Impact ejecta-induced melting of surface ice deposits on Mars

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**Abstract**

Fluvial features present around impact craters on Mars can offer insight into the ancient martian climate and its relationship to the impact cratering process. The widespread spatial and temporal distribution of surface ice on Mars suggests that the interaction between impact cratering and surface ice could have been a relatively frequent occurrence. We explore the thermal and melting effects on regional surface ice sheets in this case, where an impact event occurs in regional surface ice deposits overlying a regolith/bedrock target. We provide an estimate for the post-impact temperature of martian ejecta as a function of crater diameter, and conduct thermal modeling to assess the degree to which contact melting of hot ejecta superposed on surface ice deposits can produce meltwater and carve fluvial features. We also evaluate whether fluvial features could form as a result of basal melting of the ice deposits in response to the thermal insulation provided by the overlying impact ejecta. Contact melting is predicted to occur immediately following ejecta emplacement over the course of hundreds of years to tens of kyr. Basal melting initiates when the 273 K isotherm rises through the crust and reaches the base of the ice sheet ~0.1 to ~1 Myr following the impact. We assess the range of crater diameters predicted to produce contact and basal melting of surface ice sheets, as well as the melt fluxes, volumes, timescales, predicted locations of melting (relative to the crater), and the associated hydraulic and hydrologic consequences. We find that the heat flux and surface temperature conditions required to produce contact melting are met throughout martian history, whereas the heat flux and surface temperature conditions to produce basal melting are met only under currently understood ancient martian thermal conditions. For an impact into a regional ice sheet, the contact and basal melting mechanisms are predicted to generate melt volumes between $10^{-3}$ and $10^{4}$ km$^3$, depending on crater diameter, ice thickness, surface temperature, and geothermal heat flux. Contact melting is predicted to produce fluvial features on the surface of ejecta and the interior crater walls, whereas basal melting is predicted to produce fluvial features only on the interior crater walls. Before basal melting initiates, the ice-cemented cryosphere underlying the crater ejecta is predicted to melt and drain downwards through the substratum, generating a source of water for chemical alteration and possibly subsurface clay formation. These candidate melting processes are predicted to occur under a wide range of parameters, and provides a basis for further morphologic investigation.

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1. Introduction

The wide array of fluvial features present on the surface of Mars, despite its current below-freezing surface temperatures, has raised many questions regarding the climate of Mars throughout its early history. Fluvial channels incised onto impact crater walls, rims, and ejecta (e.g., Craddock and Maxwell, 1993; Craddock and Howard, 2002; Howard, 2007; Morgan and Head, 2009; Mangold, 2012; Mangold et al., 2012b; Hobbs et al., 2016) are particularly interesting because these features may offer insight into the conditions of the ancient martian climate, and its relationship to the impact cratering process. Fluvial channels have been reported around the rims and ejecta facies of impact craters spanning the entire martian geologic record. These include (1) Amazonian- and Hesperian-aged craters which exhibit fluvial channels superposing the crater walls and ejecta facies (Fig. 1A) (Morgan and Head, 2009; Howard and Moore, 2011; Jones et al., 2011; Mangold, 2012; Mangold et al., 2012a; Schon and Head, 2012; Hobbs et al., 2016). These channels are generally isolated with poor connectivity, and range from locally sinuous to wide and braided (Mangold et al., 2012a). (2) Relatively older early Amazonian- and Hesperian-aged closed-basin-lakes (CBLs) (Cabrol and Grin, 1999), which are craters that exhibit inlet channels superposed on the rim-crest (Fig. 1A). These inlet channels typically exhibit an amphitheater-shaped headwall...
Fig. 1. Examples of fluvial channels associated with impact craters on Mars. (A) A 26 km diameter closed basin lake (CBL) exhibiting inlet channels on the rim (red and white arrows) and fluvial channels superposing the ejecta (blue and white arrows), and the ejecta of a younger nearby 12 km diameter crater characterized by Mangold (2012) (6.6° E, 35.3° N). (B) A highly degraded Noachian-aged crater exhibiting numerous fluvial channels along the rim characterized by Mangold et al. (2012b) (59.1° E, 18.8° S). (C–F) Fluvial features from (A). (A) CTX images B01_009892_2148, P21_009325_2160, D15_032968_2170, and D15_033047_2171 superposed on THEMIS IR global day. (B) THEMIS IR global day. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
with one short (~10 km long) channel that drains into the crater interior (Fig. 1A) (Goudge et al., 2015). (3) The ancient and highly degraded Late Noachian highland craters, which exhibit fluvial features superposing highly subdued rims (Fig. 1B) (Masursky et al., 1977; Craddock and Maxwell, 1993; Craddock et al., 1997; Craddock and Howard, 2002; Forsberg-Taylor et al., 2004; Mangold et al., 2012b; Grant et al., 2015; Hobbs et al., 2016). With the exception of degraded craters in close proximity to, or in contact with valley networks (e.g., Fassett and Head, 2008; Fig. 3 in Hoyle and Hynne, 2009), the fluvial features associated with the degraded Late Noachian highland craters (including some CBLs with longer, branching tributaries (Fig. 1B) (Goudge et al., 2015) appear to drain into the crater interior (Mangold et al., 2012b).

These different fluvial features have been variously explained through several processes. For example, the relatively younger Amazonian- and Hesperian-aged fluvial features associated with impact crater ejecta have generally been attributed to contact melting, wherein hot ejecta deposited on surface or near-surface icy deposits generates melting (Morgan and Head, 2009; Jones et al., 2011; Mangold, 2012; Mangold et al., 2012a; Schon and Head, 2012). The fluvial features present on the rims of Amazonian- and Hesperian-aged CBLs remain of uncertain origin, but have been proposed to form from overland flow generated by regional flooding events (Goudge et al., 2015). The older Late Noachian fluvial features associated with impact craters have typically been attributed to rainfall and fluvial erosion in a warmer and wetter early martian climate (Craddock and Maxwell, 1993; Craddock et al., 1997; Craddock and Howard, 2002; Forsberg-Taylor et al., 2004). Noting that a warm early martian climate (and sustained rainfall) is not predicted by recent 3D global climate models (Forget et al., 2013; Wordsworth et al., 2013, 2015), these fluvial features have been alternatively explained by snowmelt and fluvial erosion (Hobbs et al., 2016), from snow deposition on hot ejecta (Kite et al., 2011), or top-down melting during peak seasonal or daytime temperatures (Head and Marchant, 2014) in a cold and icy Late Noachian climate. Recent work (Weiss and Head, 2015) has proposed that impact-induced contact and basal melting of surface ice are candidate processes which may contribute to some of the ancient impact-related fluvial features in such a cold and icy ancient martian climate.

Here we investigate the quantitative characteristics of impact ejecta-induced melting of surface ice deposits. We examine the mechanisms of contact melting (Mangold, 2012) and basal melting (Weiss and Head, 2015) of surface ice deposits in order to assess their roles in the formation of impact-associated fluvial channels. In the contact melting scenario (Fig. 2), ejecta is at elevated temperatures due to a combination of pre-impact geothermal heating at the depth from which it is excavated, and shock heating during the impact. When the ejecta is emplaced on the surface snow and ice deposits (Fig. 2B), the hot ejecta radiates heat outwards and conducts heat downwards into the icy deposits, thereby generating meltwater. In the basal melting scenario (Fig. 2), ejecta deposition on top of regional surface snow and ice deposits inhibits geothermal heat diffusion through the ice. As a result, following the impact event the 273 K ice melting isotherm within the shallow crust is predicted to rise to the base of the ice sheet given sufficient ejecta thicknesses (Fig. 2D and E). This causes the ice sheet to melt from the bottom-up, supplying a potential source of liquid water for fluvial erosion proximal to the impact crater. These mechanisms appear attractive within the constraints of the current 3D climate models for the Late Noachian (e.g., Wordsworth et al., 2013) because they do not require warm atmospheric temperatures (e.g., the rainfall hypothesis for Late Noachian craters (Craddock and Maxwell, 1993; Craddock et al., 1997; Craddock and Howard, 2002). For example, contact and basal melting could operate as a background landscape/crater degradation processes in a cold and icy early Mars (Weiss and Head, 2015) even in the absence of punctuated warming events (e.g., Haley and Head, 2014; Wordsworth et al., 2015).

The source of water (surface snow and ice) in the contact and basal melting hypotheses has been shown to be readily available throughout martian history. This includes, for example:

(1) The Amazonian-aged ~10 m thick latitude-dependent mantle (LDM) (Mustard et al., 2001; Kreslavsky and Head, 2002; Head et al., 2003), up to ~1 km thick lobate debris aprons (LDAs) (e.g., Peirce and Crown, 2003; Chuang and Crown, 2005; Head et al., 2006a; Plaut et al., 2009; Baker et al., 2010; Fastook et al., 2014), lineated valley fill (LVF) (Head et al., 2006b; Holt et al., 2008; Kress and Head, 2008; Morgan et al., 2009; Baker et al., 2010), concentric crater fill (CCF) (Levy et al., 2010; Dickson et al., 2010; Fastook and Head, 2014), and other buried ice deposits (e.g., Viola et al., 2015; Bramson et al., 2015) interpreted to be debris-covered glacial deposits that are remnants of regional ice sheets (Fastook et al., 2014; Fastook and Head, 2014) formed in the mid-high latitudes during periods of higher martian obliquity (Madeleine et al., 2009).

(2) Amazonian-aged pedestal craters (Barlow, 2006; Wrobel et al., 2006; Kadish et al., 2008, 2010) and double-layered ejecta (DLE) craters (Weiss and Head, 2013, 2014) hypothesized to form in ~20-200 m thick regional surface ice sheets in the mid-high latitudes during periods of higher martian obliquity.

(3) Tropical mountain glacier deposits (Head and Marchant, 2003; Shean et al., 2005; Kadish et al., 2014; Head and Weiss, 2014; Scanlon et al., 2015).

(4) Evidence for a Late-Noachian-Early Hesperian-aged expanded south-polar cap (Kargel and Strom, 1992; Head and Pratt, 2001; Ghatan and Head, 2002; Fastook et al., 2012; Kress and Head, 2015; Scanlon and Head, 2014; Scanlon et al., 2016).

(5) The potential presence of hectometers-thick regional surface snow and ice deposits in the southern highlands during the Late Noachian period (Head and Marchant, 2014; Fastook and Head, 2015), proposed on the basis of recent 3D global climate models (Forget et al., 2013; Wordsworth et al., 2013, 2015).

Although sources of surface snow and ice in the contact and basal melting scenarios are not lacking, the degree to which both contact and basal melting of surface ice may contribute to fluvial erosion is not yet clear from a physical standpoint. Previous work which assessed contact melting of near-surface icy deposits by conduction from hot ejecta (Mangold, 2012; Mangold et al., 2012a) did so under a wide range of ejecta temperatures. Here, we attempt to provide more precise estimates for ejecta temperature based on established shock physics principles. In this contribution, we provide a quantitative treatment of both the contact and basal melting mechanisms in order to assess whether impact ejecta-induced melting of surface ice deposits could have played a role in forming impact crater-associated fluvial channels during the history of Mars.

2. Heat flow modeling

Could an impact event into any of the various martian surface ice deposits discussed above generate melting and contribute to the observed fluvial erosion around the rim and ejecta of some martian impact craters? At what crater diameters might this process occur? What surface temperatures, geothermal heat fluxes, and ice thicknesses are required to generate basal melting? How
Immediately after the impact event
Hot rocky/icy ejecta emplaced on top of surface ice sheet
Contact melting at ejecta-ice interface produces meltwater
Channels form in ejecta if meltwater from near the rim-crest encounters an impermeable layer.

