External Sources of Water for Mercury’s Putative Ice Deposits

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Received January 30, 1998; revised July 23, 1998

Radar images have revealed the possible presence of ice deposits in Mercury’s polar regions. Although thermal models indicate that water ice can be stable in permanently shaded regions near Mercury’s poles, the ultimate source of the water remains unclear. We use stochastic models and other theoretical methods to investigate the role of external sources in supplying Mercury with the requisite amount of water. By extrapolating the current terrestrial influx of interplanetary dust particles to that at Mercury, we find that continual micrometeoritic bombardment of Mercury over the last 3.5 byr could have resulted in the delivery of \((3–60) \times 10^{16}\) g of water ice to the permanently shaded regions at Mercury’s poles (equivalent to an average ice thickness of 0.8–20 m). Erosion by micrometeoritic impact on exposed ice deposits could reduce the above value by about a half. For comparison, the current ice deposits on Mercury are believed to be somewhere between ~2 and 20 m thick. Using a Monte Carlo model to simulate the impact history of Mercury, we find that asteroids and comets can also deliver an amount of water consistent with the observations. Impacts from Jupiter-family comets over the last 3.5 billion years can supply \((0.1–200) \times 10^{16}\) g of water to Mercury’s polar regions (corresponding to ice deposits 0.05–60 m thick), Halley-type comets can supply \((0.2–20) \times 10^{16}\) g of water to the poles (0.07–7 m of ice), and asteroids can provide \((0.4–20) \times 10^{16}\) g of water to the poles (0.1–8 m of ice). Although all these external sources are nominally sufficient to explain the estimated amount of ice currently at Mercury’s poles, impacts by a few large comets and/or asteroids seem to provide the best explanation for both the amount and cleanliness of the ice deposits on Mercury. Despite their low population estimates in the inner solar system, Jupiter-family comets are particularly promising candidates for delivering water to Mercury because they have a larger volatile content than asteroids and more favorable orbital and impact characteristics than Halley-type comets.

1. INTRODUCTION

Recent ground-based radar observations of Mercury have revealed the presence of unusually bright radar-reflective regions at the planet’s north and south poles (Slade et al. 1992, Harmon and Slade 1992). The similarity of both the strength and polarization behavior of these radar echoes to the observed radar characteristics of the icy Galilean satellites (e.g., Campbell et al. 1978) and the residual south polar cap of Mars (e.g., Muhleman et al. 1995) prompted the observers to conclude that water ice or some other volatile material might be present at the poles of the otherwise hot planet (Slade et al. 1992, Harmon and Slade 1992, Butler et al. 1993).

High-resolution Arecibo radar maps of Mercury’s polar regions (Harmon et al. 1994) further support the conclusion that volatiles are responsible for the radar signatures—the areas that contain the unusual radar properties are highly correlated with known craters imaged by the Mariner 10 spacecraft. Because the planet’s spin axis is almost exactly perpendicular to its orbital plane (Harmon et al. 1994), the floors of high-latitude craters on Mercury can remain in permanent shadow. Thermal models indicate that temperatures in these permanently shaded regions remain cold enough for water ice to be stable over geological time scales (Paige et al. 1992, Ingersoll et al. 1992, Salvail and Fanale 1994). The observed radar anomalies seem correlated especially with craters that are “pristine” (e.g., have sharp rims and are not highly degraded) and that have a small depth-to-diameter ratio. The floors of smaller bowl-shaped craters receive too much indirect radiation from reflected sunlight and thermal emission from the sunlit crater walls to keep temperatures below the \(~110\) K required to prevent the ice deposits from sublimating over time (Paige et al. 1992, Ingersoll et al. 1992, Salvail and Fanale 1994).
The observed correlation of the radar anomalies with high-latitude craters at both poles has removed any doubt that these features are caused by volatile deposits, but the nature of the deposits is still under debate. Water ice has the required radar properties (see Muhleman et al. 1995) and is the most abundant ice-forming volatile in the Solar System. More volatile ices such as CO₂ or SO₂ would have difficulty remaining stable against thermal evaporation, even in the permanently shaded regions. Less volatile materials such as sulfur (S₈), alkali sulfides, or volatile metals might be stable at many other locations across the planet such as poleward-facing slopes at middle latitudes or high-latitude plains; therefore, these less volatile materials might not be so obviously correlated with high-latitude craters on Mercury (cf. Butler 1997). Sprague et al. (1995) have suggested that the anomalous Mercury radar signatures are due to the presence of elemental sulfur at the poles rather than to water ice. They suggest that meteorite impacts on a sulfide-rich planetary surface could have released atomic sulfur, which could then slowly migrate and be cold-trapped at the poles. A proper evaluation of this claim must await experimental studies of the radar reflectivity and polarization properties of elemental sulfur—in particular, elemental sulfur in any stable or metastable form (particularly S₈ in the orthorhombic-α crystalline form) must be shown to have the same coherent-backscattering properties as water ice.

We will proceed with the assumption that water-ice deposits are responsible for the polar radar features on Mercury. The question of the ultimate source of the water remains open. Butler et al. (1993) discuss some of the possible sources and sinks of water on Mercury; they also examine the mechanics of the migration of water molecules from their source region (wherever that might be on the planet) to the poles (see also Butler 1997). Two broad categories of sources, endogenic and exogenic, need to be considered. For the endogenic sources, water could be directly released from major volcanic episodes or from slower outgassing from the planet’s crust and mantle. In addition, oxygen-containing surface rocks could release water through sputtering by protons from the solar wind or from Mercury’s magnetosphere (Arnold 1979, Potter 1995). Exogenic sources include impacts from interplanetary dust particles (IDPs), comets, and asteroids. None of these sources, with the exception of chemical sputtering from solar-wind or magnetospheric protons, would preferentially occur at high latitudes, and any water released during these processes would have to make its slow way to the poles before being destroyed by various mechanisms.

Outgassing may not be a major source of water for Mercury. The planet’s high density and large inferred Fe/Si ratio suggest that Mercury has lost much of its silicate mantle, perhaps as a result of a giant impact early in its history (Smith 1979, Benz et al. 1988). The crust and mantle of the remnant planet would be highly devolatilized and is unlikely to be able to supply Mercury with the amount of water needed to explain the polar ice; however, this conclusion will remain speculative until we can directly sample material from Mercury.

Exogenic sources also pose difficulties in supplying Mercury with volatiles. Because of Mercury’s proximity to the sun, impact velocities are high, and most of the vaporized debris from an impactor promptly escapes from the planet. In a preliminary study, Rawlins et al. (1995) suggested that although long-period comets are generally traveling too fast to allow much water to be retained upon impact, short-period comets and asteroids will deposit a small fraction of their mass during each impact. Rawlins et al. conclude that over long time scales, numerous impacts of asteroids or short-period comets may supply the necessary amount of water for Mercury’s polar regions, and continual impacts from micrometeorites might also supply Mercury with the requisite amount of water.

Killen et al. (1997) have presented the most detailed discussion of the possible sources and sinks of water on Mercury to date. They conclude that the loss rate due to meteoritic bombardment of exposed ice deposits will dominate over sublimation, sputtering, degassing (e.g., diffusion through overlying layers), and absorption of solar Ly α photons scattered from heliospheric hydrogen atoms. They determine that the steady-state influx of water from meteoroids (both micrometeoroids and asteroids) is approximately equal to the loss rate of exposed ice deposits at the poles due to meteoritic bombardment. Therefore, they conclude that any additional water retained from cometary impacts represents an excess of water that can contribute to the polar ice deposits. Killen et al. (1997) suggest that the nuclei of extinct short-period comets provide the largest source of water to Mercury.

The meteoritic erosion rate quoted by Killen et al. (1997) is an upper limit; the loss rate due to meteoritic impact vaporization can never exceed the meteoritic influx rate unless more target material is vaporized than the incoming projectile mass. Given the small surface area of the polar deposits compared with the planetary surface area, this scenario is unlikely. In their calculations, Killen et al. assume that virtually none (i.e., only 1%) of the water vapor that remains on the planet after impact vaporization of the ice deposits can make it to permanently shaded cold regions, so 99% of this vapor is eventually lost due to photolysis. This survival fraction is considerably less than the ~10% survival rate calculated by Butler (1997) and Butler et al. (1993) for water molecules emplaced anywhere on the planet. Because water released from the polar regions will not migrate on average more than ~25° from the polar regions, the survival rate of vaporized water molecules is likely to be higher than Killen et al.’s 1% (e.g., B. J. Butler, personal communication 1998, estimates that ~40% of the molecules vaporized under these conditions would survive to reach the permanently shaded regions). Thus, micrometeoritic impact will always represent a net source of water to Mercury (albeit a potentially small net source). This conclusion is reinforced if the ice deposits at the poles become covered by a protective dust layer. Killen et al. (1997) have also made some questionable assumptions about the asteroidal and cometary distributions and impact rates (see the discussion in Section 6), and a more detailed analysis of these potential water providers is warranted.
To better examine the statistical likelihood of external sources supplying Mercury with enough water to account for the radar observations, we have developed theoretical models to determine the amount of water retained from cometary, asteroidal, and micrometeoroid impacts on Mercury. By combining these results with models that estimate the amount of water lost during migration to the poles (Butler et al. 1993, Butler 1997) and the loss rate of the ice once it is deposited at the poles (Killen et al. 1997), we can determine whether impacting bodies are likely culprits for delivering water to Mercury.

2. OBSERVATIONAL CONSTRAINTS

Before we begin our discussion of the statistical modeling, we should discuss the constraints that the radar observations help place on the abundance and properties of the current ice deposits and on the possible timing of the deposition.

The fact that the ice is located in high-latitude craters tells us something about the age of the deposits. Because the stability of ice in the polar regions depends on the long-term presence of permanently shaded areas at high latitudes, the ice deposits probably post-date Mercury's tidal evolution that left the planet with a negligible obliquity and stable spin-orbit resonance or commensurability. The very fact that the deposits lie within craters also suggests that the deposition occurred sometime after most of the large craters were emplaced, i.e., during the later stages or after the late heavy bombardment of the inner Solar System. Note that large impacts can erode ice deposits as well as supply volatiles.

One observational constraint imposed by the radar observations (Slade et al. 1994, Harmon and Slade 1992, Butler et al. 1993, Harmon et al. 1994) arises from the fact that the ratio of same-sense-to-opposite-sense circular polarization is greater than unity for the anomalous polar radar-reflective regions; this fact combined with the derived high radar reflectivities indicates that the radar signal is being multiply scattered through a relatively loss-less medium several radar wavelengths thick. Therefore, Butler et al. (1993) and Harmon et al. (1994) emphasize that the ice deposits must be relatively clean (i.e., free of substantial amounts of dust or other radar-absorbing materials) and must be at least a meter or two thick. The thickness constrains the amount of ice currently on the planet while the cleanliness may indicate that the ice was deposited quickly.

Butler et al. (1993) also demonstrate that the radar signatures cannot rule out ice deposits that are covered by a thin regolith or dust layer. They suggest that a porous dust layer up to a meter thick could still be consistent with the observations; however, the higher-resolution data of Harmon et al. (1994) will require tighter constraints on the thickness of any layer of covering dust.

