Rheological constraints on martian landslides

Keith P. Harrison\textsuperscript{a,*} and Robert E. Grimm\textsuperscript{b}

\textsuperscript{a} Laboratory for Atmospheric and Space Physics (LASP), University of Colorado, Campus Box 392, Boulder, CO 80309-0392, USA
\textsuperscript{b} LASP and Blackhawk Geoservices, Inc., 301 B Commercial Road, Golden, CO 80401, USA

Received 4 September 2002; revised 20 December 2002

Abstract

We use a dynamic finite-difference model to simulate martian landslides in the Valles Marineris canyon system and Olympus Mons aureole using three different modal rheologies: frictional, Bingham, and power law. The frictional and Bingham modes are applied individually. Fluidized rheology is treated as a combination of frictional and power-law modes; general fluidization can include pore pressure contributions, whereas acoustic fluidization does not. We find that general fluidization most often produces slides that best match landslide geometry in the Valles Marineris. This implies that some amount of supporting liquid or gas was present in the material during failure. The profile of the Olympus Mons aureole is not well matched by any landslide model, suggesting an alternative genesis. In contrast, acoustic fluidization produces the best match for a lunar slide, a result anticipated for dry crust with no overlying atmosphere. The presence of pressurized fluid during Valles Marineris landsliding may be due to liquid water beneath a thin cryosphere (1–2 km) or flash sublimation of CO\textsubscript{2}.

© 2003 Elsevier Science (USA). All rights reserved.

Keywords: Mars, surface; Surfaces, planets; Terrestrial planets

I. Introduction

Landslide features on Mars bear an unambiguous resemblance to terrestrial landslides. They occur mainly in the equatorial Valles Marineris canyon system, in both open and closed depressions (Lucchitta, 1979). The aureole deposits on the lower flanks of the largest martian shield volcano, Olympus Mons, may also be landslides (Lopes et al., 1980, 1982), although slower processes such as gravity spreading may have been responsible for their formation (Francis and Wadge, 1983; Tanaka, 1985). In many Valles Marineris landslides a curved scar cut into the canyon rim indicates the site of the original failure. The scar is typically a few tens of kilometers wide and recedes a few kilometers into the canyon wall. Landslides occur in walls as high as 8 km and cover horizontal distances (runouts) as great as 80 km if unimpeded by obstacles. The volume of individual landslide deposits ranges from tens to thousands of cubic kilometers.

Of all the processes that have shaped the Valles Marineris, landsliding is among the most recent. The landslide scars are smooth and the deposits are sparsely cratered and show little evidence of deflation (Lucchitta, et al., 1992), suggesting a late Hesperian to Amazonian genesis. Slide scars cut into preexisting spur-and-gully formations, and the slide material overlies interior deposits. There is, however, some indication that tectonic movement postdating landslide deposition has occurred along faults closely related to the formation of the Valles Marineris (Lucchitta, 1979; Lucchitta et al., 1994).

Although reentrants are generally smooth, their upper elevations are typified by dark vertical banding which, on close inspection of Mars orbital camera (MOC) images, is seen to consist of small, rugged spur formations, sometimes horizontally striated, that are consistent with the volcanic nature of the plateaus surrounding the Valles Marineris (Scott and Tanaka, 1986). Emanating from the lower boundaries of these miniature spurs are lighter colored, linear, overlapping talus deposits blanketing the lower elevations of the scar. The influence of material strength heterogeneity on landslide dynamics is discussed in the following.

\* Corresponding author.
E-mail address: keith.harrison@colorado.edu (K.P. Harrison).
The landslide deposits themselves have small-scale features, some of which are difficult to explain. These include longitudinal grooves, rather than transverse ridges like those observed on terrestrial slides. Lucchitta (1979) suggested that longitudinal grooves result from differential forward velocities that can be explained by the presence of a fluid. Many large deposits, both martian and terrestrial, have runouts that imply lower internal friction angles than expected. Processes thought to produce long runouts typically require fluidization by water (Habib, 1975; Goguel, 1978) or air (Kent, 1966; Shreve, 1968). Arguments have been given in support of both the presence (Lucchitta, 1987) and absence (McEwen, 1989) of water in the Valles Marineris slides. The thin martian atmosphere is unlikely to be dense enough to provide lubrication, but outgassing of carbon dioxide from the soil on failure could play an important role. Some fluidization processes do not require an actual fluid: In dispersive grain flow (Bagnold, 1954), the weight of the material is supported by impacts between its constituent rock particles, whereas for acoustic fluidization (Melosh, 1979), an acoustic field transmitted through the material is sufficient to support for brief periods the static overburden pressure without requiring separation of particles.

Features of landslide morphologies may be used to infer their dynamic behavior. Rheology is an important influence on the final configuration of a landslide: Given a particular runout path and initial mass configuration, it is largely the rheology that determines key features such as runout length and final deposit thickness and slope. Here we report the results of numerical models of martian landslides designed to constrain rheological parameters. Comparisons are made with terrestrial and lunar models.

II. Model

Dynamical model

We use a dynamic analysis (DAN) model (Hungr, 1995) which, given the runout path and initial mass profile, computes the time-varying shape and velocity distribution of the slide. DAN has been successful at modeling terrestrial landslides with a wide range of rheologies, including sand and coal failures (Hungr, 1995) and, where physical parameters are not available for comparison, it successfully reproduces other numerical analyses (Hungr, 1995), including that of the Madison Canyon, Montana, rock avalanche (Trunk et al., 1986).

A finite-difference Lagrangian approach is used to track individual mass blocks as they move together down the runout path. Each slide is modeled as a single row of blocks (20 in our models) whose initial heights depend on the prescribed initial shape of the failed mass. Quantities that vary normal to the runout path are represented by their average values. For each block, the component of the gravitational force along (tangential to) the runout path works against a pressure contribution and a basal resistance force. The pressure contribution arises from the differing heights of adjacent blocks. Material tends to shift laterally to reduce high, unstable surface slopes. The resulting communication of pressures throughout the landslide is controlled by a lateral pressure coefficient whose lower and upper bounds reflect the tensional and compressional strengths of the material respectively. The pressure coefficient has a small influence on the final profile and is not considered to be of primary importance.

