Role of impact excavation in distributing clays over Noachian surfaces

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[1] Spectrometers have detected clay-bearing units in and on much of the ancient Martian crust. Geothermally heated aquifers in basaltic rock provide conditions conducive to forming Fe/Mg phyllosilicates at depth. Throughout the Noachian period, a high flux of km-scale bolides excavated buried materials and distributed them over the surface. We use the Maxwell Z-model to quantify the volume and final location of excavated clay-bearing material. We focus on two potentially detectable properties: the volume of clay-bearing material ejected as a fraction of total ejected volume and the volume percentage of clay-bearing material in the ejecta as a function of distance from the crater’s rim. Generally, the volume percentage of clays in the ejecta is greatest for craters less than 25 km in diameter. Larger crater sizes incorporate a higher fraction of clay-poor material because they excavate to greater depths at which clays are likely absent. Specific trends in bulk clay volume fraction and the distribution of clay fraction across the ejecta deposit as a function of crater size depend on the depth to the clay-rich layer and its thickness. Impact excavation likely explains clays associated with ejecta deposits and may reveal clues about the volatile content and stratigraphy of the upper Noachian crust.

Applying our model to the Mawrth Vallis region suggests that a clay layer a few hundred meters thick is buried at the ~3000 m elevation contour. Given that clay layers are likely thin and buried in the upper 3 km of the crust, we predict that small to mid-sized craters (<25 km) will best exhibit detectable amounts of clays and that these clays will be most abundant in the crater wall and rim and less so in distal ejecta.


1. Introduction

[2] The detection of minerals formed by aqueous alteration of basaltic rock complements geologic evidence for water’s presence on the surface of ancient Mars [Craford and Howard, 2002; Howard et al., 2005; Moore and Howard, 2005; Poulet et al., 2005; Bibring et al., 2006; Bishop et al., 2008; Mustard et al., 2008; Ehlmann et al., 2008; Murchie et al., 2009]. These aqueous minerals are indicators that help reveal the geochemical, atmospheric, and hydrologic nature of early Mars. Of the 10 mineral classes detected so far, the so-called “deep phyllosilicates” are the most abundant, with the Fe/Mg-O phyllosilicates smectite and nontronite being the most common assemblages [Mustard et al., 2008; Murchie et al., 2009]. Mustard et al. [2008] argued that the hundreds of detections of Fe/Mg phyllosilicates in rims, ejecta, and central peaks of craters in the southern highland Noachian cratered terrain indicate excavation of altered crust from depth. On the basis of the number of identifications in multispectral mapping coverage examined to date, and extrapolating over the Noachian-aged southern highlands, there may be 5,000 to 10,000 locations where deep phyllosilicates are detectable [Mustard et al., 2008]. Questions concerning the formation, geologic placement, and preservation of these clays remain unanswered.

[3] The logic suggesting that clays formed at depth and were then excavated to the surface by impacts is as follows. First, the Martian crust likely contains significant quantities of ice and water [e.g., Clifford, 1993]. Second, the formation of Fe/Mg phyllosilicates suggests a relatively low level of alteration and possibly low-grade metamorphism [Bishop et al., 2008; Murchie et al., 2009]. High-temperature alteration driven by volcanic activity is not required. Third, most Fe/Mg phyllosilicate detections are associated with impact-related features: ejecta deposits, rims, blocks, breccias, and megabreccias [Mustard et al., 2008; Ehlmann et al., 2008; Bishop et al., 2008; McKeown et al., 2009; Mustard et al., 2009b]. Fourth, the Noachian period experienced a high flux of impact cratering [Hartmann and Neukum, 2001]. The most abundant clay class detected on Noachian surfaces, the Fe/Mg phyllosilicates, need not have formed at the surface.

[4] This paper quantifies the hypothesis of impact excavation for the delivery of preexisting clay-rich material to the surface; we do not assess clay formation itself. We use an...
impact excavation model, detailed in section 2, to quantitatively follow clay-bearing materials from various subsurface locations to their final ejected location. Our results, described in section 3, focus on how much clay-rich material is excavated, where on the ejecta deposit it lands, and in what quantity. We make predictions for the detection of clays in section 4. Our conclusions are stated in section 5.

2. Model Setup

We seek to quantify the redistribution of material from the subsurface to the surface by impact excavation for different crater sizes, clay layer thicknesses, and clay layer burial depths. We employ the Maxwell Z-model, an analytical model informed by observations from explosion craters [Maxwell and Seifert, 1974; Roddy, 1977] which describes the subsurface flow field for vertical impacts [Maxwell, 1977].

Although computational hydrodynamic models provide a more detailed description of impact excavation [e.g., Senft and Stewart, 2008], the Maxwell Z-model describes many experimentally observed features [Stewart and Valiant, 2006; Richardson et al., 2007]. Figure 1, a schematic of our model, illustrates many of the important concepts: the transient and final crater radii, subsurface layers rich with volatiles and/or clays, and the excavation cavity populated with streamlines that follow ballistic trajectories and build the ejecta deposit.