The areal coverage of the ice currently at Mercury’s poles can be estimated from the high-resolution radar observations of Harmon et al. (1994). By adding up all the major anomalous radar spots in the images of Harmon et al. (1994), including those spots on the hemisphere that were not imaged by Mariner 10, we estimate that the current ice deposits at both poles cover a \(3 \pm 1 \times 10^{14} \text{ cm}^2\) area. If we further assume that the ice deposits have an average thickness of somewhere between 2 and 20 m and a bulk density of 1 g cm\(^{-3}\), then a rough estimate for the amount of ice currently located at both poles of Mercury is \(4 \times 10^{16} \text{ to } 8 \times 10^{17} \text{ g (or } 40-800 \text{ km}^3)\). Note that the radar observations themselves do not allow us to estimate an upper limit to the ice thickness, and the 20-m maximum value was chosen arbitrarily. The heights of the crater rims and the angle of the shadows they cast could help us impose constraints on the maximum ice thickness, but no information on crater morphology on the unimaged hemisphere currently exists.

Any potential source of water on Mercury must be able to supply an amount much greater than is currently “observed.” Unless the water is catastrophically deposited in such a large abundance that the planet temporarily has a collisionally thick atmosphere—a likely possibility for large impacts—the individual water molecules will migrate by a series of random hops until they reach regions in which they are thermally stable. Each time they hop, they have a chance of being lost from the planet. Using a simple model of thermally accommodated adsorbed water molecules that are released from the surface at the RMS velocity of a Maxwellian gas at the surface temperature and hop ballistically, Butler et al. (1993) determine that only \(10\%\) of the water molecules will survive to reach the cold polar regions before being dissociated by solar ultraviolet radiation. This value is also confirmed by more sophisticated three-dimensional modeling: Butler (1997) concludes that 5–15% of the water molecules emplaced randomly across the planet will survive to reach permanently shaded regions at the poles.

We should point out (as also noted in Butler 1997), however, that photolysis of a water molecule is not sufficient to remove it from Mercury, and this survival estimate is a lower limit. Although the hydrogen atoms released from water photolysis can move fast enough to escape Mercury’s gravitational field, OH radicals or oxygen atoms generally do not. For instance, a solar H Ly \(\alpha\) photon of 1216-Å wavelength (energy 10.2 eV) can dissociate a water molecule (H-OH bond energy 5.12 eV) leaving 5.08 eV of excess energy to be partitioned between the translational energies of the products and the rotational and vibrational energies of OH. For the translational energies, conservation of momentum and energy suggest that most of the excess energy will be partitioned to the translational motion of the H atoms (maximum possible OH translational energy of 5.08/18 = 0.28 eV), preventing the OH radicals from gaining the 1.59 eV they need to escape Mercury. The heavier photolysis products have a better chance of hitting Mercury’s surface, where they might be thermally accommodated and released or might adsorb or chemically attach to the surface rocks and remain until sputtering events or meteoritic impacts may reintroduce hydrogen and reform the water molecules. Alternatively, the O or OH could perform their own random walk to high latitudes where interaction with solar-wind protons could allow \(\text{H}_2\text{O}\) molecules to reform. In either case, the oxygen would not be permanently removed from the planet. If the water molecule...
is ionized, the charged products have a chance of being swept away by the solar wind, but even in that situation, the planet’s magnetic field can act as a shield and prevent loss before recombination can occur. Therefore, a survival rate higher than 5–15% might be indicated.

Once at the poles, several mechanisms act to erode the ice. The rate of erosion is difficult to calculate because it depends on unknown properties of the ice deposits such as their effective temperature or the presence of a protective dust covering. Killen et al. (1997) find that sublimation of fresh, exposed deposits will dominate the loss rate provided that the exposed ice deposits are above ~75 K for amorphous ice or ~100 K for cubic ice (according to their Fig. 1). Deposits of amorphous ice thicker than a few millimeters may not be stable under their own weight, causing any amorphous ice to transform to cubic ice (e.g., Butler et al. 1993). For the <110 K average temperatures predicted for the permanently shaded regions at high latitudes on Mercury (Salvail and Fanale 1994, Ingersoll et al. 1992, Paige et al. 1992, Butler et al. 1993), the results of Killen et al. (1997) indicate that meteoritic bombardment rather than sublimation will dominate the erosion rate of surface ice deposits. Once the ice becomes covered with fine layers of dust, it can be protected from sputtering, micrometeorite impacts, and sublimation, and Killen et al. (1997) find that “degassing” (e.g., diffusion through and sublimation at the surface) of the dust layer controls the loss and operates at a much reduced rate.

For lack of more definitive information, we will use the Butler (1997) estimate of the survival rate (5–15%) for water molecules to reach the permanently shaded regions at high latitudes. The results can be easily scaled to any desired survival fraction. Once the molecules condense at the poles, we assume that the ice deposits are readily covered by thin dust layers generated from micrometeorite bombardment or ejecta from large impacts, and further losses are held to a minimum. Thus, in order to be considered viable sources of water for Mercury and to counteract the losses due to photolysis or ionization during migration, external sources must supply an amount of water that is a factor of 7–20 times larger than the current estimates of the amount of ice at Mercury’s poles. In other words, comets, asteroids, and IDPs must bring in approximately (8–50) × 10^{17} g of water over the past 3.5 billion years or so (since the approximate end of the late heavy bombardment period; see Table I).

### 3. EXOGENIC DELIVERY OF WATER TO MERCURY

Impacts by comets, asteroids, and interplanetary dust particles provide the three main external sources of water for Mercury. Small dust particles with diameters less than 1 cm are continuously raining down on the Earth and other planets. These micrometeoroids represent an important but currently not well-determined fraction of the total mass influx to the Earth (cf. Grün et al. 1985, Love and Brownlee 1993, and Ceplecha 1996). If we assume that the current influx of IDPs has remained constant for the past ~3.5 billion years and extrapolate the current terrestrial influx rate to that at Mercury, we can estimate the amount of micrometeoroidal material that has impacted Mercury over that time frame (see Cintala 1992, Morgan et al. 1988). Further assumptions about the amount of hydrated minerals in IDPs and the fraction of the vaporized dust particle that is retained during an impact can let us estimate the amount of water that has been delivered to Mercury by IDPs.

Comets and asteroids, although infrequent impactors, are more massive, and a small number of favorable impacts could deliver a significant amount of water. The amount of water retained during an impact depends not only on the size of the impactor, but also (critically) on the impact velocity. The impact velocity, in turn, depends on the orbital parameters of the potential impactor. The problem of water delivery by comets and asteroids is most easily handled by statistical methods. We will use a Monte Carlo simulation to generate fictitious comets and asteroids with masses and orbital elements selected randomly from appropriate distributions. For each object whose orbit intersects that of Mercury, the collision probability, impact velocity, and retainable vapor fraction is calculated. After integrating over the age of the Solar System, we can determine how much water would have been supplied to our fictitious Mercury, and by running the calculation numerous times, we can generate many potential Mercury’s to statistically compare our results with the “observed” amount of ice at Mercury’s poles.

### 3.1. Interplanetary Dust Particles

Estimates of the micrometeoritic impact flux on Mercury have been presented by Cintala (1992) and Morgan et al. (1988). Their technique, which was developed in part by Zook (1975), is used here to extrapolate the current terrestrial influx rate to that at Mercury. First, we determine the micrometeorite mass flux at Earth using the most recent direct measurements of the terrestrial influx of IDPs as determined by the Long Duration Exposure Facility (LDEF) satellite (Love and Brownlee 1993). The recent LDEF results indicate that the terrestrial influx is 2–3 times greater than previous estimates (cf. Love and Brownlee 1993,

### TABLE I

<table>
<thead>
<tr>
<th></th>
<th>Estimated mass of water delivered (10^{17} g)</th>
<th>Estimated ice mass (10^{16} g)</th>
<th>Estimated average ice thickness (m)</th>
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<tr>
<td>Observed</td>
<td>8–50</td>
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<td>2–20</td>
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<td>0.8–20</td>
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<tr>
<td>Delivered by asteroids</td>
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<td>Delivered by Jupiter-family comets</td>
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<td>Delivered by Halley-type comets</td>
<td>0.4–10</td>
<td>0.2–20</td>
<td>0.07–7</td>
</tr>
</tbody>
</table>

*Note.* The values do not include any estimates of the erosion rate of ice deposits at the poles.
Grün et al. 1985). Then, we determine the velocity distribution of meteoroids entering the Earth’s atmosphere. This distribution is not well known. We use the information presented by Southworth and Sekanina (1973), who derived an “unbiased” velocity distribution from radar observations of over 14,000 meteors, to obtain the terrestrial velocity distribution. Finally, following the technique of Cintala (1992) and Morgan et al. (1988), we transform the terrestrial flux to that at Mercury by (1) gravitationally “defocusing” the observed atmospheric velocity distribution to get the IDP flux at 1 AU but away from the Earth’s gravitational influence, (2) extrapolating the IDP flux to Mercury’s distance from the Sun (assuming certain properties of the meteoroid orbits and the spatial density of meteoroids with heliocentric distance), and then (3) gravitationally focusing this population onto Mercury. Details of these calculations are presented in Appendix A and Fig. 1. We find that the current IDP flux at Mercury is roughly \(4 \pm 3 \times 10^{-16} \text{ g cm}^{-2} \text{ s}^{-1}\) (a factor of \(\sim 2\) greater than the current terrestrial IDP flux), corresponding to a total mass influx to Mercury’s surface of roughly \(1 \pm 0.8 \times 10^{10} \text{ g yr}^{-1}\).

Dust particles that impact Mercury with high velocities will be vaporized upon impact (along with a certain amount of surface material). A significant fraction of the molecules within the impact-derived vapor cloud will be released at velocities that exceed Mercury’s escape velocity. Details of impact vaporization are poorly understood. The internal energy of the vapor cloud is a complicated and currently unknown function of impact velocity, impact angle, and composition of both the impactor and target material. In Appendix B, we discuss our strategy for determining the fraction of vapor that is retained during an impact and compare our formalism with other theoretical methods and laboratory experiments. When we apply the impact vaporization calculations discussed in Appendix B to the problem of IDP impacts on Mercury, we find that approximately 63% of the IDP mass will remain on the planet after an impact—100% of the particles impacting with velocities less than \(~15 \text{ km s}^{-1}\) will be retained, but the fraction retained decreases rapidly with increasing impact velocity (see Figs. 1 & B2).

Roughly half of the IDPs collected in the terrestrial atmosphere are anhydrous and half contain \(\sim 20\%\) water, mostly in the form of hydrated minerals (M. E. Zolensky, personal communication, 1995). Therefore, we assume that IDPs contain a mass fraction of water of roughly 10%. Combining this information with the estimates of the amount of vapor retained during an impact, we determine that Mercury might have collected \(2.2 \pm 1.7 \times 10^{18} \text{ g}\) of IDP-derived water over the past 3.5 billion years (since the approximate end of the late heavy bombardment period). Note that this range of values only takes into account uncertainties in the terrestrial IDP flux, in the extrapolation of the IDP spatial density to Mercury’s orbital distance, and in uncertainties in the velocity distribution at Earth, but does not consider such effects as possible fluctuations in the meteoritic flux with time or the uncertainties in the eccentricities of dust particles near Mercury’s orbit.

Only 5–15% of the water delivered from IDPs will survive migration to the poles, so IDPs can supply \((0.3–6) \times 10^{17} \text{ g}\) of water to the high-latitude permanently shaded regions, corresponding to an ice layer of average thickness \(~0.8\) to 20 m (see Table I). Killen et al. (1997) have derived a similar result.