The basal resistance force for each block is determined by one of several available rheologies, which include frictional, Bingham, and power law. In the frictional rheology, the basal resistance offered by a block is proportional to the normal stress it exerts on the runout path. The constant of proportionality is the friction coefficient, often expressed as the tangent of the friction angle $\phi$, set to 20° in all models unless where otherwise specified. By default, DAN uses the same static and dynamic friction angles. In reality, the dynamic friction angle is less than the static value (the latter is typically around 30° for fractured crustal rock; Brace and Kohlstedt, 1980) and may change with the velocity of the material (Jaeger and Cook, 1979). Other workers (e.g., Wang, 1997) have used models with dynamic friction coefficients as low as 0.6 times the static value (i.e., a static angle of 30° corresponds to a dynamic value of 19°). There is also a dependence on clay content (Skempton, 1964) that could reduce the dynamic angle by a few degrees. Small quantities of clay minerals have been observed indirectly in spectra (e.g., Soderblom, 1992) and directly in the SNC meteorites (e.g., Gooding et al., 1991; Treiman et al., 1993). Our choice of friction angle thus represents a physically plausible lower limit. The frictional law used in DAN is such that a change in friction angle will have the same affect on dynamics as a corresponding change in pore fluid pressure. A lower limit for friction angle thus precludes the interpretation of pore pressure as artificial compensation for an exaggerated friction angle.

The basal resistance also depends on the pore-pressure coefficient $r_p$, defined as the ratio of pore pressure exerted by an interstitial fluid to the total normal stress at the base of the block. Pore fluid carries some of the weight of the overlying material, reducing basal friction. We stress that a given pore-pressure coefficient describes conditions at the failure surface only and heterogeneities in permeability may allow other parts of the failed wall section to remain dry. Consequently, pore-pressure coefficients that correspond to hydrostatic or even superhydrostatic pressures do not necessarily imply total saturation at all depths. The basal resistance is also affected by centrifugal forces, since these increase or decrease the effective normal stress according to the curvature of the path.

The Bingham rheology, used successfully to model rock avalanches (Hungr, 1995, and references therein) consists of a Newtonian fluid with a finite yield strength $\tau_0$. Assuming that shear stress increases linearly with depth, the constitu-
tive equation (which relates shear stress to velocity gradient) may be integrated to yield an expression for the basal resistance in terms of fluid velocity \( v \), yield strength, and dynamic viscosity \( \mu \). (The basal resistance is simply the shear stress at the base of the slide.) The yield strength gives rise to a solid cap which rides on top of the flow. As the deposit extends itself along the runout path, its tensile strength slows it down, causing the cap to thicken until it consumes the entire flow, bringing the deposit to a halt. This effect is exacerbated by the lowering of the cap as the deposit thins. Landslides with higher yield strengths have thicker caps and so form thicker deposits with shorter runouts.

A power law may be used to model flows in which most of the lateral shear is concentrated in a thin basal layer (Melosh, 1987), such as the Blackhawk landslide in California (Shreve, 1968). The Elm landslide in Switzerland was simulated successfully by Hsu (1975) in a scaled down model using a bentonite suspension, a fluid with power-law behavior (Besq et al., 2000). For a power-law fluid, the shear stress is

\[
\tau = \tau_0 + \mu \left( \frac{dv}{dz} \right)^n,
\]

where \( z \) is depth and \( n \) is a unitless index. In Eq. (1) the symbol \( \mu \) does not have units of real viscosity and is called the “apparent” viscosity. We adopt \( n = 0.125 \) [suggested by Melosh (1987) for acoustic fluidization] for all slides and for both fluidization processes. A value of 1 represents a Bingham fluid, while a value of 0 corresponds to a rigid sliding block. We investigated the sensitivity of runout to \( n \) for a martian slide (Slide 3 in Fig. 2) and found that within the range of 0 to 1, halving or doubling the value of \( n \) increases or decreases runout by 2 to 2.5 km for a slide with a runout of 72 km at \( n = 0.125 \). The uncertainty in our choice of \( n \) is thus largely immaterial.

We consider next a general fluidization process in which the slide is frictional on failure but becomes fluidized when sufficiently energetic, thus behaving more like a power-law fluid. As already noted, pore pressure may be specified explicitly in the frictional rheology. With the power law, however, water content is implicit and its influence is assumed to manifest itself through the determination of yield strength and apparent viscosity.

A special case of general fluidization is acoustic fluidization (Melosh, 1979), which can occur without the presence of a liquid or gas and whose frictional stage may thus be associated with zero pore pressure. When acoustic waves of sufficient strength are transmitted elastically through the individual constituents of the landslide material, intraparticle rarefactions produce short periods of decreased overburden pressure, temporarily reducing basal resistance. The acoustic field builds up to the required strength some time after the initial failure, so fluidization does not occur immediately. Although acoustic fluidization may be possible in the presence of an interstitial fluid, we address here the case of dry material only. This is, in fact, the fundamental distinguishing feature of this rheology and since it includes no explicit dependence on the acoustic processes described here, it may encompass other processes in which frictional and power-law behaviors occur in the absence of fluid pressure, if others indeed exist.

For clarity we note that the general and acoustic fluidization processes are not modal rheologies, but sequential combinations of two rheologies: frictional and power law. In general, the yield strength and apparent viscosity of the power-law stage for a particular slide will not be the same for acoustic fluidization and general fluidization. The values of these parameters depend on the position at which the frictional rheology ends and the power law begins, and also on the velocity and stress distribution of the material at the transition. Transitions occurring further along the runout path have a frictional stage ending with different velocities, and the power-law stage must thus have a different yield strength and/or viscosity to conserve runout.

Inputs common to all rheologies include unit weight \( \gamma \), volume yield rate \( E \) (a measure of material deposition or entrainment, described subsequently), and the lower and upper bounds of the lateral pressure coefficient. Unit weight was taken to be 10.00 kN \( \cdot \) m\(^{-3}\) for martian slides, 26.44 kN \( \cdot \) m\(^{-3}\) for terrestrial slides, and 4.371 kN \( \cdot \) m\(^{-3}\) for lunar slides, corresponding in all cases to a density of 2700 kg \( \cdot \) m\(^{-3}\).

**Landslide geometry**

Landslide deposit volume \( V \) was estimated for martian slides by summing 1/32 by 1/32 degree Mars Global Surveyor (MGS) MOLA elevation data over the area covered by the material (with Viking images used to determine the location of deposit boundaries). For these calculations, the base of the deposit was assumed to be horizontal and its elevation was inferred from the surrounding canyon floor, or

![Fig. 1. Conceptual model of initial landslide shape. Symbols correspond to those defined in the text. Runout occurs in the positive z direction.](image-url)
matched to the toe of the deposit. Our method of calculating volumes differs from previous approaches (e.g., McEwen, 1989), which used the scar geometry to estimate the initial volume of the failed section of canyon wall. Since the configuration of the pre-slide wall is unknown there is considerable uncertainty in measuring volume this way; a direct measurement of the final deposit volume is likely to be more accurate.

