sun angle at high latitudes that in our model defines the surface slope of the deposit. However, given the estimate of ~50 m for the current deposits (with >1000 m required for even the modeled low velocities) we expect the deposits are supply limited, and that they are basically stagnant unmoving deposits, reflecting the extreme efficiency of the cold-trapping mechanism.

10. Implications for lunar polar volatiles: In a manner similar to Mercury, the lack of an atmosphere on the Moon prohibits atmospheric volatile deposition as snow and net ice accretion in the accumulation zone. Therefore, lunar polar volatiles should also migrate to cold traps via thermal processes (e.g., Crider and Vondrak, 2002), after possible delivery from water-rich impactors (e.g., Cintala, 1992) or volcanic outgassing (e.g., Needham and Kring, 2017), or through in-situ reactions from solar wind implantation (e.g., Crider and Vondrak, 2000). Given that radar measurements of lunar polar deposits show that they are more tenous than mercurian polar deposits (e.g., Lawrence et al., 2017), and that lunar polar deposits appear to be much more spatially heterogeneous on the surface (e.g., Hayne et al., 2015) in comparison to the polar deposits on Mercury, it is plausible that polar ice deposits on the Moon are relatively less thick and more degraded than those on Mercury. Therefore, given the results of our modeling, we consider that the flow velocities of any lunar polar deposits to be negligible, even when considering the effects of lateral conduction. However, if ice were ever to accumulate in such amounts as to approach the permanently shadowed terminator regions on the Moon, as modeled here for Mercury, then we would expect the style of glaciation of these deposits to be cold-based as well. If the ice on the Moon is relatively ancient (e.g., Siegler et al., 2016), then the ice stratigraphy within lunar polar craters may show layering correlated with the accumulation sequences, representing both the stochastic episodic supply events, as well as the intermittent deposition of ejecta from impact events (e.g., Hurley et al., 2012).

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