The Maxwell Z-model approximates the flow field using three basic assumptions [Maxwell and Seifert, 1974; Croft, 1980; Stewart and Valiant, 2006; Richardson et al., 2007]: (1) independent ballistic trajectories can be calculated for ejecta, (2) flow below the ground plane is incompressible, and (3) the radial velocity below the ground plane is given by \( u_r = \alpha (r/r_0)^Z \), where \( r \) is the radial distance from the effective origin of flow, \( \alpha \) is a measure of the strength of the flow field, and \( Z \) determines the shape of the flow field and the change of velocity with increasing radial distance. As we discuss below, these assumptions lead to a definition of the transient crater radius, the excavation cavity, and descriptions of individual streamlines. Materials moving along a particular streamline follow the same path and are ejected at the same radial distance from the crater’s center with the same velocity and trajectory.

2.1. Limitations, Assumptions, and Caveats

The Z-model is fundamentally limited by its neglect of interactions between the streamtubes [Melosh, 1989]. It is an approximate model and should not be used to resolve finer details. We used a simple form of the Z-model by (1) placing the flow field’s origin at the surface, (2) setting \( Z \) constant for all streamlines, (3) setting an \( \alpha \) constant for all streamlines associated with a particular crater size, and (4) assuming a vertical impact angle of 90° from the surface. This allows a full description of crater flow but fails to conserve energy [cf. Melosh, 1989]. Velocities of streamlines near the
crater’s center are unrealistically fast but constitute a small volume. Holding \( \alpha \) and \( Z \) constant allows one to explicitly evaluate the flow field, calculate the angle of ejection, and explicitly calculate the ejecta deposit [Croft, 1980]. In spite of its faults, the \( Z \)-model gives a reasonably accurate representation of the gross features of the cratering flow [Maxwell, 1977; Croft, 1980; Melosh, 1989]. Several further refinements can be made to tailor the \( Z \)-model to specific impact events to increase the accuracy in describing the excavation flow and the emplacement of the rim and ejecta deposit [Richardson et al., 2007].

[9] The best match between numerically calculated flow fields and the predicted \( Z \)-model flow fields occur when the \( Z \)-model flow field origin is placed beneath the target surface [Anderson et al., 2001; Thomsen et al., 1980; Austin et al., 1980]. The origin for our model is at the surface. This may mischaracterize flow for streamtubes that leave the ground surface plane near the center of the crater but still provides a reasonable quantitative approximation for particle movements for \(-70\text{–}90\% \) of the ejected volume [Croft, 1980]. Recreating a realistic excavation depth as a function of final crater diameter was our primary concern: The excavation depth controls the extent to which the subsurface is sampled, and the final crater diameter ties surface observations to the impact excavation and distribution hypothesis.

We find that our model, with \( Z = 2.71 \) and \( \alpha \) determined as a function of crater size (see equation (3) below), yields model excavation depths that overestimate theoretical excavation depths (\( h_{\text{exc}} = 0.1 \, D; \text{Melosh [1989]} \)) by less than a factor of 2.5\% at all crater diameters.

[10] In addition to the simple form of the \( Z \)-model employed here, we also neglect several additional physical effects that might be important. These effects are discussed below and include the role of subsurface volatiles, the atmosphere, and mixing with surface materials. Some of these effects are most pronounced in distal ejecta deposits, which we explicitly excluded from our calculations. None of the effects are likely to qualitatively change our results, though the quantitative details may be affected.

2.1.1. Volatiles

[11] Our model does not include the effects that subsurface volatiles may have on the crater formation process. Theoretical considerations [Clifford and Hillel, 1983], fluvial and pluvial landforms [Carr, 1996], and neutron spectrometer data [Feldman et al., 2002] indicate significant quantities of water and ice within the Martian crust. The Martian crust is likely heterogeneous with basalt-rich, volatile-rich, and clay-rich layers [Edgett and Malin, 2004; Beyer and McEwen, 2005]. These materials’ geologic strength, damage, density, and dilatancy have a significant effect on crater formation processes [O’Keefe et al., 2001; Collins et al., 2004; Osinski, 2006; Senft and Stewart, 2008]. Targets with layers composed of differing materials further complicate crater formation due to shock impedance, density contrasts, and strength mismatch between layers [O’Keefe et al., 2001; Senft and Stewart, 2008].

[12] Numerical simulations of crater formation showed that the presence of an ice layer modifies the excavation volumes, ejection angles, and ejection velocities of ejecta [Senft and Stewart, 2008]. In these simulations, adding a surface ice layer does not change the average launch angle of ejecta with distance; ejection angles range from 60° to 30°, with the angle decreasing with distance [Senft and Stewart, 2008]. A buried ice layer, however, leads to much higher ejecta angles on the order of \(-70^\circ\text{–}80^\circ \) [Senft and Stewart, 2008]. This causes most of the ejecta to be deposited near the rim. It is important to differentiate between ice layers, ice-rich layers, and nonicy weak layers. All layer types modify the excavation and collapse process, but the effects of ice layers are much more dramatic than weak sediments [Senft and Stewart, 2008]. This behavior is related to the density and coefficient of friction of ice (1 kg/m\(^3\), 0.2) being roughly a factor of 3 less than that of sediment or rock (\(-3\) kg/m\(^3\), 0.6) [Holsapple, 1993; Senft and Stewart, 2008]. Despite these complications, however, the general pattern of redistribution of material is unchanged: Inner streamlines deliver material to distal reaches of the ejecta deposit, and outer streamlines deliver material to proximal locations. Thus, although the quantitative details of our model results may be affected by the presence of subsurface volatiles, the qualitative results are unlikely to change.