We assumed initially that volume was conserved during the slide event (i.e., no net deposition or entrainment of material). Approximate measurements of scar width $W$, scar slope $\alpha_1$, and typical nearby wall slope $\alpha_2$ allowed us to infer a suitable initial configuration for the failed material, with the required volume (Fig. 1). Our use of volume as the primary measurement used to reconstruct the initial configuration is in line with experimental evidence (Hsu, 1975) showing that volume, rather than geometry, has the stronger influence on runout. Nonetheless, the initial configuration was adjusted to match the shape of the unfailed wall adjacent to the landslide scar, where possible. Similarly, the
runout path was shaped according to nearby wall and floor geometry. The coordinate system (Fig. 1) was positioned such that the back edge of the top surface of the initial shape was at \( x = 0 \). The vertical coordinate represents elevations relative to the MOLA datum.

MOLA data were used to obtain vertical cross sections of the landslide deposits along the direction of motion. (All cross sections mentioned henceforth refer to this plane.) These were compared to DAN results to assess the suitability of the chosen rheological parameter values. Values were iteratively adjusted and the model rerun until a suitable final profile was achieved.

In DAN, different sections of the runout path may be assigned different rheologies. We took advantage of this feature to simulate general and acoustic fluidization. The dynamics of the transition from frictional to power law is not well established and we chose the bottom of the initial steep section of the runout path as the default transition point. The transition position was varied about this point from model to model depending on the characteristics of the
flow (such as velocity) and structural features observed in the final deposit profile (such as abrupt decreases in slope). Rather than an abrupt change in rheology, this transition should be seen as the point at which the power-law rheology becomes better than the frictional rheology at describing the motion.

The conceptual model of landslides described thus far is somewhat limited by its essentially two-dimensional nature since changes in runout path width (dimension y in Fig. 1) may influence the final cross section of the deposit. In most cases, the landslide scar is curved in the cross-slope direction; i.e., each contour line is crescent-shaped. (In the downslope direction the scars, where exposed, are remarkably linear.) This trend, which supposedly continues into the buried parts of the scars, suggests that material converges during the initial, steep part of its journey. On reaching the level, unconfined canyon floor, most of the energy in the slide has been channeled directly forward, resulting in a toe which travels further than the laterally spreading material on either side. The resulting deposit is shaped like an arrowhead pointing in the direction of motion. Despite lateral spreading, the average final deposit width is frequently smaller than the average initial width. Indeed, initial cross-sectional areas of the slides studied here are frequently smaller than final cross-sectional areas (Table 2), implying an average decrease in width to conserve volume. One can specify the required deposit width at each point along a DAN runout path. This feature is intended to simulate the effects of a laterally confining channel and is thus suitable for the curved landslide scar in our models. Unconfined spreading is best modeled by fixing the path width on the canyon floor to match the observed deposit.

Another way to account for changes in cross-sectional area is to relax constraints on volume conservation. Landslides typically accumulate new material during the initial downhill phase. In DAN, one can specify the desired amount of entrainment or deposition over segments of the runout path. Individual block volume is updated at each time step as (Hungr and Evans, 1997)

\[ V_j = V'_j + E\Delta s \frac{V'_j}{V}. \]

where \( V \) is the total slide volume, \( V'_j \) and \( V_j \) are the old and new volumes of the \( j \)th block, respectively, and \( \Delta s \) is the displacement of the \( j \)th block during the time step. Parameter \( E \) is a volume per unit runout path length and specifies how much material is deposited or entrained (depending on

### Table 1
Landslide deposit dimensions

<table>
<thead>
<tr>
<th>Slide number</th>
<th>( V ) (km(^3))</th>
<th>Horizontal coverage (km(^2))</th>
<th>( L_s ) (km)</th>
<th>( L_f ) (km)</th>
<th>( W ) (km)</th>
<th>( \alpha_1 ) (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.283</td>
<td>15</td>
<td>2.5</td>
<td>9.8</td>
<td>1.7</td>
<td>15.3</td>
</tr>
<tr>
<td>2</td>
<td>2100</td>
<td>1500</td>
<td>5.5</td>
<td>79</td>
<td>29</td>
<td>25.9</td>
</tr>
<tr>
<td>3</td>
<td>1700</td>
<td>1600</td>
<td>5.5</td>
<td>72</td>
<td>26</td>
<td>24.4</td>
</tr>
<tr>
<td>4</td>
<td>67</td>
<td>1300</td>
<td>3.8</td>
<td>51</td>
<td>26</td>
<td>32.5</td>
</tr>
<tr>
<td>5</td>
<td>14</td>
<td>170</td>
<td>6.7</td>
<td>26</td>
<td>7.1</td>
<td>26.4</td>
</tr>
<tr>
<td>6</td>
<td>460</td>
<td>2000</td>
<td>13</td>
<td>58</td>
<td>29</td>
<td>16.4</td>
</tr>
<tr>
<td>7</td>
<td>1300</td>
<td>1000</td>
<td>9.0</td>
<td>60</td>
<td>33</td>
<td>22.0</td>
</tr>
<tr>
<td>8</td>
<td>36,000</td>
<td>420,000</td>
<td>160</td>
<td>690</td>
<td>500</td>
<td>25.9</td>
</tr>
<tr>
<td>9</td>
<td>0.200</td>
<td>21</td>
<td>5.0</td>
<td>8.8</td>
<td>4.2</td>
<td>24.8</td>
</tr>
</tbody>
</table>

*Note.* Values for the terrestrial slide (1) were calculated from maps and cross sections from Shreve (1968), values for martian slides were measured from MOLA data, and values for the lunar slide (9) were calculated from data and cross sections from Howard (1973).

