The geology of the Viking Lander 2 site revisited

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Abstract
Reevaluating the geologic history of the prior Mars landing sites provides important ground truth for recent and ongoing orbital missions. At the Viking 2 Lander (VL2) site, topographic measurements of relict landforms indicate that at least 100 m of sedimentary mantle material has been stripped away. The observed paucity of impact craters \(<\ 100 \text{ m in diameter suggests that resurfacing processes (likely in the form of the recent deposition and removal of thin } 1–10 \text{ m mantle layers) continue up to the present. A dearth of craters in the } 100–500 \text{ m diameter range, however, also necessitates erosion of a thicker mantle layer. Partially inverted chains of secondary craters from nearby Mie Crater indicate that the mantle was already in place when the impact occurred. The density of craters superposed on Mie ejecta is consistent with a Late Hesperian age and provides a minimum age constraint for the mantle’s emplacement. The thermophysical properties of the surface around VL2 as observed with Thermal Emission Imaging System (THEMIS) data indicate that the landing site occurs in an intracrater region that may typify mid to high northern latitude sites. Elevated thermal inertias of a pedestal crater superposed atop a larger pedestal crater suggest that rocky or indurated material can be created by impacts into sedimentary targets. Rock abundances at VL2 are consistent with the addition of impact-emplaced material from the missing small impact crater population documented in this study. Thus, the VL2 site may be a reasonable proxy for the landscape expected at the upcoming Phoenix Lander site.

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1. Introduction

Although recently acquired remotely sensed data of Mars have contributed greatly to our understanding of global-scale characteristics, difficulty remains in linking Mars as observed from orbit with the martian surface as viewed by landed spacecraft. Two critical issues are the spatial and time scales involved. Orbital imaging and spectral data sets with global coverage typically have relatively coarse spatial resolution, i.e., \( \geq 100 \text{ m/pixel}^{-1} \) (higher spatial resolution imaging exists, but it is limited in terms of the percentage of surface area covered). The geologic processes inferred from orbit necessarily operate on the macro-scale and tend to reflect long timescales. Yet surface properties observed in situ can be highly variable over short time scales at the cm- to m-scale. Spectral observations are controlled by physical interactions on even smaller spatial scales (e.g., \( \sim 10^{-6} \text{ m for visible light} \)). It remains a challenge to link the processes inferred from the micro-structure observed with landed spacecraft to macro-scale processes inferred from orbital observations. Landed missions have successfully returned data from only five locations on the surface (Viking Lander 1, Viking Lander 2, Pathfinder, Spirit, and Opportunity). Therefore, re-evaluating the prior landing sites with newly returned orbital data will help integrate orbital and surface observations and improve knowledge of both. Establishing a firmer connection between the processes inferred from combined surface and orbital observations also allows for better predictions of the environments at sites where no ground truth exists. Hence, future landing missions can be better targeted to address specific science objectives. The goal of this analysis is therefore to use recently acquired orbital data sets in order to reassess the geologic processes that have shaped the landscape at the Viking Lander 2 (VL2) site and in so doing strengthen the link between orbital and landed data sets.

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Of the five sites where missions have successfully landed, VL2 is located the farthest from the equator at 47.7° N (Mayo et al., 1977), lying within the northern plains of Mars near the eastern edge of Utopia Basin (Fig. 1). Measurements of buried craters indicate that the underlying basement of the northern plains dates back to the Noachian period (Frey et al., 2002). This underlying heavily cratered, ancient surface has been overprinted by a unit containing numerous wrinkle-ridges (e.g., Thomson and Head, 2001; Withers and Neumann, 2001). This unit is inferred to be volcanic based upon its morphologic similarity to lunar volcanic plains and other plains on Mars of probable volcanic origin (such as Hesperia and Lunae Plana), but could also represent a layered sedimentary sequence since wrinkle ridges are not diagnostic of unit origin (e.g., Schultz, 1972; Chicarro et al., 1985). Atop the wrinkle-ridged layer is a mantle of younger sedimentary material. Several studies interpret this mantle as volatile-rich (e.g., Schultz and Lutz, 1988; Lucchitta, 1993; Carr and Head, 2003; Tanaka et al., 2003), although there is no consensus regarding the origin or modification history of this material. This material has been proposed to be eolian debris derived from the circumpolar layered deposits (Soderblom et al., 1973), air-fall (loess) deposits that
are remnants of circum-polar expansion due to obliquity variations (Schultz and Lutz, 1988), eolian deposits of unspecified provenance (Zimbelman et al., 1989), oceanic/lacustrine deposits (e.g., Lucchitta et al., 1986; Scott et al., 1992; Parker et al., 1993), glacial deposits (e.g., Kargel et al., 1995), deposits from volcanic-ice interactions (e.g., Chapman, 1994), and mass-wasting deposits (e.g., Jöns, 1985; Tanaka et al., 2001).

Regardless of the origin of this sedimentary mantle material, substantial differential erosion has modified the terrain around the VL2 site (Mutch et al., 1977; Arvidson et al., 1979). Unarmored surfaces around the VL2 site have been deflated, leaving resistant features standing in relief. In this study, topographic measurements of relict landforms (such as inverted secondary crater chains and pedestal craters) establish the vertical magnitude of eolian deflation. Second, high-resolution images from the Mars Orbiter Camera (MOC) and the Thermal Emission Imaging System (THEMIS) are used to measure the size distribution of impact craters down to crater sizes that are an order of magnitude smaller than previously reported for this area (Greeley and Guest, 1987; Hartmann et al., 2001). Since smaller craters are more sensitive to the dynamics of the uppermost surface layers than larger ones, examining the cratering record over a full range of diameters helps constrain the complex depositional and erosional history of the mantle layer. Although additional high-resolution images of the VL2 site have been acquired by the HiRISE (High Resolution Imaging Science Experiment) camera on Mars Reconnaissance Orbiter (McEwen et al., 2007), they were not publicly available prior to the submission of this work and await further analyses in a future publication. Third, thermal data from the Thermal Emission Spectrometer (TES) and THEMIS instruments are used to characterize the thermophysical properties of the upper surface layer of the mantle and open possible connections between the abundance of rocks and thermal signature. Finally, we incorporate all of the insights gained from this study to place constraints on the nature, emplacement, and resurfacing history of the sedimentary mantle layer in the vicinity of the Viking 2 Lander. The results of this study provide a stratigraphic framework for interpreting the recent geological history of the northern plains of Mars.