2.1.2. Atmospheric Interactions

[13] Does the simple ballistic formulation used to trace back ejecta to the \( Z \)-model streamtubes provide reasonable estimates of the distribution of clay-rich material on the ejecta deposit, given possible interactions with an atmosphere? Studies involving laboratory experiments and numerical simulations have shown that the presence of an atmosphere affects crater ejecta [Schultz and Gault, 1979; Barnouin-Jha and Schultz, 1996; Barlow and Perez, 2003; Barnouin-Jha et al., 2005]. Strong winds form as a result of a vortex generated by an ejecta curtain that advances into the surrounding atmosphere [Barnouin-Jha and Schultz, 1996]. The winds preferentially entrain fine-grained material and deliver it to distal reaches and may possibly form ejecta lobes [Barnouin-Jha and Schultz, 1996; Baratoux et al., 2005]. However, ballistic emplacement probably dominates deposition in the proximal ejecta blanket [Barnouin-Jha et al., 2005]. Since the proximal region contains the majority of the ejecta mass, we neglect the effects of atmospheric effects in our calculations.

2.1.3. Mixing

[14] In our estimates of the bulk volume fraction of clay-rich material within the ejecta deposit, we ignore contributions from mixing between the ejecta deposit and surface materials. The proportion of primary crater ejecta within the ejecta blanket decreases with greater radial distances [Pieters et al., 1985]. A study of the mixing of surface material with ejecta at the Ries crater found that the proportion of primary ejecta within the blanket dropped below 50\% at a radial distance of roughly three times the excavation cavity radius [Hörz et al., 1983]. To avoid the complications of mixing and atmospheric interactions, below we limit our estimates of volume fractions within the ejecta deposit to \(-2.5 \, R_e \).

2.2. Defining the Transient Crater Radius

[15] Given a particular final crater radius, \( R_f \), we define the transient crater radius, the point at which radial growth ceases during an impact event before gravitational collapse, as \( R_t = 0.65 \, R_e \). Traditionally, crater scaling relationships cite \( R_t = 0.6 \, R_e \) [Melosh, 1989]. However, the Maxwell \( Z \)-model assumes no postcrater relaxation; a slight increase
in this ratio, up to \( R_t = 0.7 \, R_i \), generates an ejecta deposit that better reproduces the geomorphology of Martian craters [Stewart and Valiant, 2006]. Using a simple form of the \( Z \)-model, we set the depth of the effective center of \( Z \)-model flow (EDOZ) to 0 [Croft, 1980] and held \( \alpha \) and \( Z \) constant. We adopted the descriptions of Maxwell [1977] and Croft [1980] for the transient crater radius as the location, \( r_1 \), that particles ejected at ranges both smaller and larger than \( r_1 \) return to the original surface plane at ranges larger than \( r_2 \).

\[
\begin{align*}
  r_1 &= \left( \frac{4Z(Z - 2)}{g^2} \right)^{1/3} \\
  r_2 &= r_1 \left[ 1 + \frac{1}{2Z} \right]
\end{align*}
\]  

(1)

The gravitational acceleration, \( g \), on Mars is 3.71 m s\(^{-2}\), and \( Z \) is chosen to be 2.71. Our choice of \( Z = 2.71 \) corresponds well with observed excavation cavities [Croft, 1980; Stewart and Valiant, 2006] and is consistent both with laboratory-scale impacts [Austin et al., 1980] and with ejecta blanket studies [Hörz et al., 1983]. Assuming that \( \alpha \) and \( Z \) are independent of time, we can now solve for \( \alpha \) by setting \( r_1 \) to the transient crater radius \( R_t \) and rearranging equation (1) as follows:

\[
\alpha = \left( \frac{gR_t^{2Z+1}}{4Z(Z - 2)} \right)^{1/2}
\]  

(3)

2.3. Populating the Excavation Cavity with Streamlines

[16] Figure 1 illustrates how we define streamlines in polar and axisymmetric coordinate systems. We populate the interior of the excavation cavity with 100 streamtubes—a value that yields adequate resolution and high computational efficiency. These 100 streamlines are defined by evenly spaced intersections, \( R_n \), with the horizontal. For example, a crater with a final diameter of 45 km has a transient radius, \( R_t \), of 14.6 km and streamlines spaced every 145 m. In polar coordinates, all streamlines interior to the excavation streamline are defined as follows:

\[
s_r = R_n(1 - \cos \theta)^{1/2} 
\]  

(4)

where \( \theta \), the angle from the vertical downward axis is varied from 0 to \( \pi \) [Croft, 1980; Anderson et al., 2001; Stewart and Valiant, 2006]. The streamline that intersects the horizontal surface plane at the transient crater radius, \( R_n \), defines the boundary of the excavation cavity. We assume that stratigraphic layers containing clays or volatiles are horizontal and convert the streamlines into axisymmetric coordinates as defined below:

\[
\begin{align*}
  r &= s_r \sin \theta = R_n \sin \theta(1 - \cos \theta)^{1/2} \\
  z &= s_r \cos \theta = R_n \cos \theta(1 - \cos \theta)^{1/2}
\end{align*}
\]  

(5)

2.4. Volume Calculations

[17] As shown in Figure 1, the lowest streamline, the streamline that intersects the horizontal at \( r_1 \), determines both the excavation depth and the total volume of material ejected. The total volume of material ejected, \( V_{\text{ejected}} \), can be calculated analytically as follows [Croft, 1980]:

\[
V_{\text{ejected}} = \frac{2\pi(Z - 2)}{3(Z + 1)} \, R_t^3
\]  

(9)

which for \( Z = 2.71 \) implies \( V_{\text{ejected}} = 0.4 \, R_t^3 \). For reference, Melosh [1989] indicated that \( V_{\text{ejected}} \approx R_t^3 \), and the 85 m diameter Prairie Flat TNT-formed explosion crater exhibited a relationship of \( V_{\text{ejected}} = 0.46 \, R_t^3 \) [Roddy, 1977; Croft, 1980]. Figure 1 shows how two adjacent streamlines define a streamtube. We calculate the total volume of each streamtube, as well as the volume of various layers intersected by that streamtube iteratively, in axisymmetric coordinates.