**FIG. 1.** The IDP velocity distribution at the Earth and Mercury. The distribution at the Earth is represented by the dotted line and is normalized to unity. The dashed line represents the IDP velocity distribution at Mercury before vaporization of the impactors. The solid line shows the fraction of the vaporized debris that is retained by the planet after impact.
3.2. Asteroids and Comets

Asteroids and comets require a different approach. These massive objects can carry a great deal of water (e.g., a large comet of radius 14 km, density 1.0 g cm$^{-3}$, and 50% water ice by mass contains the required amount of water); however, they impact infrequently and sporadically. The impactor flux depends on the probability that an object on a Mercury-crossing orbit will collide with the planet. A few chance impacts of particularly large objects or particularly favorable geometries could contribute a substantial amount of water. On the other hand, comets and asteroids tend to reside in eccentric orbits, causing impact velocities to be high and vapor retention to be low. Stochastic methods are appropriate.

To examine the problem of water delivery to Mercury by larger objects, we have developed a Monte Carlo model to simulate the impact history of Mercury over the past 3.5 billion years. In the program, fictitious objects are generated with orbital parameters and masses chosen at random from appropriate distributions. For each fictitious object whose orbit intersects that of Mercury, we calculate the object’s collision probability and impact velocity using a method developed by Kessler (1981) and outlined further in Steel and Baggaley (1985). With this technique, the “spatial density” of an orbiting object, or the probability of finding the object in any particular volume of space, can be calculated for any small volume. The collision probability then becomes the probability that an asteroid or comet will occupy the same volume of space at the same time as Mercury. We use the approximate (average) form given by Kessler (1981) for the spatial density of an object in a small finite volume element and typically divide our integration range into 100 volume elements. Note that unlike collision probability estimates based on the Öpik algorithm (Öpik 1951), the Kessler method treats Mercury’s eccentric orbit properly. The collision probabilities calculated for known objects that have been observed in Mercury-crossing orbits are shown in Tables II and III. These values compare well with other impact probability calculations (e.g., Steel and Baggaley 1985, Olsson-Steel 1987).

Given a threshold diameter (lower limit) for our impacting objects, we calculate the number of Earth-crossing objects currently thought to exist in the Solar System with this diameter or larger (up to some maximum diameter), using appropriate
The primary source of the near-Earth asteroids is believed to be the main asteroid belt between the orbits of Jupiter and Mars. Collisions between main-belt asteroids can produce debris fragments that might evolve into orbits near a secular or jovian mean-motion resonance zone; subsequent chaotic orbital evolution can send these asteroid fragments into the inner Solar System (Wetherill 1988). Another potential source of the near-Earth asteroids is extinct or dormant nuclei of short-period comets that were also transported to the inner Solar System through interactions with Jupiter. Because extinct comets do not by definition display any standard cometary activity, they are hard to distinguish from asteroids. Binzel et al. (1992) have surveyed the physical properties of the near-Earth asteroid population and conclude that extinct or dormant comets can account for only 0–40% of the observed population. On the other hand, calculations of the frequency of collision and fragmentation of main-belt asteroids by Bottke et al. (1994) indicate that the main belt alone cannot supply the observed population of near-Earth asteroids, but that a cometary or other source is required. In any case, some fraction of what we are describing as asteroids may be extinct cometary nuclei that could contain a

<table>
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<th>Name</th>
<th>Type</th>
<th>q (AU)</th>
<th>e</th>
<th>i (°)</th>
<th>$P_e$ (yr$^{-1}$)</th>
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<tr>
<td>Encke</td>
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Note. Orbital parameters are from Marsden and Williams (1995). Htc = Halley-type comet, Jfc = Jupiter-family comet, Etc = Encke-type comet.
larger fraction of volatile material than is generally associated with asteroids.

The distributions of orbital parameters and sizes used in our asteroid calculation are shown in Fig. 2. The inclination distribution is taken from the Earth-crossing asteroid population of Rabinowitz (1993), whose data include recent discoveries from the Spacewatch Telescope. Rabinowitz has attempted to reverse the process of observational bias to find theoretical "unbiased" distributions for orbital elements of Earth-crossing asteroids with diameters 1 km or larger. We adopt his unbiased inclination distribution throughout this work; however, we use actual histogram distributions from a catalogue of the ~150 known potentially Earth-crossing asteroids discovered before 1994 for the other orbital parameters (e.g., \(q\) and \(Q\) or \(e\)). For asteroids discovered before June 1991, our values for the aphelion distance \(Q\) and perihelion distance \(q\) at the time of Earth orbital crossing are taken from the mean orbital elements listed in Table I of Rabinowitz et al. (1994); for asteroids discovered between June 1991 and July 1993, we use the orbital parameters for the epochs listed in the appendix of Rabinowitz et al. (1994).

\(Q\) and \(q\) appear to be mostly uncorrelated in the observed population of Earth-crossing asteroids. Aphelion distances generally do not exceed 5 AU (i.e., only one Jupiter-crossing object exists in this population) or fall below 1 AU (e.g., there are no known asteroids with orbits completely within that of the Earth). The former result probably reflects the short dynamical lifetime of Jupiter-crossing bodies while the latter may be (at least partially) an observational bias or may be a real dynamical constraint (e.g., Gladman et al. 1997). Clearly absent are objects with both low \(e\) and low \(q\). Therefore, Mercury-crossing asteroids will tend to have eccentricities higher on average than those of Earth-crossing asteroids, and we consider the aphelion distance \(Q\) to be a more reliable parameter with which to work than the eccentricity \(e\). For simplicity, we will assume that \(q\) and \(Q\) are independent (uncorrelated), and we allow each to vary across the range of observed values; the probabilities that an object is found within a certain size or parameter bin within that range are shown in Fig. 2.

The asteroid size distribution is taken from Rabinowitz et al. (1994) and incorporates population estimates by Shoemaker et al. (1990) as well as recent Spacewatch results. Rabinowitz et al. (1994) predict that there are a total of 135,000 Earth-crossing asteroids with diameters larger than 0.1 km and 1500 with diameters larger than 1 km currently in the Solar System. These numbers and the corresponding size distribution are uncertain (especially for smaller objects) because they include estimates of the numbers of undiscovered asteroids for each size bin. For our minimum asteroid diameter of 0.5 km, the predicted number of Earth-crossing asteroids (5800) is uncertain by a factor of 2 or so (see Fig. 1 of Rabinowitz et al. 1994). Note also that the slope of the size distribution adopted by Rabinowitz et al. for the largest currently known Earth-crossing objects may not accurately reflect the steady-state size distribution. For instance, Rabinowitz et al. use a power-law slope of \(-5.4\) for asteroids with diameters greater than 3.5 km while the power-law slope of the main-belt asteroids (e.g., Jedicke and Metcalfe 1998) is much shallower (i.e., \(\sim -3\)). Because the catastrophic asteroid collisions that would send very large collisional fragments into Earth-crossing orbits are rare, the likelihood of seeing very large Earth-crossing asteroids at any particular point in time is low.

Such large Earth-crossing objects do exist, however, and will be an important component of the delivery of mass to the terrestrial planets. The Rabinowitz et al. distribution may underestimate the steady-state population of these very large asteroids on million-year time scales. For instance, orbital simulations of the Mars-crossing asteroid Eros performed by one of the authors (L.D.) indicate that Eros (diameter \(\sim 22\) km) is Earth-crossing \(-10\%\) of the time for the next 10 million years. Thus, Eros by itself represents 0.1 Earth-crossing asteroid greater than 22 km diameter; the distribution of Rabinowitz et al. (1994) provides for only \(8 \times 10^{-3}\) such objects. To prevent the underestimation of large potentially Mercury-crossing asteroids, we have adopted a shallower slope of \(-3.5\) for asteroids larger than \(-3.4\) km (see Fig. 2).

### 3.2.2. Comet parameter distributions.

Short-period comets have historically been defined as comets with orbital periods less than 200 years. Within this population are two distinct groups of comets, Halley-type comets and ecliptic comets. Following Levison and Duncan (1997) and others, we will use the Tisserand parameter with respect to Jupiter,

\[
T = \frac{a_J}{a} + 2 \cos i \sqrt{(1-e^2) \frac{a}{a_J}},
\]

to distinguish between the two populations (where \(a_J\) and \(a\) are the semimajor axes of Jupiter and the comet, respectively). Ecliptic comets, i.e., those with orbital parameters leading to values of \(T > 2\), are generally believed to have evolved from the Kuiper belt. Halley-type comets, with \(T < 2\) and a more nearly isotropic distribution of inclinations, are believed to derive in large part from the Oort cloud. As Levison and Duncan (1997) emphasize, further subdivisions within the ecliptic-comet population are often useful. For example, ecliptic comets with \(2 < T < 3\) are dynamically dominated by Jupiter and are termed Jupiter-family comets, while comets with \(T > 3\) and \(a < a_J\) cannot cross Jupiter’s orbit and are called Encke-type comets.

Only \(-28\) short-period comets have ever been observed in Earth-crossing orbits (Marsden and Williams 1995; statistics are for discoveries before 1995). Of these, 15 are Halley-type comets, 11 are Jupiter-family comets, and 2 (2P/Encke and 107P/Wilson-Harrington) are Encke-type comets. Because of the large differences in their orbital properties, we separate the ecliptic comets from the Halley-type comets in this paper; however, due to the small number of Encke-type comets, we include these comets within the Jupiter-family population.

The orbital parameter and size distributions adopted for our comet calculations are shown in Fig. 2. The cometary size distributions are particularly uncertain. Our adopted size distributions
FIG. 2. Orbital parameter and size distributions for the Earth-crossing populations; some of these parameters are used as input to the Monte Carlo code (see text): (a) cumulative number of objects with diameters greater than diameter $D$, (b)–(f) probability that the object’s inclination, perihelion distance, aphelion distance, eccentricity, and Tisserand parameter falls within a particular histogram bin. See text for an explanation of the sources of these distributions. Asteroid distributions are shown by a thin solid line, Halley-type comet distributions by a thick solid line, and Jupiter-family comet distributions by a dotted line. Following Shoemaker et al. (1990) and Rabinowitz et al. (1994), we have assumed a bulk density of $1.2 \text{ g s}^{-1}$ for the comets and $3.5 \text{ g s}^{-1}$ for the asteroids.
are primarily from the population estimates of Shoemaker et al. (1994). Because they believe that the discovery of short-period Earth-crossing comets larger than 1 km is not complete, Shoemaker et al. (1994) use observational selection effect arguments to estimate the current number of Earth-crossing comets. The word “current” here apparently means within the past two centuries because they include in their list several comets that have disappeared from view and are assumed lost. Shoemaker et al. (1994) estimate that \( \sim 40 \pm 2 \) active (i.e., displaying cometary activity) Earth-crossing short-period Jupiter-family comets with nucleus diameters larger than 1 km currently exist in the Solar System. Since such comets may spend a large portion of their lifetime in orbits that would preclude visible activity (i.e., orbits with a relatively large perihelion distance) and since dormant or extinct comets are much more difficult to observe, the total number of active plus extinct comets is presumably much greater.

Shoemaker et al. (1994) believe the extinct/active ratio for Jupiter-family comets is \( \sim 18 \) based on arguments of the active physical lifetime versus dynamical lifetime of short-period comets and on statistical arguments of the rate of discovery (and biases in the discovery) of extinct short-period comets. Theoretical investigations of Kuiper-belt objects (Levison and Duncan 1997) suggest that the ratio of extinct/active Jupiter-family comets could be much smaller (i.e., between 2.0 and 6.7, with a favored value of 3.5). For our estimate of the number of Earth-crossing Jupiter-family comets with diameters greater than 1 km currently residing in the Solar System, we use the lower limit of 20 active comets from Shoemaker et al. (1994) along with their extinct/active ratio of 18 to obtain a nominal value of 380 of these comets. For our estimated lower limit to the number of Earth-crossing Jupiter-family comets with \( D \geq 1 \) km, we assume 20 active comets and an extinct/active ratio of 3.5 to get 90 such comets, and our upper limit of 1140 is obtained by assuming that 60 active comets of that type exist along with an extinct/active ratio of 18.