2. Background and methods

2.1. Landing site location

Although the general position of the Viking 2 Lander is well known from radio tracking (Folkner et al., 1997), the lander’s exact location relative to nearby surface features has been difficult to establish. The locations of VL1 and MPF have been triangulated using enhanced super-resolution composites of horizon topography (e.g., Parker and Kirk, 1999). This method has been difficult to apply to the VL2 site due to the extremely flat local topography: the surface elevation varies less than 1 m within a radial distance of 100 m from the lander (Mutch et al., 1977). Three possible lander locations have been proposed that are clustered with a few km of each other (Stooke, 1997, 2004; Parker and Kirk, 1999; Oberst et al., 2000). Though not publicly available at the time of this writing, the lander has been directly observed from orbit by the HiRISE camera on Mars Reconnaissance Orbiter (HiRISE image PSP_001501_2280; Parker et al., 2007). This location is close to those identified previously and places the lander 12.3 km southwest from the pedestal crater Goldstone (Fig. 1). The rim of the 104 km diameter crater Mie is over 200 km to the east of the site. About 13 km southeast of the lander is a lobe of Mie ejecta that is elevated more than 50 m relative to its surroundings. Thus, VL2 does not rest directly upon a lobe of Mie ejecta, but is instead situated on smooth terrain that is morphologically similar to the surrounding intracrater plains material.

Lander panoramas at the VL2 site reveal a field of loose rocks in a matrix of fine-grained sediment. Most rocks are pitted, but it is unclear if these pits are primary vesicles, secondary erosional features, or some combination of the two (Thomson and Schultz, 2003). Rock shape analyses suggest that the rocks are glassy, aphanitic fragments that have not been subjected to fluvial transport (Garvin et al., 1991). Fine materials at the site exhibit variable cohesion and are present both as loose drifts and as weakly bonded aggregates that form either clumps or platy horizons of material (duricrust) a few cm below the surface (Moore et al., 1987). One of the more interesting macroscopic features observed in the soil-like materials is a network of small trenches that form a polygonal network. These trenches are about 1 m across and 10–15 cm deep, are relatively rock-free, and are partially infilled in places by small eolian ripples (Mutch et al., 1977). Thermal modeling indicates these polygonal troughs are the result of repeated seasonal cooling and thermal contraction of ice-rich soil (Mellon, 1997). Given the wide distribution of small-scale polygonal features and other patterned ground in the middle to high latitudes of Mars (e.g., Mangold, 2005), the VL2 site provides a unique opportunity to observe these terrains up close.

2.2. Craters as measures of deflation

Craters larger than about 2 km in diameter near VL2 site can be classified into three categories—fresh craters, filled craters, and differentially eroded craters—based on their morphology and topographic expression. Reliable classification of craters smaller than a few km in diameter into these categories met with limited success due to the cross-track gaps in MOLA profile coverage. Fresh impact craters (Fig. 2A) have crisp rims, little to no mantling or infilling, and commonly have distinct thermal signatures. Topographic profiles across fresh craters show deep, unfilled cavities whose floors lie well below the level of the surrounding terrain. Most fresh craters tend to be relatively small in size, which is likely a reflection of the frequency of mantling and resurfacing events. Filled craters (Fig. 2B) are not pristine but do not appear to have been subject to substantial differential erosion. Many appear to have acted as sedimentary debris traps and have mantled and muted exteriors as well as interiors.

Differentially eroded craters (Figs. 2C and 2D) are characterized by crater floors that are higher in elevation that
Fig. 2. Examples of crater types around the Viking 2 Lander site. (A) Fresh impact crater located east of Mie Crater near 51.7° N, 188.0° W. Portion of THEMIS daytime infrared image I04151002 band 9 (12.57 µm). (B) Example of a filled crater near 46.2° N, 224.3° W. Portion of THEMIS visible image V05825007 band 3 (0.654 µm). (C) Example of a full pedestal crater located east of Mie Crater near 41.5° N, 206.8° W. Portion of THEMIS daytime infrared image I04601009 band 9 (12.57 µm). White dashed region indicates margins of area given in Fig. 8D. (D) Differentially eroded crater (“heavily modified” or “partial” pedestal crater) near 41.6° N, 206.6° W. Portion of THEMIS visible image V13575007 band 3 (0.654 µm). The pedestal craters shown in Figs. 8C and 8D also appear to be partially infilled. Accompanying each image subset is a portion of a MOLA topographic profile: (a–a′) 17884, (b–b′) 12802, (c–c′) 18532, (d–d′) 14395.

the surrounding terrain. Craters in this category express a substantial range of forms can be further divided into “full” pedestal craters and “heavily modified” or “partial” pedestal craters. Pedestal craters, initially recognized with Mariner 9
data (McCauley, 1973), are mesa-like impact structures that lie perched above the surrounding terrain (e.g., McCauley, 1973; Arvidson et al., 1976; Guest et al., 1977; Schultz, 1988; Schultz and Lutz, 1988). In these craters, the impact process has produced an irregular, roughly annular region around the crater that is more resistant to deflation than the surrounding material. Subsequent differential erosion has stripped away the surrounding unprotected material, leaving these craters standing in relief. Given its susceptibility to erosion, the host material appears to be weakly consolidated (McCauley, 1973) and possibly volatile-rich (Schultz and Lutz, 1988). The height of the pedestal therefore provides a direct measure of the former surface elevation and can be used to infer the minimum thickness of material that has been removed. Distinctive examples of full pedestals (Fig. 2C) possess a roughly annular region of high-standing, armored terrain that extends up to 4+ crater radii from the raised central cavity (e.g., Schultz, 1988; Schultz and Lutz, 1988; Wrobel et al., 2006). The upper surface of the pedestal often has fine radial striations, and the outer edge of the pedestal commonly terminates in an outward-facing scarp with scalloped edges. Heavily modified or partial pedestals are a more morphologically diverse group that includes craters with a narrow, armored pedestal that extends only 1–2 crater radii (Fig. 2D). Further differential erosion may result in more enigmatic forms, some of which may superficially resemble volcanic structures (Schultz, 2000, 2003).

