Formation of an eroded lava channel within an Elysium Planitia impact crater: Distinguishing between a mechanical and thermal origin

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A lava channel identified on the wall of an Elysium Planitia impact crater is investigated to identify the dominant erosion mechanism, mechanical vs. thermal, acting during channel formation. Observations of channel morphology are used to supplement analytical models of lava channel formation in order to calculate the duration of channel formation, the velocity of the lava flowing through the channel, and the erosion rate in each erosion regime considered. Results demonstrate that the channel observed in the Elysium Planitia impact crater formed primarily due to mechanical erosion. In a more general sense, results of this study suggest that lava channels can form primarily due to thermal erosion in the presence of more gradual slopes and more consolidated substrates whereas lava channels can form primarily due to mechanical erosion in the presence of more energetic flows on steeper slopes and more poorly consolidated substrates. Therefore, both erosion regimes must be considered when analyzing origins of eroded lava channels that cut through strata of different strengths.

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1. Introduction

Sinuous channels observed on Mars, the Moon, and Venus have commonly been interpreted to have formed when large volumes of lava flowed rapidly across a surface (Hulme, 1973; Head and Wilson, 1980; Wilson and Head, 1980; Leverington, 2004; Bleacher et al., 2007; Coleman et al., 2007; Mangold et al., 2008; Hauber et al., 2009). The detailed origins of various specific channels have been widely debated, however, and uncertainty still exists about how the mechanical behavior of the flowing fluid influences individual channel formation and development. Potential volcanic channels have been identified that initially formed either as constructed features with marginal levees that constricted flow (e.g., Greeley, 1971; Hulme, 1974; Spudis et al., 1988; Gregg and Greeley, 1993) or as eroded features where the fluid cut into the underlying substrate (e.g., Hulme, 1973; Hulme and Fielder, 1977; Pinkerton et al., 1990; Wilson and Head, 2010), such as the eroded channel observed in an impact crater within the lava plains of Elysium Planitia (Fig. 1).

A channel formed by erosion might have developed either through mechanical erosion as the fluid wore away the substrate (e.g., Gregg and Greeley, 1993; Greeley et al., 1998; Sklar and Dietrich, 1998; Siewert and Fertlito, 2008) or through thermal erosion as hot lava of a given composition flowed over and melted the underlying substrate (e.g., Hulme, 1973; Greeley et al., 1998; Williams et al., 1998; Kerr, 2009; Wilson and Head, 2010). A third erosion regime, thermo-mechanical erosion, differs from thermal erosion in that only a fraction of the substrate must be partially melted before it is carried downstream (e.g., Hulme, 1973; Williams et al., 1998, 2001).

The flow regime of fluids is controlled by their dimensionless Reynolds number defined as $Re = \frac{av}{gd}$ (see Table 1 for definition of parameters). Flow is fully laminar if $Re < \approx 500$ and fully turbulent if $Re > \approx 2000$, with a transition zone between (Crowe et al., 2009). Lava s have been observed or inferred to flow in both laminar (Dawson et al., 1990; Fagents and Greeley, 2001; Greeley et al., 1998; Pinkerton et al., 1990) and turbulent (Hulme, 1973, 1982; Hulme and Fielder, 1977; Wilson and Head, 1980, 2010; Wilson et al., 1985; Williams et al., 1998) regimes. Fluid flowing in the laminar regime will transfer heat via conduction whereas fluid flowing in the turbulent regime will transfer heat via convection, keeping hot material closer to the substrate while also losing heat to both the surface and the atmosphere more efficiently. In both cases an insulating crust eventually forms that extends the distance hot lava travels downstream, enabling more erosion than by open channel flow.