Crater radii (R) from rim-crest
Crater radii (R) from rim-crest
Crater radii (R) from rim-crest

Fig. 2. Post-impact melting configuration used in our models (with 55X vertical exaggeration). (A) The pre-impact target is composed of a surface ice layer overlying ice-cemented regolith/rock. The pre-impact ice-melting isotherm (273 K) (dashed red line) defines the base of the cryosphere (the zone cold enough for pore-ice stability). (B) The impact occurs, and hot ejecta is deposited on top of the surface ice; contact melting of the surface ice begins. (C) Contact melting continues and meltwater drains out of the ejecta; meltwater derived from near the topographically high rim-crest may form channels within the ejecta facies if the meltwater encounters an impermeable layer (e.g., a spring). (D) The surface ice sheet may flow, enhanced by the weight of the overlying ejecta. (E) The thermally insulating ejecta layer inhibits heat conduction, which raises the melting isotherm (273 K) (dashed red line) up through the cryosphere; the melted pore-ice then drains downward and is a source for groundwater recharge. The 273 K isotherm is raised up to the base of the ice sheet near the rim, where the ejecta is thickest. This allows for basal melting of the ice sheet; the meltwater is predicted to be transported up the crater rim (blue arrows) and towards the crater interior due to the pressurization from the overlying ejecta and ice. (F) The meltwater transported into the crater interior could form fluvial channels on the crater walls. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
long after impact would contact and basal melting of the surrounding surface ice occur, and over what period of time would it continue? What volume of melt is expected, and what are the predicted melting rates? During which martian periods could this process have operated?

In order to address these questions, we implement thermal models to test whether the presence of ejecta on top of surface ice can produce substantial contact melting at the ice sheet surface, or raise the geotherm sufficiently to induce melting at the base of an ice sheet (Fig. 2).

2.1. Contact melting model

To determine the amount of heat transferred from the ejecta into the underlying ice in the contact melting scenario, we solve the one-dimensional heat conduction equation \( \frac{\partial^2 \Delta u}{\partial x^2} = \frac{1}{\rho c_p} \frac{\partial T}{\partial t} \) following Wilson and Head (2007) and Cassanelli and Head (2016). We also take heat transfer only in the vertical direction. We hold the temperature at the top of the ejecta constant at a given surface temperature \( T_S \), and the temperature at the base of the ejecta (the ejecta-interface temperature \( T_I \)) at the melting point of ice (273 K). Under these boundary conditions, solving the heat equation over the thickness of the ejecta \( E_I \) yields the following analytic solution:

\[
T(z, t) = T_S + (T_B - T_S) z / E_I + \sum_{j=1}^{n} A_j \sin(j \pi z / E_I) e^{-k j^2 \pi^2 t / E_I^2}
\]

\[
A_j = T_B - T_S + \left( \frac{T_B - T_S}{L} (1 - z) / L \right) \sin(j \pi z / E_I)
\]

where \( T(z, t) \) is the temperature (in K) at depth \( z \) within the ejecta at time \( t \), \( k \) is the thermal diffusivity, \( A_j \) is the Fourier coefficient, which sets the initial temperature distribution, and \( T_B \) is the initial ejecta temperature. We solve Eq. (1) for \( n = 20 \), which ensured solution convergence at small values of time.

In order to determine the thickness of ice melted through time (the melting rate), we must first determine the heat flux from the ejecta into the underlying ice, which is: \( Q_E = \frac{\partial \Delta u}{\partial z} \), where \( \kappa \) is the thermal conductivity of ejecta, and \( \frac{\partial \Delta u}{\partial z} \) at the ejecta-interface is found by extrapolating the temperature gradient from 0.982 to 0.9992. The melting rate (in m/s of ice melted per m²) is thus: \( R_k = Q_E / (\rho_L + \rho_C \Delta T) \), where \( \rho_L \) is the density of the ice (917 kg/m³), \( L \) is the latent heat of fusion of ice (3.34 × 10⁸ J kg⁻¹), \( \Delta T \) is the temperature difference between \( T_B \) and the ambient ice temperature \( T_{AI} \) from Eq. (3), and \( \rho_C \) is the specific heat capacity of ice; we use the temperature-dependent relationship for \( \rho_C \) by Giaquie and Stout (1936). We assume that no heat energy is lost towards warming the meltwater because it is expected to drain out of the ejecta. The thermal diffusivity is found as: \( k = \frac{\kappa}{\rho c_p} \), where \( \rho_e \) is the density of the ejecta (2500 kg/m³); in the range of lunar impact breccias; Kiefer et al., 2015, and \( C_0 \) is the specific heat capacity of the ejecta (800 J/kg K; in the range of lunar impact breccias and basalts; Hemingway, 1973).

2.2. Ejecta thermal conductivity

Our models differ from previous work on ejecta thermal profiles (Mangold, 2012; Mangold et al., 2012a) which used an ejecta \( \kappa \) similar to the uncommitted target rock (~1.7–2.1 W/m K). We instead adopt an ejecta thermal conductivity value of an impact breccia, since impact ejecta is expected to be largely composed of comminuted material. Previous investigators have found that the \( \kappa \) of lunar breccia is ~0.3 W/m K over a wide range of temperatures and atmospheric pressures (Horai and Winkler, 1980; Warren and Rasmussen, 1987; Warren, 2011), and so we use this value for dry martian ejecta.

The ejecta of martian craters is expected to host pore-ice from a global ice-cemented cryosphere (e.g., Clifford, 1993; Clifford et al., 2010), and so we calculate \( \kappa \) of the ejecta facies as a linear mixture of the volume of rocky ejecta and pore-ice, and ejected surface ice. In our models, the excavated rocky ejecta hosts 15% pore ice by volume within an ice-cemented cryosphere, the thickness of which is the depth of the 273 K isotherm determined from Eq. (3). We use a temperature-dependent \( \kappa \) of the pore-ice from Eq. (4) where \( T_e < 273 \) K. When \( T_e > 273 \) K, the \( \kappa \) of water was adopted for the pore H₂O as a constant 0.569 W/m K (the temperature dependence of water \( \kappa \) over the temperatures in our analysis is minimal) (Ramires et al., 1995). The ejecta \( \kappa \) is calculated from this mixture of breccia and pore ice by approximating the volume of the excavation cavity as a paraboloid with a diameter equal to that of the transient cavity, \( D_T \), and a depth equal to 0.1D_T (Croft, 1980; Melosh, 1989), where D_SC, the average martian simple-complex crater transition diameter is 6 km (Robbins and Hynek, 2012). Although the ejecta is predicted to mix with the target material during ballistic sedimentation (Oberbeck, 1975; Senft and Stewart, 2008), it remains quantitatively unclear how this process will affect the thermal conductivity of the ejecta, and so we ignore this process in determining the ejecta \( \kappa \).

2.3. Ejecta temperature

In order to solve Eq. (1) under a realistic range of temperatures, we first evaluate the average temperature within the ejecta as a function of crater diameter. Previous work has assessed contact melting of near-surface ice deposits by conduction from hot ejecta under a range of ejecta temperatures (Mangold, 2012; Mangold et al., 2012a) from 473 K to 1073 K. This temperature range was applied on the basis of the high temperatures and shock pressures within the outer suevite deposit (up to ~1020 K, up to 80 GPa) that superposes the primary Bunte Breccia ejecta facies of the ~26 km diameter terrestrial Ries crater (Newsom et al., 1986; Engelhardt, 1995). The outer suevite deposit (typically ~5–25 m in thickness; Stöffler et al., 2013), however, is not analogous to the primary ejecta facies (e.g., Stöffler et al., 2013; Artemieva et al., 2013). The primary ejecta facies (the Bunte Breccia) is instead interpreted to have exhibited ambient temperatures upon emplacement (Stöffler et al., 2013). We apply a different approach, and calculate ejecta temperature based on thermal parameters of the target and shock physics principles. The average ejecta temperature, \( T_e \), is evaluated as the sum of: (1) the pre-impact temperature of the target material within the excavation cavity; and (2) the post-shock temperatures of the ejecta following shock heating by the impact event. We note that ejecta temperature should be generally independent of radial distance from the crater because excavation streamtubes cross shock pressure isobars (e.g., Melosh, 1989), which results in ejecta of a variety of shock pressures present at all radial distances; this assumption is supported by more recent shock physics modeling (Collins et al., 2008). We find the pre-impact temperature of the ejecta within the excavation zone using the one-dimensional steady state heat equation:

\[
T(z) = T(z; 1) + \frac{Q \Delta Z}{K(z)}
\]

where \( T_s = T(z; 0) \) and \( Q \) is the geothermal heat flux (in W/m²). Previous work has shown that over the range of temperatures involved in this analysis, the thermal conductivity of basalt given by this relationship is generally equivalent to that of water ice (Clifford, 1993; Clifford et al., 2010):

\[
\kappa = \frac{488.19}{T + 0.4685}
\]

Therefore, a basaltic mega-regolith saturated with pore ice is also predicted to share this thermal conductivity (Clifford, 1993;
Clifford et al., 2010), and so we use this relationship for the thermal conductivity of the target. We also use this relationship for the thermal conductivity of the surface snow/ice deposits, which is representative of the fully densified portion of an ice sheet (Cassanelli and Head, 2015). The geometry of the excavated zone of the crater is estimated using the Maxwell Z model (Maxwell, 1977; Croft, 1980). The temperature distribution within the excavation cavity of a 50 km diameter crater is shown in Fig. 3A (Ts = 215 K, Q = 60 mW/m²). In this example, the pre-impact material within the excavation zone of the crater is predicted to have a volumetric average temperature of 249 K.

We then find the post-shock change in ejecta temperature (ΔT̄) based on experimentally determined equations of state for mafic rocks (Stöffler, 1982; Trunin et al., 2001) following Artemieva and Ivanov (2004) and Fritz et al. (2005), where post-shock temperature is related to the residual energy after the shock event. The post-shock temperature of ejecta ΔT̄ = (Ê_H − Ê_K)/Cp̄, where Ê_H is the Hugoniot total energy (Ê_H = U^2 / 2; U is particle velocity), Ê_K is the energy released during decompression (Ê_K = −∫dV), and Cp̄ is the specific heat capacity of the target (1000 J/kg K) (Artemieva and Ivanov, 2004). Because shock pressure ∝ U, peak shock pressure can then be related to post-shock temperature (e.g., Fig. 7 in Fritz et al., 2005). We find the peak shock pressures using the planar impact approximation (Melosh, 1989, p. 54, 2012) and the semi-analytic Gamma model (Croft, 1982; Pierazzo et al., 1997) for a chondritic impact into basalt, where the impact velocity (10 km/s) is multiplied by a factor of sin(45°) to better represent the shock pressure reduction expected from an oblique impact. Impactor diameter is found from π scaling (Holsapple and Schmidt, 1982). For simplicity, we do not account for the near-surface shock pressure reduction within the near-surface interference zone. Using an interference zone geometry from Melosh (1984) of constant 1 GPa shows that neglecting interference will overestimate our average ejecta temperature estimates by only a small amount: < 2°C for D = 10 km, < 7°C for D = 50 km, < 13°C for D = 100 km (D is crater diameter). The peak shock pressures within the excavated zone using this method are shown in Fig. 3A for a 50 km diameter crater. We then convert these shock pressures into post-shock temperature following the methods of Fritz et al. (2005), and add these values to the pre-impact temperature from Fig. 3A (left panel) to find the post-impact ejecta temperature provenance within the excavation cavity (Fig. 3B).

Fig. 3. (A) Excavation cavity of a 50 km diameter crater showing the pre-impact ejecta temperature (left panel) from geothermal heating (Ts = 215 K, Q = 60 mW/m²) and the peak shock pressure within the excavation cavity (right panel) from the planar impact approximation (Melosh, 1989, p. 54, 2012). Excavation cavity geometry is defined by the Maxwell Z model (Maxwell, 1977). (B) The post-impact ejecta temperature is found by volumetrically averaging the sum of the pre-impact ejecta temperature (A) with the post-shock temperatures derived from peak shock pressures (B).

