Evidence of Exposed Dusty Water Ice within Martian Gullies

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Abstract  Midlatitude slopes on Mars are mantled by deposits proposed to contain H2O ice and dust, overlaid by a desiccated lag. However, direct evidence of their volatile content is lacking. Here we present novel evidence of light-toned materials within midlatitude gully alcoves eroded into these mantles. The appearance and Lambert albedo of these materials suggests that they are either dust or H2O ice. We interpret them to be H2O ice because it is unlikely for a short-term, localized dust deposit to form only within the mantle walls. The temperatures are generally too warm (>∼240 K) for the ice to be a frost in equilibrium. Therefore, this ice is likely similar to the dusty ice documented within midlatitude scarps, but with more dust, and exposed in smaller patches by slumping. It has been proposed that CO2 frosts remove the overlying lag, causing the exposed H2O ice to sublimate, liberate dust within the ice for transport, and erode gullies in the mantle. But we observe gullies eroded in wall rock that continue into the mantle, implying that the same process erodes both substrates. H2O ice melt can explain gullies eroded in the wall rock and the mantle. Numerical models show that relatively dense H2O snow on Mars melts only when it contains small amounts of dust. The observed exposure of dusty ice provides a mechanism for it to melt under some conditions and form some gullies. Access to liquid water within this ice could provide potential abodes for any extant life.

Plain Language Summary  Midlatitude, Martian pole-facing slopes are covered by smooth deposits (“mantles”) thought to contain dusty ice, covered by desiccated material. However, there is a lack of evidence that these deposits contain dusty ice. We present evidence of light-toned materials being exposed within gullies eroded into these smooth deposits. Based on the materials’ brightness, we interpret them to be water ice because it is unlikely for a short-term, localized dust deposit to form only within these smooth deposits. It is too warm for the ice to be water frost, so the ice is likely dusty ice exposed when the overlying desiccated material slumps down. Previous studies propose that evaporating Martian carbon dioxide frosts can mobilize the overlying desiccated material to expose dusty ice, free up dust for transport and erode gullies by evaporating the water ice. But we observe gullies eroded in rock that continue into the smooth deposits, which are readily explained by ice melt. Computer simulations show that Martian snow melts only when it is dusty. The observed exposure of dusty ice means it could, under certain conditions, melt to form gullies and provide abodes for any extant life.

1. Background

1.1. H2O Ice on Mars

The presence of H2O ice and liquid water is key to the evolution of Martian geology, with implications for the potential for past or extant life, and the future of robotic and human exploration on Mars. H2O ice reservoirs hold key records of Mars’ climate history, and understanding the state and evolution of subsurface ice can provide insight on past and present-day processes on Mars.

The obliquity of Mars is thought to have oscillated over recent timescales (105–106 years), leading to the cycling of H2O ice from the poles to the midlatitudes (Head et al., 2003; Jakosky et al., 1995). The poles contain the largest known reservoirs (~2.6 ×105 km3) of ice (Byrne, 2009; Smith et al., 2001). The seasonal polar caps are composed mostly of CO2 ice, and are present at the surface of each pole during the winter (Kieffer, 1979; Leighton & Murray, 1966). During the summer, the retreating seasonal caps reveal underlying perennial ice caps. The northern perennial ice cap is composed of H2O ice (Farmer et al., 1976; Kieffer et al., 1976), and is the principal source and sink of water vapor on Mars (Smith, 2004). The southern perennial cap consists
of a ~10 m thick CO₂ ice layer over H₂O ice (Bibring et al., 2004; Byrne, 2009; Kieffer et al., 2000; Titus et al., 2003). Stacks of layered H₂O ice deposits that are several kilometers thick extend outwards from each pole. Like the perennial caps, these polar layered deposits also show hemispherical differences in morphology and structure, representing their different ages and formation histories (Byrne, 2009).

Ice is also present at the midlatitudes, but it is usually within the subsurface. Models of ice stability predicted that water ice is stable below <~1 m of desiccated material poleward of ~50° in either hemisphere for flat ground (Leighton & Murray, 1966; Mellon & Jakosky, 1995; Mellon et al., 2004; Schorghofer & Aharonson, 2005). This prediction was supported by data from the Mars Odyssey Neutron Spectrometer, High Energy Neutron Detector, and Gamma Ray instruments (Boynton et al., 2004; Feldman et al., 2002; Mitrofanov et al., 2002), that showed high water-equivalent hydrogen in the top ~1 m of the surface, poleward of ~50° in both hemispheres. In situ evidence of this shallow, subsurface ice was provided by the Phoenix lander at ~68°N (Smith et al., 2009).

Geomorphic evidence for ice-rich materials has been documented through much of the midlatitudes, ranging from viscous flow features (Milliken et al., 2003), lobate-debris aprons (Head et al., 2005; Mangold, 2003; Pierce & Crown, 2003; Squyres, 1979), concentric crater fill (Dickson et al., 2010; Levy et al., 2010; Squyres & Carr, 1986), lineated valley fill (Dickson et al., 2012; Levy et al., 2007), scalled compressional features (e.g., Zanetti et al., 2010), expanded craters (e.g., Viola & McEwen, 2018), and thermal contraction polygons (Heeschen et al., 2009a). Subsurface radar measurements have detected H₂O ice within some of these landforms (e.g., lobate-debris aprons; Holt et al., 2008; Plaut et al., 2009) and large regions of Utopia and Arcadia Planitia (Bramson et al., 2015; Stuurman et al., 2016). Additionally, a 1–10 m thick mantling unit thought to contain H₂O ice and dust over much of the midlatitudes has been observed (Mustard et al., 2001). This unit covers >23% of Mars, representing an ice reservoir of ~1.5–6 x 10⁴ km³ with very few craters, suggesting an age between 0.15 and 10 Ma (Mustard et al., 2001). Regional layering of this mantling unit has been observed and inferred to be due to multiple cycles of emplacement (Schon et al., 2009).