Parameters such as laminar vs. turbulent flow regime, constructional vs. erosional channel origin, and mechanical vs. thermal erosion regime influence the subsequent morphology of an observed channel. For example, a constructed channel generally has detectable levees along the margins of a relatively shallow channel while...
an eroded channel would have little to no topographic expression along the margins of a relatively deep channel that cuts into underlying material. Numerical approaches that simulate mechanical and thermal erosion processes in the presence of both laminar and turbulent flow regimes have been developed to help understand observations of channel morphology (Hulme, 1973; Keszthelyi and Self, 1998; Williams et al., 1998, 2005; Siewert and Ferlito, 2008; Kerr, 2009; Wilson and Head, 2010, in preparation). The current study applies these models to a channel with a morphology consistent with an erosional origin observed on the wall of an Elysium Planitia impact crater (Fig. 1). Results from these models can be used to compare the mechanical and thermal erosion rates required to form the observed channel. The erosion rate responsible for the formation of this channel can then be used to determine the duration of channel formation and to place channel formation into geologic context with the associated volcanic eruption.

2. Theory of lava channel formation by erosion

The two erosion regimes considered in this study are mechanical erosion, which occurs as interactions between flowing lava and underlying material physically wear down the substrate, and thermal erosion, which occurs as hot lava, at a temperature that exceeds the melting temperature of the underlying material, flows over and partially melts the substrate before carrying melted substrate down-slope. The depth of a channel formed under a mechanical erosion regime depends on the shear stress imposed on the substrate by the lava, a factor dependent on the amount and speed
The erosion rate of a channel formed under a thermal erosion regime is given by

$$\frac{d (d_{\text{chan}})}{dt} = K \rho g Q \sin \alpha,$$

where $Q$ is the average lava volume flux per unit width through the channel in $m^2 s^{-1}$, $\rho$ is the lava density (see Table 1 for parameter values), $g$ is the acceleration due to gravity on Mars, $\alpha$ is the ground slope, and $K$ is a dimensional ratio (units of Pa$^{-1}$) between the erodibility of the substrate $b$ and the strength of the substrate $Y_s$ that represents the efficiency of incision into the substrate (Sklar and Dietrich, 1998). Eq. (1) can be thought of conceptually as erosion rate as a function of the erodibility of the substrate and unit stream power $\Omega$, where $\Omega = \rho g Q \sin \alpha$ (Sklar and Dietrich, 1998). The erodibility of the substrate is significantly dependent on substrate composition; for example, a regolith substrate that contains a pore-water ice cement would be expected to be more erodible than a substrate of pure basalt. An alternative approach has been employed previously by Siewert and Ferlito (2008), where the erosion rate is taken to be proportional to the vertical load of the lava on the substrate ($\rho g \cos \alpha$) as opposed to the shear stress of the lava on the bed as used by Sklar and Dietrich (1998). Siewert and Ferlito (2008) predict that mechanical erosion will be most efficient when the regional slope is at a minimum, or when the vertical load is greatest. Erosion rates predicted by the Siewert and Ferlito (2008) model overestimate the amount of erosion observed in the upper and middle segments of the Elysium channel as described in Section 3, and thus the Sklar and Dietrich (1998) stream power formulation that assumes that shear stress is the dominant force responsible for erosion is preferred in this study. Eq. (1) indicates that a mechanically eroded channel will increase in depth faster as a higher flux of lava flows over a more poorly consolidated substrate.

The erosion rate of a channel formed under a thermal erosion regime also depends on the flux of lava through the channel. However, the erosion rate under this regime also depends on the temperature $T$ and viscosity $\mu$ of the lava. A frequently cited model for the rate of change in channel depth in a thermal erosion regime and a turbulent flow regime (Hulme, 1973) is given by

$$\frac{d (d_{\text{chan}})}{dt} = \frac{0.017K^{0.6}}{\rho_s} \left[ \frac{1}{2} \rho_s \frac{g Q}{C_t} \left( \frac{T_e - T_m}{T_e} \right)^{1/3} \left( \frac{T_e}{A} \right)^6 \right],$$

where the subscript $s$ denotes the substrate, $K$ is the thermal conductivity of lava (parameter values as used in Wilson and Head (in preparation); see Table 1), $c$ is the specific heat of the lava and substrate, $q$ is the volume fraction of rock which must be melted before structural integrity is lost, $l$ is the latent heat of fusion required to melt the substrate, $T_e$ is the average temperature of the surface of Mars, $T_m$ is the temperature of the erupted lava, $T_{\text{me}}$ is the melting temperature of a primarily basaltic surface, $A$ is a reference temperature determined experimentally, and $C_t$ is a friction factor that is dependent on $Re$ and is defined in Keszthelyi and Self (1998) by

$$C_t = \left( \frac{1}{32} \right) \log_{10} \left[ 6.15 \left( \frac{2 Re + 800}{41} \right)^{0.92} \right]^{-2}.$$ 