2.5. Ballistics and Building the Ejecta Deposit

[18] Each streamline is ejected from the horizontal at the same angle, which is set by the value of \( Z \) [Maxwell, 1977]. Given our choice of \( Z = 2.71 \), the ejection angle, \( \phi \), is 35.4°.

\[
\phi = \tan^{-1}(Z - 2)
\]  

(10)

In the Maxwell Z-model, the ejection velocity depends on the distance from the center of the crater, \( r \) (equation (1)), with the greatest velocities at the crater’s center [Maxwell, 1977]:

\[
\begin{align*}
  u_r &= \frac{\alpha}{r^2} \\
  u_z &= (Z - 2)u_r \\
  u &= \sqrt{u_r^2 + u_z^2}
\end{align*}
\]  

(11-13)

where \( u_r, u_z, \) and \( u \) are the radial, vertical, and total magnitudes of the ejection velocities. The ballistic range of each streamtube is defined by the velocities and trajectories of its...
Figure 2. A schematic of the ballistic transport and emplacement of excavated material as described in section 2.5. Two bounding streamlines define each streamtube. The choice of $Z$ determines the ejection angle of streamlines. Material contained within a given streamtube is ballistically ejected and contributes to the thickness of the ejecta deposit over the annulus on which it lands.

Bounding streamlines. We define the ballistic range by using a familiar arrangement of equation (2):

$$r_2 = \frac{u^2 \sin(2\phi)}{g}$$  \hspace{1cm} (14)

Inner streamlines land further downrange than outer streamlines. Although our choice of $Z = 2.71$ provides an accurate estimate for crater excavation depth, it yields an ejection angle (34.5°) that is roughly 5° shallower than average ejection angles observed in laboratory and exploration experiments [Croft, 1980; Austin et al., 1980; Cintala et al., 1999; Anderson et al., 2003]. An increase of ejection angles up to 45° will distribute material farther away from the crater and result in a more gradual reduction in ejecta thickness with distance. To bring our model into closer agreement with these results, and because there will in reality be a range of ejection angles, we introduce an ad hoc spread in the ejection angle, $\Delta \phi$, of 5°. Correspondingly, each streamline has two ballistic ranges as illustrated in Figure 2. The second range can be shown to be:

$$r'_2 \approx r_2 \left(1 + \frac{2\Delta \phi}{\tan 2\phi}\right)$$  \hspace{1cm} (15)

We define the annulus over which the volume of a particular streamtube lands by the minimum range ($r_2$) of the outer streamline and the maximum range ($r'_2$) of the inner streamline.

[19] The volume of ejecta is calculated in 15 discrete bins. Each bin width is 0.5 ($R_f - R_t$). We placed the bins side-by-side from the transient crater radius out to an arbitrary but adequate extent of 2.625 $R_f$. The volume of clay-poor material and the volume of clay-rich material contained within a streamtube is calculated. We assume that this material is deposited evenly over its annulus. Figure 2 shows that this annulus may cover one or more ejecta bins. The material is allocated to each bin within its range annulus as a percentage of the area that each bin or bin portion contains. The thickness of the ejecta deposit is calculated by dividing the total volume of collected ejecta in each bin by the area the bin spans assuming axisymmetry. Figure 1 shows how the volume contained within streamtubes is ballistically delivered downrange and builds the ejecta deposit. Material that lands between $R_t$ and $R_f$ is assumed not to contribute to the ejecta deposit because of postimpact slumping; its thickness is shown in Figure 1 as a dashed line.

[20] Our method only calculates the thickness of the ejecta deposit; we neglect subsurface contributions to the ejecta deposit near the rim due to structural uplift. Material that lands near the rim may maintain its stratigraphy in an overturned flap. We did not include calculations for postballistic radial mass movement, a phenomenon observed at rampart-type craters [Barnouin-Jha and Schultz, 1996]. We also assume that the ejecta is internally well-mixed because of its high velocities [Melosh, 1989].

[21] Our calculated volume fractions for clay-bearing materials in the ejecta deposits are upper limits because we do not include mixing of surface materials. This becomes more important at distal reaches of the ejecta deposit. Drill core samples obtained from the Ries impact crater ejecta deposit (26 km diameter, south Germany) indicate that the volume fraction of surface materials incorporated into the ejecta increases with distance from the rim [Hörz et al., 1983].