Earth-crossing Halley-family comets should be even more numerous; Shoemaker et al. (1994) estimate that 140–270 active and 1400–5400 extinct Earth-crossing Halley-family comets with diameters larger than 1 km may inhabit the Solar System. Therefore, we adopt a value of \( 3280^{+1740}_{-2290} \) such comets in our simulations. We use these population estimates combined with the derived slope of the cumulative size distribution of observed short-period comets (–1.97 according to Shoemaker and Wolfe 1982) to obtain the size distribution shown in Fig. 2.

The cometary inclination distributions shown in Fig. 2 are taken from tables of the 159 known Jupiter-family comets and 25 known Halley-type comets listed in the catalog of Marsden and Williams (1995). It should be noted that observational biases are probably inherent within this population. Shoemaker et al. (1994) estimate that the unbiased mean inclination of extinct Jupiter-family comets could be as high as 32° rather than the observed 19° for active Earth-crossing Jupiter-family comets. The perihelion distribution for the Jupiter-family comets is taken from Fig. 3 of Fernández (1984) and represents the computed perihelion distances of a theoretical ensemble of short-period comets under gravitational control of Jupiter. Note that small perihelion distances have a low probability. The perihelion distribution of the Halley-type comets is assumed to be uniform in \( q \) (based on orbital integrations by Dones and Dassanayake 1997).

The distribution of aphelion distances for the Jupiter-family comets is taken from known comets listed in Marsden and Williams (1995), but we consider only those 139 Jupiter-family comets with \( a < a_1 \), because the Halley-type comets possess a wide range of aphelion distances, we use an eccentricity distribution instead; the \( e \) distribution shown in Fig. 2 is taken from the 15 known Earth-crossing Halley-type comets from the catalog of Marsden and Williams (1995). However, if the Tisserand parameter \( T \) is greater than 2 after selecting \( i, q, \) and \( e \) in our Halley-type comet simulation, we reject those comets and continue to select an eccentricity until \( T < 2 \). Thus, our population of Mercury-crossing Halley-type comets will have an average eccentricity higher than that of the Earth-crossing Halley types, and we can avoid unrealistic orbits for these objects.

As Fig. 2 indicates, both types of comets tend to have higher eccentricities than asteroids, leading to generally higher impact velocities and less vapor retained for cometary impacts than for asteroidal impacts. However, more large comets with diameters greater than 10 km may exist on orbits that take them into the inner Solar System than large asteroids; this fact combined with the greater volatile content of comets suggests that comets may be more important than asteroids in supplying Mercury with water.

4. RESULTS

Once appropriate parameter distributions are defined (see Fig. 2 and discussion above), we then use a Monte Carlo code to create a population of asteroids or comets with masses and orbital parameters representative of those distributions. McGrath and Irving (1975) provide an algorithm for generating random numbers from a probability distribution in histogram form. Using this algorithm along with a random number generator, we create \( \sim 5800 \) Earth-crossing asteroids with diameters between 0.5 and 10 km (i.e., the current number of Earth-crossing asteroids believed to be within this size range). Within this population, we find that roughly 13% of the Earth-crossing asteroids also cross Mercury’s orbit. Figure 3 illustrates the accuracy of the Monte Carlo code with respect to reproducing the desired asteroid parameter distributions.

For each of the Mercury-crossing asteroids generated by the model, we calculate its collision probability with Mercury, its average impact velocity, the fraction of its mass that is retained by Mercury after the impact, and the amount of water (g yr\(^{-1}\)) potentially supplied to Mercury from such an object (see Eq. (1)). The total amount of water supplied to Mercury from each object of that type since the end of the late heavy bombardment
period is determined by multiplying the latter result by 3.5 byr. Because the ~800 Mercury-crossing asteroids generated by the Monte Carlo code are an insufficient statistical sample, we perform 300 different iterations to generate 300 potential impact histories for Mercury. The average properties of our computer-generated asteroid population are shown in Table IV. Note that we have assumed that 5% of the asteroid mass can be converted to water vapor after the impact. This value takes into account the estimate that 0–40% of the Earth-crossing asteroids may actually be extinct short-period comets (Binzel et al. 1992).

The results with regard to the amount of water supplied over the past 3.5 byr are shown in Figs. 4a and 4b. Each Mercury symbol in Fig. 4a represents one of the 300 potential Mercury’s. Although the number of Mercury-crossing asteroids varies by only a small percentage between each iteration (run) of the model, the total amount of mass incident on Mercury covers a much wider range of values. This outcome arises from the power-law shape of the size distribution for the asteroids (see Figs. 2 and 3).

<table>
<thead>
<tr>
<th>Property</th>
<th>Asteroids</th>
<th>Jupiter-family comets</th>
<th>Halley-family comets</th>
</tr>
</thead>
<tbody>
<tr>
<td>Inclination (deg)</td>
<td>26</td>
<td>12</td>
<td>61</td>
</tr>
<tr>
<td>Eccentricity</td>
<td>0.75</td>
<td>0.90</td>
<td>0.97</td>
</tr>
<tr>
<td>Perihelion distance (AU)</td>
<td>0.34</td>
<td>0.29</td>
<td>0.24</td>
</tr>
<tr>
<td>Mean collision probability (yr⁻¹)</td>
<td>1.6 × 10⁻⁹</td>
<td>7.8 × 10⁻¹⁰</td>
<td>1.1 × 10⁻¹⁰</td>
</tr>
<tr>
<td>Median collision probability (yr⁻¹)</td>
<td>1.1 × 10⁻⁹</td>
<td>6.7 × 10⁻¹⁰</td>
<td>6.8 × 10⁻¹¹</td>
</tr>
<tr>
<td>Impact velocity (km s⁻¹)</td>
<td>37</td>
<td>39</td>
<td>65</td>
</tr>
<tr>
<td>Fraction of vapor retained</td>
<td>10%</td>
<td>7.1%</td>
<td>1.5%</td>
</tr>
<tr>
<td>Assumed water content</td>
<td>5%</td>
<td>50%</td>
<td>50%</td>
</tr>
<tr>
<td>Number of Mercury-crossers</td>
<td>780</td>
<td>390</td>
<td>5800</td>
</tr>
<tr>
<td>Mass of water delivered (g)</td>
<td>2.9 × 10¹⁷</td>
<td>5.0 × 10¹⁷</td>
<td>2.1 × 10¹⁷</td>
</tr>
</tbody>
</table>

* Number with diameter greater than 0.5 km.

* Average mass of water retained by the planet after 3.5 billion years of impacts; this value does not include loss during migration to the poles or erosion/sublimation at the poles.
and from the fact that a handful of large objects can control the amount of water supplied to the planet (see Tremaine and Dones (1993) for an analytic confirmation of the dominance of high-mass objects for cumulative size distributions with indices > 3). For the 300 different iterations, the average amount of water delivered to Mercury from asteroid collisions over 3.5 byr is $2.9 \times 10^{17}$ g (median $2.5 \times 10^{17}$ g). Only 5–15% of this water survives the migration to the permanently shaded regions at high latitudes (Butler 1997). Considering the entire range in the Monte Carlo simulations, asteroids can supply $\sim (0.4–20) \times 10^{16}$ g of ice to Mercury’s polar regions, corresponding to ice deposits of average thickness $\sim 0.1$ to 8 m (see Table I). Note that the range of uncertainty here only includes the statistical variation due to different orbital and impact properties of the adopted impactor population; it does not include the uncertainty in the amount of hydrated minerals or other volatile water-generating material within the asteroids, the uncertainty in the slope of the size distribution of large asteroids, or the uncertainty in the vapor-retention calculations.

If the population estimates of Shoemaker et al. (1994) are correct, short-period comets appear to be even more promising candidates for supplying Mercury with water. For the cometary impact calculations, we generate $\sim 1500$ Earth-crossing Jupiter-family comets with diameters between 0.5 and 50 km (i.e., the current number of Earth-crossing Jupiter-family comets believed to be within this size range) and with orbital parameter distributions that are constrained by those shown in Fig. 2. Roughly 26% of these computer-generated comets cross Mercury’s orbit (compare this percentage with 2 of the 13 discovered Earth-crossing Encke-type and Jupiter-family comets also have or had Mercury-crossing orbits). We also generate $\sim 13,000$ Earth-crossing Halley-type comets with diameters between 0.5 and 50 km; $\sim 45$% of these comets are Mercury crossers (compared with the 3 out of 15 discovered Earth-crossing Halley-type comets also being Mercury crossers).

For those comets that are Mercury-crossers, we calculate the collision probability with Mercury, the impact velocity, the fraction of cometary mass retained by Mercury after the impact, the potential amount of water supplied to Mercury by the comet, and ultimately the amount of water delivered to Mercury by short-period comets over the past 3.5 byr. The average properties of the cometary populations are shown in Table IV. Figures 5 and 6 show how the computer-generated parameters for the Mercury-crossing comets compare with the input parameters. The final results for 300 different iterations of the Monte Carlo code are shown in Figs. 7 and 8.

Each Mercury symbol in Figs. 7a and 8a represents one 3.5-byr history of cometary impacts onto a fictitious Mercury. Figures 7b and 8b represent similar results in histogram form. We find that the average amount of water delivered to Mercury by Jupiter-family (and Encke-type) cometary impacts for our 300 different iterations is $5.0 \times 10^{17}$ g with a median value of $1.9 \times 10^{17}$ g. Considering just the statistical variation in the Monte Carlo simulation, Jupiter-family (and Encke-type) comets bring in $(0.3–100) \times 10^{17}$ g of water to Mercury. If we also consider the fact that the Shoemaker et al. (1994) size distribution (adopted in our nominal model) might overestimate the number of large extinct comets in current Earth-crossing orbits, then a lower limit of $7 \times 10^{15}$ g might be more appropriate. After considering loss due to migration, the polar regions of Mercury could contain $(0.2–200) \times 10^{16}$ g of comet-derived water, or an ice layer of average thickness 0.05–60 m (see Table I).

The average amount of water delivered to Mercury by Halley-type comets over the past 3.5 byr in our simulations is $2.1 \times 10^{17}$ g with a median value of $1.6 \times 10^{17}$ g. Considering the entire range of the simulations, Halley-type comets could have provided $(0.2–20) \times 10^{16}$ g of ice for Mercury’s polar regions, corresponding to an average ice thickness of 0.07–7 m. Although Halley-type comets with Mercury-crossing orbits are much more numerous in the Solar System than Jupiter-family comets, they tend to have higher eccentricities, correspondingly higher impact velocities, longer periods, and lower impact probabilities than Jupiter-family comets, causing less water to be retained on Mercury over long time scales from Halley-type comets as opposed to Jupiter-family comets (see Table IV).
FIG. 5. The results of one run of the Jupiter-family comet Monte Carlo simulations compared with the size and orbital parameter distributions for Earth-crossing Jupiter-family comets (see Fig. 2): (a) diameter, (b) inclination, (c) perihelion distance, (d) aphelion distance, (e) eccentricity, and (f) Tisserand parameter. The input distributions are shown by a thick solid line and the output distributions by thin histogram bins. As with the asteroids, the eccentricities of the Mercury-crossers are in general higher than those of the Earth-crossers. The fact that the Tisserand parameters of our simulated Jupiter-family comet population are in general lower than those of the Earth-crossing population reflects these higher eccentricities. Note that Encke-type comets are rare in our simulation.