Numerous examples of “pasted-on,” smooth mantles are observed on pole-facing, midlatitude slopes (Carr, 2001; Christensen, 2003; Conway & Balme, 2014). These mantling units have been found to increase in thickness towards the pole in both hemispheres (from a mean of ~10 m at 30° to ~40 m at 60° latitude; Conway et al., 2018). Often, a raised curvilinear edge marks the upslope boundary of these mantles; this edge can be present at varying heights on the same slope (Bleamaster III & Crown, 2005; Carr, 2001; Christensen, 2003). At some locations, arcuate, moraine-like ridges are present downslope of the mantles (Arfstrom & Hartmann, 2005; Berman et al., 2005; Conway et al., 2018). While ice stability models for flat ground predict near-surface ice poleward of ~50° in both hemispheres, near-surface ice is expected to be stable on cold, pole-facing slopes poleward of ~30° (Aharonson & Schorghofer, 2006; Vincendon et al., 2010b). Indirect, morphometric measurements have suggested that these mantles contain 46–95% H₂O ice (Conway & Balme, 2014; Gulick et al., 2019), but no direct evidence of their ice content exists at present.
A wide variety of gully activity has been observed over the last two decades, with constrained activity occurring during winter and spring (Diniega et al., 2010; Dundas et al., 2010, 2012, 2015b, 2019; Harrison et al., 2009). The often-contemporaneous presence of seasonal/diurnal frosts has led to some authors (Diniega et al., 2010; Dundas et al., 2010, 2012, 2015, 2019) suggesting that these frosts might play a key role in this present-day activity. Most of this activity has been observed to take place on crater walls and sand dunes. However, the coarse, decameter-sized to bedrock materials (hereafter referred to as wall rock), and ice-rich units often present on crater walls may not be able to be eroded by frost sublimation processes alone (Dundas et al., 2019).

2. Evidence for Dusty H$_2$O Snow and Ice on Mars

While there is a plethora of evidence for H$_2$O ice at the midlatitudes, the nature and origin of this ice is unclear. Although other theories (e.g., Fisher, 2005; Sizemore et al., 2015) have been proposed, two primary contrasting theories of widespread ice deposition at the midlatitudes exist: (1) ice deposition into soil pores by the diffusion of water vapor (Mellon & Jakosky, 1995; Mellon et al., 2004) and (2) ice deposition by precipitation as dusty snow (Christensen, 2003; Jakosky & Carr, 1985; Jakosky et al., 1995; Madeleine et al., 2014).

To date, evidence for H$_2$O ice (~50 vol%) within soil pore spaces has only been observed at the Phoenix landing site (Cull et al., 2010; Smith et al., 2009). However, despite pore-filling ice dominating volumetrically at the Phoenix landing site (Mellon et al., 2009), evidence for H$_2$O ice exceeding the soil pore space has been found at numerous locations throughout the midlatitudes. For example, ice exposed by impact craters (Byrne et al., 2009), at Phoenix (Cull et al., 2010; Smith et al., 2009) and in the form of large-scale erosional scarps (Dundas et al., 2018) is composed of about 99% ice and 1% dust/sediment. These observations indicate that they represent exposures of dusty ice (<1% dust), rather than icy regolith (<25% ice). Therefore, the midlatitude mantle units are likely composed of dusty ice, with dissected portions of the mantle (Mustard et al., 2001) representing ice-rich units whose upper few meters have been desiccated (Christensen, 2003). Although variability in the precise ice contents at each location is expected, smooth, pasted-on mantles on pole-facing slopes are likely also remnants of this once widespread, continuous mantle composed of dusty ice stable under the layer of desiccated sediment (Christensen, 2003; Conway & Balme, 2014; Schon et al., 2009). In this paper, we present evidence of this dusty ice currently being exhumed in midlatitude gully alcoves.

3. Theory and Models of H$_2$O Snow and Ice

3.1. Snow Metamorphism on Mars

Freshly fallen snow typically changes its physical state due to water vapor diffusion, overburden pressure, and the formation of liquid water. These effects cause snow grains to grow, thereby reducing the air between grains and increasing the bulk density of the snow. When individual snow grains are no longer distinct, the snow has transformed into glacier ice (Cuffey & Paterson, 2010). While it is difficult to accurately determine the rates of snow metamorphism on Mars, estimates of Martian snow metamorphism rates imply timescales of decades/centuries (Clow, 1987; Kieffer, 1990) to millions of years (Bramson et al., 2017) for fresh snow (~50 kg/m$^3$ bulk density) to transform into solid, glacier ice (~917 kg/m$^3$ bulk density). The significantly longer timescales predicted by Bramson et al. (2017) are likely due to an underestimation of temperature gradients caused by radiative heating within exposed snow (see sections 3.2 and 3.3). However, when snow is buried by a dust lag, the relatively low thermal conductivity of the overlying dust will reduce the temperatures of the snow underneath, leading to lower rates of snow metamorphism. Thus, the true rates of metamorphism are likely to vary through the snow’s lifetime, and will depend greatly on local atmospheric and thermal conditions. Nevertheless, analysis of the only current in situ data of Martian excess ice at Phoenix (~68°N) suggests average grain radii of ~1–1.3 mm (Gyalay et al., 2019). This average grain radius corresponds to coarse-grained snow (not glacier ice), and similar results have been obtained in the northern midlatitudes (~38–52°N), where modeled ice densities of ~640 kg/m$^3$ (coarse-grained snow) match radar measurements (Bramson et al., 2017). In the absence of knowledge of the exact physical properties of Martian ice, we review the theoretical behavior of pure snow, dusty snow, and dense ice on Mars in the following sections.
3.2. Behavior of Pure H₂O Snow on Mars

Pure H₂O snow at the surface of Mars sublimates and cannot melt. For melting to occur, the snow must be at or above the melting point temperature (273 K) and there must be sufficient energy to replace the latent heat lost from vapor diffusion (Clow, 1987; Farmer, 1976; Ingersoll, 1970). Radiative heating can contribute toward satisfying the latent heat requirements for melting to occur. In the solar spectrum (∼0.2–2.5 μm), for clean H₂O snow (with no dust), the absorption coefficients of ice are very high for wavelengths >1.4 μm (Warren & Brandt, 2008). This absorption leads to near-surface (top few mm) heating rates that are higher than at visible wavelengths (0.4 < λ > 0.75 μm) by over an order of magnitude, despite only 15% of the incident flux occurring for λ > 1.4 μm (Clow, 1987). However, for visible wavelengths (accounting for ∼50% of the incident flux), although radiation can penetrate the snow very deeply (few tens of cm), the majority (~99%) is scattered back out (Clow, 1987; Warren & Brandt, 2008). This means that despite accounting for radiative heating rates, there is insufficient energy available to melt clean H₂O snow at any surface pressure below 600 mbar, even at temperatures above the triple point (Clow, 1987; Dundas & Byrne, 2010; Ingersoll, 1970; Mellon & Phillips, 2001).