Some have alternatively interpreted pedestal craters as primary, unmodified structures (e.g., Mutch and Woronow, 1980). In presenting the case that pedestal craters are unmodified impacts, Mutch and Woronow (1980) observed that the armored region around many of these craters is often roughly azimuthally symmetric. If these craters are products of erosion, this would seem to require an isotropic erosive agent. Other erosion scour features such as yardangs are typically oriented parallel to the direction of the prevailing wind (e.g., Ward, 1979).

The symmetry of pedestal deposits may instead be indicative of the extreme contrast in friability between the material comprising the pedestals and the surrounding host terrain. Erosion of the surrounding mantle material may be enhanced by the breakdown of a volatile-rich cementing medium (Schultz, 1988; Schultz and Lutz, 1988; Wrobel et al., 2006). Topographic profiles of pedestal craters (Figs. 2C and 2D) demonstrate that the floors of many pedestal craters lie at elevations substantially above the surrounding host terrain—a fact that is difficult to reconcile with a non-erosional origin.

Others have interpreted pedestal craters in this region as possible volcanic structures: either as small shield volcanoes (Mutch et al., 1977) or akin to Icelandic subglacial volcanoes (Chapman, 1994; Hodges and Moore, 1994). Although there are some morphologic similarities between some pedestal craters and certain volcanic structures, “a demonstrably broad spectrum of impact crater morphologies on Mars, however, lends uncertainty to volcanic interpretations” (Hodges and Moore, 1994, p. 139). Although the existence of other small volcanic structures cannot be entirely ruled out, all of the observed characteristics of the pedestal crater-like structures in this study area are fully consistent with a modified impact origin.

In this study, we measure the pedestal height from the outermost edge of the pedestal, which typically has a sharp break in slope. All topographic measurements in this study were made from Mars Orbiter Laser Altimeter (MOLA) topographic profiles (Smith et al., 1999a). Individual MOLA data points have a precision of <40 cm on smooth flat surfaces that may increase up to ~10 m on 30° slopes (Smith et al., 1999b). Since the height measurements reported in this study represent the difference between two elevation levels, an estimate of the measurement uncertainty is indicated by crossover analysis of MOLA profiles, which have a rms residual error of ~2 m (Neumann et al., 2001). In practice, both the pedestal surface and surrounding plains have a somewhat undulatory topography, yielding a measurement uncertainty that is likely on the order of several meters (with ±5 m being a reasonable uncertainty estimate).

An alternative method to determine the former surface elevation using pedestal craters is based on the difference between observed and calculated rim heights (Bleacher et al., 2003), a technique similar to that used to estimate lunar mare basalt thickness using partially flooded craters (De Hon, 1979). Topographic measurements of martian craters are used to define empirical relationships between crater diameter and rim height for simple and complex craters (e.g., Garvin et al., 2003). This method assumes that pedestal craters retain the rim morphology of pristine craters. Major sources of uncertainty remain, however, such as crater degradation and differences in crater geometries due to different target material strengths and/or different impactor velocities in primary versus secondary craters. Therefore, a direct measurement of pedestal scarp heights above the surrounding surface is preferred, although the rim height difference technique is useful for craters where the pedestal itself has been eroded back to the crater’s rim. Both of these methods provide only minimum estimates of the former mantle thickness, for no constraints are placed upon the possible maximal mantle thickness.

2.3. Areal density of impact craters

In the absence of direct sampling, the principal method for determining the relative ages of planetary surfaces is to compare the areal density of impact craters: in general, older surfaces have accumulated more craters than younger surfaces (Shoemaker and Hahnman, 1963). Since we lack age dates for surface samples of known geologic provenance on Mars, the lunar cratering record has been adapted as a calibration standard for estimating absolute ages (Ivanov, 2001; Neukum et al., 2001). In addition to providing surface exposure ages, crater size frequency distributions also can provide age constraints on resurfacing and modification processes.

The relative importance of primary versus secondary craters at small (<1 km) crater diameters has again become the subject of much debate (McEwen and Bierhaus, 2006; Quantin et al., 2007). Dating youthful surfaces of limited surface area becomes especially problematic in this regard due to inherent
problems in the statistics associated with small numbers of impacts. Here the aim is not to resolve this debate, but instead focus on understanding the nature of the processes that have affected the distribution of small craters in this study. This contribution attempts to avoid the problem of small surface areas by examining apparently recent surface modification over a large distribution attempts to avoid the problem of small surface areas by examining apparently recent surface modification over a large area. Thermal inertia, which is a measure of a material’s resistance to changes in temperature, has units of J m⁻² s⁻¹ K⁻¹ and is formally defined as

\[ I \equiv \sqrt{kc}, \]  

where \( k \) is the thermal conductivity, \( \rho \) is the bulk density, and \( c \) is the specific heat capacity of a material. For geologic materials, variations in \( I \) are dominated by variations in the thermal conductivity \( k \), which can vary over two orders of magnitude. The product of \( \rho c \), in contrast, typically varies by only a factor of 3 for most geologic materials (Palluconi and Kieffer, 1981).