2.6. Parameter Space: Crater Size, Clay Layer Thickness, and Burial Depth

[22] In this treatment, the Maxwell Z-model is only appropriate for gravity-dominated, complex craters [Richardson et al., 2007]. On Mars, crater diameters in the range of 5–8 km mark the transition diameter from simple to complex [Pike, 1980]. We model the effects for crater diameters at 1 km intervals from 5 to 200 km. We do not model crater diameters larger than 200 km because the application of the Maxwell Z-model and associated ballistic calculations become inappropriate in a nonuniform gravitational field [Melosh, 1989].

[23] The formation of Fe/Mg phyllosilicates depends on pressure, temperature, and water availability [Griffith and Shock, 1997]. Although the Martian crust likely contains significant quantities of ice and water [e.g., Clifford, 1993], vadose zones and cryospheric regions would inhibit clay formation. However, in the past (4.1 Gyr B. P.), higher concentrations of radiogenic elements would have maintained a higher geothermal flux at roughly 60 mW m$^{-2}$ [Nimmo and Tanaka, 2005]. With sufficiently high permeabilities (10$^{-12}$ m$^2$), this background geothermal heat alone may drive hydrothermal convection [Travis et al., 2003], thinning the cryosphere and increasing chemical reaction rates.

[24] Assuming that lithostatic pressure reduces pore space exponentially with depth and closes pores beyond depths of ~10 km [cf. Toon et al., 1980; Clifford and Hillel, 1983], water-rich regions are constrained to the upper 10 km of crust. Higher water-to-rock ratios are likely further constrained to the upper ~7 km of crust, where higher porosities allow higher permeabilities and enhanced flow [Barnhart et al., 2010]. Given these loose constraints, we model a wide range of reasonable clay layer thicknesses and burial depths.

[25] We set clay layer thicknesses, $\Delta z$, to 10, 33, 100, 333, and 1000 m and set clay layer burial depth, $d$, to 10, 33, 100, 333, 1000, and 3333 m below the surface. The parallel dashed lines in Figure 1b define a 1 km thick layer residing...
3. Results

Impact excavation is capable of delivering significant percentages of clay-rich material from the subsurface to the surface of Mars. Here we show how crater diameter, clay layer thickness, and clay layer burial depth affect the total volume fraction of clay-rich material ejected and the volume fraction present throughout the ejecta deposit. We discuss the issue of shock heating and its consequences for clay preservation in ejected material in section 4.2.

3.1. The Volume Percentage of Clay-Rich Material Ejected

Figure 3 plots the volume fraction of clay-rich material ejected by an impact event as a function of final crater diameter, clay layer thicknesses, and burial depth. The volume fraction of clay-rich material increases with increasing clay layer thickness (color-filled envelopes) and decreases with increasing layer burial depth (solid lines marked by depths).

The dotted line in Figure 3 shows an example with a 333 m thick clay-rich layer buried 1000 m below the surface. Excavation depth increases with crater size (equation (8)), so the volume fraction for a given burial depth and thickness remains at zero until a crater excavates to that layer depth. In the example case, no clays are excavated by craters smaller than 14 km. For craters larger than this value, the volume fraction increases with increasing crater diameter until the excavation depth slightly exceeds the base of the clay-rich layer. Beyond this crater size (Df ≈ 28 km in this case), the volume fraction asymptotically decreases as increasing amounts of clay-poor material contribute to the total volume of ejected material.

3.2. The Effects of Crater Size

Crater size is the most critical parameter affecting the delivery of subsurface clay-rich material to the surface. Figure 4 shows the excavation cavity, rim location, and volume fraction of clay-rich material in the ejecta deposit as a function of distance from the rim for four crater sizes: 6, 9, 30, and 60 km. A hypothetical clay-rich layer with a thickness Δz = 333 m resides below 100 m of clay-poor material. The excavation field of small 6 km diameter craters reaches the shallowly buried clay-rich layer (Figure 4a). When a craters’ excavation depth reaches but does not exceed the clay-rich layer, outer streamtubes intersect significant amounts of clay-bearing material. These streamtubes, ejected at lower velocities than inner streamtubes, travel shorter distances and build the proximal ejecta deposit. Consequently, proximal ejecta exhibits a higher fraction of clay-rich material than the distal ejecta (Figure 4a).

As crater size increases, the excavation depth exceeds the clay-rich layer (Figures 4a–4d). The total amount of clay-rich material ejected continues to increase, but the volume fraction decreases. The region of the ejecta deposit with the highest fraction of clay-rich material shifts radially outward, away from the rim. The deepest, outermost streamtubes become less enriched with clays because they sample increasing amounts of deep clay-poor material. The volume fraction of clay-rich material at the rim becomes reduced at greater crater diameters.

3.3. The Effects of Clay-Rich Layer Depth

The depth to the clay-rich layer affects the volume of clays delivered to the surface and the distribution of clay-rich material across the ejecta deposit. Figures 3 and 4 show that, for a given crater size, the total amount of clay-rich material delivered to the surface decreases with increased burial depth. As discussed above, the depth of the clay-rich layer...
layer with respect to the excavation depth affects the volume distribution of clays on the ejecta deposit. If the layer is deeper than the excavation depth, then no clays are delivered to the surface. Figure 4b demonstrates a special case: The excavation depth is slightly deeper than the base of the clay-rich layer. This causes the volume fraction of clay-rich material to peak neither at the rim nor at distal reaches, but between the crater rim and one crater radius beyond the rim.