3.3. Behavior of Dusty H₂O Snow on Mars and Issues with Martian Ice Models

Exposed snow/ice on Mars is not clean and typically contains <~1% of dust in it (Byrne et al., 2009; Dundas et al., 2018). The presence of these small amounts of dust within snow (<1%) increases the energy absorbed at visible wavelengths by 3–4 orders of magnitude, resulting in enhanced radiative heating at depth by a factor of up to 10 (Clow, 1987; Dang et al., 2015; Warren & Wiscombe, 1980). Larger snow grains (1–4 mm grain radius; 400–500 kg/m³ density) also increase the likelihood of photon absorption (Dang et al., 2015), which leads to greater radiative heating at depth, more than offsetting conductive losses (Mellon & Phillips, 2001) to the surface (Clow, 1987). Thus, the latent heat energy required to melt dusty snow is provided by increased radiative heating at depth caused by the presence of dust within relatively coarse-grained snow.

However, the critical role of dust within ice has not been explicitly accounted for in most models of ice on Mars (e.g., Bramson et al., 2017; Dundas & Byrne, 2010; Hecht, 2002; Ingersoll, 1970; Mellon & Phillips, 2001; Schorghofer, 2010; Schorghofer & Forget, 2012), despite being included in Earth ice models for decades (e.g., Dozier et al., 2009; Warren, 1984). The failure to account for absorption by dust within ice leads to an underestimation of heating rates at depth, and therefore insufficient energy for melting to occur. Additionally, Kite et al. (2011) point out that it is possible that the latent heat losses predicted by these kinds of models are overestimates (particularly at the scale of Martian snowpacks, when the effects of atmospheric circulation and near-surface air resistance are explicitly accounted for; Clow & Haberle, 1990, 1991), which also leads to less energy available for melting to occur.

Models that do explicitly account for radiative heating effects (Clow, 1987; Williams et al., 2008) predict that a dust content of only 0.1% is sufficient for melting to occur for a wide range of snow properties and atmospheric pressures. Melting is expected to occur 10–20 cm below the surface, despite surface temperatures far below the freezing point of water (Christensen, 2003; Clow, 1987). However, for liquid water to be stable, the local H₂O partial pressure must be at, or above the triple point pressure (611 Pa), which is much greater than the average Martian H₂O partial pressure of 1 Pa at the surface. How is this subsurface liquid water stable? Within the snow, the gas within the pores is saturated with water vapor (Neumann et al., 2009), and overlying snow/ice and dust can act as barriers against vapor diffusion to the atmosphere (Christensen, 2003; Clow, 1987; Farmer, 1976; Hecht, 2002; Hudson et al., 2007; Richardson & Mischna, 2005). Thus, despite evaporative latent heat losses, subsurface temperatures can reach the melting point (273 K), where the local H₂O partial pressure is saturated at 611 Pa, allowing liquid water to be stable. As a result, exposed, coarse-grained dusty snow can melt and sublimate in the midlatitudes today under these conditions (Christensen, 2003; Clow, 1987; Williams et al., 2008).

3.4. Behavior of Denser H₂O Ice

As noted in sections 3.2 and 3.3, the heating rates in snow are governed by the balance between incoming solar radiation being absorbed at depth and conductive heat losses. These heating rates can change when the density of snow increases through metamorphism. In particular, snow metamorphism causes the thermal
conductivity (also dependent on temperature) to rise, leading to greater conductive losses (Text S1 in the supporting information). For example, at temperatures between 200 and 220 K, the thermal conductivity of 550 kg/m$^3$ snow is about half (∼1.2–1.5 W/m·K) that of glacier ice (∼2.8–3.1 W/m·K; Figure S1). However, between ∼255 and 273 K, the thermal conductivity of 550 kg/m$^3$ snow (∼2.2–4 W/m·K) is essentially equal to, or greater than that of glacier ice because of vapor diffusion/latent heat contributions present in snow that diminish in solid glacier ice (Clow, 1987; Williams et al., 2008), whose conductivity drops from ∼2.29 to 2 W/m·K in this temperature range. But despite the elevated thermal conductivity of coarse-grained, 550 kg/m$^3$ snow, models predict that it can melt under current Martian summer conditions at the midlatitudes (e.g., Clow, 1987; Williams et al., 2008). This result suggests that when ground surface temperatures exceed ∼250 K (e.g., at the midlatitudes during summer), snow with densities >550 kg/m$^3$ (and glacier ice) can also melt, because denser snow (and glacier ice) has a lower thermal conductivity than less dense snow at these temperatures (Figure S1), meaning less heat is lost to conduction and more energy is available for melting.

An increase in snow density is also accompanied by a greater penetration depth for solar radiation, due to denser ice's significantly lower extinction coefficient at visible wavelengths, leading to radiative heating occurring deeper (∼1–5 m) in glacier ice than in snow (10–20 cm) (Brandt & Warren, 1993; Chinnery et al., 2020; Schwerdtfeger, 1969; Schwerdtfeger & Weller, 1967; Weller, 1969). The magnitude of this radiative heating is enhanced for denser ice because of its lower albedo (discussed in section 3.5). Because the radiative heating occurs deeper for ice than in snow, conductive heat transfer to the surface is less efficient in removing heat, resulting in greater amounts of energy being available for melting (Brandt & Warren, 1993; Liston & Winther, 2005). The presence of small amounts of dust (<1%) within the ice can further increase subsurface solar heating rates (as summarized in section 3.3). Although there are currently no models of Martian glacier ice that explicitly account for the key radiative effects outlined above, these results suggest that if Martian snow has metamorphosed into glacier ice, it can melt a few meters below the exposed ice surface, at temperatures well below freezing (Brandt & Warren, 1993; Liston et al., 1999; Liston & Winther, 2005).

### 3.5. Albedo of Dusty H$_2$O Snow and Ice

The spectral reflectivity, albedo, and appearance of snow can vary tremendously with grain radius, structure and the incorporation of dust (Dozier et al., 2009; Kayetha et al., 2007). Larger H$_2$O ice grains cause greater forward scattering and are more absorptive. Thus, the spectral albedo of snow decreases at all wavelengths as ice grain radii (and bulk densities) increase toward becoming glacier ice (Dang et al., 2015; Warren, 2019; Wiscombe & Warren, 1980). We computed the spectral variation of ice-dust mixtures between wavelengths of 0.5 and 1.5 µm (see Text S2 for details) using Mie theory (Matzler, 2002), a delta-Eddington radiative transfer model. The effect of dust (gray) on the spectral albedo of pure snow (black) is shown in Figure 1 (a) between 0.5 and 1.5 µm wavelengths and (b) sampled at HiRISE color filter center wavelengths. These computations are for a semi-infinite, 550 kg/m$^3$ (40% porosity) snowpack with snow and dust grain radii of 1,000 and 1.8 µm, respectively, and a solar zenith angle of 49.5°.
transfer model (Dang et al., 2015; Warren & Wiscombe, 1980) and refractive index data for H₂O ice (Warren & Brandt, 2008) and Martian dust (Wolff et al., 2009). Figure 1 shows that the inclusion of very small amounts of dust (<1%) within snow results in a reduction in albedo by about 50% at visible wavelengths. The addition of greater amounts of dust (>1%) can make dusty snow essentially indistinguishable from pure dust at visible wavelengths (Figure S2; Dang et al., 2015).