5. DISCUSSION

The details of the results for one iteration of the asteroid and comet calculations are shown in Fig. 9. The objects that deliver most of the water in each iteration are indicated by stars rather than circles in the plots. Figure 9a illustrates the correlation between the mass of the impactor and the amount of water delivered during the impact. Note that for both the asteroid and
comet calculations, most of the water delivered to the fictitious Mercury’s in our Monte Carlo calculations derives from a small number of large objects. This result is consistent with the observed cleanliness of the ice deposits on Mercury (Harmon et al. 1994, Butler et al. 1993). Similarly, Fig. 9b illustrates the correlation between the collision probability and the amount of water delivered. Unfavorable orbital parameters produce unfavorable collision probabilities and correlate with such values as the impact velocity to inhibit the amount of water delivered during these impacts. The objects that deliver most of the water to Mercury have average-to-high collision probabilities, and average-to-low impact velocities, inclinations, and eccentricities. Perihelion distance seems to be less of a factor.

The single most important parameter for determining the amount of water delivered during any particular impact appears to be the mass of the impactor. Therefore, the assumed slope of the size distribution at large sizes is critical to our calculations. Unfortunately, due to the statistics of small numbers, the steady-state slope for Earth-crossing asteroids and comets is difficult to determine. The second most important parameter is the collision probability, which depends on the orbital parameters of the asteroid or comet population.

Our method for determining the average (and perhaps more importantly, the median) collision probability for the Mercury-crossing asteroid and comet population should be accurate to within a factor of ~2. However, because we assume that aphelion and perihelion distances are uncorrelated for the asteroids and Jupiter-family comets, we may miss a segment of the population that produces higher-than-average collision probabilities. For instance, our simulations rarely produce Encke-type comets (i.e., those with Tisserand parameters greater than 3): a typical iteration of the Monte-Carlo code will produce 1–2 Encke-type comets out of 300 Mercury-crossing Jupiter-family comets. Compare this percentage with one comet out of five known Mercury-crossers being an Encke-type comet, or one out of two known Mercury crosses with $T > 2$ being an Encke-type comet (see Table III). It appears that we may underestimate the number of Encke-type comets in our simulation compared with reality. Because Encke-type comets have higher-than-average collision probabilities with Mercury, it is important to know how common Mercury-crossing Encke-type comets are in the Solar System. Standard numerical simulations of Kuiper-belt objects do not produce Encke-type comets, perhaps because nongravitational forces (e.g., jetting), close encounters with terrestrial planets, or

**FIG. 6.** Same as described in the legend to Fig. 5 except for Halley-type comets: (a) diameter, (b) inclination, (c) eccentricity, and (d) perihelion distance.
other processes not included in typical numerical simulations are responsible for decoupling these comets from Jupiter (Levison and Duncan 1997). The problem of the production of Encke-type comets needs to be studied further before we can determine the contribution of these objects to volatile delivery on Mercury and the other terrestrial planets.

The average properties of the impactor population for all the Monte Carlo iterations are shown in Table IV. The mean amount of water delivered to Mercury from these external bodies is similar for all three sources, although Jupiter-family comets have a larger mean value and a greater potential for contributing very large amounts of water. Mercury-crossing asteroids tend to be smaller and less numerous than the estimated population of Halley-type comets; however, asteroid collisions are more probable, and asteroid impact velocities tend to be lower, leading to a much larger fraction of vapor being retained after an asteroid impact. The advantage gained by the lower impact velocities for asteroids is offset by the assumption that the water content of asteroids or more specifically, $f_{H_2O}$, the mass fraction within the impactor that can be released as water vapor after the impact, is an order of magnitude smaller for asteroids than for comets. Thus, asteroids and Halley-type comets supply a similar mean amount of water to Mercury. Jupiter-family comets, on the other hand, have orbital parameters, impact velocities, collision probabilities, and vapor retention fractions that are more favorable than those of Halley-type comets, and Jupiter-family comets can supply more water to Mercury despite their smaller population estimates.

A comparison of water delivery from all the different external sources is shown in Table I. All four sources are capable of contributing the observed amount of ice to Mercury to within the limits of uncertainty of the observed amount of ice currently on the planet. Given the cleanliness of the observed deposits (Butler et al. 1993, Harmon et al. 1994), IDPs are less likely to be the main water providers than asteroids or comets. Although the mean predicted combined cometary delivery value is higher than that for asteroids, both results are uncertain. In particular, the actual slopes of the cometary and asteroid power-law distributions as shown in Fig. 2 are not well known. Because the largest objects are particularly important in providing Mercury with water, we need more reliable estimates of the slope of the size distributions for objects with diameters larger than $\sim$3 km before we can make any firm conclusions regarding the importance of asteroids as compared with comets.

The assumptions we have made with regard to IDPs may have caused us to inflate the IDP delivery source as compared

**FIG. 7.** Same as described in the legend to Fig. 4 except for Jupiter-family comets.

**FIG. 8.** Same as described in the legend to Fig. 4 except for Halley-type comets.
FIG. 9. The amount of water delivered to Mercury from individual impactors in one simulation as a function of (a) mass of impactor, and (b) collision probability. Asteroids are in green, Jupiter-family comets are in red, and Halley-type comets are in blue. Each impactor is marked by an open circle; stars depict the objects that deliver the most water in each simulation. Note that most of the water delivered to Mercury is supplied by objects with high masses and average-to-high impact probabilities.
with asteroids and comets. For the Earth, the total mass influx from all external sources is estimated to be \( \sim 2 \) to 15 times the IDP mass influx (see discussions by Love and Brownlee (1993), Ceplecha (1996), Rabinowitz 1993), with the bulk of the terrestrial mass influx being supplied by objects larger than 1 m. For our calculations of the micrometeorite influx at Mercury, we have assumed that the eccentricities of the IDPs at 0.39 AU are similar to the low eccentricities of IDPs at 1 AU as determined by Southworth and Sekanina (1973). Thus, the dust particles are impacting at low velocities on Mercury and have a high fraction of vapor retained after the impact. The eccentricities of Mercury-crossing asteroids (see Table II) are on average much higher than the eccentricities of the Earth-crossing asteroid population so that the asteroid-Mercury collision probability is reduced, the average impact velocity is increased, and less vapor is retained after impact for asteroid collisions at Mercury than would be the case if the eccentricities at 1 AU were similar to those at 0.39 AU. Rather than being many times the IDP mass influx at Mercury, the asteroid mass influx is thus comparable to the IDP mass influx, and due to the large fraction of vapor retained after a low-velocity IDP impact, more water is supplied by IDPs than by asteroids. If dust particle eccentricities are actually higher at 0.39 AU than at 1 AU, the relative importance of asteroids as compared with IDPs could shift.

Several uncertainties inherent in the calculations need to be discussed more fully. As discussed above, the comet and asteroid mass distributions are poorly known. Strong observational selection effects distinguish the discovery of Earth-crossing asteroids and comets. We are forced to rely on statistical estimates of the number of undiscovered objects, extinct comets in particular. Many checks on the model are available: the lunar and terrestrial mass influx being supplied by objects larger than 1 m. For our calculations of the micrometeorite influx at Mercury, we have assumed that the eccentricities of the IDPs at 0.39 AU are similar to the low eccentricities of IDPs at 1 AU as determined by Southworth and Sekanina (1973). Thus, the dust particles are impacting at low velocities on Mercury and have a high fraction of vapor retained after the impact. The eccentricities of Mercury-crossing asteroids (see Table II) are on average much higher than the eccentricities of the Earth-crossing asteroid population so that the asteroid-Mercury collision probability is reduced, the average impact velocity is increased, and less vapor is retained after impact for asteroid collisions at Mercury than would be the case if the eccentricities at 1 AU were similar to those at 0.39 AU. Rather than being many times the IDP mass influx at Mercury, the asteroid mass influx is thus comparable to the IDP mass influx, and due to the large fraction of vapor retained after a low-velocity IDP impact, more water is supplied by IDPs than by asteroids. If dust particle eccentricities are actually higher at 0.39 AU than at 1 AU, the relative importance of asteroids as compared with IDPs could shift.

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A possible way to improve our calculations or reduce uncertainties would be to use the cratering record on Mercury to derive the planet’s impact history. This approach is also fraught with uncertainties. Aside from the fact that only one hemisphere of Mercury has been imaged in sufficient detail (by Mariner 10) to get accurate cratering statistics, resurfacing events such as those that formed the intercrater plains on Mercury and the possible saturation of Mercury’s surface with craters of smaller sizes could cause us to underestimate the amount of mass delivered to Mercury by crater-causing bodies. Other uncertainties in deriving impacting populations from cratering records include the need to relate crater diameter to impactor size and the need to translate current crater morphology to transient crater diameter. In addition, it is difficult or impossible to distinguish craters caused by cometary impacts from those caused by asteroids, and most of the craters on Mercury seem to have formed during the late heavy bombardment period (e.g., Strom and Neukum 1988), whereas we are mainly concerned with impacts from the period that postdates this violent episode in the planet’s history. For these reasons, we have chosen to estimate Mercury’s impact history from a statistical analysis of the current impactor population rather than from Mercury’s cratering record. If the terrestrial-planet-crossing population of asteroids and periodic comets has not been in approximate steady state for the past 3.5 byr, our results will not be correct.

An additional uncertainty in our Monte Carlo modeling involves impact vaporization. Very few experiments or numerical models have attempted to elucidate the details of impact vaporization. For instance, the degree of mixing between the vaporized target and impactor material in the impact-derived vapor plume is uncertain, as is the energy partitioning into the plume or the effect of impact angle on this partitioning. We have also neglected the thermochemical processes that could occur within the vapor plume and could cause H2O to be converted to CO or other oxygen-containing compounds (see Zahnle et al. 1995). Further experimental and numerical work on impact vaporization is urgently needed, especially considering the role of impacts in the evolution of terrestrial planet atmospheres.

The degree of loss during migration to the poles may be overestimated by a considerable degree considering that H2O photolysis might not represent a permanent loss of oxygen. Interactions with protons from the solar wind could allow water molecules to reform, or oxygen atoms could perform their own random walk to the poles. Large impacts will also create a temporary collisionally thick atmosphere, and the behavior of water molecules in such an atmosphere would differ from that of molecules hopping in a random walk across Mercury’s surface. All the comets and asteroids in our simulations are large enough to create a collisionally thick atmosphere. For instance, a Jupiter-family comet of diameter 0.5 km (mass \( \approx 8 \times 10^{13} \) g) with 7% vapor retained after impact will create a temporary water–vapor atmosphere of average surface density \( \sim 3 \times 10^{10} \) cm\(^{-3}\) with a mean free path of \( \sim 90 \) m (compared with an average scale height of \( \sim 40 \) km). The vapor released from large impacts will flow supersonically to colder regions on Mercury (i.e., the poles and dark side), where water and other less volatile vapor can condense. Molecules can then be released from the condensed phase when they are rotated onto the day side again. The survival rate of water molecules in such a situation is not known, and this problem deserves further study.

If many of the asteroids in Mercury-crossing orbits are really extinct cometary nuclei, then our estimate of the water content of these objects might be low. In all, our estimates should be considered good only to within an order of magnitude.
6. COMPARISONS WITH OTHER ESTIMATES

In Table V, our estimates of the amount of water delivered to Mercury’s permanently shaded polar regions by different external sources over the past 3.5 byr are compared with the estimates of Killen et al. (1997). As discussed in Section 3.1, our two results for the delivery of IDP-derived water are similar; however, the results for asteroids and comets differ noticeably.