2.4. Ejecta thickness

In order to find the ejecta thickness (E_t), we use a simplified ejecta thickness function for martian craters (Garvin and Frawley, 1998; Garvin et al., 2000). We approximate the ejecta thickness as a function of radial distance (r) from the rim-crescent as the total topography of the ejecta minus the structural uplift height function from Stewart and Valiant (2006), such that

\[ E_t = h \left( \frac{r}{R} \right)^b - hC \left( \frac{r}{R} \right)^n \]

where R is crater radius and h is rim-crescent height. Rimcrest heights and ejecta topography display exceptional variability (Barnouin-Jha et al., 2005), so we examine both the global average fresh rim-height least-squares fit function (h_{avg} = 0.025D^{0.820}, in km) (Robbins and Hynek, 2012) and a maximum (h_{max} = 0.13D^{0.573}) (Craddock et al., 1997), both of which are valid for rim heights post-crescent collapse. For the fraction of the rim-crescent composed of the structural uplifted target (denoted by C), we use a value of 0.5 (e.g., Stewart and Valiant, 2006; Mouginis-Mark and Boyce, 2012; Mouginis-Mark, 2015) (the C value of ~0.2 for lunar craters, Sharpton, 2014, has not been demonstrated for Mars), and for the structural uplift decay constant, we use an n value of ~3.5 (Stewart and Valiant, 2006; Black and Stewart, 2008). Garvin et al. (2000) found that the average ejecta decay constant, b, is −3.73 for polar craters, and −2.30 for nonpolar craters; we adopt an intermediate b value of ~3. To assess our models under the wide range of observed crater rim-heights, we present the average rim-height (h_{avg}) models as generally representative of typical craters, and the maximum rim-height (h_{max}) models as an upper-limit, but not necessarily rare configuration (e.g., Craddock et al., 1997, Stewart and Valiant, 2006).
2.5. Basal melting model

In order to evaluate the basal melting mechanism, we find basal ice temperatures from the 1D steady-state heat equation (Eq. (3)) using the parameters and conditions outlined in Section 2.1. Our preliminary analysis shows that this approach is appropriate because the cooling timescales of the warm ejecta are substantially lower than the time required for the 273 K isotherm to melt through the cryosphere and rise to the base of the ice sheet. We calculate total melting timescales (where melting begins at the base of the ice-cemented cryosphere and evolves upwards) by solving for the melt rate (in m/s of ice melted per m²):

\[ R_b = \left( Q - \frac{2}{Z} K_z z (T_m - T_s) \right) \frac{1}{\rho \lambda} \]  

(6)

where \( T_m \) is the ice-melting isotherm (273 K). We set the ice-cemented cryosphere thickness at the initial depth of the 273 K isotherm. The final stratigraphy used in our thermal models is shown in Fig. 2B.

We run the models under parameters illustrative of the entire course of martian geologic history: we use surface heat fluxes ranging from 20 mW/m² to 100 mW/m² (Montési and Zuber, 2003; McGovern et al., 2004; Solomon et al., 2005; Ruiz et al., 2011; Plesa et al., 2015), ice thicknesses ranging from 10 to 1000 m (Kadish et al., 2010; Fastook and Head, 2014), and crater diameters up to 150 km. We implement surface temperatures of 215 K and 235 K, which are within the range of mean annual surface temperatures between the Amazonian and those predicted by Late Noachian global climate models with a thicker CO₂ atmosphere and 100% humidity (Forget et al., 2013; Wordsworth et al., 2013). We also explore a warmer Late Noachian scenario with a surface temperature of 255 K.

3. Discussion

In this section, we evaluate our contact and basal melting model results. Our models assume the presence of a continuous ice sheet, although we note that local, patchy ice deposits may also have occurred, and could also be subject to melting (albeit with lower melt volumes). The crater diameters shown represent the diameter upon impact rather than the reduced diameter following surface ice removal (Fig. 5). We first discuss the general predictions made by the models (Sections 3.1 and 3.2), and then assess the fate of the meltwater in chronological order: We first examine the fate of...
of the meltwater produced at the top of the ice sheet by contact melting immediately following the impact (Section 3.3), and then examine the fate of the meltwater produced at the base of the cryosphere as the 273 K isotherm rises through the martian subsurface (Section 3.4). We then discuss the fate of the basal meltwater generated after the 273 K isotherm has reached the base of the ice sheet (Section 3.5). Finally, we examine the possibility for the icy surface layer to glacially flow in order to evaluate its role in changing the model geometry and geomorphological predictions (Section 3.6).

3.1. Contact melting model results

In our contact melting models we calculate (1) the volumetric average ejecta temperature for a given crater diameter (Fig. 4) (Section 2.3), (2) the thickness of ice melted as a function of radial distance from the rim crest (Fig. 6), (3) the total melt volumes produced (Fig. 7), and (4) the total contact melting duration (Fig. 7).

Our ejecta temperature model results (Fig. 4) show that average martian ejecta temperatures are substantially lower than assumed by previous work (473–1073 K; Mangold, 2012). We find that the average ejecta temperatures typically range from ∼220 K to ∼500 K under typical martian conditions, and may range up to ∼600 K for exceptionally high heat fluxes (100 mW/m²). Our results show that the average ejecta temperature increases with increasing crater diameter due to the combination of increasing average peak shock pressures, and increasing the excavation depth of the target (thereby excavating deeper, hotter material). Increasing the surface temperature has an almost linear effect of increasing the ejecta temperature (Fig. 4). For example, increasing the surface temperature to 235 K from 215 K increases the average ejecta temperature by a nearly constant 20 K over the entire crater diameter range. At larger crater diameters, the temperature increase is slightly greater due to the temperature-dependent \( \kappa \) of the subsurface (Eq. [4]) in concert with the increased excursion depth. Increasing the surface heat flux has a much larger effect, and can increase surface temperature by tens to hundreds of degrees (Fig. 4), an effect that is amplified at larger crater diameters due to the increased excavation depth.

Our contact melting models which show the ice thickness melted as a function radial distance from the crater rim-crest are shown in Fig. 6 for surface temperatures of 215 K (left panels; Fig. 6A and D), 235 K (middle panels; Fig. 6B and E), and 255 K (left panels; Fig. 6C and F), and for surface heat fluxes of 40 mW/m² (top panels; Fig. 6A–C) and 60 mW/m² (bottom panels; Fig. 6D–F). Here we do not place a limit on the ice sheet thickness, and thus show the maximum amount of ice that can be melted for a given crater diameter. Our models show that under martian conditions, hot ejecta is able to melt tens to hundreds of meters of ice depending on the surface temperature and heat flux. Although \( T_e \) is constant with radial distance from the rim-crest in our models, the thicker ejecta near the rim (Eq. (5)) is able to melt a substantially greater thickness of ice (Fig. 6). For a surface temperature of 215 K and heat flux of 40 mW/m², a 100 km diameter crater (orange line; Fig. 6A) can melt ∼120 m of ice at 0.18R from the rim-crest, but only ∼20 m of ice 1.2R from the rim-crest. Larger craters have hotter ejecta (Fig. 4), and so are able to melt surface ice out to greater distances relative to smaller craters (Fig. 6). Increasing the surface temperature and heat flux results in melting of modestly thicker amounts of ice at greater distances from the rim-crest for any given crater diameter (Fig. 6). Increasing the ejecta thickness also modestly increases the thickness of ice predicted to melt (see smaller panels for maximum-rim height case) (Fig. 6).

We also present the total volume of surface ice predicted to melt by conduction from hot ejecta in Fig. 7 for surface temperatures of 215 K (top panels; Fig. 7A and D), 235 K (middle panels; Fig. 7B and E), and 255 K (bottom panels; Fig. 7C and F). The contact melt volumes are shown for surface heat fluxes of 40, 60,
and 100 mW/m². We also present contact melt volumes for a lower (Amazonian) heat flux of 20 mW/m² for $T_s = 215$ K (Fig. 6A and D). The contact melting volumes increase substantially as crater diameter increases, but there is only a modest increase in melt volume as surface temperature increases. The most important variables controlling the melt volume are (1) thickness of surface ice available for melting (i.e., the supply limit; Fastook and Head, 2014); (2) the crater diameter; and (3) the surface heat flux.

In all cases, meltwater volumes are predicted to range between $10^6$ and $10^8$ km³, and melting timescales range from ~100 years to ~80 kyr (Fig. 7). We find that contact melting meltwater fluxes range from 0.003 to 0.3 m/yr per m² ($10^{-11}$ to $10^{-9}$ m/s per m²). The large variation in meltwater volumes is largely due to crater diameter and ice thickness, and to a lesser extent, surface heat flux. At small crater diameters, the ejecta facies is colder and thinner, and so cools below 273 K before it can melt substantial volumes of ice. The thickness of the surface ice at the time of impact further acts as a supply limit for the volume of meltwater produced. For example, the melting of a 10 m thick ice sheet can produce up to $10^7$ km³ of meltwater (e.g., dashed white line in Fig. 7A), whereas the ejecta may be hot enough to melt substantially greater thicknesses of ice (see solid white line in Fig. 7A for 100 m thick ice) and the underlying ice-cemented regolith. Geothermal heat flux also plays a large role in meltwater production because modest variations in the heat flux can generate large variations in the temperature distribution within the pre-impact target, and thus the post-impact ejecta (Fig. 3).

The major parameter that controls the wide range in melting timescales (from hundreds of years to ~80 kyr) is ice thickness, because thicker ice sheets provide a larger volume available for melting. For example, for $T_s = 215$ K and $Q_s = 40$ mW/m², a 120 km diameter crater would exhibit contact melting of a 10 m thick ice sheet for hundreds of years (Fig. 6A). For the same thermal conditions, the 120 km diameter crater would exhibit contact melting of a 100 m thick ice sheet for a few tens of kyr. Additionally, ejecta thickness affects the melting timescale. At large ice thicknesses, thicker ejecta facies increases the melting timescales because cooling is slower, and thus melting is prolonged (e.g., Fig. 7A and D). For thinner ice sheets, thicker ejecta instead reduces melting timescales because the heat flux into the surface ice is increased, and the entire thickness of the ice sheet can be fully melted (e.g., Fig. 7A and D).
In summary, contact melting of surface ice sheets overlain by impact ejecta is predicted to occur under a wide range of surface temperatures, heat fluxes, and ice thicknesses, and is thus predicted to operate on a large range (\( \sim 10-30 \) km diameter) of martian impact craters that impacted into surface ice deposits on Mars. These model results are broadly consistent with our current knowledge of the size-distribution of impact craters that host fluvial channels on the ejecta (Morgan and Head, 2009; Jones et al., 2011; Mangold, 2012; Schon and Head, 2012) (discussed in Section 5), but now place quantitative limits on the ejecta temperatures and expected meltwater volumes. We discuss predictions for contact melting through martian geologic history in Section 4.

3.2. Basal melting model results

In our basal melting models (Figs. 8-10), we calculate the total melt volumes produced (Fig. 8A), and because basal melting is driven entirely by the geothermal heat flux (which is low compared with the heat flux from contact melting), we also present the time-averaged melt fluxes (Fig. 8B). We calculate the maximum radial-extent of basal melting from the crater rim-crest (Fig. 8C), as well as the total basal melting durations (Fig. 11). Our basal melting models use the ice thickness outputs from the contact melting models to set the initial ice thickness as a function of radial distance that is available for basal melting.

In a manner similar to contact melting, in the case of basal melting, increasing crater diameter, ice thickness, and surface heat flux all lead to greater volumes of melt, and melting at larger radial distances. We now briefly present one illustrative example: consider a 50 km diameter crater formed during a period where the surface temperature is 235 K, the surface heat flux is 60 mW/m², and a regional ice sheet is present that is 100 m thick (Fig. 9). In this scenario, a substantial amount of surface ice has already been melted through contact melting (Fig. 6E), and so the average rim-height case would not predict basal melting to occur, except for craters between 85 and 90 km in diameter (Fig. 9A). In the maximum rim-height models, however (Fig. 9D), the thermal insulation from the ejecta is sufficient to generate \( \sim 45 \) km² of meltwater (black star in Fig. 9D). The average melt flux in this case is predicted to be \( \sim 10^{-3} \) m³/yr (black star in Fig. 9E), and basal melting
is predicted to extend out to 0.2R from the rim-crest (black star in Fig. 9F). Basal melting of the surface ice begins ∼500 kyr following the impact (black star in Fig. 11K), and continues for ∼80 kyr (black star in Fig. 11E).