4. Methods

In order to search for potential occurrences of exposed, dusty ice we examined visible image data (~30 images) that had overlapping 1-km-wide central color swaths from the High Resolution Imaging Experiment (HiRISE; McEwen et al., 2007a) onboard the Mars Reconnaissance Orbiter during local spring/summer seasons at gully locations (Harrison et al., 2015) using JMARS (Christensen et al., 2009). At the Gully 1 location (Table S1), we also examined available Context Camera (CTX; Malin et al., 2007) data to document the behavior of the light-toned materials with time. The HiRISE (~0.25 m/pixel) filters are centered at 536, 694, and 874 nm (McEwen et al., 2007a), whereas the CTX (~6 m/pixel) filter is centered at 611 nm (Malin et al., 2007). At each location, we analyzed available images, calculated the Lambert albedo of the light-toned materials to assess their brightness as described below, and made qualitative comparisons with known ice (Brown et al., 2008; Byrne et al., 2009; Dundas et al., 2018; Table S2) and dust (e.g., Dundas, 2020; Table S3) exposures on Mars. In addition, Thermal Emission Imaging System (THEMIS; Christensen et al., 2004) infrared (~100 m/pixel) images (Table S4) were analyzed at all light-toned material gully locations to check for the potential presence of frosts near the time of HiRISE/CTX observations.

HiRISE radiometrically calibrated, map-projected Reduced Data Records (RDRs) were converted to I/F (measured intensity divided by solar flux at the top of the atmosphere) by applying scale factors to pixel values (Delamere et al., 2010). CTX Experiment Data Records (EDRs) were processed and radiometrically calibrated to I/F values using ISIS3 (Bell III et al., 2013; Gaddis et al., 1997). I/F values were then converted to Lambert albedo by dividing by the cosine of the incidence angle over the entire image (Hapke, 2012). This assumes that the surface is Lambertian (Bell III et al., 2013; Rice et al., 2018; Soderblom et al., 2006). For each region of interest (known ice exposure/light-toned material/dust) in HiRISE Infrared-Red-Blue (IRB) products, averaged pixel data from 5 × 5 pixel boxes (as shown in HiView http://www.uahirise.org/hiview/) were used. Albedo values are not topographic slope-corrected but only pixels near light-toned regions with similar slope characteristics were used for comparisons. No atmospheric correction or ratios were performed on the Lambertian albedo data.

The estimated uncertainty of HiRISE’s absolute calibration is ±20%, with a spectral calibration uncertainty of ~5% (Milazzo et al., 2015). The uncertainty in CTX measurements ranges from 10% to 20% (Bell III et al., 2013). In addition to the instrument calibration uncertainties, changing dust cover, standard image noise filtering, differences in viewing geometry and atmospheric conditions can cause variations in inferred Lambert albedo (Bell III et al., 2008; Daubar et al., 2016). However, despite these potential effects, HiRISE and CTX albedo estimates have been found to agree within 15% of measurements made in situ at the surface of Mars, suggesting that their radiometric calibration is validated (Rice et al., 2018). Additionally, our use of 25-pixel averages for HiRISE albedo values improves the signal to noise and reduces the probability of random noise effects. Regardless, we interpret all albedo estimates as approximate, and do not draw precise quantitative conclusions from the albedo data alone.

5. Observations

5.1. Gully/Mantle Morphology

Pasted-on mantles are present at a variety of heights from the top of the slope wall and often appear relatively smooth (Figure 2). The surface of these materials often contains blocks interpreted to be boulders (Dundas et al., 2018; McEwen et al., 2007b). The upslope boundary is typically curvilinear, and can appear lobate. Often, at the edges of the mantle, a depression within the wall rock is observed, suggesting that the pasted-on material lies in a shallow depression that has been eroded into the underlying substrate (Figure 2).
Gullies are formed in the mantle material in some cases (Figure 3), and if the mantles are indeed volatile-rich, then little true erosion (i.e., weathering and transport) would be required; the gullies could form simply by the sublimation of icy material (i.e., transport by vapor diffusion). However, gullies are also observed to have formed in the underlying wall rock (Figures 3b and 3d; note the transition from the mottled, rockier material above the mantle, referred to as 'texturally altered bedrock' by Conway et al. (2018), to the relatively smooth mantle downslope). Gullies eroded in wall rock materials appear shallower, and their U-shape is consistent with erosion into equally resistant topsoil (wall rock) and subsoil (wall rock) (Weidelt, 1976). Gullies eroded in the mantle are often V-shaped (Figures 2 and 3), consistent with a subsoil (wall rock) that is more resistant than the topsoil (mantle) (Conway et al., 2018; Weidelt, 1976). This observation is especially pronounced in smaller, earlier-stage gullies, before gully alcoves grow into broader, more developed alcoves (Figure 4). These channels that are eroded into the wall rock can be seen continuing into the mantle (see Figure 14 in Harrison et al. (2015) for additional examples of similar wall rock-mantle gullies), implying that the same process eroded the wall rock and the mantle (Figure 3).

5.2. Context for Light-toned Material

Figure 4 shows examples of mantle association with wall rock and gullies that host light-toned materials. Gullies in Dao Valles are eroded into two different substrates (Figure 4a). Towards the west, gullies can be seen eroded into wall rock. However, to the east, there is a depression within the smooth mantle with gullies emerging from within the depression. One set of gully channels can even be seen eroded into the wall rock below the mantle, where the mantle is the thinnest, before continuing into the thicker mantle downslope. Multiple channels can also be seen eroded into the mantle in Figure 4b, with some channels cross-cutting each other and combining to form larger gullies. In Figure 4c, some alcoves to the west appear to be coalescing to form one larger alcove. Gullies generally begin at a relatively uniform height at these locations.