For asteroids, Killen et al. (1997) assume that the mass influx of asteroids in the size range from 1 cm to 10 km is roughly equal to the mass influx from IDPs. Although this assumption may not be true for the Earth (e.g., Rabinowitz 1993; Shoemaker et al. 1990, 1994; Ceplecha 1996), we find that it should be roughly correct for Mercury (unless the eccentricities of Mercury-crossing dust particles are much higher than the eccentricities of IDPs in Earth-crossing orbits). However, Killen et al. (1997) adopt a higher average IDP mass influx than ours, so their assumed asteroidal mass influx is ~2 times ours. In addition, they estimate that on average 37% of the vaporized impact debris will remain on the planet after asteroid impacts, while we estimate a vapor retention fraction that is a factor of ~4 smaller than theirs. Therefore, we derive an average amount of water delivered by asteroids that is a factor of 7–9 times smaller than theirs. We believe our assumed vapor-retention fraction (10%) to be more reliable. However, the total mass influx of asteroids to Mercury will remain uncertain until we can get a reliable estimate of the steady-state size distribution of large Earth-crossing asteroids averaged over million-year time scales.

Our estimates for the delivery of comet-derived water to Mercury also differ markedly from Killen et al. (1997), and we believe that Killen et al. (1997) have made some unreasonable assumptions in deriving their results. First of all, for the steady-state population of active short-period comets, Killen et al. (1997) assume that there are ~10 active short-period comets in the entire Solar System at any time (no size range given), with only 0.4% of these having Mercury-crossing orbits. Thus, only 4% of a comet exists in a Mercury-crossing orbit at any time (cf. Table III). The two main problems with these assumptions are that (a) they violate observations of known Earth-crossing and Mercury-crossing comets (see Shoemaker et al. (1994) and Marsden and Williams (1995)), and (b) Killen et al. consider comets to be Mercury crossing only if they have perihelion distances less than 0.3 AU while comets with q out to Mercury’s aphelion distance of 0.467 AU should be included. Secondly, Killen et al. assume that the ratio of extinct-to-active short-period comets is ~78, a factor of ~4 larger than the highest estimate in the literature (Shoemaker et al. 1994), and they use their (mistaken) estimate of 0.4% in Mercury-crossing orbits to hold true for Earth-crossing extinct comets rather than for all extinct comets. The actual fraction of Mercury-crossing comets within the Earth-crossing population as determined by Fernández (1984) should be closer to 25% rather than 0.4%.

Killen et al. (1997) also assume that the collision probability per short-period comet is $2 \times 10^{-9}$ yr$^{-1}$, and they quote Ip and Fernández (1988) for this result. In reality, Ip and Fernández (1988) discuss long-period comets, not short-period comets, and they derive an impact probability with Mercury that is $2 \times 10^{-9}$ per orbit for a new long-period comet injected into the Solar System, not an impact probability per year. Due to their long orbital periods, such comets will have a collision probability per year that is about a million times smaller than $2 \times 10^{-9}$ yr$^{-1}$. Fortunately, the number adopted by Killen et al. is not unreasonable for short-period comets (see Tables III and IV and Olsson-Steel (1987)).

Killen et al. (1997) also determine that on average 20% of the vaporized comet will be retained by the planet after impact while we derive much lower values of 7% for Jupiter-family comets or 1.5% for Halley-type comets. More importantly, Killen et al. (1997) use an average value for cometary mass when calculating the total mass influx to Mercury, while we determine that the largest objects bring in most of the mass. The net result of our different assumptions is that Killen et al. (1997) derive a comet-supplied water abundance that is on average a factor of ~2.5 smaller than ours. Due to the statistical variation in our comet simulations and to the fact that the largest comets are the most important, our maximum value for the estimate of the amount of water delivered to Mercury’s poles in the past 3.5 billion years is a factor of ~30 higher than the maximum estimate of Killen et al. (1997).

<table>
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<th>TABLE V</th>
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<td>Estimates of the Mass of Water Delivered to Mercury’s Permanently Shaded Polar Regions over the Past 3.5 Billion Years (in $10^3$ g)</td>
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*Note. The values of Killen et al. (1997) were derived from their estimates of the mass influx rate, the fraction of vapor retained, the fraction of water in the object, and the fraction of the vaporized water molecules that survive to reach the polar regions; the numbers were normalized to 3.5 billion years for comparison with our estimates.*

7. SUMMARY AND CONCLUSIONS

The correlation of the radar-bright regions observed by Harmon et al. (1994) with high-latitude craters on Mercury leaves little room for doubt that the material responsible for the unusual radar signatures is a volatile solid. Carbon dioxide or sulfur dioxide ices may be too volatile to remain thermally stable over geologic time scales. Sulfur and other more refractory “volatiles” might be stable in many regions across the planet’s surface and would not be so obviously correlated with high-latitude craters (e.g., Butler (1997); note that because the vapor pressure of sulfur at low temperatures is poorly known, further investigations into possible sulfur deposits at the poles might
still be warranted). Water is the most abundant volatile in the Solar System, found everywhere from comets and interplanetary dust particles to atmospheres and surfaces of planets. In addition, water ice has the desired radar properties and would be stable in the regions in which the unusual radar signatures are found. Therefore, water was the focus of our study.

The radar features imply that the ice deposits in the permanently shaded regions on Mercury are substantial; from information given in Harmon et al. (1994), we estimate that as much as (4–80) \times 10^{16} \text{g} of water ice could currently be residing at high latitudes on Mercury. Because much of the original inventory of water on Mercury could have been lost during the migration of the water molecules from wherever they were emplaced on the planet to the cold polar regions (Butler et al. 1993; Butler 1997) or could have been eroded or sublimated once they condensed at the poles (Killen et al. 1997), 85–95% of the original source must have been lost before arriving in the permanently shaded regions at the poles. In other words, the original source must have supplied (8–50) \times 10^{17} \text{g} of water to Mercury. Given the possible but uncertain dryness of Mercury’s crust and mantle (e.g., Benz et al. 1988), interior sources such as volcanic outgassing might have difficulty releasing such large amounts of water, even over billions of years. Chemical sputtering of surface rocks might release some water (Potter 1995), but we feel that external sources such as cometary and asteroid impacts have the greatest potential for delivering large amounts of water to Mercury. The only problem with exogenic sources is that Mercury’s low gravity prevents much of the energetic vaporized debris released after a high-velocity impact from being retained by the planet.

We have extrapolated the current influx of IDPs at the Earth to that at Mercury assuming that the flux has not changed in the past 3.5 byr and that the size distribution and orbital properties of the dust particles do not change with heliocentric distance. If impact vaporization of the dust particles occurs as is delimitated in Appendix B, we find that IDPs could have delivered roughly (5–40) \times 10^{17} \text{g} of water to Mercury since the end of the late heavy bombardment period (approximately 3.5 byr ago). Although this value is consistent with the estimated amount of water needed to explain the radar anomalies on Mercury, the IDP source cannot explain the cleanliness of the ice deposits on Mercury (see below).

Nearly parabolic or hyperbolic comets will impact Mercury at high velocities, usually in excess of 70 km s^{-1}, and little water would be retained from such objects (see Fig. B1). In addition, the collision probabilities of long-period comets with Mercury are low (e.g., Olsson-Steel 1987); therefore, we do not consider impacts from long-period comets in this paper. If the population estimates or sizes of long-period comets are considerably greater than those of Halley-type comets, long-period comets might be worth reconsidering as sources of water for Mercury.

To determine how much water may have been delivered to Mercury by short-period comets (both Halley-type and Jupiter-family comets) and asteroids since the end of the late heavy bombardment period, we have developed a Monte Carlo model to simulate the impact history of Mercury over the past 3.5 byr. We use current estimates of the number and size distributions of Earth-crossing comets and asteroids (Shoemaker et al. 1994, Rabinowitz et al. 1994) as well as orbital parameters from the known Earth-crossing objects (see Fig. 2) to constrain our impacting population. In the model, we select orbital elements and impactor sizes from these input distributions and identify those objects that potentially cross Mercury’s orbit. If we assume that the asteroid and short-period comet population has been in approximate steady state over the past ~3.5 byr (see Shoemaker et al. 1979), we can use this statistical sample to predict collision probabilities, impact velocities, and the mass fraction of the impactors that will be retained by the planet after impact vaporization. Since the steady-state number of Mercury-crossing objects is small (~780 asteroids, ~390 Jupiter-family comets, and ~5800 Halley-type comets larger than 0.5 km diameter), we perform the Monte Carlo calculations 300 separate times to derive 300 possible impact histories for Mercury.

We find that within the possible lower limits to the mass of the current ice deposits on Mercury, impacts from both asteroids and comets appear to be sufficient to supply Mercury with the required abundance of water (see Table I). Jupiter-family comets, in particular, have the potential for delivering large amounts of water to the planet. Our Monte Carlo simulations indicate that the polar deposits could contain (0.4–20) \times 10^{15} \text{g} of asteroid-derived ice, (0.2–20) \times 10^{15} \text{g} of Halley-type comet-derived ice, or (0.1–200) \times 10^{16} \text{g} of Jupiter-family comet-derived ice. Although asteroids have generally higher collision probabilities and lower impact velocities than both types of comets (and thus retain more of their vaporized mass after impact), they are thought to contain a volatile fraction smaller than that of comets (even than that of extinct comets). If many of the Mercury-crossing asteroids are extinct cometary nuclei rather than true main-belt asteroid collisional fragments, or if we have overestimated the slope of the size distribution for large asteroids, then the relative importance of asteroids versus comets in supplying Mercury with water could change. Mercury-crossing Jupiter-family comets, although less numerous than Halley-type comets, have higher collision probabilities, lower impact velocities, and a higher fraction of the vaporized debris retained by the planet after an impact; thus, Jupiter-family comets have a higher potential for delivering large amounts of water to Mercury.

The Monte Carlo simulations also indicate that a small number of massive objects deliver most of Mercury’s water inventory. This result can help explain the cleanliness of the polar deposits. The strength and polarization behavior of the radar echoes implies that the ice layers must be relatively free from dust or other radar-absorbing materials (Butler et al. 1993, Harmon et al. 1994). If a continuous source such as IDP impact or chemical sputtering were responsible for providing Mercury with most of its water, the ice deposits might be expected to be well mixed with dust and other debris kicked up by micrometeorite impacts (e.g., Killen et al. 1997). However, the time for migration of
single water molecules to the poles is relatively short (Butler et al. 1993, Butler 1997); if one or more large events were responsible for the water being supplied to the planet then the water could arrive at the poles fast enough to remain relatively free from dust contaminants.

The comets that supply most of the water to Mercury tend to have Tisserand parameters with respect to Jupiter that are between 2 and 3. Halley-type comets, with $T < 2$, have low collision probabilities, and high eccentricities and impact velocities, and thus retain less water after impact. Encke-type comets, with $T > 3$, are rare in our simulations despite their presence in the observed Earth-crossing comet population. Encke-type comets would be expected to have high collision probabilities and low impact velocities and thus would have favorable properties for delivering water to Mercury. The origin of Encke-type comets is still an unsolved problem (see Levison and Duncan 1997). A theoretical determination of the expected fraction of Encke-type comets within the Jupiter-family comet population would aid in the determination of the amount of water supplied to Mercury by short-period comets.