### Table 2
Information inferred from vertical cross-sections parallel to the direction of motion

<table>
<thead>
<tr>
<th>Slide number</th>
<th>( H ) (km)</th>
<th>( H/L_s ) (unitless)</th>
<th>( \alpha_1 ) (°)</th>
<th>( D ) (km)</th>
<th>Initial cross-sectional area (km(^2))</th>
<th>Final cross-sectional area (km(^2))</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1.1</td>
<td>0.112</td>
<td>19.0</td>
<td>0.26(^b)</td>
<td>0.17</td>
<td>0.15</td>
</tr>
<tr>
<td>2</td>
<td>8.0</td>
<td>0.173</td>
<td>24.2</td>
<td>8.6</td>
<td>74</td>
<td>120</td>
</tr>
<tr>
<td>3</td>
<td>7.7</td>
<td>0.0980</td>
<td>24.2</td>
<td>8.6</td>
<td>67</td>
<td>84</td>
</tr>
<tr>
<td>4</td>
<td>5.0</td>
<td>0.0966</td>
<td>14.1</td>
<td>4.0</td>
<td>27</td>
<td>26</td>
</tr>
<tr>
<td>5</td>
<td>4.5</td>
<td>0.122</td>
<td>26.4</td>
<td>0.47</td>
<td>19</td>
<td>2.2</td>
</tr>
<tr>
<td>6</td>
<td>5.6</td>
<td>0.107</td>
<td>16.4</td>
<td>3.1</td>
<td>16</td>
<td>17</td>
</tr>
<tr>
<td>7</td>
<td>7.3</td>
<td>0.101</td>
<td>22.0</td>
<td>7.4</td>
<td>64</td>
<td>105</td>
</tr>
<tr>
<td>8</td>
<td>10</td>
<td>0.0145</td>
<td>3.31</td>
<td>86(^b)</td>
<td>720</td>
<td>720</td>
</tr>
<tr>
<td>9</td>
<td>2.4</td>
<td>0.273</td>
<td>24.8</td>
<td>0.022</td>
<td>0.048</td>
<td>0.048</td>
</tr>
</tbody>
</table>

\(^a\) In some Valles Marineris slides, the initial mass was modeled after nearby unfailed canyon wall and the value for \( \alpha_2 \) represents an average, rather than an exact, value.

\(^b\) Average value for a wedge-shaped initial shape.
its sign). If the entire slide passes over a segment of the runout path with length \( a \) and entrainment rate \( E \), the total change in volume is given approximately by \( aE \). We assigned a finite entrainment rate to the initial, steep section of the runout path, and a value of zero elsewhere. Values of \( E \) for our models range from \( 10^4 \) to \( 10^7 \) m\(^3\) m\(^{-1}\) depending on the initial volume of the slide and the required change therein.

We ran models in which changes of cross-sectional area were produced exclusively by either width or volume variations, or a combination of both. In the last case, measurements of the landslide geometry were used where possible to constrain the width variations, and volume variations were adjusted to account for the remaining change in cross-sectional area. The models presented here are exclusively of the last kind, although we note that models with only one kind of variation did not produce significantly different results.

### III. Results

We present detailed models of a terrestrial landslide, six Valles Marineris landslides, part of the Olympus Mons aureole, and a lunar landslide. Four of the Valles Marineris slides have been described in detail by Lucchitta (1979) and we refer to her landslide numbering scheme for easy reference. Images of Slides 1 to 9 are shown in Fig. 2, and initial and final deposit dimensions are given in Tables 1 and 2.

#### Slide 1

The California Blackhawk landslide (Fig. 2) is a terrestrial example of a long-runout landslide (Shreve, 1968). With \( V = 0.28 \) km\(^3\), it is the smallest slide modeled here. Detailed profiles (Shreve, 1968) provided useful information on the shape of the runout path. The lack of an abrupt change in angle between canyon wall and floor typical of the Valles Marineris slides makes a wedge-shaped initial mass more appropriate than the block shape of Fig. 1.

The general fluidization and Bingham models produce equally satisfactory results (Fig. 3). The rock mass does not fail frictionally when dry and thus the starting conditions for acoustic fluidization are not met. The factor of safety (FS), defined as the ratio of resisting to driving forces, has an average value of 1.33 (failure is expected to occur at FS = 1), although local values of FS may be lower (Záruba and Mencš, 1982). Therefore the slide may be considered only marginally stable for a friction angle of 20°. A more appropriate measure of failure probability in the context of acoustic fluidization is the highest friction angle that still produces movement at the toe of the initial shape: 18.9° for this model, probably close enough to the baseline value (given its uncertainty) that acoustic fluidization cannot be ruled out altogether if a suitable explanation can be found for such a low static friction angle.

A pore-pressure coefficient of 0.27 corresponds to hydrostatic pressure (with rock density of 2700 kg m\(^{-3}\) and water density of 1000 kg m\(^{-3}\)), so the values of 0.65 and 0.31 (Table 3) required for the frictional and general-fluidization models, respectively, are superhydrostatic. Experiments conducted by Iverson et al. (1997) show that, on failure, pore pressures may increase by as much as 100% due to soil compaction. Such an increase raises the pore-pressure coefficient to 0.42, above that required for the general-fluidization model that best matches this slide. Fluid at lithostatic pressure has \( r_a = 0.5 \), so values in excess of this limit are physically impossible, ruling out the viability of purely frictional dynamics for this slide.

We define rear runout \( L_r \) as the minimum \( x \) coordinate of the final deposit configuration and front runout \( L_f \) as the maximum \( x \) coordinate. The rear runout of all Bingham slides is zero because the tapered upper end of the initial mass is, in general, too thin to produce stresses above the yield value. The relative motion of the lower portion of the initial mass thus produces a thin, stretched tail. In DAN, this causes the rear elements of the slide to lengthen significantly and the tail may not always be smoothly resolved.

#### Slide 2

One of the largest landslides considered here, Slide 2 (Fig. 2), is situated against the north wall of Ophir Chasma. Three distinct deposits are visible below this wall, and Slide 2 is the furthest west. All three deposits have clearly defined tails, but further along their lengths they overlap and are difficult to distinguish. They are, nonetheless, good examples of unconfined long-runout landslides. Slide 2 has a

---

Fig. 3. DAN results for Slide 1, the Blackhawk landslide. In each panel, the dashed line is the final profile [measured from Shreve (1968) data] and the thin black line is the inferred runout path. The heavy line represents the DAN results using the rheology indicated. The short line at right angles to the runout path in the general fluidization panel marks the position at which the frictional rheology switched over to a power law.
front runout of 79 km, a distance 10 times that of the initial height of the failed material (Table 1).