Partially exposed, light-toned materials are present within gullies eroded into the mantle in summertime imagery (Figures 4 and 5). The light-toned materials are present in patches that range from ~1 to 40 m in length and generally appear on west-facing alcove walls that show evidence of slumping. Although gullies eroded into wall rock are often nearby (Figure 4a), light-toned materials are not observed within wall rock gullies, or elsewhere in the scene. While there are few polygonal fractures within the alcove walls where the
Figure 3. Channels that erode the bedrock extend downslope into the pasted-on mantle. Erosion within the wall rock is significantly narrower than in pasted-on materials, likely due to the difference in mechanical strength between the two materials. (A and B) HiRISE image PSP_007526_1435, centered at 36.3°S, 178.6°E. (C and D) HiRISE image ESP_014329_1435, centered at 36°S, 199.4°E.
Figure 4. Context for light-toned materials within some gullies. Light-toned materials are partially exposed only within the mantle. (A) Gullies are eroded in the mantle adjacent to gullies eroded in the wall rock at 32.9°S, 93.2°E (ESP_013067_1470). (B) Multiple gullies are eroded in the mantle, with some channels continuing into the wall rock at 36°S, 199.4°E (ESP_014329_1435). (C) Gullies eroded to varying depths into the mantle at 37.8°S, 217.9°E (ESP_048824_1420). Black boxes indicate locations shown in Figure 5.
light-toned materials are exposed (suggesting that the material is relatively fresh), in some cases fracturing is visible in the mantle nearby (Figure 5b).

5.3. Light-Toned Material Albedo

In stretched, false-color IRB images, known exposures of subsurface H₂O ice on Mars (Figures 6a–6c and S3) generally appear blue/white in relation to nearby lithic material although there are areas of the icy scarps that resemble the color of nearby lithic material (Figures 6a and b). These icy materials have higher albedo values at each HiRISE filter wavelength relative to their surroundings (corresponding to an increase in brightness in the stretched images), although the magnitude of the difference can vary from relatively low (Figure 6a) to high (Figure 6c). The ice spectra generally show a slight reduction in slope between the 694 and 874 nm filters relative to the nearby material. Apart from this difference in spectral slope, the overall shape of the ice spectra resembles that of the nearby materials, but with greater albedo values. Note that only pixels near with similar slope characteristics were used for spectral comparisons between different groups of materials.

The light-toned materials within gullies (Figures 6d–6f and S4) generally also appear blue/white relative to their surroundings. Like the icy scarps, some parts of the light-toned material can resemble their surroundings in color. These light-toned materials also have higher albedos in comparison with adjacent areas, with varying relative magnitudes. The spectral shape of light-toned materials relative to their surroundings resembles the known ice exposure spectra, although the albedo increase at the 536 nm filter can sometimes be relatively small (e.g., Figure 6f).

In contrast, relatively dusty surfaces (Figures 6g–6i and S5) appear more yellow/gray than their darker surroundings. While the presence of dust also increases the albedo at visible wavelengths, the spectral slope is relatively steep. This steep spectral slope caused by dust is consistent with terrestrial-based experiments by Wells et al. (1984) showing increases in surface reflectance at longer visible wavelengths by dust deposition, i.e., the albedo difference at the 536 nm filter is usually negligible, suggesting that the light-toned materials observed within gullies (Figures 6d–6f and S4) are not dust. It is interesting to note that in all cases (Figures 6, S3–S5), despite the differences in color between the surroundings and the region of interest (ice, light-toned material, and dust), the spectral shapes of the surroundings are somewhat similar to the region.
of interest, but lower in albedo. This result suggests that each region of interest might contain some portion of the surrounding material.

Assessing the behavior of known exposures of ice, light-toned materials within gullies and relatively dusty areas qualitatively using HiRISE data suggests that the light-toned gully materials are more similar to the known exposures of ice than they are to pure dust, and potentially represent a mixture of ice and dust. The evolution of ice exposed at the Phoenix landing site lends support to this ice-dust mixture possibility (Figure 7). After the ice was first exposed on Sol 9, its albedo on Sol 28 (Figure 7a) was significantly higher than that of nearby soil and there was a reduction in spectral slope between 694 and 874 nm (dashed blue line in Figures 7c and 7d). As noted in sections 2 and 3, the ice spectrum is consistent with 1–1.3 mm ice grains, with <1% dust. The observed spectral behavior is similar to the known ice exposures seen in HiRISE data, despite potential differences in instrumentation and viewing conditions. Fifty-four sols later, the ice was not as bright (Figure 7b), and its spectral shape resembled that of soil nearby (solid blue line in Figures 7c and 7d). Furthermore, the reduction in spectral slope was no longer present, probably because the amount of dust/lithic materials within the ice-dust mixture had increased. Qualitatively similar spectral behavior is observed in the light-toned materials present within Gully 12 (Figure 8) over six Mars Years. Initially, the light-toned materials had a significantly higher albedo than nearby materials (Figure 8a). The color of the light-toned materials was white/blue with a slight yellowish tint, and both

Figure 6. HiRISE Lambert albedos of (A–C) known exposures of ice on Mars, (D–F) light-toned materials within gullies, and (G–I) dusty materials compared to nearby less dusty regions. (A and B) Icy scarps (Dundas et al., 2018); ESP_040772_1215 and ESP_022389_1230. (C) Ice-exposing impact crater (Byrne et al., 2009); PSP_010625_2360. (D–F) Light-toned materials within Gully 5, Gully 10, and Gully 1; ESP_032012_1415, ESP_048824_1420, and ESP_013067_1470. (G) Slope streak exposing dark substrate, likely from dust avalanching (Dundas, 2020; Sullivan et al., 2001); ESP_058424_2035. (H) Mars Science Laboratory (bright blue dot to the west in the image) landing site a few sols after landing. The landing process blows off dust, exposing a darker substrate around the rover near a relatively dusty area to the east; ESP_028335_1755. (I) Slope streak exposing dark substrate, likely from dust avalanching (Dundas, 2020; Sullivan et al., 2001); ESP_046740_2175. All images are HiRISE false-color (IRB) and individually contrast-stretched. Note the different y axis limits in Figures 6c and 6i. Approximate locations of spectra are shown in boxes (not to scale). White arrows are shown in A and B to help locate blue box locations.
types of materials showed an increase in spectral slope between 694 and 874 nm. These color and albedo results suggest that dust was present throughout the scene (e.g., Figure 6). After six Mars Years, the light-toned materials were still present, but in a smaller area. The difference in albedo between the light-toned materials relative to the nearby materials decreased by a factor of $\sim 3$, which is comparable to the factor of $\sim 4$ decrease in the relative albedo at the Phoenix site. In addition to these similarities in albedo decrease over time, the spectral shape of the light-toned gully materials (e.g., Figures 6d, e, 8b and S4a, S4e, S4h) is consistent with that of the relatively dusty ice at Phoenix (solid blue line in Figure 7d). Hence the gully materials could represent exposures of H$_2$O ice that is relatively dustier than previously recognized Martian ice exposures.