The models show that the most important factor in impact ejecta-related basal melting is how much surface ice remains following contact melting. Consequently, the initial ice thickness plays a large role in determining whether or not a particular crater will exhibit basal melting; for example, in Fig. 9D, craters above ∼100 km in diameter are not predicted to exhibit basal melting for an ice thickness of 100 m (Q = 100 mW/m²; dashed red line in Fig. 9D) because there is no ice left to be basally melted. For an ice thickness of 1000 m, however, basal melting is predicted to occur for all craters > 5 km in diameter (solid red line in Fig. 9D) because contact melting is not able to melt the entire 1 km thick surface ice.

The geothermal heat flux also plays a major role in the melt volume, and it directly controls the melt fluxes and timescales. For example, Fig. 9A and B shows that an impact crater forming on a 1000 m surface ice layer (for the case of an average rim-height and $T_s = 235$ K) generates several orders of magnitude greater melt volumes and melt fluxes for the 100 mW/m² heat flux compared with the 40 or 60 mW/m² heat fluxes. Fig. 9H shows how the 100 mW/m² heat flux causes basal melting to begin in ∼100 kyr, compared with ∼400 kyr for the 60 mW/m² heat flux and ∼1 Myr for the 40 mW/m² heat flux. The geothermal heat flux also controls the basal melting duration: Fig. 9E shows how the 100 mW/m² heat flux causes basal melting to continue for ∼200 kyr for a 1000 m thick ice sheet, compared with ∼300 kyr for the 60 mW/m² heat flux and ∼600 kyr for the 40 mW/m² heat flux.

Thicker ejecta, higher geothermal heat fluxes, and larger crater diameters all lead to greater melting volumes (e.g., Fig. 10A), and melting at increasing distance from the rim-crest (e.g., Fig. 10C). The timescale (following the impact) required for initiation of basal melting of the surface ice decreases with increasing ejecta thickness due to the greater insulation. Consequently, basal melting is predicted to initiate closest to the rim-crest, and continue progressively further from the rim-crest with time.
We now discuss the chronology of ice melting and the predictions for the hydrologic fate of the meltwater.

3.3. Contact melting

Contact melting is predicted to initiate immediately following the deposition of hot ejecta, and may last for hundreds to tens of thousands of years (Fig. 7). Ice melting begins at the interface between the overlying ejecta and the surface ice. Consequently, the meltwater will follow the local topographic gradient of the surface ice, which, in the absence of complex surface topography, will be controlled by the outward-facing slope of the rim structural uplift (Fig. 12A). Thus, once generated, the meltwater is predicted to flow downhill, away from the crater, on the top of the ice layer, and could contribute to erosion of the ejecta through sapping (Fig. 12). Exceptions to this are for craters forming in $\sim 10^2$ m thick ice deposits (as thick as the structural uplift; Fig. 12C), where melting of the near-rim ice is predicted to reverse the topographic slope and lead to contact meltwater flowing into the crater (Fig. 12D). Importantly, fluvial channels interpreted to result from contact melting are largely found at the ejecta surface (Morgan and Head, 2009; Jones et al., 2011; Mangold, 2012; Schon and Head, 2012). As discussed in Mangold (2012), if the meltwater is over-pressureized, it may be able to breach the ejecta vertically to generate bursts of fluvial activity and flow down the crater walls and/or down the slope of the ejecta away from the crater interior. Alternatively, fluvial channels at the surface of the ejecta could form in an analogous mechanism to terrestrial springs. In this scenario, meltwater generated at the ice surface close to the rim will flow downslope. If the meltwater meets an impermeable layer, it will flow along this layer laterally until it breaches the ejecta surface (blue arrows in Fig. 12A) and can incise the ejecta, ultimately forming fluvial channels.

3.4. Cryospheric meltwater

Following chronologically from contact melting is cryosphere melting. We predict that the 273 K isotherm will rise through the cryosphere and melt the subsurface pore-ice due to the thermally insulating effects of the ejecta. Indeed, in order for basal melting of the surface ice to occur, the 273 K ice-melting isotherm must first rise through the ice-cemented cryosphere to the base.
of the ice sheet (Fig. 2D and E). Consequently, any impact crater predicted to exhibit basal melting of an ice sheet is also predicted to have melted through the entire thickness of the ice-cemented cryosphere (Fig. 2E). However, the inverse is not necessarily always true, that is, a crater can experience melting of the ice-cemented cryosphere without associated basal melting of an ice sheet. This can occur in cases where contact melting of ejecta has already melted the emplaced surface ice sheet, or in cases where the 273 K isotherm is not raised all the way to the surface. Fig. 11G–L shows the crater diameter and thermal conditions under which the entire cryosphere is predicted to be melted through.

The meltwater generated by bottom-up melting of the pore-ice within the cryosphere (due to the thermally insulating ejecta) is predicted to drain downwards through the dry permeable subsurface (Fig. 2E), and thus offers a source of water for groundwater recharge below impact craters. This is analogous to the “heat-pipe drain pipe” groundwater recharge mechanism produced by local volcano thermal effects on the cryosphere (Cassanelli et al., 2015). Furthermore, as the pore-ice within the cryosphere melts and drains downwards, the combination of percolating water and high subsurface temperatures in the subsurface could cause rock-water interactions that might plausibly chemically alter crustal rocks into clays (e.g., Tosca and Knoll, 2009).

Clay minerals requiring water-rock interactions have been detected near-globally on the surface of Mars (Poulet et al., 2005; Bibring et al., 2006; Mustard et al., 2008; Murchie et al., 2009; Ehlmann et al., 2009; Loizeau et al., 2012; Quantin et al., 2012; Carter et al., 2013, 2015; Sun and Milliken, 2015). Their identification primarily within crater central peaks, walls, and ejecta has
led to the interpretation that these clay minerals were excavated from depth. These clay minerals are variously interpreted to reflect weathering at the surface and subsequent burial (Loizeau et al., 2012; Poulet et al., 2005; Carter et al., 2015), or formation in the subsurface by interaction with crustal fluids (Ehlmann et al., 2011; Loizeau et al., 2012; Carter et al., 2013; Sun and Milliken, 2015). Indeed, the observation of increasing chloritization with depth (between ~1 and 7 km) requires the presence of fluids at these depths within the martian crust (Sun and Milliken, 2015). Craters which excavate these clays are primarily Noachian and Hesperian in age, although some Amazonian examples are also present (Sun and Milliken, 2015), raising the possibility of younger, localized regions of subsurface clay formation. The aqueous environment which formed the clays has previously been proposed to relate to impact-induced hydrothermal activity (i.e., Abramov and Kring, 2005; Schwenzer and Kring, 2009, 2013; Schwenzer et al., 2012) or burial diagenesis/metamorphism in the subsurface (e.g., Ehlmann et al., 2011; Loizeau et al., 2012; Carter et al., 2013; Sun and Milliken, 2015), or the remnants of a buried ancient Noachian-aged surface which experienced warmer and wetter conditions (e.g., Loizeau et al., 2012; Poulet et al., 2005; Bibring et al., 2006; Carter et al., 2015).

Due to the widespread distribution of martian impact craters throughout both space and time, we suggest that the impact-induced basal melting scenario (Fig. 2E) provides another candidate mechanism for generating the subsurface liquid water needed to form clays in the martian subsurface. Full melt-through of the cryosphere through basal melting is predicted to have occurred at progressively smaller crater diameters at higher heat fluxes (and is thus predicted to be more common during the Noachian and Hesperian periods) (see Fig. 11G), but partial melting of the cryosphere is predicted in the Amazonian as well. This is generally consistent with the inferred ages associated with clay formation (Sun and Milliken, 2015). Importantly, surface ice is not required to be present on the pre-impact target in this scenario. Given sufficiently large impactors and the presence of a global ice-cemented cryosphere (Clifford, 1993; Clifford et al., 2010), this process is predicted to be virtually ubiquitous throughout the subsurface of Mars. After an impact event, the insulating effects of the ejecta are predicted to elevate the geotherm. As the 273 K ice melting isotherm rises through the ice-cemented cryosphere, pore-ice will melt and percolate downwards through the crust. This mechanism could plausibly contribute to subsurface clay formation, and warrants further quantitative investigation to determine if the subsurface temperatures, volumes of water, and timescales of water-rock interaction (e.g., Tosca and Knoll, 2009) predicted by this mechanism are consistent with the observed clay maturity and mineralogy.

3.5. Basal meltwater

After the 273 K isotherm rises through the cryosphere and reaches the base of the ice sheet, basal melting of the ice is predicted to initiate. What is the fate of the meltwater generated at the base of the ice sheet? In which direction will the basal meltwater be transported, and is it predicted to produce surface runoff, or infiltrate and drain through the substrate?

The infiltration capacity (I in m/s) of the target material (i.e., the rocky substrate underlying the surface ice) can be calculated as (Hendriks, 2010):

\[ I = \frac{K}{x} \left( x + S + \frac{h}{x} \right) \]

where \( x \) is the thickness of the porous medium, \( S \) is the wetting front soil suction head, \( h \) is the head of infiltrating ponded water (all in m), and \( K \) is the hydraulic conductivity (in m/s). In order to estimate a minimum infiltration rate, we conservatively assume that \( S \) and \( h \) are equal to zero (e.g., Cassanelli and Head, 2016). Consequently, \( S \) and \( h \) in Eq. (7) reduce to zero and the infiltration rate in the basal melting scenario can simply be approximated by the hydraulic conductivity:

\[ K = \frac{k_{\rho_w g}}{\mu} \]

where \( K \) is the intrinsic permeability (in m²), \( \rho_w \) is the density of water (1000 kg/m³), \( g \) is gravity (3.71 m/s²), and \( \mu \) is the dynamic viscosity of water (1.793 × 10⁻⁵ Pa s at 273 K). In our calculations, basal melt fluxes range from 3 × 10⁻⁵ to 0.009 m/yr per m² (~10⁻12 to ~10⁻10 m/s per m²) (Fig. 13A). On the basis of these melt fluxes, substrate permeability (\( k \) from Eq. (8)) is required to be \( < \) 10⁻18 to 10⁻16 m² (i.e., crystalline or consolidated sedimentary bedrock; Fig. 13A) for melting rates to exceed infiltration rates. If the melt fluxes exceed substrate infiltration rates, the melt may be transported laterally and lead to runoff (and thus potentially erosion). Higher permeabilities (such as expected for regolith) enhance infiltration rates, and are predicted instead to allow for the meltwater to infiltrate and drain through the substrate (Fig. 13A).