This friction factor formulation is valid only for moderately turbulent lava flows, where $10^3 < Re < 10^5$ (Keszthelyi and Self, 1998). A thermally eroded channel is expected to increase in depth faster as a larger flux of hotter and less viscous lava flows over the substrate. It should be noted that the Hulme (1973) model presented in Eq. (2) probably predicts somewhat greater erosion rates than other models of thermal erosion developed because it underestimates the thickness of the thermal boundary layer and thus overestimates heat flux to the ground (i.e., Williams et al., 2000). It should also be noted that the original formulation for thermal erosion rate presented in Hulme (1973) uses an independent friction factor (developed by McAdams (1954)) that represents flow through an enclosed tube and must be solved iteratively to arrive at a solution. The current study employs the friction coefficient shown in Eq. (3) (developed by Goncharov (1964) and used by Shaw and Swanson (1970) and Keszthelyi and Self (1998)) because it represents sheet flow and open channel flow, a more relevant representation of the

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Symbol</th>
<th>Value</th>
<th>Units</th>
<th>Certainty/value range</th>
<th>Model sensitivity</th>
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<tr>
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<td></td>
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<tr>
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<td>Pa s</td>
<td>0.1–1000 Pa s</td>
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<td>$Y_s$</td>
<td>$1 \times 10^6$</td>
<td>Pa</td>
<td>1 kPa–25 MPA</td>
<td>Very sensitive, constrained by observations, discharge ($Q$), or duration</td>
</tr>
<tr>
<td>Erodibility factor$^c$</td>
<td>$b$</td>
<td>$10^{-3}$</td>
<td>–</td>
<td>$10^{-4}$–$10^{-1}$</td>
<td>Extremely sensitive, not well constrained, most reasonable results for $b = 10^{-3}$</td>
</tr>
<tr>
<td>Proportionality constant for the erodibility of the substrate</td>
<td>$K$</td>
<td>$1 \times 10^{-9}$</td>
<td>Pa$^{-1}$</td>
<td>$10^{-18}$–$10^{-4}$</td>
<td>A ratio of $b/Y_s$, dependent on these factors</td>
</tr>
<tr>
<td>Thermal conductivity of basalt$^d$</td>
<td>$k$</td>
<td>2</td>
<td>W m$^{-1}$ K$^{-1}$</td>
<td>Fairly certain</td>
<td></td>
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<tr>
<td>Specific heat capacity of basalt</td>
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<td>J kg$^{-1}$ K$^{-1}$</td>
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<tr>
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<td>J kg$^{-1}$</td>
<td>Very certain</td>
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<tr>
<td>Fraction of rock melted prior to mechanical erosion</td>
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<td>–</td>
<td>0–1</td>
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<td>K</td>
<td>1335–1450 K</td>
<td>A 100 K change in $T$ changes thermal erosion rate by one order of magnitude</td>
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<td>K</td>
<td>Experimentally derived</td>
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</table>


$^b$ Value range compiled from Arvidson et al. (2004) and Tanaka and Golombek (1989).

$^c$ Value range defined in Zum Gahr (1998).

$^d$ Values for the following constants ($k$, $c$, $l$, $q$, $T_e$, $T_m$, and $A$) are presented in Wilson and Head (2010, in preparation).

$^e$ Value derived in Hulme (1973).
scenario observed in the Elysium Planitia lava channel. In practice, the use of $C_t$ in place of the independent friction factor results in an insignificant quantitative difference of a factor of $\sim 2$, while qualitative interpretations of the transition from thermal erosion to mechanical erosion at a given slope remain unchanged.