The exact proportion of ice to dust is difficult to determine because numerous factors can affect the spectral behavior of small areas in HiRISE data (as discussed in section 4) and there are too many unconstrained parameters to accurately model the ice-dust ratios from the observed albedos. However, a qualitative assessment based on spectral shape can be made. For example, the overall shape of the known ice spectra (Figures 6a–6c and S3) is similar to the modeled ice spectra containing 0.1% dust (violet curve in Figure 1). This result is consistent with Compact Reconnaissance Imaging Spectrometer for Mars (CRISM; Murchie et al., 2007) data suggesting dust contents of <1% within scarp ice (Dundas et al., 2018). The overall shape of the light-toned gully material spectra can vary between that of the icy scarps (0.1% dust in modeled spectra) in some cases (e.g., Figures 6d and 6e) and modeled spectra containing <1% dust (Figures 1 and S2) at other locations (e.g., Figure 6f), suggesting that the light-toned gully materials are also consistent with high ice contents.

Figure 7. Spectral behavior of exposed H$_2$O ice at the Phoenix landing site compared to nearby soil over time. (A) Approximate locations of ice (blue box) and nearby soil (red box) spectra on Sol 28 of the mission. (B) Approximate locations of ice (blue box) and nearby soil (red box) spectra on Sol 82 of the mission. (C) Spectra of exposed ice on Sol 28 (dashed blue line) and Sol 82 (solid blue line), compared to nearby soil spectra (red line). (D) Spectra shown in (C), sampled at HiRISE filters. All spectra shown are from Blaney et al. (2009). Surface Stereo Imager (SSI) color images (RGB = 672, 533, 485 nm) derived from ss028rad896899390_132r1r1m1 and ss082rad903481988_196c0r1m1 are shown in A and B, respectively. Observations were collected at the same time of day to minimize viewing geometry, illumination, and calibration differences. The soil spectra did not change (Blaney et al., 2009).
5.4. Changes in Gullies with Light-toned Material

The evolution of light-toned materials within Gully 1 is documented with CTX imagery ∼250 sols before they were exposed (Figure S6a), during the exposure (Figure S6b), and after they have faded (Figure S6c) over three Mars Years (Table S4). Ratios of the CTX Lambert albedo of the light-toned material to that of nearby materials for each image show that the light-toned material was 5% brighter than nearby materials when it was exposed. The change between the three images is localized and discrete, suggesting that the changes are not due to atmospheric effects or instrument noise.

Figure 9 and supplementary animations 1–3 show numerous changes observed in Gully 1 over six Mars Years (MYs). First, the light-toned materials present in MY 29 (Figures 9a, 9c, and 9e) are no longer visible in MY 35 (Figures 9b, 9d, and 9f). Second, there are multiple locations within the gully alcove where slumping has occurred, with the largest (∼50 m in length) slump at the top of the alcove (black arrow in Figures 9a and 9b). To check whether the appearance of the ∼50 m long slump is caused by differences in lighting to first order (see section 4.1 in Cushing et al., 2015), we measured the width of the slump in Figure 9b. We
obtained a slump width of ∼10 m. Assuming that the slump is actually a shadow cast by a ridge, we obtain a ridge height of ∼10 meters for an incidence angle of 46.1° (Figure 9b), which would cast a ∼9 m shadow for an incidence angle of 41.8° (Figure 9a). However, no such ∼9 m shadow is apparent in Figure 9a, indicating that this ∼50 m slump represents a topographic change of ∼10 m; the effect of a ∼4° difference in emission
angle between the two images is small. Additional, smaller-scale (∼10 m length) slumps might also have occurred at various points within the mantle (black arrows in Figures 9e–9f). Alternatively, existing slopes and pit edges may have moved slightly. Third, topographic depressions (Figure 9e) that seem to be present in MY 29 appear to have deepened, although it is possible that slightly different solar illumination angles cause this apparent deepening effect in some cases.

6. Discussion

6.1. Is the Light-toned Material Dust or Ice?

One hypothesis for the light-toned material having a higher albedo than nearby materials is that it is dust. Dust likely deposits onto all of these surfaces, but it seems unlikely that a highly localized, short-term deposit of dust would form. In addition, this light-toned material occurs only within alcoves eroded into snow-rich mantles, and not nearby rocky walls or surfaces. We suggest that a more plausible explanation is that the light-toned materials are exposed H₂O ice. It may be possible for dust to accumulate within these alcoves as a lag being left behind after the ice disappears (by sublimation and/or melting). However, the appearance, and then subsequent disappearance of these light-toned materials suggests that they are some form of volatile, such as dusty ice, rather than dust alone. Since these light-toned materials are exposed in relatively small (∼20 m) patches, they cannot be confidently resolved by sparsely available CRISM data. However, the appearance (in HiRISE color data) of these light-toned materials is similar to the >100 m thick, light-toned ice deposits exposed by steep midlatitude scarps (Dundas et al., 2018), indicating that these materials are probably also ice, with some amount of dust on, and within the ice.

6.2. Is the Ice Frost or Exposed Subsurface H₂O Ice?

Light-toned materials present within and around midlatitude gully alcoves have been attributed to surficial frosts (H₂O and CO₂) (e.g., Dundas et al., 2019; Levy et al., 2009b), especially during winter and spring imagery. These frost observations have been confirmed by CRISM data at numerous southern midlatitude active gully sites (Vincendon, 2015). However, the exposed ice seen in Figures 4–8, S4, and S6 is unlikely to be persistent frost in equilibrium, due to several observations: (1) The HiRISE data were obtained during late spring/summer afternoon, when frost is not expected at these latitudes (Carrozzo et al., 2009; Schorghofer & Edgett, 2006; Vincendon et al., 2010a). (2) The morphologies of the exposed ice are reminiscent of exhumation rather than surficial frosts, similar to the recent discovery of much larger-scale, exposed H₂O ice sheets by Dundas et al. (2018). (3) THEMIS infrared data (100 m/pixel) indicate that the gully alcoves containing exposed ice generally have late-afternoon spring/summertime temperatures >240 K (Table S4), which are greater than the predicted frost points for CO₂ (150 K; Piqueux et al., 2016) and H₂O (210 K; Schorghofer & Aharonson, 2005). Thus, the ice must be present in the subsurface, and is exposed by slumping of overlying material. Based on the morphology, albedo and setting of the ice, we conclude that dust is likely to be within and overlying the ice. These observations of exposed subsurface H₂O ice therefore represent the lowest latitude (e.g., 32.9°S) detections of H₂O ice to date, and agree with model predictions that near-surface ice is stable on cold, pole-facing slopes poleward of ~30° (Aharonson & Schorghofer, 2006; Vincendon et al., 2010b).