In order to assess if the melt can be transported up the rim structural uplift, we calculate the hydraulic head of the basal melt \( (H_{\text{melt}}) \) generated by the overlying ejecta:

\[ H_{\text{melt}} = \frac{z_i + z_e \rho_e}{\rho_w} \]

where \( z_i \) and \( \rho_i \) are the ice thickness and density, and \( z_e \) is the ejecta thickness. In order to examine the hydraulic head predicted independent of surface ice thickness, we do not include any surface ice \((z_i = 0)\). Including surface ice would increase the hydraulic head, and so these estimates should be considered lower bounds. Fig. 13B shows the predicted hydraulic head as a function of crater diameter and distance from the rim-crest. If the rim structural uplift height (black line in Fig. 13B) exceeds the hydraulic head, basal meltwater will stall below the surface ice and ultimately drain downwards into the substratum. Conversely, if the hydraulic head exceeds the rim structural uplift height, then the meltwater will be sufficiently pressurized to flow up the rim structural uplift and drain into the crater interior (Fig. 13B). Assuming a continuous ice-sheet underlying the ejecta, which acts as an impermeable barrier to vertical ascent of meltwater (i.e., it prevents meltwater from diffusing into the overlying ejecta), Fig. 13B shows that all melt within ~0.6 m from the rim-crest is predicted to be transported up the rim structural uplift slope towards the crater interior due to the overburden pressure of the overlying ejecta. The pressurization and transport of meltwater up the slopes of the crater rim structural uplift is made possible by the decreasing ejecta thickness with radial distance from the crater rim-crest. The decreased thermal insulation provided by relatively thinner distal ejecta prevents the 273 K melting isotherm from rising to the base of the ice sheet (Fig. 2F). Consequently, the ice sheet is predicted to be frozen to the bed far from the rim-crest (generally ~0.1–2.5 m from the rim-crest), which prevents meltwater drainage away from the crater. Exceptions to this may exist in the case of patchy rim-ice deposits, where the discontinuous extent of surface ice allows meltwater to drain away from the crater through the porous ejecta. Meltwater ascent up the rim structural uplift is only possible as long as the substrate permeability is low enough such that melt rates exceed infiltration rates (Fig. 13A). If the rim structural uplift is highly fractured by the impact (which increases infiltration rates), or in the case where contact melting has melted the entire thickness of near-rim ice, the meltwater may instead infiltrate downwards through the substratum. Thus, thicker ice layers, which are less likely to be fully melted (Fig. 6) favor transport of melt up the rim and into the crater interior, whereas thin ice layers generally prevent melt transport into the crater interior. If a hydrothermal
system (Abramov and Kring, 2005; Schwenzer and Kring, 2009) is still present within the crater (predicted to last tens to hundreds of kyrs; cf. Fig. 11G–L) and the cryosphere directly adjacent to the crater (Fig. 2E) is melted during basal melting, the meltwater may flow laterally into the crater interior (e.g., Schwenzer and Kring, 2009) independent of ice thickness.

Is the meltwater expected to form subglacial fluvial channels on the outer slopes of the structurally uplifted rim (the rocky rim below the surface ice; Fig. 2F)? An analogy may be drawn with other Martian fluvial features interpreted to result from basal melting of surface ice deposits. For example, the relatively young valley networks present on several Martian shield volcanoes (e.g., Hecates Tholus, Ceraunius Tholus, Alba Patera) are interpreted to result from basal melting of snowpack (Fassett and Head, 2006, 2007) during periods of high surface heat flux associated with magmatic intrusions. Fassett and Head (2006, 2007) show that these valley networks are present primarily on steep slopes, downslope of the predicted location of the snowpack. They thus concluded that surface slope (i.e., flow velocity) is a major controlling factor on the formation of basal melt-generated fluvial channels (Fassett and Head, 2006, 2007). We note that in the impact crater-induced basal melting scenario explored here, the velocity of the meltwater which flows up the rim structural uplift will be controlled solely by the rate of meltwater production. Because our models predict extremely low melt-fluxes (∼10⁻⁴ to ∼10⁻³ m/yr per m²), the velocity of the meltwater as it flows up the rim structural uplift is predicted to be comparably low; consequently, the meltwater is generally not predicted to incise subglacial fluvial channels into the rocky substrate during flow up the rim structural uplift. While the presence of fluvial channels that flow up topography would provide strong evidence for a subglacial origin, these channels are unlikely to be observed due to their formation beneath the ejecta. Fluvial incision is predicted instead to initiate when the meltwater begins to flow down the slopes of the interior crater walls (Fig. 2F). This prediction may not hold in the case of impacts into relatively thick surface ice, where the rim structural-uptilt may be partially accommodated by the surface ice, leading to relatively lower (or entirely absent) bedrock rim-slopes (Weiss and Head, 2015) that are more likely to form fluvial channels. In this case, the removal of a thick surface ice layer at a later time would serve to reduce the currently observed crater diameter (Fig. 5) (Weiss and Head, 2015) such that channels currently observed at large radial distances from the rim may in fact have been located proximal to the rim-crest.

3.6. Ice flow

Could the glacial flow of the surface ice layer modify the model geometry and morphological predictions? It is important to consider the effects of ice flow because the weight of the overlying ejecta will increase the shear stress within the icy layer. In the geometry explored in this study, ice is primarily predicted to flow down the slopes of the rim structural uplift (Fig. 2D); however, because ice flow follows pressure gradients, a small amount of ice may also extrude into the crater (Fig. 2D) We assess the speed at which the icy layer beneath the ejecta will flow, and examine how ice flow is predicted to effect the contact or basal melting process. We adopt a steady-state parallel-sided flow model (e.g., Paterson, 1981, p. 86) modified to include the effects of the overlying debris weight (Konrad and Humphrey, 2000) in order to explore the range of speeds the ice could flow downslope of the rim-crest. In the parallel-sided flow model, an ice sheet of infinite width rests on an inclined plane of constant slope (θ). The ice sheet has a thickness (z_i) and density (ρ_i) and is superposed by a debris layer of thickness (z_e) and density (ρ_e). In this case, the steady-state depth-averaged horizontal velocity of the ice sheet (Konrad and Humphrey, 2000) is:

$$U_{avg} = \frac{2A(\rho g \sin \theta)^n}{(n+1)(n+2)z_i} \left[ z_i (n+2) \left( \frac{\rho_e}{\rho_i} z_e \right)^{n+1} \right.$$

$$+\left( \frac{\rho_e}{\rho_i} \right)^{n+1} \left( z_i + \frac{\rho_e}{\rho_i} z_e \right)^{n+2}$$

$$\left. \right]$$

(10)

where the flow law constant (n) is 3, and the constant rate factor, A, (in s⁻¹ Pa⁻¹) varies as a function of temperature. Contact melting (and ejecta cooling) durations are predicted to be on the order of hundreds of years to a few tens of kyrs (Fig. 7) which is typically much lower than our initial glacial flow model timescales, and so we calculate the icy layer flow speeds following contact melting, and apply ambient temperature conditions for the ejecta temperature. We adopt the temperature-dependence of A values from Paterson (1981, p. 39), so that the depth-averaged A values within the ice sheet (using the temperature profile from Eq. (3)) ranges between 2.8 × 10⁻²⁷ for Ts = 215 K, and 1.3 × 10⁻²⁴ for Ts = 255 K. For the parameters θ (from Eq. (5)), z_e, and z_i in Eq. (10), we adopt the values averaged between 6R and 0.6R from the rim crest (the area of interest for basal melting and the area of high slopes; Fig. 13B). As in the basal melting models, the ice thickness inputs are taken from the contact melting model outputs (Fig. 6). This allows us to address how ice flow could modify the initial basal melting conditions. In this case, the 273 K isotherm has not yet risen through the ice-cemented cryosphere; the ice is still frozen to the bed, and so we neglect basal sliding. Even in the case of a wet-based ice sheet during melting, we do not consider basal sliding to be a major contributor to the flow speed because the speed of the ice will be buffered by the ice downslope from the crater that remains frozen to the bed (Fig. 6F). Fig. 14 shows the ice flow speed model results for Ts = 215 K (Fig. 14A), Ts = 235 K (Fig. 14B), and Ts = 255 K (Fig. 14C). The ice layer within 0.6R exhibits a wide range in average flow speeds between 0.1 μm/yr and 200 m/yr depending on crater diameter, ice thickness, surface temperature, and the ejecta thickness (average vs. maximum rim-height functions). Glacial flow speed generally decreases with increasing crater diameter (Fig. 14) because larger craters melt thicker sequences of the ice layer (Fig. 6). Thicker ice layers (particularly 1000 m) are excluded from this trend because they retain a substantial thickness even after contact melting (Fig. 6). The thick blue lines in Fig. 14 show the speed required in order for the ice to flow 0.1R in 100 kyrs and 1 Myrs, which is the range in timescales required for the 273 K isotherm to rise from depth to the base of the ice sheet (Fig. 11). We find that ice flow speeds are generally well below these values (blue lines in Fig. 14), indicating that minimal ice flow ( < 0.1R flow distance) is expected on the timescales of cryospheric and basal melting, except in the case of 1 km thick ice sheets (100 m ice sheets and Ts = 255 K for the maximum rim-height case). We conclude from the results that ice flow is generally not a major process in the impact-induced ejecta melting scenario, and is not expected to modify the model geometry except in the case of thick ice sheets (on the order of ~1 km) and high surface temperatures ( > 235 K).

3.7. Summary of results

In summary, following the impact (Fig. 2), hot ejecta is predicted to melt tens to hundreds of meters of surface ice (Fig. 6) over the course of ~100 years to 80 kyr through contact melting, corresponding to 10²⁰–10³⁰ km³ of meltwater (Fig. 7). Larger craters and higher surface heat fluxes generate substantially greater volumes of meltwater. Contact melting initiates throughout the entire radial distance of ejecta after ejecta emplacement, but terminates earlier for the thinner, faster cooling distal ejecta. Meltwa-
Fig. 12. Contact melting scenario for thin ice (left panels; A, B) and thick ice (right panels; C, D) for early times (top panels) and late times (bottom panels). (A) For the thin ice case, melt will flow away from the crater during the early times when contact melting initiates. If the near-rim melt encounters an impermeable layer within the ejecta, spring-like breaching of the ejecta may generate fluvial channels on the ejecta facies. (B) In the later times, the near-rim ice has melted, reversing the slope of the topography. Melt may run off toward the crater, but because the ice is lower than the structurally uplifted rim-crest, it is predicted to pool within the ejecta. (C) For the thick ice case, melt will flow away from the crater during early times when contact melting initiates. As in the thick ice case, breaching of the ejecta by a spring would form fluvial channels on the ejecta. (D) In the later times, a substantial thickness of near-rim ice has melted, which reverses the topographic slope. In this case, meltwater is predicted to flow into the crater interior, and could produce fluvial channels on the crater walls. (For interpretation of the references to color in this figure, the reader is referred to the web version of this article.)

The ice-melting isotherm is predicted to reach the base of the ice sheet within ~100 kyr to ~1 Myr after the impact (Fig. 11). If the ice sheet is thick and has not been previously removed by contact melting, this can lead to basal melting of the ice sheet that lasts for ~1 kyr to ~1 Myr. Larger crater diameters and higher geothermal heat fluxes lead to increased melt volumes (~10^9 – 10^10 km^3) and fluxes. Melting initiates closest to the rim and continues progressively outward, generally out to ~0.1 to 2.5R from the rim-crest. If the rocky substrate below the surface ice has a relatively low permeability (i.e., crystalline or consolidated sedimentary bedrock), the basal melt fluxes will exceed the infiltration rates. In this case, the pressurization from the overlying ejecta and ice may transport the basal meltwater up the slopes of the crater rim structural uplift, whereupon the meltwater is predicted to drain into the crater interior and down the crater walls.

The ice layer is predicted to flow away from the crater, but the flow speeds are so low that this process is not likely to have a large effect on the basal melting geometry or melt volumes. With these guidelines in place, we now specifically examine the conditions under which basal melting may occur in each of the different martian geologic periods.

4. Ice melting throughout martian geologic history

In this section, we discuss the results of our thermal models in the context of the different martian geologic periods (ages from Werner and Tanaka, 2011 using the Hartmann, 2005 production function) and their potential surface temperature (Wordsworth et al., 2013, 2015) and heat flux ranges (from Montési and Zuber, 2003; McGovern et al., 2004; Solomon et al., 2005; Ruiz et al., 2011; Plesa et al., 2015) for crater diameters up to 150 km. These results are summarized in Fig. 15.
4.1. Amazonian period (0.0–3.0 Ga; $T_s = 215$ K; $Q = \sim 20$–40 mW/m$^2$)

Under Amazonian conditions, post-impact ejecta temperatures range between 220 K and 360 K for craters between 5 km and 150 km in diameter (Fig. 4). Craters larger than $\sim 30$–40 km in diameter are predicted to exhibit contact melting (blue star in Fig. 15A), and produce melt volumes ranging from $10^6$ to $10^3$ km$^3$ (pink and white lines in Fig. 7A and D).