If nominal values for the bulk density and specific heat are assumed, thermal inertia can be related to the effective or mean particle size. Low thermal inertia regions are straightforwardly interpreted as being covered with loose silt-sized or smaller particles that exceed one diurnal skin depth in thickness. Extremely high thermal inertias are also unambiguous in that they require exposures of non-particulate surfaces such as bedrock or a similarly competent, slab-like material (i.e., an indurated surface). Intermediate to high thermal inertias, however, are more difficult to interpret because they could represent sand-sized or coarser particles, exposures of cobbles and/or boulders, indurated material, or some combination of the above. Thus, the non-uniqueness of thermal inertia data necessitates cautious interpretation.

Ideally, temperature measurements taken over a full diurnal cycle would be used to give the most accurate view of the thermal properties of a surface. In lieu of broad temporal coverage for most regions, thermal models have been developed for use with a single temperature observation (see Appendix I in Kieffer et al., 1977). In this analysis, THEMIS thermal inertias were derived with a model similar to that used with IRTM data. This is a simplification from the coupled surface-atmosphere model used for the TES thermal inertia (Mellon et al., 2000), although the elevation correction of Bridges (1994) was included. A simplified model is justified because a more involved thermal model might not necessarily provide more accurate results given the inherent uncertainties in THEMIS-derived thermal inertia values. Higher noise levels in the THEMIS instrument and the absence of spatially and temporally coincident albedo and dust opacity measurements (among other factors) contribute to an absolute uncertainty in THEMIS-derived thermal inertia values that is estimated to be ~25%, compared to ~6% for TES-derived thermal inertia values (Jakosky et al., 2005).

### Table 1a

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Area surveyed with MOC images = 3.2 × 10² km²

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Area surveyed with THEMIS visible images (excluding overlap) = 7.2 × 10³ km²
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Nighttime infrared images used in crater counts

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<tr>
<td>I07941016</td>
<td>103.0</td>
<td>4.51</td>
<td>269.9</td>
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<td>I08041032</td>
<td>102.5</td>
<td>4.43</td>
<td>275.0</td>
</tr>
<tr>
<td>I10175011</td>
<td>103.5</td>
<td>4.18</td>
<td>12.5</td>
</tr>
</tbody>
</table>

Area surveyed with THEMIS night IR images = 2.5 × 10\(^5\) km\(^2\)

3. Results

3.1. Vertical extent of deflation

Fig. 3A gives the measured pedestal crater scarp heights for all pedestal craters greater than 1.0 km in diameter within 50 km of the landing site. These pedestal scarp heights, which are inferred to represent the minimum thickness of material removed, range from 62 to 102 m with a mean value of 79 m (Fig. 3B). Inferred mantle thickness values are slightly less than but generally consistent with the “several hundred meters” value estimated previously using shadow measurements of pedestal craters and an a priori assumption of their exterior slopes (Arvidson et al., 1979).

The largest value of pedestal crater scarp relief measured in this study, 102 m, occurs at the 3.1 km diameter pedestal crater Canberra. The southern edge of Canberra’s pedestal lies 82 m above the level of the surrounding terrain. Of particular interest is a second smaller pedestal crater that is superposed atop the eastern edge of the main pedestal (see Fig. 4). The top of this smaller unnamed pedestal crater is an additional 20 m above the main pedestal, providing a total relief of 102 m. This set of “nested” pedestal craters is not unique on Mars; other examples have previously been documented in both the polar and equatorial regions (Schultz and Lutz, 1988).

3.2. Impact crater density

Further details about the nature and timing of regional resurfacing processes can be provided by examining the number density of craters measured across a range of crater sizes. Impact crater densities were measured with THEMIS infrared
Fig. 3. (A) Topographic shaded relief map of the VL2 region using the MOLA 128 pixels per degree gridded data. Pedestal heights of large craters in the immediate vicinity of the landing site (indicated by filled star) are reported in meters. The artificial illumination direction is from the NW (az. 315°, el. 30°), and the contrast was adjusted to enhance local topographic features. (B) Histogram of pedestal crater scarp heights given in (A).

and visible images, and with Mars Orbiter Camera (MOC) narrow-angle images. The results are presented in two formats in Figs. 5a–5b: a cumulative plot (showing the total number of craters larger than a given diameter) and an incremental plot (showing the number of craters observed within a given diameter range). Large craters (>1 km in diameter) are consistent with a production population of craters. The 2- and 5-km-crater densities \( N(2) \) and \( N(5) \) are \( 5.7 \times 10^2 \) and \( 1.2 \times 10^2 \) craters > D per \( 10^6 \) km\(^2\), respectively, which correspond to a Late Hesperian surface according to the crater-density boundaries of Tanaka (1986).

Smaller craters (<500 m), in contrast to larger ones, are severely depleted. For example, less than 10% of the predicted original population of craters 200 m in diameter remains. Craters 100 and 50 m in diameter have been reduced in number by over two orders of magnitude such that less than 1% of the presumed original population remains. This deficiency indicates an extreme loss of small craters by erosion and/or infilling. The fact that even very small decimeter-diameter craters are not in production indicates that the resurfacing process continues to be active.

Additional counts were also made of the density of superposed craters on the ejecta of the crater Mie (Fig. 5c) and at a location ~100 km further west of the VL2 site (Fig. 5d). Although smaller measured craters in Fig. 5c have been affected by erosion, results from the largest-sized crater diameter bins are consistent with a Late Hesperian age for the Mie impact. In the region east of VL2 (Fig. 5d), there is a notable lack of small craters <500 m in diameter. This deficiency is similar to the cratering record observed at VL2. The relevance of these additional count areas is discussed further in subsequent sections.

3.3. Thermophysical characteristics

A wide diversity of small-scale surface types are evident in the THEMIS infrared mosaic (Fig. 6). VL2 is situated on a terrain similar whose thermophysical properties are broadly similar to the surrounding intracrater plains material. The THEMIS-derived thermal inertia of this material is 190 J m\(^{-2}\) s\(^{-1/2}\) K\(^{-1}\) (Fig. 6B), compared to the TES-derived value of 234 J m\(^{-2}\) s\(^{-1/2}\) K\(^{-1}\) for this region (Putzig et al., 2005). Although VL2 did not come to rest upon a lobe of continuous ejecta, a subdued discontinuous component of distal Mie ejecta may still be present either at the surface or buried beneath more recent mantling deposits.