The frictional and general fluidization rheologies provide comparable fits to the final configuration (Fig. 4). General fluidization produces better toe and tail shapes but is less accurate between $x H_{1005}$ 20 and 40 km. Required pore pressures in both models are high ($r_u/ H_{1005}$ 0.49 and 0.56 for the frictional and general-fluidization models, respectively); the small difference between the two does, however, span the lithostatic value, perhaps marginally favoring the general-fluidization model. Note that the position of the transition from frictional to power law for general fluidization was chosen in this case to coincide with an abrupt decrease in profile slope. The transition position for the acoustic fluidization rheology is, however, further back along the runout path since the initial frictional stage is less energetic without pore pressure and reaches its maximum velocity sooner.

**Slide 3**

This landslide is the furthest east of the north Ophir Chasma slides (Fig. 2). Like Slide 2, the scar in the wall of the chasm is clearly contrasted with its surroundings, as is the tail of the deposit, and most deposit dimensions are similar to those of Slide 2 (see Table 1). Although both the frictional and general-fluidization rheologies match much of the profile (Fig. 5; $x = 20$ and 40 km), only the latter matches the slide tail and the long, thin runout to $x = 72$ km. Furthermore, significantly lower pore pressure is required for general fluidization compared to friction ($r_u/ H_{1005}$ 0.35 vs 0.47). Note that the position of the transition from frictional to power law in the general-fluidization model was also chosen here to coincide with an abrupt decrease in profile slope.

**Slide 4**

This relatively small landslide (number 1 in Lucchitta, 1979) is located against the south wall of Gangis Chasma. It is the only slide of those considered here for which MGS MOC data are available. An image covering the length of the slide (Fig. 6) reveals detailed structure, including the scar features already described. The deposit has longitudinal ridges and grooves, and transverse ripple-like structures reminiscent of a turbulent fluid.

General fluidization produces the best fit (Fig. 7) with only modest pore pressure (Table 2). The Bingham fluid reproduces the bulbous toe shape, but it fails to mimic the tail. Because the structure of the runout path is somewhat

### Table 3

Summary of DAN results

<table>
<thead>
<tr>
<th>Slide number</th>
<th>Frictional $r_u$ ($\mu$)</th>
<th>Bingham $\tau_0$ (kPa) $\mu$ (kPa·s)</th>
<th>General fluidization $\tau_0$ (kPa) $\mu$ (kPa·s)</th>
<th>Acoustic fluidization $\tau_0$ (kPa) $\mu$ (kPa·s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.65 24.5 10</td>
<td>13.5 5.0 0.31</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>2</td>
<td>0.56 870 200</td>
<td>135 20 0.49</td>
<td>245 20</td>
<td>157 20</td>
</tr>
<tr>
<td>3</td>
<td>0.47 790 50</td>
<td>49.5 10 0.35</td>
<td>0.35 157 20</td>
<td>0.49 245 20</td>
</tr>
<tr>
<td>4</td>
<td>0.58 265 40</td>
<td>79.0 50 0.29</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>5</td>
<td>0.58 75.0 20</td>
<td>21.5 5.0 0.00 ($\phi = 22.8^\circ$)</td>
<td>21.8 5.0 20</td>
<td>21.8* 5.0* 20</td>
</tr>
<tr>
<td>6</td>
<td>0.71 170 40</td>
<td>65.5 10 0.48</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>7</td>
<td>0.58 525 200</td>
<td>145 50 0.41</td>
<td>146 50</td>
<td>--</td>
</tr>
<tr>
<td>8</td>
<td>0.94 50.0 20</td>
<td>49.2 5.0 0.98</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>9</td>
<td>0.41 0.50 0.30</td>
<td>1.8 0.50 0.00</td>
<td>1.8* 0.50* 20</td>
<td>0.50* 0.50* 20</td>
</tr>
</tbody>
</table>

**Note.** Rheologies other than general fluidization produced unequivocal best fits for only two of the nine slides, and these are indicated by asterisks. The frictional phases of all slides had $\phi = 20^\circ$, except where indicated, and for acoustic fluidization $r_u = 0$ in all cases. Where no data for acoustic fluidization are given, this rheology did not cause the initial mass to fail.
simplified, the apparent deepening of the deposit near the toe might also be interpreted as a constant-thickness deposit overlying a bulge in the runout path. Despite the steep initial slope of the runout path, acoustic fluidization does not initiate failure. This is because much of the material lies initially over the level section of the runout path and its weight does not contribute to the downslope driving force. The highest friction angle that produces motion in the toe of the initial shape is 14.5°, significantly less than the baseline value.

Since the initial and final cross-sectional areas are very similar (Table 2), the decrease in area caused by the great degree of widening of the slide (Fig. 2) has to be compensated by a high (10² m³·m⁻¹) entrainment rate, which may imply that the canyon wall in this region is weaker than that of the other slides.

Slide 5

Situated against the north wall of the eastern reaches of Gangis Chasma, Slide 5 (number 8 in Lucchitta, 1979) is the smallest martian landslide modeled (V = 14 km³). The shape of the deposit toe is influenced by its contact with another slide emanating from the opposite canyon wall (Figs. 2 and 8). Consequently, it is not surprising that none of the rheologies, except perhaps the Bingham fluid (which was run with 30 slide elements to improve tail resolution), reproduces the narrow hump at the deposit toe. Overall, the acoustic fluidization rheology provides a comparably accurate fit; note that the friction angle had to be increased to 22.8° from the baseline value of 20°. The initial height H (Table 2) of the failed material is not significantly less than that of other martian landslides, but because the deposit volume is so small, the depth D of the deposit (Fig. 1; Table 2) is also small. This gives rise to a greater rear runout and hence reduces the need for pore pressure that allows acoustic fluidization to model this slide successfully. In the available Viking images of Slide 5 no obvious scar is visible on the canyon wall and H was thus chosen to correspond to the full height of the wall, a rule of thumb supported by observations of most other large-scale Valles Marineris slides.

Slide 6

This landslide (Fig. 2; number 9 in Lucchitta, 1979) is located at the west end of Coprates Chasma. It is a relatively small slide, with a distinctively level, square tail (Fig. 9). This may be due to the wholesale slipping of the topmost section of the original mass. DAN does not account for the motion of large rigid blocks, so the main priority of Slide 6 models was to match front runout. None of the models produces a satisfactory fit. Acoustic fluidization does not initiate failure; the highest friction angle that produces motion in the toe of the initial shape is 16.3° (FS = 1.24).