The primary parameters needed to calculate the erosion rates represented by Eqs. (1) and (2) are $Q$ and $Re$, where both parameters are dependent on the channel slope $\alpha$, the depth of the lava $d_{lava}$, flowing within the channel (a parameter that must be inferred from observations), and the velocity $v$ of the lava flowing through the channel. The lava velocity in a turbulent flow regime is given by

$$v_{lava} = \frac{gd_{lava} \sin \alpha}{C_t}$$

where $C_t$ is given in Eq. (3) (Keszthelyi and Self, 1998) and the other factors are measured directly or estimated. This equation is only applicable to moderately turbulent flows ($10^3 < Re < 10^5$), but lava flows investigated in this study are shown later to have $Re < 10^5$ so the application is appropriate to this scenario.

An additional parameter that is of interest for an eroded lava channel is the time required for the channel to form. The formation time is given by

$$t = \frac{V_{lava}}{Q_{vol}}$$

where $V_{lava}$ is the volume of lava that flowed through the channel as observed at the channel terminus, and $Q_{vol}$ is the average volume flux ($\text{m}^3\text{s}^{-1}$) of lava flowing through the channel, determined using the models in Eqs. (1)–(3). Solving these equations requires detailed observations of the channel considered, specifically of the channel depth, width, length, and slope. In the following sections, dimensions of the observed channel are distinguished from dimensions of the lava flow within the channel, parameters that are estimated from observations and are not necessarily equal to the dimensions of the final channel observed. Results from the models will be used to place the observed channel into its geologic context, generating a more complete understanding of how the channel formed.

3. Geologic setting/observations

The models introduced in the previous section can be applied to the lava channel observed within an impact crater in the Elysium Planitia lava plains (Fig. 1a). This crater has a diameter of 15 km and is located at the eastern extent of low-viscosity lava flows (160.6°E, 0.9°N) that originated at Cerberus Fossae dikes and flowed through Athabasca Valles, covering nearly 250,000 km$^2$ of Elysium Planitia (Fig. 1b; Fig. 6 in Jaeger et al. (2010)). This lava flowed more than 800 km across the martian surface before completely eroding a ~50 m wide canyon through the crater rim crest and flowing into the crater, leaving troughs preserved in the exterior lava plains adjacent to the crater. As lava flowed into the crater, it accumulated at the base of the crater rim crest, where remnants of lava appear as hummocky deposits outside the currently observed channel (Fig. 2b; black arrow). The upper channel segment itself is at first poorly formed on the steep slope of the crater rim crest, then develops better-defined banks at the base of the crater rim crest. At this location, the channel has a rectangular cross-section, with nearly vertical walls that indicate the presence of more consolidated material such as bedrock. This channel segment has a length of 725 m, a width of 90 m, and it formed on a slope of ~6.2°. The channel depth is unresolved by MOLA topography data as this portion of the channel was not locally crossed by MOLA tracks, but shadow measurements were used with spacecraft incidence (54.1°) and azimuth (143.4°) angles to estimate a channel depth of 32 m in the upper channel segment.

The middle segment of the channel is defined as the portion of the channel that formed parallel to the crater rim (Fig. 2c). This channel segment is also characterized by a rectangular cross-section, indicating that the channel continued to cut more consolidated material such as impact melt that ponded behind crater wall slump terraces. The down-slope wall of the channel displays distinct layers of consolidated material, and the up-slope wall of the channel displays boulders of more consolidated material that fell from farther up the channel wall. Sand dunes can be seen in the thalweg of the channel, indicating that some infilling has occurred since channel formation. This channel segment has a length of 1040 m, a width of 100 m, and it formed on a slightly more gradual slope of ~4.2°. This channel segment was also not crossed by MOLA tracks, but shadow measurements were again used to estimate a channel depth of 28 m.