For typical Amazonian parameters and the average rim-height case, basal melting is not predicted to occur in our models (Fig. 8A and D), except perhaps for craters larger than 150 km in diameter (e.g., Head et al., 2016), which we do not explore in this study. This
Fig. 14. Average glacial flow speeds of the icy layer within 0.6R from the crater rim-crest for (A) surface temperatures of 215 K, (B) 235 K, and (C) 255 K. Model results are shown for the average rim-heights (black lines) and maximum rim-heights (red lines) for surface ice thicknesses of 1000 m (solid lines), 100 m (dashed lines), and 10 m (dotted lines). Surface heat flux in these models is 60 mW/m²; these results do not vary significantly from using heat fluxes of 40 or 100 mW/m². The thick blue lines show the flow speeds required for the ice to flow 0.1R within 100 kyr to 1 Myr. These low flow speeds represent the cold-based ice flow prior to basal melting. Ice flow speeds are generally predicted to be extremely low, with minimal movement of the icy layer. Where model lines are not present (e.g., at larger diameters for the 10 and 100 m ice sheets), all ice has been melted within 0.6R from the rim-crest, and so no ice flow is predicted to occur. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
is due to contact melting following the impact, which removes any surface ice that would have undergone basal melting under typical Amazonian conditions (blue star in Fig. 15B).

For a surface heat flux of 40 mW/m² ($Ts = 215$ K) the 273 K isotherm depth is ~3.6 km, and the full thickness of the cryosphere is not predicted to fully melt through.

### 4.2. Hesperian period (3.0–3.6 Ga; $Ts = 215$ K; $Q = 30–60$ mW/m²)

Under Hesperian conditions ($Q = 40–60$ mW/m²), ejecta temperatures range between 220 K and 430 K for craters between 5 km and 150 km in diameter (Fig. 4). Craters larger than ~30 km in diameter are predicted to exhibit contact melting (green star in Fig. 15A) and produce melt volumes ranging from $10^3$ to $10^4$ km³ (Fig. 7A and D).

Basal melting is predicted to occur only for $Q \geq 60$ mW/m², and only for surface ice thicknesses of ~1 km; smaller ice thicknesses are almost entirely removed by contact melting, and thus prevent ice from being preserved for basal melting. For a heat flux of 60 mW/m², basal melting will initiate for craters between ~85 km in diameter (maximum rim-height) and 140 km in diameter (average rim-height) (Fig. 8A and B). Melt volumes range from $10^3$ to $10^4$ km³ (Fig. 8A and B), and basal melting is predicted to extend from 0.1 to 0.4R from the rim-crest (Fig. 8C and F).

For a surface heat flux of 60 mW/m² ($Ts = 215$ K), the 273 K isotherm depth is ~2.4 km. Basal melting will begin at
least \( \sim 550 \text{ kyr} \) after impact (Fig. 11J), and will last up to \( \sim 400 \text{ kyr} \) in order to entirely melt-through a 1000 m ice sheet (Fig. 11D).

### 4.3. Late Noachian period (3.6–3.8 Ga; \( T_s = 215–255 \text{ K} \);
\( Q = \sim 40–65 \text{ mW/m}^2 \))

Under Late Noachian conditions (\( T_s = 215–255 \text{ K} \), \( Q = 40–60 \text{ mW/m}^2 \)), ejecta temperatures range between 220 K and 490 K for craters between 5 km and 150 km in diameter (Fig. 4). Craters larger than \( \sim 10–25 \text{ km} \) in diameter are predicted to exhibit contact melting (yellow star in Fig. 15A), and produce melt volumes ranging from \( 10^3 \) to \( 10^4 \text{ km}^3 \) (white and black lines in Fig. 7) for ice sheet thicknesses (Fastook and Head, 2015) between 10 and 1000 m.

Basal melting is not predicted in a cold Late Noachian period (\( T_s = 215 \text{ K} \)), except for heat fluxes \( \geq 60 \text{ mW/m}^2 \) and ice thicknesses of \( \sim 1000 \text{ m} \). Representative melting volumes, distances, and timescales are identical to those discussed above for the Hesperian case with a 60 mW/m² heat flux.

In the case of a slightly warmer Noachian period (\( T_s = 235 \text{ K} \); e.g., Wordsworth et al., 2013) onset diameters for basal melting range from \( \sim 30–130 \text{ km} \) for heat fluxes between 40 and 60 mW/m² and an ice thickness of 1000 m (yellow star in Fig. 15B; Fig. 9A and D). Melt volumes range from \( 10^6 \) to \( 10^7 \text{ km}^3 \) (Fig. 9A and D), and basal melting may extend from 0.1 to 1.0R from the rim-crest (Fig. 9C and F). Basal melting is predicted for 100 m thick ice sheets only for the maximum rim-height case and 60 mW/m² heat flux for a restricted crater diameter range between 35 and 85 km in diameter (Fig. 9D), which may produce melt volumes between \( \sim 4–65 \text{ km}^3 \). For this case, basal melting extends between 0.1 and 0.5R from the rim.

To simulate an even warmer Late Noachian period, we also examine the thermal models for a surface temperature of 255 K (Fig. 10). For the average rim-height case, onset diameters for basal melting are \( \sim 35–55 \text{ km} \) for a 100 m thick ice sheet, or 5 km for a 1000 m thick ice sheet (Fig. 10A). Craters larger than 125 km in diameter are not predicted to exhibit basal melting in the case of 40 mW/m² with 100 m thick ice because contact melting has already removed the surface ice. Melt volumes range from \( 10^1 \) to \( 10^2 \text{ km}^3 \) for a 100 m thick ice sheet, and from \( 10^6 \) to \( 10^7 \text{ km}^3 \) for a 1000 m thick ice sheet (Fig. 10A). Basal melting may extend from 0.1 to 0.9R from the rim-crest for 100 m thick ice, and 0.4 to 2.5R for 1000 m thick ice (Fig. 10C).

For the maximum rim-height case, onset diameters for basal melting are \( \sim 15–25 \text{ km} \) for a 100 m thick ice sheet, or 5 km for a 1000 m thick ice sheet (Fig. 10A). Craters larger than 125 km in diameter are not predicted to exhibit basal melting in the 40 mW/m² case with 100 m thick ice because contact melting has already removed the surface ice. Melt volumes range from \( 10^1 \) to \( 10^2 \text{ km}^3 \) for a 100 m thick ice sheet, and from \( 10^6 \) to \( 10^7 \text{ km}^3 \) for a 1000 m thick ice sheet (Fig. 10A). For a 10 m thick ice sheet, there is a restricted range of crater diameters between 10 and 15 km in diameter that exhibit basal melting, but melting volumes are negligible (0.2 km³). Basal melting may extend from 0.2 to 1.2R from the rim-crest for 100 m thick ice, and 1.1–2.5R for 1000 m thick ice (Fig. 10F).

For a surface heat flux of 60 mW/m² (\( T_s = 235 \text{ K} \)), the 273 K isotherm depth is \( \sim 1.2 \text{ km} \). Basal melting will begin at least \( \sim 325 \text{ kyr} \) after impact (Fig. 11K), and will last \( \sim 10–100 \text{ kyr} \) for a 100 m ice sheet, and \( \sim 200 \text{ kyr} \) for a 1000 m ice sheet (Fig. 11E). For \( T_s = 255 \text{ K} \), basal melting will begin \( \sim 150 \text{ kyr} \) after impact (Fig. 11L), and continue for at least \( \sim 5 \text{ kyr} \) for 10 m thick ice, \( \sim 30 \text{ kyr} \) for 100 m thick ice, and \( \sim 200 \text{ kyr} \) for 1000 m thick ice (Fig. 11F).

### 4.4. Early-Mid Noachian period (3.8–4.5 Ga; \( Q = \sim 45–100 \text{ mW/m}^2 \))

Atmospheric temperatures during this period are not yet well constrained, and so we evaluate this period with a temperature range of 215–235 K, noting that warmer or colder temperatures are possible based on warming (or lack thereof) from impacts and volcanism (e.g., Toon et al., 2010; Head and Head, 2014). This period exhibits higher geothermal heat fluxes, and is thus also applicable for younger periods during martian history in areas of anomalous heat flux, such as in volcanic terrains (e.g., Fassett and Head, 2006, 2007; Cassanelli et al., 2015). Results for heat fluxes of 60 mW/m² are discussed above, and so the following discussion is for a heat flux of 100 mW/m².

Under potential Early-Mid Noachian conditions (\( T_s = 235 \text{ K} \), \( Q = 100 \text{ mW/m}^2 \)), ejecta temperatures range between 250 K and 550 K for craters between 5 km and 150 km in diameter (Fig. 4). Craters larger than 15 km in diameter are predicted to exhibit contact melting (red star in Fig. 15A) and produce melt volumes ranging from \( 10^6 \) to \( 10^7 \text{ km}^3 \) (Fig. 7B and E).

For a surface temperature of 215 K and the average rim-height case, surface ice 1000 m thick is predicted to exhibit basal melting for crater diameters as low as 50 km in diameter, and 100 m thick ice is predicted to exhibit basal melting for a restricted crater diameter range between 80–85 km in diameter (Fig. 8A). Melt volumes range from \( 10^1 \) km³ for the 100 m thick ice sheet, and from \( 10^6 \) to \( 10^7 \text{ km}^3 \) for the 1000 m thick ice sheet (Fig. 8A and D), and basal melting may extend out to 0.1–0.8R from the rim-crest (Fig. 8C). For the maximum rim-height case, surface ice 1000 m thick is predicted to exhibit basal melting for crater diameters as low as 25 km in diameter, and 100 m thick ice is predicted to exhibit basal melting for a restricted crater diameter range between 35 and 90 km in diameter (Fig. 8D). Melt volumes range from \( 10^3 \) km³ for the 100 m thick ice sheet, and from \( 10^6 \) to \( 10^7 \text{ km}^3 \) for the 1000 m thick ice sheet (Fig. 8D), and basal melting may extend out to 0.1–1.1R from the rim-crest (Fig. 8F).

In a slightly warmer environment (\( T_s = 235 \text{ K} \), and the average rim-height case, a 10 m thick ice layer will still not exhibit basal melting because it has already been fully contact melted. Basal melting is predicted to occur for crater diameters between 45 and 105 km for a 100 m thick ice layer (red star in Fig. 15B; Fig. 9A). In the case of a 1 km thick surface ice sheet, basal melting will occur even in the absence of an impact crater and insulating ejecta (Fig. 9A). Basal melting may extend out to 0.1–0.6R from the rim-crest for 100 m thick ice, or 2.5R for 1000 m thick ice. For the maximum rim-height case, a 10 m thick ice layer will produce basal melting for crater diameters between 15–20 km, and a 100 m thick ice layer will produce basal melting for crater diameters between 20 and 100 km (red star in Fig. 15B; Fig. 9D). Basal melting may extend up to 0.2–0.9R from the rim-crest for 10–100 m thick ice sheets, and 2.5R for 1000 m thick ice sheets. Melt volumes range from \( 10^1 \) to \( 10^3 \text{ km}^3 \) (Fig. 9A and D).

In summary, contact melting is predicted to occur throughout martian history for all craters greater than \( \sim 40 \text{ km} \) in diameter that impacted into surface ice sheets. In contrast, basal melting is not predicted in the Amazonian period, except perhaps for craters larger than 150 km in diameter which we did not explore in this study. Higher heat fluxes in the Hesperian period allow basal melting to occur under a relatively small parameter range, only for large craters >85–140 km in diameter for the upper-end predicted surface heat flux of 60 mW/m² and 1000 m thick ice sheets. Due to the potential for higher heat fluxes and higher surface temperatures in the Noachian period, basal melting is predicted to occur under a relatively broader range of crater diameters, potentially for craters as small as \( \sim 5–30 \text{ km} \) depending on the surface temperature and ice thickness. Because contact melting removes a substan-
tial portion of surface ice prior to the onset of basal melting, the ice thickness (Fastook and Head, 2015) plays a major role in limiting the crater diameters which are predicted to exhibit basal melting. Thus, ice sheet thicknesses in the 10 to 100 m range are generally predicted to produce basal melting under a more restricted crater diameter range, even during the Noachian period.