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**Figure 4.** (a–d) Crater size exerts the strongest influence on the delivery of clay-rich material to the surface. Here four final crater diameters—6, 9, 30, and 60 km—excavate a region with a 333 m thick clay-rich layer (parallel dashed lines) buried 100 km below the surface or (e and f) a 33 m layer buried 10 m. The final radius is marked by a solid vertical line. In the upper right corner of each image, the ejecta deposit thickness and the relative thickness of clay-rich material are marked by the solid and dashed lines, respectively. The volume fraction of clay-rich material in the ejecta deposit as a function of distance from the rim is indicated by the dashed-dotted line in the lower right corner of each image. When the excavation depth terminates in the clay-rich layer, the rim has the highest volume fraction of clay-rich material in the excavation deposit (Figure 4a). The volume fraction peak in the ejecta deposit shifts radially away from the rim as crater diameter and corresponding excavation depth increases (Figures 4b–4d). Figure 4b shows a special case: The excavation depth of the 9 km crater, about 100 m greater than the base of the clay-rich layer, leads to the greatest volume fraction of clay-rich material on the ejecta deposit. Crater diameters with excavation depths that penetrate well beyond the clay-rich layer exhibit a reduced volume fraction that is greatest in the distal reaches of the ejecta deposit (Figures 4c–4f). Ejecta-related length scales and volumes scale with crater size (Figures 4d and 4e). Note: vertical exaggeration varies between panels.
3.4. The Effects of Clay-Rich Layer Thickness

The thickness, \( \Delta z \), of the clay-rich layer determines the total amount of clay-rich material available for delivery to the surface. Figures 4a and 4e and Figures 4d and 4f demonstrate how \( \Delta z \) affects the total and volume fractions of clay-rich material that build the ejecta deposits for 6 and 60 km diameter craters. The total volume of clay-rich material delivered to the surface for 60 km diameter crater increases by an order of magnitude as \( \Delta z \) increases from 33 to 333 m.

Figures 4d and 4e illustrate a consequence of the geometry of the Maxwell model: Many aspects of impact excavation scale with crater size. Figure 4d shows a 60 km crater excavating a crust with a 333 m thick clay layer buried 100 m below the surface. Figure 4e shows a 6 km crater excavating a crust with a 33 m thick clay layer buried 10 m below the surface. While crater length scales (excavation depth and ejecta deposit thickness) shrink by a factor of 10, volumes are reduced by a factor of 1000. The total volume of clay-rich material excavated decreases from \( 3.7 \times 10^{11} \) to \( 3.7 \times 10^8 \) m\(^3\), but the volume fraction of clay-rich material, 13%, remains the same.

3.5. Maximum Clay-Rich Material Percentages in Ejecta Deposits

As discussed throughout section 3, the volume fraction of clay-rich material is greatest for some cases at the rim and for others at the distal reaches of the ejecta deposit. Figure 5 plots the maximum volume fraction of clay-rich material calculated in the ejecta deposit, regardless of location, as a function of crater diameter for burial depths of 33, 100, 333, and 1000 m for clay-layer thicknesses of 33, 100, 333, and 1000 m. The maximum volume fraction of clay-rich material increases with crater size and clay-layer thickness but decreases once the excavation depth penetrates beyond the clay-rich layer.

4. Discussion

What role did impact excavation play in delivering clays to Noachian surfaces? Crater size, clay layer thickness, and clay layer burial depth affect both the volume fraction and ejected location of clay-rich material. Observations of the variation in volume fraction across the ejecta deposit for craters without significant postemplacement radial flow may thus provide constraints on the nature of subsurface clay-
rich layers. If no clays are detected in an ejecta deposit, then (1) there simply are no clays in the subsurface or (2) the crater failed to penetrate into a clay-rich layer. If a rim is enriched with clays compared to the rest of the ejecta deposit, then the base of the clay-rich layer exceeds the excavation depth (Figure 4a). If the outer reaches of an ejecta deposit have a higher fraction of clays than a rim, then the excavation depth is deeper than the base of the clay-rich layer (Figures 4c–4e). The bulk volume fraction of clays across the ejecta deposit provides some indication of the thickness of the clay-rich layer, with thicker layers producing much higher fractions (Figure 3). Of course, surface modification following the impact event—aolian, fluvial, and depositional—may obfuscate and frustrate detection. Lack of detection does not imply that no clays are present.

37] Craters smaller than 25 km in diameter will likely deliver higher concentrations of clays to the surface than larger craters. Buried clays likely reside in the upper few kilometers of crust, assuming that either (1) clays were formed at the surface and then buried by subsequent geologic processes or (2) they were formed in subsurface aquifers. Therefore, as Figure 3 shows, the volume fraction of clay-rich material is highest for smaller craters. Furthermore, the volume of material that experiences significant shock-heating during an impact event is less than 10% for craters less than 25 km in diameter (see section 4.2).

38] Regional variations in clay presence as a function of crater diameter may provide important clues about the subsurface of Mars. As spectral coverage increases, the bulk volume fraction and radial distribution of clays may be used in conjunction with Figures 3 and 5 to estimate clay layer depths and thicknesses.