Slide 7

Located against a west-facing wall in the region between Ophir and Melas Chasmas, Slide 7 (Fig. 2; number 11 in Lucchitta, 1979) is similar in volume and dimension to the north Ophir Chasma slides (Table 1) but has a shorter runout. General fluidization, which produces the best fit (Fig. 10), thus requires a high yield strength (145 kPa). The required pore-pressure coefficient (0.41) is between hydrostatic and lithostatic values. The frictional rheology also yields a good fit to this profile but requires superlithostatic fluid pressure (r_u = 0.56).

Slide 8

We ran a model of the westward-facing Olympus Mons aureole deposit. In a bitmap generated from MOLA data (Fig. 2), where each picture element has a brightness proportional to its elevation, the deposit can be seen overlying a darker (and therefore lower elevation) deposit. The scarp near the tail of the deposit is assumed to be the landslide scar, and the initial mass is wedge-shaped to simulate the pre-slide shield topography. Postemplacement plains and surface deposits covering the rear of the aureole make it difficult to infer the deposit configuration in this region, and we thus assume a uniform deposit thickness. Three different profiles were constructed to get an overall picture of the aureole topography. The runout path dips toward Olympus Mars proximally, and away from it distally, agreeing with observations by Francis and Wadge (1983). Neither frictional, Bingham, nor general-fluidization models produce particularly good fits (Fig. 11) and most notably the required pore-pressure coefficients must exceed 0.9 to achieve the
690-km front runout. Failure is not initiated for the dry frictional precursor to acoustic fluidization: $FS = 2.10$ and the highest friction angle that produces motion in the toe of the initial shape is $3.8^\circ$. We conclude that the Olympus Mons aureole is not a landslide deposit but the result of an alternative process such as gravity sliding.

**Slide 9**

We modeled a large lunar landslide near the Apollo 17 landing site whose volume is similar to that of the Blackhawk slide (Table 1). Although detailed topographic data are unavailable, approximate rear and front runouts (5.0 and 8.8 km) and estimates of average proximal and distal deposit thicknesses (20 and $<10$ m) are known (Howard, 1973). Since no obvious scar is visible, we assumed that the landslide material initially covered the full height of the mountain massif down which it slid. We did not include width or volume variations in this model. The initial shape is very shallow: With $D = 22$ m the depth normal to the runout path is only $22 \times \sin 24.8^\circ = 8.9$ m (Table 2). Since the lunar regolith is 5- to 10-m deep (Quaide and Oberbeck, 1973) it is unlikely that bedrock failure was involved.

Acoustic fluidization produces the most accurate combi-
nation of rear and front runouts (Fig. 12). This conforms to the clear requirement that lunar landslide “fluidization” must proceed in the absence of a fluid. The introduction of even small pore pressures produces a longer rear runout than required. The frictional rheology requires substantial fluid pressure to achieve the desired front runout \( r_u = 0.41 \), confirming the need for a fluidizing process to describe its dynamics.

The small depth of the initial shape requires low yield strengths to promote failure. A simple sliding-block model can be used to estimate yield strength \( \tau_0 \) (e.g., McEwen, 1989) as

\[
\tau_0 = \rho gd \sin \beta,
\]

where \( \rho \) is density, \( g \) is gravitational acceleration, \( d \) is the deposit toe thickness, and \( \beta \) is the slope of the runout path at the deposit toe. Applying appropriate values of these variables, including a slope of 5° (a reasonable upper limit) and a toe depth of 10 m, we obtain a yield strength upper bound of 3.8 kPa, which compares well with values from DAN models (Table 3).

As a result, the Bingham rheology produces a deposit thinner than the estimate from Howard (1973) of 10 to 20 m. On the other hand, the frictional rheology produces a short, bunched deposit with a maximum depth of 50 m. Only acoustic fluidization produces the desired depths.

We repeated the simulation with \( H \) reduced to half its original value. With the initial shape occupying only the lower half of the mountain slope, depth \( D \) was doubled to conserve volume. Despite the lower potential energy of the

Fig. 7. DAN results for Slide 4.

Fig. 8. DAN results for Slide 5. The best fit was acoustic fluidization: Introduction of pore pressure made the results less accurate, hence the optimized acoustic and general fluidization results coincide.

Fig. 9. DAN results for Slide 6.

Fig. 10. DAN results for Slide 7.
initial mass, acoustic fluidization remained the best rheology. Higher yield strengths were required to maintain the same runout (1.20 kPa for the Bingham rheology and 1.27 kPa for acoustic fluidization).

IV. Discussion

Best overall rheology

The general fluidization rheology is the obvious best fit for Slides 3 and 4 and can be considered a comparable best fit for Slides 1, 2, and 7. None of the selected rheologies produces good fits for Slides 6 and 8, which we attribute to the effects of large, discrete blocks and a nonlandslide origin, respectively. Acoustic fluidization is the best fit for Slide 9. For Slide 5 the Bingham rheology is clearly superior to general fluidization, but this slide is compromised by longitudinal confinement.

The advantages of general fluidization can be described by considering the shortcomings of the other rheologies. The frictional rheology has a friction angle (20°) close to the failure angle in many cases (Table 1). This limits the reach of the slide, producing small rear and front displacements. Lower friction angles would produce greater runouts but the current choice already constitutes a physically reasonable minimum. Instead, large pore pressures are required to achieve the desired runouts and these frequently exceed the likely maximum (corresponding to \( r_u = 0.42 \) as already described). With the exception of the fluidization rheologies in the Olympus Mons aureole model, only the frictional rheology requires superlithostatic pore pressures (\( r_u > 0.5 \)), emphasizing the inability of this rheology to explain long runout landslides (Cleary and Campbell, 1993).

The frictional profiles are typified by a linear decrease in deposit thickness in the positive \( x \) direction (if pore pressure is zero, the surface of the deposit comes to rest at the friction angle of the material). Indeed, some of the observed deposit surfaces (e.g., Slides 2, 3, and 7) have linear sections, suggesting that the governing rheology does, at times, exhibit frictional behavior, pointing toward the use of two-stage fluidization processes such as those considered here.

The Bingham rheology consistently fails to produce the correct tail shape, even when model resolution is increased. Toes are frequently too deep, indicating that the yield strength is greater than values calculated directly from deposit thickness (McEwen, 1989). Actual yield strengths are likely to be smaller, probably similar to the power-law yield strengths obtained by the general-fluidization rheology.