3.4. Rock abundance

The surface area covered by rocks >2 cm at the VL2 site as estimated by direct visual observation ranges from 16 to 19%, depending on the boundaries of the count area used (see Table 2; Moore and Jakosky, 1989; Moore and Keller, 1990). Rock counts made using HiRISE images (Golombek et al., 2007) over a wider area indicate a variable but slightly higher overall abundance of large rocks (>1.5 m), but the results are generally consistent with the exponential form of rock counts made from lander images. Surfaces within about 3 m of the lander that are accessible to the sampling arm are referred to as the “near field” and encompass a partial (~160°) annulus in front of the lander. Using the catalog of rock data compiled by Moore and Keller (1990, 1991), an estimate of the total volume of rocks in the near field can be derived using ellipsoidal (volume = \( \pi/6 l w h \)) or block (volume = \( l w h \)) approximations. Here \( l, w, \) and \( h \) represent rock length, width, and height, respectively. At VL2, this results in an “observed” volume of rocks of 0.13–0.25 m\(^3\) in the 12.75 m\(^2\) surface area of the near field, which is equivalent to a rock layer about 1–2 \( \times 10^{-2} \) m in thickness. This observed thickness can be compared to the predicted thickness of ejecta at the site based on the visible craters in the vicinity. A first-order estimate of ejecta thickness (\( t \)) can be obtained from scaling relationships (Eq. (2)) (e.g., Housen et al., 1983;
Fig. 4. Portions of THEMIS daytime infrared images I10668006 and I05252005 (band 7, centered at 11.04 μm) over the pedestal crater Canberra. Inset shows a 3D perspective view looking northwest of Canberra with smaller superposed pedestal crater (vertical exaggeration is ~20 to enhance small elevation differences). Topographic profiles labeled A–A’ and B–B’ are portions of MOLA profiles 18671 and 13289, respectively. Vertical black arrows in topographic profiles indicate the craters’ rim positions, while black horizontal arrows and dashed vertical lines indicate extent of the pedestals.

Schultz and Mustard, 2004):

\[ t = 0.032r^{-2.61}/(0.64RD)^{-3.61} \]  

In Eq. (2), \( r \) is the range from the source crater and \( RD \) is one-half of the observed rim-to-rim diameter of the crater [see Schultz and Mustard (2004) for further discussion]. Considering the size and distance to each of the visible craters within 100 km of VL2 (excluding Mie Crater), the total amount of predicted ejecta is \(~7 \times 10^{-3} \) m, in contrast to the observed thickness of \(1–2 \times 10^{-2} \) m. Hence, the amount of rocky material observed at the surface slightly exceeds the first-order predicted ejecta estimate based on visible craters by a factor of about 1.4 to 2.9. Despite the fact that many complexities of the cratering process are not addressed in this first-order ejecta thickness
estimate (including the effects of impact angle, target layering, crater rays, ejecta bulking, and atmospheric effects), the predicted rock abundance may be overestimated since it assumes that all of the ejecta is delivered as rocky debris. Thus, the actual difference between observed rocks and predicted ejecta blocks from visible craters may be larger.

4. Discussion

4.1. Mantle thickness and characteristics

The Viking imaging team proposed several possible geologic histories for the VL2 site, none of which was deemed
Fig. 6. Mosaic of nighttime brightness temperatures of the VL2 region obtained from THEMIS nighttime infrared images overlain on a shaded relief map. Calibrated band 9 (12.57 μm) radiances of each image were normalized to the image with the highest radiance to reduce effects of local time variations. Subsets indicated by red boxes give THEMIS-derived thermal inertia values in units of J m⁻² s⁻¹/² K⁻¹. Subset (A) is a portion of image I05357014 over Canberra pedestal crater; subset (B) is portion of I10175011 over the VL2 site and lobe of Mie ejecta. Individual THEMIS images are 32 km in width.

uniquely persuasive (Mutch et al., 1977). In their first proposed history, the lander is situated on a poorly sorted surface debris flow from the crater Mie. A second proposed history is that the surface is a thin, widespread basaltic lava layer that was emplaced upon a layer of eolian material, followed by erosion and disaggregation of the lava layer into the present boulder field.

A third proposed history is that the landscape is covered by an eolian sedimentary deposit with interleaved ejecta and/or volcanic deposits. These coarser deposits were concentrated into a lag deposit during subsequent deflation. A fourth proposed scenario is that the rocks are actually cemented ferricrete blocks formed in the sub-surface that were exposed by later deflation.
The results from this current study support the third proposed geologic history: a thick (>102 m), possibly volatile-rich, mantle of sedimentary material was formerly present at the VL2 site. Subsequent erosion has stripped much of this mantle away and left a lag surface of coarser debris. Deposition and removal of thin (1–10 m) mantle layers has apparently continued to the present, interspersed with polygonal crack growth in the upper few meters of ice-rich soil. Many characteristics of the landscape at VL2 can be better understood in light of this massively deflated, ice-rich eolian mantle layer.

A direct estimate of the mantle layer’s former thickness is provided by measurements of relict topographic landforms such as pedestal craters. An indirect thickness estimate can also be obtained from impact crater statistics. Crater counts in this region document a deficiency of craters <500 m in diameter (Fig. 5). On the basis of the depth-to-diameter ratio for simple martian craters (depth = 0.21D^{0.81} for \( D < 7 \) km; Garvin et al., 2003), the inferred depth of erosion is \( \sim 120 \) m. This inferred thickness value is consistent with the direct estimate made at the pedestal crater Canberra (102 m). If, however, most small craters on Mars are secondaries as suggested by McEwen et al. (2005), this indirect estimate may be reduced by a factor of about 2 due to the smaller depth/diameter ratios of secondary craters (e.g., Pike and Wilhelms, 1978; Golombek et al., 2006).