The statistical models developed in this paper can be applied to studies of volatile delivery on the Moon and other airless bodies in the Solar System. Given the larger population of Earth-crossing objects than of Mercury-crossing objects and the lower average impact velocity on the Moon than on Mercury, it seems certain that asteroidal and cometary impacts would have supplied large amounts of water to the Moon as well as Mercury. However, the erosion rate or sublimation rate of ice deposits on the Moon could be considerably different from that at Mercury, especially considering the long-term stability of Mercury’s obliquity. The recent Lunar Prospector mission has revealed the presence of water at the Moon’s poles. Preliminary results indicate that the lunar poles contain a combined amount of ice that is roughly within an order of magnitude of $\sim 5 \times 10^{14}$ g (A. Binder and W. Feldman, NASA press conference, March 5, 1998). Although the area covered by these ice deposits is similar to the area observed in the polar regions of Mercury, the cleanliness of the deposits on the two bodies is much different. The Lunar Prospector data suggest that the mixing ratio of water in the lunar regolith is only 0.3–1%. Therefore, although the results for both planets are uncertain, it appears that Mercury has at least two orders of magnitude more ice in its polar regions than the Moon. Realistic models of the fate of ice deposits on the Moon must be developed before this interesting result can be explained.

APPENDIX A: IDP INFUX ON MERCURY

We use the method of Cintala (1992), Morgan et al. (1988), and Zook (1975) to determine the current mass flux of IDPs onto Mercury (averaged over an orbit). Here, the observed terrestrial flux is extrapolated to Mercury, taking gravitational focusing and the heliocentric variation of the IDP population into account.

A.1. Mass distribution at the Earth. The LDEF satellite was ideally suited for determining the terrestrial IDP flux (e.g., Love and Brownlee 1993). The satellite exposed a large surface area to incoming debris and was stabilized so that the space-facing end of the satellite remained continually pointed toward the zenith (to within 1°). Over the course of nearly six years, 761 tiny craters were recorded on this surface. The mass distribution of IDPs at Earth was determined from an analysis of these craters (Love and Brownlee 1993), and the results are shown in Fig. A1. Note that a strong peak appears in the mass distribution at $1.5 \times 10^{-5}$ g. Love and Brownlee determine that the total mass accreted by the Earth per year within the size range sampled ($10^{-9}$ to $10^{-4}$ g) is $40 \pm 20 \times 10^9$ g per year. However, when we integrate under the curve shown in Fig. A1 (which

![FIG. A1](image-url) The mass distribution of interplanetary dust particles encountering the Earth (from Love and Brownlee 1993).
was obtained from hand-digitizing the Love and Brownlee curve and checked by working through their polynomial fit, we obtain a value of \(30 \times 10^9 \text{ g yr}^{-1}\). Although this result is within the quoted error bars, the discrepancy is puzzling and serves to emphasize that even the terrestrial influx rate is uncertain; extrapolating to Mercury just adds further sources of uncertainty. We take the actual curve shown in Fig. A1 at face value and use it throughout the later calculations.

For a lack of better information, we will assume that the size distribution of IDPs near Mercury’s orbit will have the same relative shape (so that the relative populations of the different particle sizes remain the same), but that the total mass influx differs.

If the observed velocities of incoming meteoroids can be assumed to be independent of the mass of the particles, then the total flux of small particles onto Earth can be written

\[
\Phi = \int_{v_{\text{esc}}}^{v_{\text{max}}} P_0(v_0) dv_0 \int_{10^{-9} g}^{10^{-4} g} h(m) dm,
\]  

(A1)

where \(P_0(v_0) dv_0\) is the velocity distribution (i.e., the probability that the incoming particles has a velocity between \(v_0\) and \(v_0 + dv_0\), \(v_{\text{esc}}\) is the escape velocity at the top of the Earth’s atmosphere, \(v_{\text{max}}\) is the maximum possible velocity for particles entering the Earth’s atmosphere (~72 km s\(^{-1}\) for particles in orbits that are bound to the Solar System), and \(h(m) dm\) is the distribution of particles with masses between \(m\) and \(m + dm\) (see Fig. A1). From the LDEF findings, the terrestrial influx is \(30 \times 10^9\) g per year, or \(\Phi = 1.9 \times 10^{-16} \text{ g cm}^{-2} \text{ s}^{-1}\) if the particles are coming in isotropically. Note that this value is twice the terrestrial flux rate determined by Grün (1985) and used by Cintala (1992) and is ~46% lower than the terrestrial flux adopted by Morgan et al. (1988) and Killen et al. (1997).

A.2. Velocity distribution at 1 AU. Following Cintala (1992), we take the Southworth and Sekanina (1973) observed radar-meteor velocity distribution to be representative of meteoroid velocities at the top of the Earth’s atmosphere. For the purposes of transforming the terrestrial distribution to that at Mercury, it is convenient to have the velocity distribution in a functional form. Cintala (1992) and Morgan et al. (1988) (and Zook (1975) before them) use an exponential fit to the data presented in Southworth and Sekanina (1973). Because such exponential fits might cause an underestimation of the number of moderately low-velocity impactors and an overestimation of the number of high-velocity impactors, we have chosen a more complicated analytical form using 17 basis functions (including sines, cosines, and polynomials). The actual observed velocity distribution in the Earth’s atmosphere and the different analytic fits to this distribution are shown in Fig. A2. Note that the distribution has been normalized so that an integration over all possible velocities gives a value of unity.

The Earth sweeps up an amount of dust that is larger than the flux that would be observed at 1 AU far from the Earth’s gravitational influence. Conservation of mass, angular momentum, and energy require that if we ignore the third-body effect of the Sun, this “gravitational focusing” effect causes the meteoritic flux onto the Earth to be a factor of \(v_0^3/v_{\text{esc}}^3\) times greater than the flux in free space at 1 AU (Opik 1951), where \(v_0\) is velocity at the top of the Earth’s atmosphere and \(v_{\text{esc}}\) is the relative velocity of the meteoroid with respect to the Earth when the particle is far enough away that gravitational effects are negligible. If we assume that the particle masses are not correlated with velocities, then the observed velocity distribution at the top of the atmosphere can be used to determine the relative velocities of the Earth and dust particle before gravitational focusing took place. The velocity distribution in free space \(P_0(v_{\text{esc}})\) becomes (e.g., Zook 1975)

\[
P_{\infty}(v_{\text{esc}}) = \left( \frac{v_{\text{esc}}}{v_0} \right)^3 P_0(v_0),
\]  

(A2)

where the observed velocity \(v_0\) at Earth is related to the velocity \(v_{\text{esc}}\) in free space by the relation

\[
v_0^2 = v_{\text{esc}}^2 + v_{\text{esc}}^2.
\]  

(A3)

**FIG. A2.** The velocity distribution of interplanetary dust particles entering the Earth’s atmosphere. The solid line is the observed distribution (from Southworth and Sekanina 1973), the dashed line is the exponential analytic fit of Cintala (1992), and the dotted line is our analytic fit to the Southworth and Sekanina distribution.
A.3. Transformation to Mercury. To determine the IDP flux at Mercury, we need several other pieces of information. First of all, we need to know the spatial density of IDPs near Mercury. Observations of light scattering by dust particles in the inner solar system by the Heliios spacecraft (Leinert et al. 1981) suggest that the spatial density of IDPs in the inner Solar System has an $R^{-1.3}$ dependence, where $R$ is the distance from the Sun. Thus, IDPs are more abundant at Mercury’s mean heliocentric distance of 0.387 AU than they are at 1 AU. Secondly, we need to know the free-space velocity distribution of dust particles with respect to Mercury far from the planet’s gravitational influence. Little information exists concerning the orbits of dust particles in the inner Solar System; we will simply assume as did Morgan et al. (1988) and Cintala (1992) that the orbits derived for the Earth-impacting micrometeorites have the same eccentricity and inclination in the inner Solar System as they do near 1 AU. The orbital velocities then scale as $R^{-0.5}$. We will ignore the fact that Mercury itself has a different eccentricity and inclination than the Earth and assume that this scaling holds true for the relative velocities of the dust particles with respect to Mercury. Finally, we need to take gravitational focusing onto Mercury into account. Note that we derive average values for the flux over an orbit rather than considering the different influx rate at Mercury aphelion versus perihelion.

Putting it all together, if $P_M(v_M)\, dh(m)\, dv_M\, dm$ is the flux of impactors with masses between $m$ and $m+dm$ and with velocities between $v_M$ and $v_M + dv_M$, then the IDP flux at Mercury’s surface can be written (Cintala 1992)

$$\Phi_M = \int_{v_M,esc}^{116 \text{ km/s}} P_M(v_M)\, dv_M \int_{10^{-4} \text{ g}}^{h(m)} h(m)\, dm,$$  

(A4)

where the “velocity distribution” at Mercury’s surface is

$$P_M(v_M) = R^{0.2} \left( \frac{v_M}{v_0} \right)^3 P_0(v_0).$$  

(A5)

$R = 0.387$ AU, $v_M$ is the impact velocity on Mercury and is related to $v_0$ through the relation $v_M^2 = R(v_M^2 - v_{M,esc}^2) + v_{M,esc}^2$, $v_{M,esc}$ is the escape velocity from Mercury’s surface, and $h(m)$ is shown in Fig. A1. Therefore, from the observed distribution of meteoroid masses and velocities in the terrestrial atmosphere, we can estimate the IDP flux onto Mercury.

Using the approximations discussed above, we find that the current flux of IDPs onto Mercury while the planet is at its mean distance from the Sun is a factor of 2.25 times higher than that of the Earth, implying a flux of order $4 \times 10^{-16}$ g cm$^{-2}$ s$^{-1}$. Over Mercury’s total surface, the mass influx comes to $9.8 \times 10^9$ g yr$^{-1}$. The average impact velocity of IDPs on Mercury is $\sim 19.7$ km s$^{-1}$ while the average impact velocity on Earth using the Southworth and Sekanina (1973) velocity distribution is 15.0 km s$^{-1}$.

Our value for the IDP flux at Mercury is a factor of 2 smaller than that estimated by Morgan et al. (1988) (see also Killen et al. 1997) due to a smaller assumed terrestrial flux and a velocity distribution weighted toward lower velocities.

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**APPENDIX B: IMPACT VAPORIZATION AND VAPOR RETENTION**

Because impact velocities of objects reaching Mercury’s surface tend to be quite high, impactors are often completely vaporized after colliding with the planet. A cloud of vaporized projectile and target material is generated during the impact, and the gas molecules within the cloud will have a wide distribution of molecular velocities. Although the vapor cloud may be sufficiently hot that, on average, most of the vapor can escape from the planet, a fraction of the vapor molecules will be traveling slower than the planet’s escape velocity and thus will be retained after the impact (see Fig. B1). To determine how much water vapor remains on Mercury’s surface after an impact, we need to first calculate the mass fraction of the vaporized projectile that remains gravitationally bound to the planet.

Our technique follows that of Zel’dovich and Raizer (1966, 1967) who considered the problem of the sudden expansion of a cloud of vapor into vacuum (see also Vickery and Melosh 1990, Zahnle 1990, Melosh 1989). We assume that the cloud of vaporized impactor and target material is initially represented by an ideal gas with uniform density $\rho_0$ and uniform pressure $p_0$ and occupies a hemispherical volume of radius $R_0$. Our hemisphere of vapor has total mass $M_{tot} = \frac{4}{3}\pi R_0^3 \rho_0$ and total energy $E_{tot} = \frac{1}{2}(1/\gamma - 1)\pi R_0^3 p_0$, where $\gamma$ is the ratio of specific heats of the gas. The cloud is initially at rest. At time $t = 0$, expansion into vacuum begins. The expansion is homologous (shape preserving) and isentropic because the constant initial pressure and density of the vapor ensures that all the gas molecules along the radius of the cloud have the same entropy. The average thermal velocity in the cloud is $v = \sqrt{2E_{tot}/M_{tot}}$. The cloud edge expands at a maximum rate that is determined by

$$u_{max} = \frac{2}{(\gamma - 1)\rho_0}\left(\frac{\gamma}{\gamma - 1}\right)^{1/2}$$

(Zel’dovich and Raizer 1966), where $v_0$ is the specific internal energy of the vapor plume, and the radius of the cloud at any given time is determined by $R = R_0 + u_{max} t$. Eventually, the cloud expands such that $R \gg R_0$, and the flow becomes nearly self-similar. At that point, the velocity is linear with radius $r$ such that

$$u(r) = u_{max} \frac{r}{R}.$$

(B1)

The outer regions of the cloud are traveling faster than the central portion.