On the basis of the predicted basal meltwater hydrologic geometry (Fig. 2F) and the restricted diameter range predicted to exhibit basal melting (compared with Noachian-aged craters which exhibit fluvial channels) (Fig. 15B), basal melting is not predicted to erode substantial volumes of ejecta, and is not considered a candidate process which can carve any valley network features. Contact melting, however, is predicted to occur under a broader parameter range in the Noachian period (Fig. 15A), and may contribute to the lack of observable ejecta associated with Noachian-aged craters (e.g., Craddock and Maxwell, 1993; Craddock et al., 1997; Craddock and Howard, 2002; Forsberg-Taylor et al., 2004; Mangold et al., 2012b).

5. Candidate craters for ejecta-induced ice melting

As discussed above, contact melting is predicted to occur under a wide range of impact and climatic conditions. In contrast, basal melting is theoretically plausible only under certain impact diameter and climatic conditions. The thermal models derived here can place important constraints on the crater diameters, target properties, and locations (relative to the crater) where contact and basal melting are predicted to occur. Higher surface temperatures (≥225 K) and/or heat fluxes, such as those expected during the Noachian period, reduce the crater diameter and ejecta thickness conditions necessary to generate melting of surface ice. With these guidelines, we assess if there are any observed impact crater-associated fluvial channels that are consistent with either a contact or basal melting origin.

5.1. Candidate craters for contact melting

Previous investigators have identified fluvial channels superposing the crater walls, rim, and ejecta of Hesperian- and Amazonian-aged impact craters (e.g., Hobbs et al., 2016), and interpreted these result from contact melting (Morgan and Head, 2009; Jones et al., 2011; Mangold, 2012; Mangold et al., 2012a; Schon and Head, 2012). Fig. 1A shows an example of several fresh fluvial channels associated with the ejecta of two impact craters (26 km and a 12 km diameter) characterized by Mangold (2012). The larger 26 km diameter crater exhibits inlet channels on the rim (red and white arrows in Fig. 1A) and fluvial channels on the ejecta. The 12 km diameter crater in Fig. 1A (on the right) exhibits fluvial channels on the ejecta, but also exhibits fluvial channels within the surrounding terrain, draining away from the ejecta (middle blue and white arrows in Fig. 1A). Fig. 16A and D are two more example of relatively young (Amazonian to Hesperian-aged) and small (19 and 26 km diameter) impact craters identified by Mangold (2012) which exhibit fluvial channels on their ejecta facies. The 26 km diameter crater in Fig. 16D also shows some fluvial channels in the surrounding terrain draining away from the ejecta (Fig. 16E). A larger example is shown in Fig. 17A; this 47 km diameter crater identified by Mangold (2012) exhibits fluvial channels superposing the ejecta facies (Fig. 17B and C). The 63 km diameter Sinton crater (Fig. 17D), characterized by Morgan and Head (2009) is another larger impact crater which exhibits fluvial channels on the crater wall (draining into the crater interior) (Fig. 17E), and on the ejecta facies (Fig. 17F).

Recently, Hobbs et al. (2016) characterized the fluvial channels associated with two Hesperian-aged craters. Fig. 18A shows a 74 km diameter crater characterized by Hobbs et al. (2016) which exhibits fluvial channels on the ejecta (Fig. 18B) and on the interior crater walls (Fig. 18C). Fig. 18D shows the 35 km diameter Choyr crater characterized by Hobbs et al. (2016) which exhibits channels on the interior crater walls draining into the crater interior (Fig. 18E). These particular channels may be fluvial or glacial in origin (Hobbs et al., 2016). Interestingly, a nearby pre-existing ~10 km diameter appears to be infilled and exhibits a large fluvial channel draining out of the crater (Fig. 18F).

As noted by Mangold (2012), most fluvial features observed are on the ejecta facies, while larger craters may exhibit channels draining into the crater interior. Mangold (2012) also found that channels in the surrounding terrain are not observed as frequently, and that most of the fluvial landforms associated with impact ejecta exhibit amphitheater-shaped heads and are not located on topographic highs within the ejecta. Mangold (2012) interpreted these observations to suggest that the fluvial channels were consistent with "local water outbursts" (e.g., Fig. 12C) rather than precipitation. Mangold (2012) also noted that craters with fluvial channels on the ejecta are concentrated in the mid-latitudes (25–45°), which is consistent with crater formation in a regional surface ice sheet which was deposited in the mid-high latitudes during periods of higher martian obliquity (e.g., Head et al., 2006a, 2006b; Madeleine et al., 2009; Fastook et al., 2011). Mangold (2012) concluded that these observations suggest that the fluvial channels were formed by contact melting of hot ejecta on the surface icy deposits. We find the fluvial features around the 35 km diameter Choyr crater (Fig. 18D) to be particularly consistent with a contact melting origin. For example, if surface ice deposits were present at the time of impact and infilled the smaller crater (akin to Amazonian concentric crater fill deposits; Levy et al., 2010; Fastook and Head, 2014), hot ejecta from Choyr landing on the smaller ice-filled crater would be predicted to melt a substantial volume of ice, leading to a fluvial channel draining out of the crater (e.g., Fig. 19F).

Our modeling results support these interpretations. None of these craters are predicted to exhibit basal melting (Fig. 8A and B) due to their small size, and low surface temperature and heat flux. Our models predict that the onset diameter for contact melting in the Hesperian and Amazonian periods is ~30–40 km. Six of the craters discussed above (which are 26 km, 26 km, 35 km, 47 km, 63 km, and 74 km in diameter) are broadly consistent with these results. These craters are predicted to generate substantial amounts of meltwater: ~10^6–10^7 m^3 of meltwater (26 km), ~10^4–10^5 m^3 (35 and 47 km), ~10^3–10^4 m^3 (63 and 74 km). Mangold (2012) found that fluvial channels on the interior crater walls are only found on larger craters. We note that this is a prediction of the contact melting models, which show that larger craters are able to melt substantially greater thicknesses of ice near the rim (Fig. 6), which can generate the ice slope reversal necessary for sustained inflow of contact meltwater into the crater (Fig. 12). The smaller craters (12 and 19 km in diameter) discussed above are smaller than the estimated onset diameters (~30 km) from our thermal models. As noted by Mangold (2012), however, the fluvial features associated with the 12 km diameter crater (Fig. 1A) may actually pre-date the impact. Nonetheless, of the 27 impact craters identified with fluvial landforms on the ejecta by Mangold (2012), nine are smaller than 25 km in diameter. We attribute this difference to a combination of factors: (1) our shock pressures (and post-shock temperatures) neglect the contribution from pore ice, and so our shock heating may be slightly underestimated; and (2) the removal of thick surface ice sheets will decrease the observed crater diameter (Fig. 5). For example, a 30 km diameter crater that formed in a surface ice sheet is predicted to appear to be ~26 km in diameter following the removal of a 300 m ice sheet, or 22 km in diameter following the removal of a 700 m ice sheet (Fig. 5).
Fig. 16. Impact craters which exhibit fluvial channels on the ejecta (red and white arrows), identified by Mangold (2012). These craters are candidate craters for contact melting. (A) 19 km diameter crater (39.8°N, 5.8°E). (B, C) Fluvial channels are located on the ejecta of this crater. (D, E) Sketch map of (B) and (C). CTX image B19_017184_2187. (F) 26 km diameter crater (40.1°N, 5.1°E). (G, H) Fluvial channels are located on the ejecta and within the surrounding terrain. (I, J) Sketch map of (G) and (H). CTX images P03_002139_2209. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
Fig. 17. Larger impact craters which exhibit fluvial channels on the ejecta (red and white arrows). These craters are candidate craters for contact melting. (A) 47 km diameter crater (35.6°N, 0.5°E) identified by Mangold (2012). (B, C) Fluvial channels are located on the ejecta and surrounding terrain. (D, E) Sketch map of (B) and (C). THEMIS IR global daytime and CTX images B18_016815_2151, B03_010657_2168, G23_027153_2179, P17_007519_2147, P16_007374_2183. (F) 63 km diameter Sinton crater (40.7°N, 31.7°E) characterized by Morgan and Head (2009). (G, H) Fluvial channels are located on the ejecta of this crater. (I, J) Sketch map of (G) and (H). THEMIS IR global daytime and CTX images P20_008968_2191, P03_002138_2194. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
Fig. 18. Two examples of Hesperian-aged impact craters characterized by Hobbs et al. (2016) which exhibit fluvial channels on the ejecta and interior walls (red and white arrows). These craters are candidate craters for contact melting. (A) 35 km diameter Choyr crater (32.4° S, 18.6°E). (B, C) Fluvial channels are located on the ejecta and on the crater interior walls. (D, E) Sketch map of (B) and (C). THEMIS IR global daytime and CTX images P16_007400_1414, P15_006965_1419, P16_007255_1411, B17_016340_1402. (F) 74 km diameter crater (39.6° S, 19.1°E). (G, H) Fluvial channels are located on the ejecta and interior walls of this crater. (I and J) Sketch map of (G) and (H). In (H), ejecta appears to be infilling a 10 km diameter pre-existing crater, and a fluvial channel is draining out of the smaller crater, consistent with contact melting of ice deposits within the topographic low of the pre-existing crater. CTX images B22_018186_1473, B17_016485_1458. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
5.2. Candidates for basal melting: Noachian craters

The Amazonian and Hesperian-aged fluvial features associated with impact crater ejecta (<~85 km in diameter and for ice sheets <~1 km thick) can be attributed to contact melting (see Section 4). We now examine impact craters which are candidates for basal melting. Three examples of typical impact craters located in the Noachian-aged southern highlands are shown in Fig. 19. The two craters in Fig. 19A and B exhibit fluvial channels nearly circumferentially around the rim, and all three craters exhibit subdued rimcrests and a lack of observable ejecta. On the basis of their highly degraded state (e.g., Mangold et al., 2012b), these craters are interpreted to be Noachian in age. We first describe the model predictions for craters of this size and age, and then assess whether the basal melt fluxes predicted by the models are sufficient to form fluvial channels. Finally, we assess the unique geomorphological settings that support a contact and/or basal melting origin for these two examples.

Fig. 19A shows a ~180 km diameter crater of Noachian age. A crater of this diameter is predicted to generate ~10^3 km^3 of meltwater through contact melting. A wide range of heat fluxes are sufficient to generate basal melting for a crater of this size (Figs. 8-10) for Ts between 215 and 255 K. For a heat flux of 60 mW/m^2 and an ice sheet 1 km thick, predicted meltwater volumes range up to 400 km^3 for the average rim-height case (Ts = 215 K; Fig. 8A), and up to 20,000 km^3 for the maximum rim-height case (Ts = 255 K; Fig. 9D). The 10 and 100 m thick ice sheet scenarios will not provide any meltwater because contact melting is predicted to fully melt these in the locations where basal melting could occur (except for Ts = 255, where an ice thickness of 100 m provides ~10^2 km^3 of meltwater; Fig. 10A and D). Melt fluxes are predicted to be 3 x 10^-4 to 4 x 10^-3 m^3/yr per m^2. The time-averaged melt flux values given in m/yr per m^2 can be converted to a peak melt flux into the crater by multiplying by the area of the annulus predicted to supply melt. The peak melt flux into the crater interior is estimated to be 14.5-1000 m^3/yr over the course of ~40-250 kyr. This value represents the peak flux because it portrays the case where the entire annulus around the crater is undergoing basal melting (as discussed in Section 3, basal melting is predicted to initiate at the rim where the ejecta is thickest, and continue outwards). For comparison, the mean annual discharge of the Onyx River, Wright Valley, Antarctica between 1970 and 2003 was ~3 x 10^6 m^3/yr (Gooseff et al., 2007). Fig. 19B shows a ~50 km diameter Noachian-aged crater. A crater of this diameter is predicted to generate ~10^2 km^3 of meltwater through contact melting. For Ts = 235 K and the maximum rim-height case (Fig. 9D), this crater may exhibit basal melting for a heat flux of 60 mW/m^2 for surface ice sheets between 100 m and 1000 m thick. Predicted meltwater volumes range from 45 km^3 (100 m ice sheet), up to 350 km^3 (1000 m ice sheet). The 10 m thick ice sheet (and all Ts ≤ 235 K with average rim-height) scenarios will not provide any meltwater because contact melting has already melted these deposits, and the 215 K scenarios do not predict sufficiently high basal temperatures. Melt fluxes are predicted to be 10^-3 m^3/yr per m^2 (Fig. 9E). In these cases (Ts = 235 K, 60 mW/m^2), the maximum radial distance from the rim-crest that basal melting is predicted to occur ranges from 0.2R and 0.4R from the rim (Fig. 9F), yielding peak melt fluxes into the crater of 1.2 m^3/yr to 2.3 m^3/yr over the course of ~50-300 kyr.