6.3. Implications for Gully Formation

The observation of dusty H₂O ice within midlatitude mantling deposits that often have gullies eroded into them has implications for the processes that can erode the ice, and sometimes the underlying wall rock. In the following sections, we review and present some models of gully erosion and discuss the possibility of applying them to gully erosion in the H₂O ice-rich mantle and the wall rock beneath.

6.3.1. CO₂ Frost Model

Sand dunes, gully fans and existing gully alcoves/channels (e.g., Figures 6 and 7 in Dundas et al., 2015b) are usually sites of the most significant present-day gully activity (Dundas et al., 2019). Aeolian ripples are often observed within such materials, indicative of their unconsolidated nature. Morphological studies (Diniega
et al., 2010; Dundas et al., 2010, 2012, 2015b, 2019) and laboratory experiments (Sylvest et al., 2016, 2019) have shown that CO₂ frost processes are excellent candidates for transporting these unconsolidated, granular materials. However, we consider them unlikely to erode gullies into the mantle and the wall rock beneath because of the far greater erosive power needed to erode H₂O ice or rock than transport loose, granular material. In addition, even if CO₂ frost processes can cause boulders to fall and/or be transported on steep slopes in the form of debris flows (Dundas et al., 2019), to date, these sorts of flows have not caused any observable erosion into wall rock, although it is possible that the erosion is too minor to be resolved by HiRISE. Thus, other processes seem likely to explain the erosion of H₂O ice or rock at gully locations.

First, we consider rock erosion. Experimental studies (Viles et al., 2010) and in situ data (Eppes et al., 2015) have shown that thermal cycling can potentially induce small reductions in strength (∼1%) and form cracks in Martian rocks. However, the presence of numerous large rocks on ~billion-year old surfaces (Fergason et al., 2006; Golombek et al., 2006) near the equator, where solar insolation and thermal breakdown processes are most effective (Eppes et al., 2015; Viles et al., 2010) argues against the efficacy of thermal cycling to produce unconsolidated, granular materials for CO₂ frost to transport. Thermal cycling might cause initial weakening (∼1%), after which rock surfaces reach equilibrium, with little additional weathering after initial weakening (Viles et al., 2010). At the midlatitudes, rock breakdown might also occur from CO₂ frost sublimation-freeze cycling and ice wedge formation (Sizemore et al., 2015), but the efficacy of these processes under Martian conditions is presently unknown. Another possible rock breakdown mechanism for gullies formed on crater walls is the impact itself, which can create talus for CO₂ frost to transport. Although this mechanism could be applicable in isolation, most midlatitude slopes are mantled by H₂O ice, which cannot be eroded by CO₂ frost. Additionally, about 35% of all Martian gullies have been observed on noncrater wall slopes (on isolated knobs, valley walls, etc.; Harrison et al., 2015) where the effects of impact events will not apply. Thus, although the CO₂ frost model cannot be ruled out, some additional processes might still be required to explain how gullies are eroded into H₂O ice and wall rock. In the sections below, we discuss two models for erosion into H₂O ice and wall rock, that we term the “H₂O Ice Sublimation Model” and the “H₂O Ice Melt Model.”

### 6.3.2. H₂O Ice Sublimation Model

To explain erosion into the mantle and the wall rock, some authors (Dundas et al., 2018, 2019; Forget et al., 2016) have suggested that the removal of overlying dust (e.g., by CO₂ processes) can induce the sublimation of H₂O ice below, eroding gullies into the mantle in an alternating sequence of overlying dust removal and H₂O ice sublimation (which frees up any lithics within the ice for transport, as it ablates). However, ice sublimation typically occurs unevenly due to the nature of vapor diffusion (Law & Van Dijk, 1994;
Crystal structure defects act as centers for ice sublimation and are greatly influenced by localized thermal gradients (Law & Van Dijk, 1994). Thus, if the overlying dust is removed and the H$_2$O ice begins to sublimate, then variations in local solar insolation, surface geometry, and ice exposure will dictate the surficial morphology and cause it to erode unevenly. Therefore, sublimation processes in volatile-rich materials form scalloped depressions or pits (Mangold, 2011a, 2011b). Evidence for volatile sublimation can be seen in (a) quasi-circular depressions in the south polar cap (Byrne & Ingersoll, 2003), (b) the “degraded” portions of the midlatitude mantle (Mangold, 2011a; Mustard et al., 2001; Schon et al., 2009), (c) in ice-rich midlatitude materials (Dundas et al., 2015a, 2018; Harrison et al., 2017; Viola & McEwen, 2018; Zanetti et al., 2010) as well as (d) cross-cutting fractures in the form of coalescing pits sometimes associated with gullied terrain (Dickson et al., 2015).

These sublimation processes typically progress downwards through volatile-rich materials, rather than propagating into a channel. Even on slopes, these sublimation features appear to retain their quasi-circular expression (Figure 10), suggesting that sublimation processes play a limited role in the formation of the relatively well-defined V-shaped gully incisions eroded into the mantle.

Additionally, the observation of the same channels eroded into the mantle and wall rock seems unlikely to be explained by the combination of proposed CO$_2$ frost processes and H$_2$O ice sublimation. This is because there is probably little to no H$_2$O ice within the wall rock, and if H$_2$O sublimes, it will diffuse upwards into the relatively dry and heavy (m$_{CO_2}$ > m$_{H_2O}$) atmosphere (Blackburn et al., 2010; Ingersoll, 1970; Sears & Moore, 2005) rather than downwards into wall rock. While CO$_2$ frost processes might be capable of mobilizing loose materials, and potentially trigger boulder-filled flows, we consider CO$_2$ frost unlikely to erode wall rock unless the wall rock is assumed to be unconsolidated and granular, which is not expected for most Martian slopes (see section 6.3.1). Furthermore, we propose that the sublimation of H$_2$O ice will likely lead to uneven mass loss by vapor diffusion. Thus, neither process seems likely to be able to erode both materials. However, H$_2$O sublimation could potentially still play a role in the growth of some gully alcoves (Dundas et al., 2018, 2019).