The general-fluidization parameter values are mostly reasonable, although yield strength and apparent viscosity values are difficult to judge, being poorly constrained (viscosity is probably between 10 and 1000 kPa \( \cdot \) s in dry sand, Goetz and Melosh, 1980). The pore-pressure coefficient has an average value of 0.40, a physically plausible value given additional pressurization due to soil compaction (Iverson et al., 1997). There are at least four potential scenarios in which geologically recent martian landslide formations were lubricated:

1. As noted in the introduction, pore pressures in DAN models refer to conditions at the failure surface and are not representative of the entire deposit. Thus, for a fluid to promote greater runouts, it is not necessary for the initial material to be saturated at all depths. The top 300 to 500 m of near-equatorial crust is expected to have been dessicated over the duration of current atmospheric conditions (Fanale et al., 1986; Clifford, 1993). Since most martian slides have an initial depth of a few kilometers (Table 2), a substantial part of their mass may have been saturated with ice at the time of failure. The frictionally dissipated energy produced during the slide event is, however, unlikely to be sufficient

![Fig. 11. DAN results for Slide 8, the northern Olympus Mons aureole deposit.](image-url)

![Fig. 12. DAN results for Slide 9, the Apollo 17 landslide. No measured final profile is available, but the known required rear and front runouts are marked with vertical dashed lines.](image-url)
to melt ice for lubrication. The energy produced per unit
time, per unit mass, is $v g \sin(\alpha)FS$, where $v$ is the depth-
averaged downslope velocity of the slide (Iverson et al.,
1997). We integrated this energy over time for each slide
element in dry frictional DAN models and found that on the
order of 1 kJ/kg is produced by material on the initial, steep
section of the runout path. For ice at 0°C, 334 kJ · kg
is required for melting, so the supply is two orders of magni-
tude too small.

2. Alternatively, water may have been present in the
liquid phase prior to failure. Reasonable present-day values
of crustal thermal conductivity (2.0 W · m$^{-1}$ · K$^{-1}$) and
geothermal heat flux (30 mW · m$^{-2}$) suggest a melting point
depth of 2.3 km at the equator, assuming that melting occurs
at 252 K, a likely value given pressure and solute effects
(Clifford, 1993). Higher heat flux estimates [e.g., 45 mW · m$^{-2}$
from Toksoz et al. (1978)] yield depths as low as 1.5 km,
within the depth of most Valles Marineris slides. Nota-
bly, of the Valles Marineris slides modeled, only Slide 5
has a depth less than 3 km, and acoustic fluidization is its
most successful rheology. The recent formation of a near-
surface water-related gully feature at only 27° latitude (Ma-
lin and Edgett, 2000) supports the idea of a water table only
within the depth of most Valles Marineris landslides. No-
ther sporadic features have been placed contemporaneously with the landslides (Peul-
vast et al., 2001) and near-equatorial outflow channel drain-
age may have continued well into the Amazonian (Gulick,
2001), supporting the idea of water in the Tharsis region at
the time of landsliding.

3. If lakes existed in the Valles Marineris (Carr, 1996;
Lucchitta et al., 1994), water may have infiltrated into the
walls, providing sufficient pore pressure for later landslide
fluidization. Although our principal argument here is for the
presence of a fluid for lubrication purposes, water in this
scenario might also have been partially responsible for the
failure itself. If the Valles Marineris drained in a relatively
short period (e.g., via the catastrophic outflow channels in
the case of open depressions, and by flow through perme-
able aquifers in the case of closed depressions), leaving
lacustrine deposits which correspond (at least in part) to the
interior layered deposits of the present day, then the newly
exposed canyon walls, no longer under pressure from a
large body of water and still highly saturated, would be
unstable, resulting in widespread landsliding. Lucchitta et
al. (1994) suggested that the Valles Marineris spur-and-
gully formations may have formed subaquously. They
would thus have formed prior to the landslides while the
lakes were still in place. This scenario is then in agreement
with the observed relative ages of the landslides, spurs and
gullies, and the interior layered deposits. Problems with this
means of landslide emplacement include the ability of the
crust to hold water for a sufficiently long period. With a
length scale on the order of 10 km, and a permeability of
$10^{-17}$ m$^2$ (a likely crustal lower bound; Harrison and
Grimm, 2002), a 1-myr period would be available for water-
lubricated landslide formation (this is the time scale for
diffusive drainage of fluid from an initially saturated aquifer
10-km thick). This is a reasonable lower bound only, and it is
possible that landsliding occurred over a longer period
(given the sometimes large differences in deposit age; Luc-
chitta, 1979), requiring unreasonably low permeabilities.

4. In the absence of water, CO$_2$ may have been a suitable
fluidizing agent. Hoffman (2000) proposed a “white Mars”
in which CO$_2$ is the principal fluid at work in the cata-
strophic flood channels which open into the northern high-
lands. Depressurization of subsurface CO$_2$ liquid causes it
to flash into the gaseous phase and mix broken landslide
material into a turbulent cloud of dust, rocks, ice, and gas.
Continual degassing of liquid from the landslide mass pro-
vides an ongoing source of lubrication. Hsu (1975) simul-
lated the Elm landslide with a mixture of silt and dry ice
whose degassing did indeed produce long runouts.

We conclude this section by mentioning some aspects of
landslide dynamics that may contribute to long runouts but
are not simulated in DAN. The rear areas of many of the
martian deposits contain unbroken segments of the original
landslide mass that have slumped or rolled down the runout
path, possibly producing greater rear runout than material
composed of uniformly sized small-grained particles. In
most cases, these blocks are not sufficiently isolated from
smaller grained particles to justify their exclusion from
DAN models, with the exception of Slide 6 whose tail has
a large level section, indicating that the upper part of the
initial mass slipped in one piece down the failure plane.
The dynamical influence of large blocks, and possibly other
phenomena, may mean that actual pore pressures are some-
what less than those predicted by DAN models, although
they are likely to remain nonzero. This effect will be offset
by the increase in required pore pressures if higher friction
angles are found to be more appropriate.

Other rheologies

The Bingham rheology produces the best results for
Slide 1. A yield strength of 24.5 kPa and a viscosity of 10
kPa · s are, however, significantly smaller than the values
used to model rock avalanches (300 kPa and 40 kPa · s
respectively, Hungr, 1995).