The density distribution within the evolving cloud cannot be determined uniquely. By direct analogy with the analytically solvable one-dimensional problem, Zel’’dovich and Raizer (1966), Vickery and Melosh (1990), and Zahnle (1990) suggest the following form for the density distribution:

$$\rho(r) = \frac{A}{R^3} \left(1 - \frac{r^2}{R^2}\right)^{\alpha}, \quad \alpha = \frac{5 + \gamma}{2(\gamma - 1)}, \quad R = u_{max} t.$$

(B2)

The constant $A$ can be determined from conservation of mass, i.e., by integrating the density distribution over the entire volume of the hemisphere and setting

---

**FIG. B1.** An illustration of vapor cloud expansion following a hypervelocity impact.
the result equal to \( M_{\text{tot}} \). The equations become more tractable if \( a \) is an integer; Vickery and Melosh (1990) and Zahnle (1990) suggest that \( \gamma = \frac{4}{3} \) is a reasonable choice for a high-temperature silicate or mixed silicate-water cloud after a large impact, implying that \( a = 11 \) and \( a \) is approximately 15.39 \( M_{\text{tot}} \).

The mass fraction of the plume that does not attain escape velocity \( v_{\text{esc}} \) is

\[
\frac{M(v < v_{\text{esc}})}{M_{\text{tot}}} = \int_{v_{\text{esc}}}^{v_{\text{max}}} \frac{\rho(r)2\pi r^2 dr}{\int_0^{r_{\text{max}}} \rho(r)2\pi r^2 dr},
\]

where \( v_{\text{esc}} = Rv_{\text{esc}}/u_{\text{max}} \). This latter quantity depends on the specific internal energy \( e_0 \) of the vapor plume through the expression for \( u_{\text{max}} \). Herein lies the largest uncertainty of our calculations of the fraction of vapor retained after an impact: few experimental data or numerical models are available to help determine what fraction of the projectile’s initial kinetic energy will be coupled to the expanding vapor cloud, although this topic is coming under increasing scrutiny due to its importance in problems of volatile delivery and atmospheric erosion on the terrestrial planets.

The internal energy of the vapor plume will depend in a complicated way on the impact velocity \( v_i \), impact angle \( \theta \), and the composition of both the impactor and target. The physics of impact vaporization is not well understood, especially for oblique impacts, and the partitioning of the original kinetic energy of the projectile into vaporization, displacement, comminution, melting, and vaporization of the target as well as melting, vaporization, and ricochet of the impactor is not known.

Using the planar impact approximation for vertical impacts, Melosh (1989) demonstrates that during the compresional stage of impacts of projectiles with targets of the same composition, one-quarter of the initial kinetic energy of the projectile is partitioned into kinetic energy of the target, one-quarter to internal energy of the target, one-quarter to kinetic energy of the projectile, and one-quarter to internal energy of the projectile. Vaporization begins during this stage of the impact as the material is released from high pressure. During the later crater-formation stage, more of the projectile’s energy will be distributed to the target so that as much as 90% of the initial energy will be converted to internal energy of the target. Oblique impacts are more complicated. Vaporization is more pronounced, and jetting of material and ricochet of the projectile are often observed (Schultz 1996).

To determine the specific internal energy of the vapor cloud, we will follow the procedure of Vickery and Melosh (1990) who estimate that, in general, the vapor cloud energy \( E_{\text{tot}} \) can be calculated by

\[
E_{\text{tot}} = m_p \left( \frac{u_i^2}{2} - L_p \right) + m_t \left( \frac{u_t^2}{2} - L_t \right),
\]

where \( m_p \) and \( m_t \) are the mass of the vaporized target and projectile, \( u_i \) and \( u_t \) are the maximum particle velocities in the projectile and target, and \( L_p \) and \( L_t \) are the latent heat of vaporization of the projectile and target material. For impacts of a silicate projectile onto a silicate target, \( L_p = L_t = 1 \), and the peak particle velocity in the limit of a sharp shock in either the target or projectile is approximately hal the impact velocity (Vickery and Melosh 1990), implying a maximum vapor-cloud energy of \( E_{\text{tot}} \approx (m_p + m_t)u_i^2/8 \). Because the peak pressure transmitted downward into the target during the compresional stage of the impact depends on the vertical component of velocity, we might expect the energy coupled to the vapor cloud in an oblique impact to be at most \( E_{\text{tot}} \approx (m_p + m_t)u_i^2 \sin^2 \theta / 8 \), where \( \theta \) is the impact angle relative to the horizontal. Complete vaporization of the projectile and target material will occur when the specific thermal energy left in the shocked material at the end of the compression stage is roughly twice the latent heat of vaporization (Zel’’dovich and Raizer 1967), or in our description when \( v_i \sin \theta \approx 4\sqrt{L/\rho} \). A typical value for the latent heat of vaporization of silicates is \( 1.3 \times 10^{11} \) erg g\(^{-1}\) (Melosh 1989).

Schultz (1996) finds that although the energy of the vapor plume never exceeds the limits described above, the dependence on impact angle differs from the expected \( \sin^2 \theta \). In experiments of hypervelocity impact-induced vaporization of volatile targets, Schultz (1996) finds that although peak shock pressures decrease with decreasing impact angle (from the horizontal), the amount of vaporized target material increases because of increased frictional shear heating in the target downrange from the impact site; hence, the temperature and velocity of the vapor cloud decreases with decreasing impact angle. Schultz finds that once the impact velocity exceeds some critical value, the internal energy of the vaporized target material scales as \( \cos^2 \theta \cos^2 \theta \), giving a relation for the mass of target material that is vaporized during the impact that is proportional to \( v_i^2 \sin^2 \theta \) times the mass \( m_t \) of the projectile.

In Schultz’s experiments, the projectile itself was not vaporized but was fragmented after impact, and the ricocheted fragments continued to induce vaporization downrange in the target. At the high impact velocities typical for Mercury, impactors will most likely be completely vaporized, and fragments of the impactor should not contribute measurably to vaporization of the target. If the ricochet component of vaporization is removed from Schultz’s description, the internal energy of the vaporized target material scales as \( v_i^2 \cos^2 \theta \sin \theta \).

Because the projectile was not vaporized in Schultz’s experiments, however, it is not clear how applicable these experiments are to the situation on Mercury. Does the vaporized projectile mix with the vaporized target material to form the “hemispherical” vapor plume? Are the vapors from the different sources separated in time and/or space? The answers to these questions are currently unknown and will have to await further experiments or numerical models.

For silicate-onto-silicate impacts such as would roughly occur with impacts of asteroids or IDPs onto Mercury’s surface, we will assume that the maximum particle velocities are given by \( u_i = u_t = \frac{v_i}{\cos \theta} \), and latent heats by \( L_p = L_t = L = 1.3 \times 10^{11} \) erg g\(^{-1}\). Furthermore, we will follow the formalism of Vickery and Melosh (1990) and assume that if \( u_i \) exceeds 14.4 km s\(^{-1}\), the projectile is completely vaporized and ends up well mixed with an equal mass of vaporized target so that the specific energy of the vapor cloud becomes

\[
e_{\text{tot}} = \left( \frac{v_i^2 \sin^2 \theta}{8} - L \right). \tag{B5}
\]

for velocities below 14.4 km s\(^{-1}\), we assume that the projectile and target are not vaporized. In reality, the mass of target material vaporized will scale with impact angle as discussed above; however, the actual mass produced will depend on the the compositions of the materials involved as well as \( v_i \) and \( \theta \). Schultz (1996) finds that for oblique impacts into a variety of volatile targets, the mass of vaporized target material easily exceeds the mass of the projectile, and values of \( m_t/m_p \) of \( \sim 10 \) are achieved for impact angles in the range 15°–30°. Schultz also suggests that the scaling relations will hold provided that the impact velocity exceeds some critical velocity for the target material in question. These results suggest that silicate-onto-silicate impacts at impact velocities typical for Mercury would liberate an amount of vaporized target that is greater than the mass of the impactor; however, given the uncertainties involved, we have chosen a more conservative value that may lead to an underestimation of the amount of water retained during an impact on Mercury.

As an additional approximation, we consider that all impactors strike Mercury at the most probable impact angle of 45°. The vaporization process is simply too complicated and not well documented at this stage to attempt to determine the effects of impact angle on the partitioning of energy to the vapor cloud. The largest uncertainties are for very low angle (oblique) impacts where the projectile may ricochet downrange with a velocity nearly equal to its original horizontal velocity. As a further complication, the energy tied up as latent heat of the vaporized debris may become available to the expanding vapor cloud if any of the vapor recondenses. However, the hot vapor may also radiate some of its energy away and so may not expand as rapidly as we have assumed. Judging from the experimental results of Schultz (1996), the results of these combined approximations will probably be to overestimate the specific energy of the vapor cloud and thus underestimate the amount of vapor retained in the impact.

For comet impacts on Mercury, we also assume that the impact causes an equal mass of target material to be vaporized. We assume that the comet is
FIG. B2. The mass fraction of the impact-induced vapor cloud that is retained by Mercury after an impact. The dotted line is for cometary impacts and the solid line is for asteroidal or micrometeoroidal (i.e., IDP) impacts.

50% water ice and 50% silicate by mass and follow the procedure of Vickery and Melosh (1990) to determine the resulting particle velocities and vapor-cloud energy during such an impact. The specific energy of the vapor cloud becomes

\[ E_0 = 0.128 \rho_i^2 \sin^2 \theta - \frac{1}{2}(L + L), \]  

(B6)

where the latent heat of vaporization for the silicate target \( L = 1.3 \times 10^{11} \text{ erg g}^{-1} \) and the latent heat of vaporization for the comet \( L \) is the average between that of silicates \( (1.3 \times 10^{11} \text{ erg g}^{-1}) \) and water ice \( (3 \times 10^{10} \text{ erg g}^{-1}) \). As with the case for asteroids and IDPs, we assume that the impact angle is constant at 45°.

The results are shown in Fig. B2. Here, the mass fraction of the impact-induced vapor cloud that remains gravitationally bound to Mercury is plotted as a function of impact velocity. The results for both cometary-onto-silicate (dotted line) and silicate-onto-silicate (solid line) are presented. The results for the two impactor populations are similar—the vaporized fraction that is retained by the planet decreases precipitously with increasing impact velocity beyond \( \approx 15 \text{ km s}^{-1} \).

To determine the amount of water that is retained upon impact, we assume that Mercury’s surface is dry and that impacts do not liberate water from the surface material itself. Asteroids are assumed to contribute 5% water by mass, comets 50% water by mass, and IDPs 10% water by mass. These results can easily be scaled to any desired water content. For all the impactor types, if \( E_0 < 0 \), we assume that the impactor does not become vaporized, and the entire water content of the impactor will eventually become available to migrate to the polar regions.

ACKNOWLEDGMENTS

We thank M. W. Schaefer for useful advice and suggestions, S. Hokanson for helping create Fig. B1, and Yale University and the LPI summer intern program for providing support for the second author. Conversations with B. J. Butler, R. M. Killen, T. H. Morgan, M. Zolensky, and J. B. Pollack enhanced this work and are gratefully acknowledged.

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