Canberra crater also provides significant insight into the regional stratigraphy. The presence of the small, unnamed pedestal crater superposed on Canberra (Fig. 4) demonstrates a stratigraphic succession that can be unambiguously determined. Fig. 7 gives a schematic interpretation of the inferred sequence of events responsible for this nested stack of pedestal craters. In Fig. 7A, the Canberra structure formed in a layer of unconsolidated mantle material. An additional layer of mantling material was deposited over Canberra after it was formed in Fig. 7B, followed by the impact of the smaller superposed crater in Fig. 7C. Note that this smaller crater (diameter = 1.4 km) formed entirely within the upper mantle layer. Subsequent deflation has removed all of the unarmored material and resulted in the present configuration of a smaller pedestal crater perched atop a larger one (Fig. 7D).

Table 2
Rock abundances and thermal inertias of prior landing sites

<table>
<thead>
<tr>
<th>Landing site</th>
<th>Sample field rock abundance (%)(^a)</th>
<th>Near + far field rock abundance (%)</th>
<th>IRTM rock abundance (%)(^f)</th>
<th>TES thermal inertia (J m(^{-2}) s(^{-1/2}) K(^{-1}))(^g)</th>
</tr>
</thead>
<tbody>
<tr>
<td>VL1</td>
<td>8</td>
<td>6(^b)–16(^c)</td>
<td>14 ± 4</td>
<td>283 ± 14</td>
</tr>
<tr>
<td>VL2</td>
<td>14</td>
<td>16(^c)–19(^d)</td>
<td>17 ± 3</td>
<td>234 ± 11</td>
</tr>
<tr>
<td>MPF</td>
<td>–</td>
<td>19(^e)</td>
<td>20 ± 2</td>
<td>386 ± 10</td>
</tr>
</tbody>
</table>

\(^a\) Moore et al. (1979, 1987).

\(^b\) From Golombek and Rapp (1997), Fig. 4.

\(^c\) Far-field estimates assume near-field distribution of rocks <25 cm. VL1 far-field estimate includes 4.5% area covered by rock “outcrops” (Moore and Keller, 1990).

\(^d\) Far-field estimate assumes near-field distribution of rocks >3.5 cm and <14 cm (Moore and Jakosky, 1989).

\(^e\) Golombek et al. (2003).

\(^f\) Reported values for VL1 and VL2 are 4° latitude by 5° longitude averages (Christensen, 1986). MPF value given in Golombek et al. (1999) is an average for entire landing ellipse (100 by 200 km). Values for single (1° by 1°) pixels containing landers are 16, 17, and 18 for VL1, VL2, and MPF sites, respectively (Golombek, personal comm.; MPF single pixel value given in Golombek et al. (1999)).

\(^g\) Values reported are 6 by 6 pixel averages from 1\(^2\)° binned TES data (Putzig et al., 2005).

Fig. 7. Schematic diagram of major events in the evolution of the pedestal crater Canberra. See text for details. Figures are not drawn to scale.

Nested pedestal craters illustrate several important aspects of the mantle layer and the processes that have affected it. First, the presence of nested pedestal craters implies multiple depositional episodes. Second, both the deposition and exhumation processes were gentle enough not to erase the small-scale crater structure on armored surfaces (a fact observed previously in other friable martian deposits; Schultz and Lutz, 1988; Zimbelman et al., 1989). Third, the complete removal of at least 102 m of mantle material indicates that the host mantle material consisted of weakly consolidated sediments that were highly susceptible to erosion. The extreme friability of this material seemingly precludes several proposed emplacement mechanisms of sedimentary material in this region. For example, glacial or mass-wasting processes would be expected to produce poorly sorted, clastic deposits that seem unlikely to express the advanced degree of differential erosion observed here. As previously mentioned, some authors have suggested that the extreme friability of mantle material is due to the removal of a volatile-cemented dust via sublimation during backwasting (Schultz and Lutz, 1988). Although the former presence of volatiles in these deposits is not directly implicated by
this present study, this explanation is consistent with observed lack of asymmetric eolian scour around resistant or armored features. The role of volatiles notwithstanding, all of these observations help place constraints on the origin, formation, and modification of this mantle deposit.

The observed characteristics of the mantle deposit in the VL2 region are consistent with previous interpretations of a latitude-dependent, possibly volatile-rich mantling deposit. A deficiency of fresh craters in high latitude regions was initially observed with Mariner 9 images and was attributed to a mantle of eolian debris extending down from both poles to 30°–40° N and S latitudes (Soderblom et al., 1973). Subsequent analyses with Mariner 9 (Arvidson et al., 1976), Viking (e.g., Schultz and Lutz, 1988), and Mars Global Surveyor (e.g., Mustard et al., 2001) data have further documented the presence of a latitude-dependent mantle deposit. Mars Odyssey gamma-ray and neutron spectrometers also have revealed that high-latitude regions are enriched in hydrogen, which is consistent with a significant water ice component in the near surface (Boynton et al., 2002; Feldman et al., 2002; Mitrofanov et al., 2002). The modeled volume percent of ice in polar regions exceeds the expected pore volume of loose unbound dust, thereby suggesting that the ice-dust mixture was emplaced directly in solid form (i.e., as dusty snow or frost) rather than by the downward diffusion and deposition of ice into the pore spaces of a dry dust matrix (e.g., Boynton et al., 2003). The mid-latitude boundaries of these ice-rich terrains, however, are correlated with zones of predicted ice stability (e.g., Leighton and Murray, 1966; Farmer and Doms, 1979; Mellon and Jakosky, 1995). Hence, sublimation processes may have influenced the present extent of the mantling deposits by reshaping their margins.