Fig. 19C shows a ~62 km diameter Noachian-aged crater characterized by Howard (2007). A crater of this diameter is predicted to generate ~10^3 km^3 of meltwater through contact melting. For Ts = 235 K and the average rim-height case (Fig. 9A), this crater may exhibit basal melting for a heat flux of 60 mW/m^2 for surface ice sheets of ~1000 m thick, producing ~30 km^3 of meltwater. For the maximum rim-height case (Fig. 9D), this crater may exhibit basal melting for a heat flux of 60 mW/m^2 for surface ice sheets between 100 m and 1000 m thick. Predicted meltwater volumes range from 55 km^3 (100 m ice sheet), up to 1000 km^3 (1000 m ice sheet). The 10 m thick ice sheet (and all Ts ≤ 235 K with average rim-height below 1000 m thick ice) scenarios will not provide
any meltwater because contact melting has already melted these deposits, and the 215 K scenarios do not predict sufficiently high basal temperatures. Melt fluxes are predicted to be $10^{-3}$ m/yr per m$^2$ (Fig. 9E). In these cases, the maximum radial distance from the rim-crest that basal melting is predicted to occur ranges from 0.3R and 0.5R from the rim (Fig. 9F), yielding peak melt fluxes into the crater of 3.7 m$^3$/yr to 5.7 m$^3$/yr over the course of ~40–300 kyr.

In order for a basal melting origin to be consistent with the fluvial features, the melt fluxes must be sufficient to generate fluvial incision. Although the melt fluxes predicted for basal melting are low ($\sim 10^{-3}$ m/yr per m$^2$), it is important to note that the young valley networks present around some of the martian shield volcanos (Fassett and Head, 2006, 2007) are interpreted to form by basal melting of snow pack, and have comparable heat fluxes to the basal melting scenario considered here. For example, assuming peak heat fluxes between 130 and 220 mW/m$^2$ (Fassett and Head, 2007), a constant snow thermal conductivity of 1 W/m/K, and a snow thickness of 500 m, peak melt fluxes (from Eq. (6)) are predicted to be between $\sim 4 \times 10^{-4}$ and $\sim 9 \times 10^{-3}$ m/yr per m$^2$ for the valley networks considered by Fassett and Head (2006, 2007) (compared to $\sim 10^{-3}$ m/yr per m$^2$ for the impact craters considered here). In other words, similarly low melt fluxes appear to have plausibly generated fluvial channels on some martian shield volcanos, which suggests that the basal melt fluxes predicted in this study are also sufficient to form fluvial channels.

On the basis of the circumferential interior wall channel morphology observed (red and white arrows in Fig. 19) and model results, we propose that the fluvial channels on the interior crater walls (Fig. 19A and B) could have formed through basal melting of surface ice underlying the crater ejecta. If surface ice were present at the time of impact, basal melting of this deposit is predicted to lead to meltwater transported towards the crater interior (Fig. 2E). The meltwater would then be able to exit and flow down the slopes of the crater walls, where fluvial incision and/or liquid water-assisted debris flows could contribute to channel formation (Fig. 2F). Contact melting could also be responsible for the formation of the interior wall channels (Fig. 6D) in a manner similar to the Hesperian-aged 63 km diameter Sinton crater (Morgan and Head, 2009).

While the fluvial channels circumferentially arranged around the rims of these craters are consistent with either a contact or basal melting origin, their morphology does not exclusively require contact/basal melting to form. That is, other previously proposed formation mechanisms, such as rainfall and surface runoff (Craddock and Maxwell, 1993; Craddock et al., 1997; Craddock and Howard, 2002; Forsberg-Taylor et al., 2004; Howard et al., 2005; Grant et al., 2015; Irwin et al., 2015) or surface snowmelt and runoff (e.g., Howard et al., 2005; Head and Marchant, 2014; Weiss and Head, 2015; Irwin et al., 2015; Hobbs et al., 2016) cannot be ruled out. We identified another characteristic of these typical craters that appears to be exclusively consistent with a contact/basal melting origin. The crater shown in Fig. 19A partially intersected two smaller pre-existing craters located in the south-eastern quadrant of the image (features D and E in Fig. 19A). Two anomalously large valleys associated with these impact craters (features D and E in Fig. 19A) appear to initiate at the crater center, and drain down into the larger crater. In a cold and icy Mars, impact craters are predicted to serve as topographic lows that accumulate large thicknesses of surface ice (akin to the Amazonian debris-covered glacial deposits known as concentric crater fill (e.g., Levy et al., 2010; Fastook and Head, 2014)). Upon impact, the ejecta of the 180 km diameter crater will superpose both the surrounding icy surface and the pre-existing ice-filled impact craters (i.e., CCF). Contact and basal melting of any ice that fills the pre-existing craters (e.g., Fig. 19A) will then produce higher volumes of meltwater (relative to the surrounding terrain), leading to greater degrees of channel incision. This interpretation is identical to our interpretation discussed above for some of the fluvial channels associated with the younger, Hesperian-aged Choy crater (Fig. 18F).

A similarly incised valley associated with an older (\sim 15 km diameter) pre-existing impact crater is located in the north-west quadrant of the 50 km diameter crater (feature F in Fig. 19B). In a similar manner to that shown in Fig. 19A, if this topographically low \sim 15 km diameter impact crater (feature F in Fig. 19B) were filled with surface ice at the time of the impact event that formed the 50 km diameter crater, the ejecta from this event would have infilled the older crater (feature E in Fig. 19B) with hot ejecta, leading to contact and basal melting of the thick ice deposit (i.e., CCF) and enhanced fluvial incision. The 62 km diameter crater in Fig. 19C shows a similar relationship, in which the fluvial channels are concentrated in a pre-existing \sim 38 km diameter crater (feature G) that appears to be infilled by the ejecta of the 62 km diameter crater. In a manner similar to the previous examples, if the pre-existing crater were filled with ice prior to the impact of the 62 km diameter crater, the ejecta from this event would have infilled the older crater (feature G in Fig. 19C), leading to contact and basal melting of the thick ice deposit. Unlike the previous examples, feature G in Fig. 19C appears to exhibit a branching network of channels upstream to the main valley, and is thus distinct from the Amazonian examples (Figs. 16–18) discussed above.

These observations (features D and E in Fig. 19A, feature F in Fig. 19B, and feature G in Fig. 19C) raise the possibility that surface ice deposits were present within the pre-existing craters at the time of impact (within the Noachian period), and that contact/basal melting of these deposits produced the associated channels. These fluvial features appear to be anomalously large when compared with the circumferential wall channels, but they are present on the low-sloped floors of the pre-existing craters as opposed to the steep crater walls. The features of the typical impact craters discussed above (Fig. 19) are unique (intersecting pre-existing impact craters with larger superposing fluvial features on relatively flat topography), and point to a major contribution from contact and/or basal melting. We note that a contact and/or basal melting origin is also plausibly consistent with the other observed crater-associated fluvial features (i.e., circumferential fluvial channels). This then raises the possibility that contact and basal melting could be viable candidate degradation processes for all Noachian-aged impact craters $\sim 15–30$ km in diameter (Figs. 7 and 9D) which exhibit fluvial channels circumferential to the rim (and not just those impact craters which intersect pre-existing low topography).

6. Conclusions

In summary, our calculations predict contact melting of surface ice sheets overlain by hot ejecta to occur over a range of crater diameters, surface temperatures, heat fluxes, and ice thicknesses. This process is predicted to occur for all craters larger than \sim 40 km in diameter which impacted into surface ice deposits, and may melt tens to hundreds of meters of surface ice (Fig. 6) over the course of \sim 100 years to 80 kyr, corresponding to $10^{14}$–$10^{15}$ km$^3$ of meltwater (Fig. 7). Meltwater is generally predicted to flow outwards (Fig. 12A), away from the crater. Fluvial channels on the ejecta may be formed by breaching of the ejecta, analogous to terrestrial springs (Fig. 12C). In the case of ice sheets of comparable thickness to the rim structural-uplift (Fig. 12C), meltwater may flow downwards into the crater (Fig. 12D).

Following the cooling of ejecta, the overlying ejecta facies is predicted to provide sufficient thermal insulation to raise the 273 K melting isotherm through the ice-cemented cryosphere to the base of the ice sheet (Fig. 11). Melting of the pore-ice within the cryosphere will provide a large volume of water for groundwater
recharge, and could contribute to subsurface chemical alteration to form clays in the shallow crust.

The ice-melting isotherm is predicted to reach the base of the ice sheet within ~100 kyr to ~1 Myr after the impact (Fig. 11). If the ice sheet is sufficiently thick and has not been previously removed by contact melting, this can lead to basal melting of the ice sheet, which can last for ~1 kyr to ~1 Myr. Basal melting of surface ice sheets overlain by impact ejecta is predicted to occur under a more confined range of crater diameters, surface temperatures, heat fluxes, and ice thicknesses (relative to contact melting). Basal melting is predicted to occur only at near-rim ejecta thicknesses (generally within ~1R from the rim-crest). Overburden pressure from the overlying ejecta and surface ice is predicted to transport the meltwater up the structurally uplifted rim and toward the crater interior. Basal melting is generally not predicted to occur for typical Amazonian heat fluxes, but may occur for moderately large crater diameters (~ ≥150 km in diameter) in the Early Amazonian and Hesperian (Head et al., 2016), or in areas of anomalously high heat flux. For example, in the higher-end heat flux estimates in the Hesperian period, basal melting is possible for craters as small as ~85 km in diameter. The higher heat fluxes and potentially higher surface temperatures predicted for the Late Noachian period allow basal melting to occur for craters as small as ~5–30 km in diameter. We consider the basal melting process a candidate mechanism for fluvial channels located on the interior walls of ancient impact craters on Mars. The low melt fluxes could alternatively lead to low water:rock ratio flows (e.g., water-assisted debris flows). We conclude that contact and basal melting could operate as background crater degradation processes in a cold and icy early Mars even during periods which lack punctuated warming events (e.g., Hagle and Head, 2014; Head and Marchant, 2014; Weiss and Head, 2015; Wordsworth et al., 2015).

This study offers a theoretical treatment of the conditions required for contact and basal melting to occur as well as a brief case study of several candidate craters. Our models support previous interpretations that contact melting was responsible for the formation of fluvial channels on some Amazonian- and Hesperian-aged impact craters (e.g., Mangold, 2012). We also suggest that the fluvial features associated with three Noachian-aged impact craters are consistent with a contact and basal melting origin: The colocation of large valleys with ejecta-infilled pre-existing craters (features D and E in Fig. 19A, feature F in Fig. 19B, and feature G in Fig. 19C) raises the possibility that surface ice deposits were present within the pre-existing craters at the time of impact (during the Noachian period). In this scenario, contact/basal melting of these deposits could have produced the associated channels. Additional morphologic studies (e.g., Hobbs et al., 2016) should be conducted to evaluate whether additional evidence for the presence of impact ejecta-induced melting exists, particularly within ancient terrains.

Acknowledgments

The authors acknowledge support from the NASA Mars Data Analysis Program (Grant NNX11AI81G) and the Mars Express High Resolution Stereo Camera Team (HRSC) (JPL 1488322) to JWH. We gratefully acknowledge helpful discussions with James Cassanelli, Lauren Jozwiak, and Andy Ryan and insightful comments from two anonymous reviewers.

References