The lower segment of the channel begins when the channel changes course and extends down the crater wall (Fig. 2d). The channel changes morphology substantially in this section, with a v-shaped cross-section and smoother walls that show evidence of slumping of poorly consolidated material after channel formation. The uppermost portions of the channel walls appear more consolidated than the lower portions, with a cap of steeper, rocky faces overlying the lower talus slopes. This portion of the channel is 3400 m in length and is significantly steeper in gradient, wider, and deeper than previous segments, with an observed slope of ~13.5°, a width of 400 m, and a depth of at least 150 m as indicated from analysis of local channel-crossing MOLA tracks. It should be noted that the channel width is currently greater than during channel formation as slumping has redistributed material from the channel wall to the channel floor. The channel width at the end of channel formation was more likely to be about 275 m, assuming an average angle of repose for sand on Mars of 34.2° (Matijevic et al., 1997). The observed channel dimensions indicate that the total channel volume and thus total volume of material eroded by
Fig. 2. Detailed observations of the channel observed in the wall of the Elysium Planitia impact crater, as shown in HiRISE image PSP_004151_1810. (a) The full channel observed in the crater wall. The channel has been divided into three segments, an upper segment that extends down the crater rim from the breach in the crater rim crest, a middle segment that extends parallel to the crater rim crest, and a lower segment that extends down the crater wall. Observations of channel dimensions and regional slope (Table 2) are used to calculate erosion rates in order to distinguish whether the channel formed from mechanical or thermal erosion. (b) The source region and upper segment of the channel. The source region lies in the surrounding lava plains. As the lava breached the crater rim crest, it flowed from the plains into the crater, leaving the observed troughs in the adjacent lava plains and high lava marks outside the channel (black arrow). The upper channel segment is rectangular in cross-section, consistent with erosion through more consolidated material such as bedrock that preserves initial channel morphology over time. (c) The middle channel segment is also rectangular in cross-section and displays evidence for layering within the down-grade wall, consistent with the erosion of more consolidated material such as ponded and solidified impact melt. (d) The lower channel segment is significantly wider and deeper than the upper two channel segments, and the v-shaped cross-section and slumped walls are consistent with erosion of a less consolidated substrate. Comparisons of calculated mechanical and thermal erosion rates provide insight into how these different channel morphologies developed during channel formation. (e) The terminus of the channel shows a return of channel morphology to a rectangular cross-section as the channel narrows and shallows. Remnants of lava flows outside the channel walls can be observed where this transition occurs (black arrows). This transition in morphology corresponds to a change in slope, indicating that lava overflowed the channel walls at breaks in channel slope.
the flowing lava is 0.075 km², a value that represents 1.3% of the volume of lava that flowed through the channel.

The termination of the channel (Fig. 2e) is characterized by a shallower and narrower channel with a rectangular cross-section. As the slope changes from the larger-scale v-shaped channel to the smaller-scale rectangular channel, remnants of lava flows can be observed once more outside the channel (Fig. 2e; black arrows). These flows radiate away from the channel, indicating flow down-slope towards the crater floor. The locations of the flows outside the upper and terminal channel segments suggest that the channel overflowed where the slope on which it formed changed abruptly from steep to more gradual. The low volume of these deposits suggests that these breaches occurred early in channel formation, declining in frequency as the channel eroded to greater depths.

4. Calculating erosion rates

The observations made of channel dimensions and slope can be used to determine the input parameters needed to solve for lava flow velocity (Eqs. (3) and (4)) and channel erosion rates (Eqs. (1)–(3)). Eqs. (3)–(5) are solved iteratively to calculate the volumetric lava flux and channel formation duration. The calculated formation duration is then used to match the depth of erosion that is observed in the lower channel segment because channel depth is best constrained for that channel segment. The resulting calculated lava flux is held constant in the analysis of the remaining channel segments in order to satisfy continuity of flow through the channel.