4.2. Constraints on mantle emplacement age

The relative age of the VL2 region as determined solely by the density of large visible impact craters (Fig. 5a, N(5) equal to 1.2 × 10^2 craters >5 km per 10^6 km^2) is consistent with previous age estimates made using large craters. VL2 lies on member HvK, the knobby member of the Vastitas Borealis Formation (VBF), which has a measured N(5) crater density of ~0.9 to 1.4 × 10^2 (Greeley and Guest, 1987). More recent geologic mapping with MGS data has combined the 4 members of the VBF into 2 larger members, which results in a slightly younger overall relative age but is still consistent with the age determined in this study (Tanaka et al., 2003). The age of the sedimentary mantle over this region, however, is not necessarily the same age as the underlying surface—it is likely younger. Determining how much younger is the focus of this section.

Based on the lack of unmodified impact craters >100 m in diameter, the uppermost thin mantle layers (with thicknesses ranging between 1–10 m as indicated by shadow measurements) have been interpreted to be recently formed, i.e., in the Late Amazonian (Mustard et al., 2001; Head et al., 2003). The crater size–frequency distributions presented in this current study, however, cannot be explained solely through the recent deposition and removal of a 1–10 m thick packet of mantle layers. While the deficiency of craters <100 m in diameter supports the conclusion that modification processes (such as the deposition and removal of thin mantle layers) have been recently active, the lack of craters in the 100–500 m diameter range and the presence of large pedestal craters necessitates the erosion of a much thicker mantle layer (e.g., Schultz and Lutz, 1988).

There are at least two methods for determining the emplacement age of this mantling deposit. One method involves dividing up the total crater population on the basis of morphology. Implicit in this first method is that any age estimate obtained by a subset of the crater population will (by definition) represent a shorter period of time than the overall age inferred from the entire population. A second method is to use superposition to establish the relative ages of events that have occurred in this region. If the age of a specific feature or deposit that is stratigraphically younger than the mantle layer can be determined, this necessarily constrains the age of the mantle layer to be older than the superposing feature.

The number density of differentially eroded craters indicates the minimum net length of time that the surface was covered with a mantle layer of appreciable thickness in which these craters could form. No obvious correlations between pedestal height and crater diameter, pedestal diameter, or degradation state were observed in the study area. Approximately 80% of counted craters ≥2 km are inferred to be differentially eroded craters, corresponding to a N(2) crater density of about 4.8 × 10^2. This indicates that the surface around VL2 was mantled to some extent prior to at least the end of the Late Hesperian. Of the pedestal craters ≥2 km in diameter analyzed in this study, less than 25% appear to be full pedestals. The crater density determined with this subtype of crater alone is consistent with a period of time extending back to the Middle Amazonian.

These time constraints provide an important tool for understanding the geologic history of this mantle layer. Several stages of mantle development can be inferred, including the initiation of mantle deposition, the attainment of a maximum-recorded thickness of at least 102 m, and subsequent erosion into its present state. However, due to variable spin and orbital parameters (e.g., Toon et al., 1980), this region likely experienced episodic climatic fluctuations between conditions favoring deposition, when the mantle layer became more extensive than today, and conditions favoring erosion when mantle cover was reduced. Thus, the time intervals suggested by examining various divisions and subdivisions of the crater population represent net times and not necessarily continuous time intervals. Consequently, a second and perhaps more robust method for determining the emplacement age of the mantling deposit is to determine the age and stratigraphic position of Mie crater, which is a discrete event that establishes an important stratigraphic marker.

Several chains of prominent secondary craters extend predominantly to the east of Mie crater (suggesting that the impactor came from the west). Topographic profiles across
one of these chains (Fig. 8) conclusively demonstrate post-impact erosional modification. Close to Mie, the secondary chain is expressed as a series of linear, rimmed depressions that are elongated in the downrange direction. Farther downrange, the secondary chain is inverted and is expressed as a series of linear, crater-topped mounds. Similar inverted secondaries are associated with the crater Lyot as they cross into high-latitude mantling deposits (Schultz and Mustard, 2004). The fact that distal portions of the secondary chain have been topographically inverted by differential erosion is an unambiguous indication that a mantle layer existed prior to the Mie impact. The age of this impact therefore provides a constraint on the minimum age of the co-existing mantling deposit.

The number density of impacts that are superposing Mie ejecta is difficult to assess due to both the relatively small area and the hummocky topography of the inner ejecta facies. Nevertheless, the results indicate an \( N(5) \) crater density of roughly \( 1.0 \times 10^2 \), which is consistent with a Late Hesperian age (Fig. 5c). The results from these diverse age estimation methods all converge upon the same conclusion: the lowermost layers of the mantle deposit in the vicinity of VL2 were emplaced well before the Late Amazonian. The fact that the mantle deposit has been apparently unaffected by processes other than continued impact bombardment and eolian activity since the Late Hesperian places severe constraints upon the timing of other proposed processes in the northern plains (e.g., lacustrine, oceanic, or glacial processes).
4.3. Regional thermophysical properties

The question has been asked if the Viking landing sites are representative of the surface of Mars (Jakosky and Christensen, 1986). At Viking IRTM-scale of $2^\circ2^\circ$ latitude-longitude bins, VL2 (and VL1) fall outside a general anticorrelation trend between thermal inertia and broadband albedo. It was concluded on this basis that the surfaces observed from the Viking landers “... are distinctive from most of the rest of the planet” (Jakosky and Christensen, 1986, p. 125). However, on a histogram of TES bolometric albedo versus thermal inertia mapped into 0.05° by 0.05° latitude-longitude bins, the VL2 site is near the center of a newly resolved mode (Mellan et al., 2000, Fig. 5; Putzig et al., 2005, Fig. 4). This mode occupies roughly one-quarter of the surface area of Mars and represents surfaces with relatively high thermal inertia and intermediate albedo. Given this fact and coupled with the knowledge that VL2 does not lie on a lobe of continuous Mie ejecta, it would appear that the characteristics of VL2 site are indeed representative of at least one quarter of the martian surface (including a significant portion of the northern plains).