39] Much of the ancient Noachian crust exhibits phyllosilicate alteration [Mustard et al., 2009a]. The Mawrth Vallis and Nili Fossae regions show particularly strong detections of Fe/Mg phyllosilicates by OMEGA and CRISM [Bibring et al., 2006; Bishop et al., 2008; Ehlmann et al., 2008; Murchie et al., 2009; Noe Dobrea et al., 2010]. Figure 6 shows Fe/Mg and Al phyllosilicate detections by the OMEGA spectrometer [Bibring et al., 2006] and the observed MOLA topography for the Mawrth Vallis region. In general, to date, clays are detected near the craters labeled A, C, D, and E. Clays are not detected in a region west of crater C’s ejecta deposit and south of crater A (lower left quadrant of Figure 6). High concentrations of clays are distributed on the rim and ejecta deposit of the 26 km diameter crater labeled C. Craters D (55 km) and E (57 km) exhibit sparse concentrations of clays on their ejecta deposits. Crater B (13 km) shows no clay detection on its rim or ejecta deposit. The strong clay signature in crater C, the weak signatures in craters D and E, and the absence of clays in crater B are all consistent with the 1000 m burial depth line shown in Figure 3. We therefore conclude that a clay-rich layer a few hundred meters in thickness resides in the subsurface at the −3000 m elevation contour. This finding is in agreement with observations of erosional windows that indicate that the clays are thinly bedded and ~100 m thick [Noe Dobrea et al., 2010].

40] The ejecta deposit of crater A (* in Figure 6) exhibits examples of phyllosilicate detections associated with craters smaller than 25 km in diameter. At this location, Figure 7b shows two 4 km craters that reveal continuous, layered phyllosilicates at the high reaches of their inner walls and on their rims. Scattered clay signatures are present on the ejecta deposits of these craters and are most concentrated near their rims and decay in concentration with distance from their rims.

41] At 4 km in diameter, these craters are near the strength-gravity regime transition for craters on Mars, making excavation depth estimations uncertain. Strength-regime craters approximate the transient, or bowl-shaped, crater form; the depth to diameter ratio is 1:5 [Melosh, 1989; Noe Dobrea et al., 2010]. This assumption yields excavation depths of 800 m for these craters. Alternatively, assuming that these craters formed in the gravity regime, equation (8) estimates excavation depths of 250 m. These craters formed ~10–15 km from the rim of crater A. Applying the methods described in section 2.5, crater A’s ejecta deposit is ~300–500 m thick at this location. It is therefore likely that the majority of material excavated by the two 4 km craters came from the ejecta deposit of crater A.

42] The layered nature of the clays at the rims of the 4 km craters suggests that they may have formed from near-surface and possibly exposed material in and on the ejecta deposit of crater A following its emplacement. That stratigraphy may have been preserved in the 4 km craters’ overturned flaps. The scattered detections of Fe/Mg phyllosilicates distributed on the ejecta deposits of the 4 km craters are likely a record of deeper, unexposed material contained within the ejecta deposit of crater A that was excavated by the 4 km craters. Figure 7a shows a clay-rich layer near the −3000 m Mars Orbiter Laser Altimeter (MOLA) contour line on the upper wall of crater A (+ in Figure 6). This layer may have been a source for the redistributed clays present in crater A’s ejecta deposit. Future spectroscopic observations of the wall of crater A may indicate a potential regionwide clay-rich layer.

4.1. Implications for Volatiles

43] Although this study was motivated by the deep phyllosilicate detections on Noachian surfaces, the results are applicable to any stratified, horizontal subsurface layer. The Martian surface reveals numerous and varied examples of past fluvial activity. Impact excavation provides a means of delivering water sequestered in subsurface aquifers and cryospheres to the surface. What crater size maximizes the delivery of volatiles to the surface? Lithostatic pressure closes pore space and limits the presence of a water-bearing region of any significance to the upper 10 km of crust [Clifford, 1981]. Crater diameters greater than ~150 km excavate material from depths greater than 10 km. All ejecta from smaller craters originates in this potentially volatile-rich region. Craters with 45 and 90 km diameters excavate 1.2 × 10^{12} and 1.0 × 10^{13} m^{3} of material from a maximum depth of ~2.9 and ~5.7 km, respectively. However, there are roughly 10 times as many craters ranging from 44 to 46 km as there are ranging from 89 to 91 km [Hartmann and Neukum, 2001]. This means that 45 km and 90 km craters delivered roughly the same volume to the surface over the course of the Noachian period. Porosity, however, decreases with depth in an exponential fashion [e.g., Clifford, 1981], so 45 km diameter craters are much more effective than 90 km craters at liberating volatiles in the subsurface of Mars.
4.2. Will Shock-Heating Denature Preexisting Clays?

[44] Hydrodynamic impact simulations suggest that over 50% of a bolide’s kinetic energy may be converted into heat as the shock wave passes through the crust [O’Keefe and Ahrens, 1982; Gisler et al., 2006]. Mixed-layer phyllosilicates become unstable at temperatures of ~370 K; 920 K can be considered the upper decomposition temperature for phyllosilicates [Byrappa and Yoshimura, 2001; Brindley and Lemaître, 1987; Fairén et al., 2009]. Simplified models of impact-induced shock heating can be constructed using analytical formulations [e.g., Melosh, 1989] We estimate the distribution of shock heating produced by an impactor of radius, $R_p$, density, $\rho_p$, and velocity, $v_p$, by employing an analytic expression for the change in internal energy as a function of distance from the impact point. Given $R_t$, we use the following scaling relationship to determine the diameter of the bolide as a function of the bolide’s density and velocity (rearranged from Schmidt and Housen [1987] and Cintala and Grieve [1998]). Note that the variables describing projectile density, $\rho_p$, target density, $\rho_t$, bolide diameter, $D_p$, velocity, $v_p$, transient crater diameter, $D_t$, and gravity, $g$, must be expressed in centimeter-gram-second units.