As expected, lava depth within the channel changed as the slope and flow velocity were varied. The estimated depths of lava that flowed through the channel are summarized in Table 2. The lava depth is greatest for the most gradually sloped middle segment and least for the steepest lower channel segment, consistent with lava thinning as slope (and thus lava velocity) increases. These depths were estimated assuming a constant lava flow width of 93 m, chosen to match the width of the breach in the crater rim crest where lava initially flowed into the channel. Changing this width does not alter the results qualitatively. Velocities calculated for these flow conditions are summarized in Table 2 and indicate that the lava velocity increased as the channel gradient increased. In each case, the calculated Re (also shown in Table 2) is within the range of values for which this model applies ($10^5$–$10^6$).

These velocities and lava depths have been used together with the measured slopes to calculate the erosion rates required to form the channel in both the mechanical (Eq. (1)) and thermal (Eq. (2)) erosion regimes. Results of these calculations are summarized in Table 2. At more gradual slopes, thermal erosion becomes increasingly more efficient than mechanical erosion, but as slope increases, mechanical erosion is consistently more efficient than thermal erosion.

The modeled eroded depths of these channel segments can be determined using the duration of channel formation (Eq. (5)), where the volume of lava that flowed through the channel is 5.8 km³ as calculated in Section 3, and using the flow parameters defined above that assumed a lower channel segment depth of 150 m. This calculation was performed separately for each erosion regime, as each regime depends on lava depth and velocity in unique ways. For mechanical erosion, the model yields a duration of channel formation of 30 Earth days and a lava flux of $\sim$2300 m³ s⁻¹, and the channel depths are predicted to be 54 m for the upper segment, 32 m for the middle segment, and 150 m for the lower segment. For thermal erosion, the model yields a duration of channel formation of 60 Earth days and a lava flux of $\sim$1200 m³ s⁻¹, and the channel depths for thermal erosion are predicted to be 116 m for the upper segment, 102 m for the middle segment, and 150 m for the lower segment. The duration of channel formation is longer for the thermal erosion model because the erosion rate of the standard lower channel is slower (see Table 2), meaning it takes a longer time period to erode the lower channel segment to 150 m. The calculated channel depths in the thermal erosion scenario are too great in the upper and middle channel segments (observed to be 32 m and 28 m, respectively), suggesting that mechanical erosion was likely the dominant erosion regime present during channel formation.

The mechanical erosion rates responsible for the formation of an eroded channel are heavily dependent on the erodibility factor $K$ of the bedrock, where $K$ is the ratio of an erodibility factor $b$ and the strength of the substrate $Y_s$, and a larger value of $K$ indicates a more easily eroded substrate. Expected values of $b$ for erosive wear by hard particles lie in the range of $10^{-4}$–$10^{-5}$ (Zum Gahr, 1998), and values of $Y_s$ range from 1 to 25 MPa for dry impact ejecta and basaltic basement rock (Tanaka and Golombek, 1989) and from 1 kPa to 1 MPa for poorly consolidated martian soil (Arvidson et al., 2004). The combination of these factors yields a range of $K$ values from $10^{-18}$ to $10^{-4}$. For a given $K$, the model determines the lava flux and time required to erode through that material to match the channel depth observed in the lower channel segment, akin to the method used for thermal erosion. As indicated earlier, steeper slopes correspond to higher flow velocities and unit stream power estimates, so as slope increases, conditions become increasingly optimal for mechanical erosion.
erosion to dominate the formation of the channel. Fig. 3 demonstrates that less consolidated substrates (high $b$, thus high $K$) transition to a mechanical erosion regime at shallower slopes than more consolidated substrates (low $b$, thus low $K$).