### 6.3.3. H$_2$O Ice Melt Model

We propose a model for gully formation based on Christensen (2003): (1) Several centimeters of dusty snow are deposited at the midlatitudes during periods of high obliquity (Jakosky & Carr, 1985; Madeleine et al., 2014). (2) As the snow sublimes and/or melts to form gullies within the snow (often characterized by a V-shaped channel), it builds up a protective lag layer of dust, as is observed on melting snowfields on Earth (e.g., Higuchi & Nagoshi, 1977). The formation of this lag, caused by snow ablation and possible dust storms (Madeleine et al., 2014; Williams et al., 2008) ceases the formation of gullies since the snow is buried and no longer exposed to sunlight. (3) Throughout these processes, the remnant dusty snow is probably metamorphosing into denser, coarse-grained ice. (4) Then, slumping within steep gully alcove walls by aeolian/frost processes leads to partial ice re-exposure. This exposed dusty ice absorbs solar radiation deeper than fresh snow, and continues to melt in the subsurface while eroding channels within the ice. If sufficient meltwater is available at the base of the ice layer, channels can be eroded into the wall rock beneath. (5) When all the ice disappears from melting and sublimation, the dust within the ice is left behind. Steps (1) to (5) repeat over multiple obliquity cycles, leading to layers of dust and snow building up within the mantle, and gullies forming deeper channels each time ice is exposed, and melts. Evidence for this ice-dust layering within the mantle is supported by modeling results (Madeleine et al., 2014), exposures of mantle stratigraphy within steep scarps (Dundas et al., 2018) and observations of layers within the mantle (Schon et al., 2009).

In this model, small amounts of runoff (~1 mm/day) at the midlatitudes can be produced during the summer (~50 sols) each year, potentially even under present conditions (Clow, 1987; Williams et al., 2008). Seasonal melting and refreezing can concentrate mechanical weathering at the locations where runoff can remove debris, leading to high rates of erosion despite small amounts of water (termed the Eisrinde effect on Earth; Baker, 1985; Büdel et al., 1982). Additionally, if Martian slope materials are already unconsolidated (as discussed in section 6.3.1; by thermal cycling, impact events, CO$_2$ frost processes, etc.), then freeze-thaw action can break them down further, leading to debris flows and/or slurries. Gullies can form
on thousand-year timescales, with ≈40 m$^3$ of liquid water available for each channel, assuming typical gully alcove size and spacing (200 × 200 m$^2$) and 1 mm of meltwater per day (Christensen, 2003).

A H$_2$O ice melt model can explain many gully and mantle characteristics. Melting is favored on pole-facing slopes at lower latitudes, and equator-facing slopes at higher latitudes due to optimal solar insolation for melting shifting to equator-facing slopes poleward of ∼45–60° latitude (Costard et al., 2002). This is consistent with gullies losing their pole-facing preference, and transitioning to being present on equator-facing slopes at higher latitudes (Harrison et al., 2015). In addition, the small amounts of melt formed in near-freezing conditions implies a low rate of chemical weathering and mineralization, which is consistent with the lack of hydrated minerals at gully sites (Núñez et al., 2016).

The observation of gullies eroded into the wall rock that continue into the mantle is also consistent with H$_2$O ice melt. If the mantle was deposited as dusty snow on slopes, it probably extended further upslope and a single process eroded the mantle and the wall rock. As the mantle was removed, the underlying wall rock gully was exposed. This underlying wall rock gully formed when meltwater percolated downwards and was available for erosion. Where wall rock erosion is not visible today, there may have been insufficient meltwater production and/or the liquid water did not reach the substrate for erosion.

However, H$_2$O ice melt is not expected to occur at all locations of snow accumulation, such as where the amount of snow accumulation is too low (near the equator) or where conditions are too cold for melting to occur (>70° latitude). Melting might also cease to occur when the snow becomes too dusty (greater than a few percent) near the surface, because light will not be able to penetrate deeply and cause subsurface heating. Furthermore, H$_2$O ice melt cannot explain erosion where there are no sources of snow, in addition to present-day gully activity occurring in the winter. Thus, the formation of midlatitude Martian gullies is probably not caused by one single mechanism and is likely dependent on a variety of factors: location, substrate, topography, season, atmospheric conditions, and obliquity period. The influence of microclimates on individual gully morphology can also play a significant role in gully formation. Gully formation and activity on Mars is therefore likely to be caused by a mixture of many different processes that may not be mutually exclusive.

7. Conclusions

Here we present novel evidence consistent with dusty, water ice deposits present within gullies eroded into midlatitude mantles. Our observations likely represent the lowest latitude (e.g., 32.9°S) detections of H$_2$O ice to date, in agreement with numerical models that predict stable near-surface ice on pole-facing slopes at latitudes poleward of 30°. These ice deposits are like the materials documented within icy midlatitude scarps, but with more dust and exposed in smaller patches by slumping. In addition, we observe gullies eroded in wall rock that continue into the mantle, implying that the same process erodes the wall rock and the mantle. We propose a H$_2$O ice melt model that can explain gullies eroded in the wall rock and the mantle. This H$_2$O ice melt model is supported by numerical simulations that show that relatively dense H$_2$O snow on Mars only melts when it contains small amounts of dust. Although the precise physical properties (grain radius, density, etc.) of H$_2$O ice on Mars are unknown, the observed exposure of dusty H$_2$O ice provides a mechanism for it to melt under some conditions and form some gullies. Subsurface H$_2$O ice melt could potentially be even occurring today, and access to liquid water within the ice could provide potential abodes for any extant life. However, H$_2$O ice melt is not expected to occur at all locations of H$_2$O ice on Mars, and Martian gullies are likely to form by a combination of many different processes (e.g., CO$_2$ frost, H$_2$O ice sublimation, H$_2$O ice melt, etc.). Future work to model the formation and evolution of H$_2$O ice in the Martian midlatitudes and refine the conditions under which Martian H$_2$O ice melt is possible is currently ongoing.

Data Availability Statements

HiRISE imagery is available at https://hirise.lpl.arizona.edu/. CTX imagery can be accessed at http://ode.rsl.wustl.edu/mars/. THEMIS data can be accessed using JMARS, available at https://jmars.asu.edu/. Phoenix Surface Stereo Imager (SSI) data is available at https://pds-imaging.jpl.nasa.gov/data/phoenix/phx-
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The authors declare that they have no competing financial interests.

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