$$D_p = \left(0.862D_t \left(\frac{\rho_t}{\rho_p}\right)^{0.44}v_p^{-0.22}\right)^{1.2821}$$

[45] We assume an asteroid-like impact event and set $\rho_p$ equal to $\rho_t$ at 2.6 g cm$^{-3}$ and $v_p$ to $7 \times 10^7$ cm s$^{-1}$. Given a bolide diameter, we determine the change in temperature as a function of distance from the impact point. We briefly outline
here the same approach described in detail in Abramov and Kring [2005] and Barnhart et al. [2010]. This approach uses an expression for specific waste heat, \( \Delta E_w \), derived from the Murnaghan equation of state by Kieffer and Simonds [1980], which depends on the peak shock pressure, \( P \), the adiabatic bulk modulus at zero pressure, \( K_0 \), and the pressure derivative of the bulk modulus, \( n \). We use the same values for \( K_0 = 19.3 \text{ GPa} \) and \( n = 5.5 \) as Abramov and Kring [2005] and Barnhart et al. [2010] as obtained from Gault and Heitowit [1963], Kieffer and Simonds [1980], and Melosh [1989].

The peak pressure, \( P \), is a function of projectile radius, \( R_p \), and varies according to

\[
P = A \left( \frac{r}{R_p} \right)^f
\]

where \( r \) is the distance from the impact, \( A \) is the pressure at \( r = R_p \), and \( f \) is the pressure falloff exponent. For distances less than \( R_p \), \( r = R_p \). Melosh [1989] estimates \( A \) as \( A = \rho_i [C_i + S_i u_0]u_0 \), where \( C_i \) and \( S_i \) are parameters in the linear shock velocity–particle relation (Melosh [1989], Appendix II). For basalt, \( C_i = 2600 \text{ m/s} \) and \( S_i = 1.58 \text{ (dimensionless)} \); the initial shock velocity is \( u_0 = 0.5 v_p \). The change in temperature is calculated by dividing the specific waste heat by the heat capacity of basalt, \( 800 \text{ J kg}^{-1} \text{ K}^{-1} \).

The falloff exponent, \( f \), dominates the temperature distribution. Motivated by temperature distributions generated by a numerical hydrocode model for a 30 km Martian crater by Pierazzo et al. [2005], Barnhart et al. [2010] found that temperature contours with depth were replicated reasonably well analytically with a value for \( f \) of \( \approx -3.0 \). The decay exponent depends on impact velocity, so we consider a reasonable bounding range from \( f = -2.5 \text{ to } -3.5 \) [Pierazzo et al., 1997].

Figure 8 shows the shock-heat deposited in 6 and 60 km diameter craters. Notice that shock-heating does not scale linearly with crater size [cf. Melosh, 1989]. As crater size increases, larger portions of the crater are shock-heated.

Figure 8c plots the volume fraction of excavated material heated by more than 333 K (black lines) and 1000 K (red lines) as a function of crater diameter for two pressure falloff exponents: \( f = -2.5 \) (dotted lines) and \( f = -3.5 \) (solid lines). The fraction of excavated material that is heated beyond 1000 K is at most 20%. A sharper falloff of \( f = -3.5 \), which may more accurately mimic shock attenuation because of pore space closure and material interfaces [cf. Senft and Stewart, 2008], yields a volume fraction heated beyond 1000 K of ~10% for crater diameters less than 100 km. This means that the craters that are most effective at delivering high-volume fractions of clays to the surface also experience the least amount of shock-heating. These impacts are unlikely to denature preexisting clays that are delivered to the ejecta deposit. Furthermore, while the shock-induced heating is concentrated at the center of the crater, Figure 8 shows that the majority of the material that builds the ejecta deposit originates more than 0.5 \( R_t \) from the center and is correspondingly less likely to be significantly shock-heated.

5. Conclusion

[47] The most abundant clays detected on Noachian surfaces are the Fe/Mg phyllosilicates. Subsurface aquifers, likely present in much of the early Martian crust, provide conditions conducive to the formation of this aqueous mineral assemblage. The high flux of impact cratersing events during the Noachian period provides a global means for distributing these clays over ancient surfaces. The amount of clay present in ejected material depends on the crater diameter, the thickness of the clay layer, and how deep that layer resides below the surface. The relative...
abundance of clays also varies across the ejecta deposit as a function of the crater size, the clay layer thickness, and the burial depth. With aquifers constrained to the upper 10 km of crust and enhanced thermodynamic activity limited to depths shallower than 3 km, craters 5–25 km in diameter maximize the delivery of clays to the surface. The abundance of clay-rich ejecta will peak closer to the rims for craters on the smaller end of this range. Conversely, the maximum abundance will be found at more distal reaches, ~1–2 crater radii out, on the ejecta deposit for craters at the larger end of this range. Thus, observations of clay distribution across the ejecta blankets of craters of different sizes may be used to infer the characteristics of the subsurface clay layer and potentially spatial variations in these characteristics. Clay alteration due to shock-heating during excavation is expected to be minimal. The impact excavation mechanism for the emplacement of Fe/Mg phyllosilicates may explain their widespread regionally and topographically independent distribution on Noachian surfaces.

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References


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