Acoustic fluidization produces the best results for
Slides 5 and 9. Although these slides are unique in having
relatively small values of $D$, models of Slide 9 with twice
the value of $D$ (and half the value of $H$) yield the same
rheology ranking, indicating that the suitability of acous-
tic fluidization is determined by other factors such as
runout slope.

A geometrical feature of significance in Slides 4 and 8
(and to a lesser extent in Slides 2, 3, and 7) is a large
initial “footprint.” The part of the initial mass which rests
on the horizontal canyon floor does not contribute to the
downhill force that promotes movement, leading to shorter
runouts which must be compensated by higher
pore pressures or lower yield strengths. Although Slide 4
has a large footprint, its runout path is steeper than that of the other slides, resulting in a relatively good performance by acoustic fluidization.

The long runout of the Olympus Mons aureole is comparable only to submarine landslides, such as those near Hawaii (Moore, 1964) which depend on a high degree of lubrication. The shoreline of a northern hemisphere ocean, if one existed at the time of failure, would probably have had a mean elevation of about $-3800$ m (Head et al., 1999) with excursions as high as $-2000$ m at Tharsis (due to subsequent uplift of the region), which only just coincides with the lowest parts of the aureole deposit. Complete submersion would have required sea levels as high as $7000$ m. The improbability of submarine formation and the extreme values of many of the DAN parameters suggest that the mode of emplacement was something other than catastrophic landsliding. Favorable alternatives include gravity spreading (Tanaka, 1985), a low strain rate process that may have acted over a period as long as 1 myr. Spreading may have occurred predominantly along a decoupling zone lubricated by ice that possibly originated from earlier lateral spreading of hydrothermal waters driven by volcanic activity within Olympus Mons (Tanaka, 1985).

**Gravity effects**

The ratio $H/L_r$ is a useful, although not perfect, measure of the extent to which a landslide runout is “long.” McEwen (1989) and Hsu (1975) provide $H/L_r$ vs volume data for Mars and Earth, respectively (Fig. 13). $H/L_r$ decreases with volume, meaning that larger landslides have longer runouts relative to their initial height. Also of note is the offset between the martian and terrestrial trends, evidence that the rheology governing the slides cannot be purely frictional. For a given set of parameter values, frictional slides produce a constant $H/L_r$ value independent of $V$ and gravitational acceleration because the basal resistance scales with gravity and volume and therefore with the downslope acceleration. For the Bingham rheology and the power-law part of the fluidization rheologies, the basal resistance does not scale with the downslope acceleration because it is based on a gravity- and volume-independent yield strength. The result is a dependence of $H/L_r$ on volume and the emergence of separate $H/L_r$ trends for different gravity environments.

McEwen (1989) recognized the need for a gravity-independent yield strength (see also Dade and Huppert, 1998) and in Fig. 13 his point is demonstrated by a set of DAN models of varying volume, run with both the Bingham and
general-fluidization rheologies in martian and terrestrial gravity. A uniform width, zero-entrainment version of the Slide 5 DAN model was scaled spatially to obtain slides of different volumes but identical aspect ratio \((D/H)\). The rheological parameter values that produced the required front runout in the original scale were used for all other scales on both planets, reflecting the assumption that yield strength and viscosity are independent of gravitational acceleration. Slide 5 has a fairly low yield strength: Experiments with higher yield strength slides (such as Slide 2) show that the curves in Fig. 12 shift upward together without significantly changing their offset. Note that the symbol representing Slide 5 (Fig. 13) falls over the intersection of the two solid curves at \(V = 14 \, \text{km}^3\), implying that both sets of rheological parameters were chosen so that \(L_f\), and therefore \(H/L_f\), coincide at the original scale.

The Bingham and fluidization rheologies reproduce the first-order features of the observed trends extremely well. The general-fluidization curves level out with decreasing volume because the efficiency of the power-law stage decreases (the yield strength remains constant while the thickness of the deposit decreases), and the frictional stage becomes the dominant influence on runout. This departure reflects the fact that fluidization is unlikely to occur at low volumes (relative to a specific gravitational acceleration). The suitability of the Bingham rheology is also somewhat restricted: It does not initiate failure for volumes below about 0.2 \(\text{km}^3\) for martian slides and 0.001 \(\text{km}^3\) for terrestrial slides. It must be remembered that all the curves in Fig. 13 were generated with one set of rheological parameter values. It is physically unlikely, however, that every martian landslide can be described by the same parameter values. There will be a dependence on grain size, rock or soil density, fracture distribution, and other local phenomena. Parameter variations are also evident in the DAN results, although these may be exacerbated by the simplified nature of the model that excludes small-scale, unobservable variations in failure geometry and other second-order effects. Overall, Fig. 13 provides further support for McEwen’s claim of a gravity-independent yield strength and extends it to general and acoustic fluidization, processes which include the frictional rheology (which alone cannot account for the observed trends).

The west Olympus Mons aureole (Slide 8, Fig. 13) lies well below the Bingham and general-fluidization trends for Mars. This reflects its unusually long runout (hence low \(H/L_f\)) which our rheologies were only able to reproduce with high pore pressures and/or low yield strengths. The Blackhawk landslide (Slide 1) is noticeable as the only data point lying close to the terrestrial trend, as expected. Its position relative to the curves generated by Slide 5 parameter values is a combination of lower yield strength (which lowers \(H/L_f\)) and higher aspect ratio (which raises \(H/L_f\)) than those of Slide 5. The Apollo 17 landslide (Slide 9) lies between the terrestrial and martian trends in Fig. 13. Because the Moon has the lowest gravitational acceleration of the three bodies, its landslides might be expected to lie above the martian trend (vertically separated by about 2.3, the ratio of martian and lunar gravities); however, Slide 9 falls below this trend. At least two factors may have affected its value of \(H/L_f\). Impacts responsible for the initial failure may have contributed energy to the subsequent motion, resulting in longer runouts than expected [this may have affected other lunar slides close to impact craters, such as the Tsiolkovsky landslide (Wu et al., 1972; Guest, 1971; Fig. 13)], and 2. the slide is probably an avalanche of regolith material (which is as deep as 15 m in places; Quaide and Oberbeck, 1973) and not a bedrock landslide.

**Acknowledgments**

We acknowledge the use of PDS Viking images from the U.S. Geological Survey and NASA, Mars Global Surveyor MOC images from the NASA Jet Propulsion Laboratory, and an image from the U.S. Department of Agriculture. This work was supported by NASA Grants NAG5-7190 and NAG5-9799.

**References**