The value of $K$ is unique for each geologic setting and must be determined experimentally. Our best estimate of $K$ is $0.5 \times 10^{-6}$ (i.e., a $b$ value of $0.5 \times 10^{-3}$ and an average $Y_s$ of 1 MPa), a value that yields a reasonable lava volume flux ($\sim 2300$ m$^3$ s$^{-1}$) consistent with estimates for the source eruption in Cerberus Fossae (Jaeger et al., 2010), a formation duration of $\sim 30$ Earth days, and a velocity range of 17–25 m s$^{-1}$. Other values of $K$ can also produce reasonable results; for example, a $K$ of $1.5 \times 10^{-6}$ (a $b$ of $1.5 \times 10^{-3}$) yields an expected volume flux of 7100 m$^3$ s$^{-1}$ and a formation duration of $\sim 10$ Earth days. A smaller value of $K$ ($0.1 \times 10^{-6}$, a $b$ of $0.1 \times 10^{-3}$) yields a lower expected flux (500 m$^3$ s$^{-1}$) and a longer formation duration ($\sim 140$ Earth days), and a larger value of $K$ ($2.5 \times 10^{-6}$, a $b$ of $2.5 \times 10^{-3}$) yields a higher expected flux (11,900 m$^3$ s$^{-1}$) and a shorter formation duration ($\sim 6$ Earth days). These scenarios are summarized in Fig. 3. While the value of $K$ changes the estimated lava flux and formation duration for each scenario, it does not change the modeled channel depths obtained for each channel segment.

5. Discussion

5.1. Model results

Observations of channel morphology suggest that as slope increased, the lava became more efficient at eroding into the substrate, carving a deeper and wider channel in the lower channel segment. Model results support these qualitative interpretations of relationships between channel morphology and mechanical erosion rate. The lowest lava flow velocities of 17 m s$^{-1}$ and 20 m s$^{-1}$ correspond with the shallower upper and middle channel segments, respectively, while the highest lava flow velocity of 25 m s$^{-1}$ corresponds to the steeper lower channel segment. Similarly, the highest erosion rates, both mechanical and thermal, occurred in the steeper lower segment, where the lava velocity was highest. The increase in mechanical erosion rate is more significant than that in thermal erosion rates (Fig. 4), indicating that mechanical erosion is more strongly influenced by changes in slope. This more significant change in mechanical erosion rate occurs because the physical energy of the flow increases with an increase in slope, enhancing the efficiency of erosion more significantly in the case of mechanical erosion than thermal erosion, which depends more heavily on the thermal energy of the flowing lava. Thermal erosion becomes more efficient than mechanical erosion at gradual slopes less than about 5°, indicating that thermal erosion is more likely to be observed in channels forming on Mars at these slopes. Channels forming on gradients of about 5° are likely to have very comparable erosion rates at these inferred fluxes in both the mechanical and thermal regimes, suggesting that both erosion regimes are comparably efficient during channel formation under these conditions on the martian surface. Channels forming at these fluxes on gradients greater than about 5° would have significantly higher mechanical erosion rates than thermal erosion rates, indicating that mechanical erosion would be the dominant process present during channel formation on steep martian slopes. The slopes observed in the...
Elysium crater channel are typically greater than 5°, an observation that is consistent with the model results that indicate that mechanical erosion is expected to be the dominant erosion process present during the formation of this channel. Our results, showing the greater importance of mechanical erosion by lava over steeper, fragmental slopes, are consistent with observational and geochemical results supporting an enhanced role for mechanical erosion in the Cave Basalt lava tubes at Mount St. Helens, Washington (Williams et al., 2004).

The validity of the calculated duration of 30 Earth days for the mechanically eroded channel can be confirmed by considering the geologic context of the lava flow source. The lava flows that formed the observed crater channel have been identified to originate from a dike in Cerberus Fossae just north of Athabasca Valles (Jaeger et al., 2010). Analysis of high lava levels within Athabasca Valles using CTX DEMs indicates that these lava flows were locally as deep as 97 m, though the lava depth was likely an average of 30 m on the 250,000 km² lava plains (Jaeger et al., 2010). Calculations using Eqs. (3) and (4) indicate that this eruption was likely to have occurred at an average discharge rate of 1–2 × 10⁶ m³ s⁻¹ (peak discharge rate of 5–20 × 10⁶ m³ s⁻¹), indicating that it would take 1–6 months to erupt the entire 5000–7500 km³ lava volume observed (Jaeger et al., 2010). Therefore, 30 Earth days is a reasonable duration for a channel forming from a lava flow with a flux of ~2300 m³ s⁻¹ that originated from these erupted lavas.

5.2. Stream power and erosion

To further understand the conditions present in the Elysium channel, it is worth considering this channel in the context of other energetic flows. Baker and Kale (1998) examined several such flows which carved deep flood channels, typically in days to weeks. As discussed above (Eq. (1)), the mechanical erosion by lava can be modeled as proportional to unit stream power (Ω = ρgQx, the energy dissipated to the bed) multiplied by a constant of proportionality that reflects the resistance of the surface to erosion. The ‘erodibility’ proportionality constant K is uncertain, and so examining flows of similar energy expenditure can be used to independently assess whether mechanical erosion of the magnitude observed is reasonable.

Table 3 reproduces depths and unit energy expended (stream power per unit width) for several examples as given in Baker and Kale (1998), as well as for the Elysium volcanic channel using channel flux and width estimated in this study. These examples demonstrate that the energy expenditure by the fluid on the Elysium crater wall is directly comparable to extremely large terrestrial floods where erosion was obviously by mechanical means alone. Since there is less buoyancy contrast between lava and surface materials than between water and sediments as well as less resistance to transport because of a lower gravity on Mars than on Earth, less energy is required for a lava flow on Mars to attain similar erosion rates. Thus, the similarity in energy expenditure between the Elysium channel and the terrestrial floods carved by purely mechanical means suggests that our parameterization of surface erodibility is reasonable (and possibly conservative). This

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Table 3
Unit stream powers and eroded depths for select channels.

<table>
<thead>
<tr>
<th>Channel</th>
<th>Depth (m)</th>
<th>Stream power (W m⁻²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chuja²</td>
<td>400</td>
<td>10²–10⁶</td>
</tr>
<tr>
<td>Rathdrum¹</td>
<td>175</td>
<td>2 × 10⁵</td>
</tr>
<tr>
<td>Grand Coulee¹</td>
<td>100</td>
<td>3 × 10⁵</td>
</tr>
<tr>
<td>Elysium volcanic channel</td>
<td>150</td>
<td>4 × 10⁵</td>
</tr>
</tbody>
</table>

¹ Flood channels carved by water on Earth, from Baker and Kale (1998).
² Parental flows of Elysium crater wall are directly comparable to extremely large terrestrial floods where erosion was obviously by mechanical means alone.
supports the view that mechanical erosion is an important mechanism to consider in channel formation.

6. Conclusions

Modeling the formation of a lava channel observed on the wall of the Elysium Planitia impact crater suggests that the channel formed over a period of ~30 Earth days as lava flowed into the crater from surrounding plains of lava that originally erupted from a dike within Cerberus Fossae. This lava flowed at velocities that range from 17 to 20 m s⁻¹ at the top of the channel, carving a relatively shallow channel of 45–70 m depth, to 25 m s⁻¹ at the base of the channel, carving a deeper channel of 150 m depth. The slopes observed along the channel were found to influence the modeled mechanical erosion rates significantly. As slope increased from ~4° to ~13.5° along the length of the channel, mechanical erosion became more efficient, resulting in the formation of the wider and deeper v-shaped channel observed in the lower channel segment as compared with the narrower and shallower rectangular channel observed in the upper and middle channel segments. The preserved rectangular cross-section and layering observed within the channel walls in the upper and middle channel segments suggest the presence of a more consolidated substrate (such as bedrock or impact melt that ponded behind crater wall slump terraces) in the upper regions of the crater wall. In contrast, the v-shaped cross-section and collapsed channel walls of the lower channel segment suggest that the materials in the lower regions of the crater wall were less consolidated and were thus more susceptible to mechanical erosion.

The results of this study demonstrate that lava channels can form primarily due to mechanical erosion in the presence of more energetic flows on steeper slopes and less consolidated substrates on Mars, though thermal erosion can also be an influential process in the presence of more gradual slopes and a more consolidated substrate. Therefore, both erosion processes must be considered when extending the investigation of erosional lava channel formation to other planetary bodies such as the Moon and Venus.

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References