The thermal inertia of the region centered on VL2 region as measured by TES is 234 J m$^{-2}$ s$^{-1/2}$ K$^{-1}$ (Putzig et al., 2005). Although multiple lines of evidence corroborate the presence of a dusty mantling deposit at high latitudes, the thermal inertia of this high latitude region does not resemble other dusty regions such as Tharsis, Arabia, and Memnonia: it is significantly higher. Higher thermal inertias are consistent with a greater abundance of coarse particles, indurated materials, or both. The thermal signature of the VL2 site, and by extension much of the northern plains, are instead consistent with a partially stripped and deflated eolian mantle.

THEMIS-derived thermal inertia determined in this study for the VL2 region is about 190 J m$^{-2}$ s$^{-1/2}$ K$^{-1}$ (Fig. 6B), which is slightly lower than the TES-derived value but still much greater than the values for fully dust-covered regions. For the product of bulk density and specific heat of 1.0 $\times$ 10$^3$ J m$^{-3}$ K, this thermal inertia corresponds to an effective particle size of about 120 m (Presley and Christensen, 1997). In accounting for the differences between the TES and THEMIS results, both the higher inherent uncertainties of the THEMIS instrument and the differences in spatial scale between the two instruments may play a role. TES thermal measurements were averaged into 1/20° bins, and the mean value for a 6 by 6 pixel block was reported for each site (Putzig et al., 2005). At the latitude of VL2, this corresponds to a region roughly 212 km$^2$ in area. At this large spatial scale, the TES results indicate surface around VL2 appears fairly homogeneous. However, the surface exhibits considerable thermal heterogeneity at THEMIS resolutions (Fig. 6).

One of the more apparent thermal contrasts in THEMIS data is between impact craters and intracrater regions. Most crater rim areas exhibit warmer nighttime temperatures, consistent with coarser particles and/or more indurated surfaces. This thermal signature is consistent with direct observations of concentrations of rock fragments around the crater Bonneville and other crater-like forms at the MER Spirit site (Grant et al., 2004). The armored distal terrains of many pedestal craters, however, exhibit cooler nighttime temperatures, which are consistent with a surface layer of fine-grained debris and/or a lack of rocks. This is the opposite of what one might expect for pedestal craters: if pedestal surfaces are indurated or covered with coarse ejecta block, they should have relatively high thermal inertias. But instead they possess relatively low thermal inertias. Based on the evidence for repeated mantling episodes up to the present, this low thermal inertia signature could be a remnant dust layer left over from a previous episode (or episodes) of mantling. Though a complete treatment of the formation and evolution of pedestal craters is beyond the scope of this paper, these thermophysical observations provide important constraints for future modeling efforts (e.g., Wrobel et al., 2006).

Nighttime temperatures in the Fig. 6A subset of Canberra crater range from 190 K near the rim down to 170 K on the ejecta blanket, which correspond to an almost factor of two difference in calculated thermal inertias between these regions. One small-scale feature of interest is the rim area of the small pedestal crater superposed on Canberra, which has a relatively high thermal inertia. This thermal signature is consistent with coarser and/or more indurated particles, and yet this smaller pedestal formed entirely in the upper layer(s) of the sedimentary mantling deposit. Therefore, it could not have excavated coarser rocks from the more competent substrate. The high thermal inertia rim supports the suggestion that some rocks observed at the surface in deflated terrains may in fact be impact-derived melt breccias (Schultz and Mustard, 1998; Thomson and Schultz, 2003) or indurated relicts.

4.4. Origin of rocks at VL2

Identifying the origin and geologic provenance of isolated and displaced fragments of rock, or floar, is a more difficult task than for intact rock strata. At the MPF location in Chryse Planitia (and by extension VL1), many of the rocks were assumed to be flood-deposited debris (e.g., Golombek et al., 1997). This is unlikely to be the case for VL2, which is situated over 6000 km from the mouth of any outflow channel system. Rocks at this site are more likely impact-emplaced and some may even be impact-derived (i.e., impact melt breccias).

Based on the number and location of currently visible craters around VL2, the first-order amount of predicted ejecta is less than the observed rock abundance by a factor of 1.4 to 2.9. A component of discontinuous distal Mie ejecta may account for this difference. The crater count evidence for a significant population of missing craters, however, also suggests that small craters since removed have contributed to the rock population. The similarity of the small crater density at VL2 (Figs. 5a and 5b) compared to regions twice as far from Mie (Fig. 5d) supports the latter interpretation. If discontinuous ejecta from Mie crater dominated the VL2 site, one would expect a different erosional history compared to other nearby regions. As previously discussed, the depletion of sub-km craters indicates that the present number of visible small craters represents only a small fraction of the total population of such impacts that have occurred. Considering the evidence for deep eolian deflation
of all unarmored surfaces in this region, impact products from missing craters are likely to have contributed to the observed rock population.

During eolian deflation, coarser materials are concentrated at the surface as finer material is removed. Theoretical models of desert pavement formation suggest that the surface continues to rapidly accrete coarser material until the lag deposit exceeds an areal abundance of about 40% (e.g., Orndorff, 1998). The fact that <19% of the surface at VL2 is covered with rocks suggests that the surface is still immature in terms of lag development. Cementing agents such as duricrusts can decrease or arrest rate of removal of fine sediments, but the observed deflation of over 102 m of material suggests that whatever duricrusts were present in the past at this site were insufficient to seal the surface from eolian erosion. It is also possible that shifts in the net sediment budget over time have resulted in former rocky lag horizons in this region that have been buried by later mantling episodes.

Small-scale polygons also bear on the issue of erosion since their presence implies thermal contraction of ice-rich soil. The required rigidity of an ice-rich surface layer may seem incompatible with eolian deflation, but the formation timescale of polygons is short compared to the potential depositional and erosional timescales of ice-rich mantle layers. For example, thermal modeling (Mellon, 1997) suggests that repeated seasonal cooling drives polygon formation. Small cracks in terrestrial ice-rich soils have been observed to grow 1–10 mm or more in a single year (Mackay, 1999). While the total time required to form 10–15 cm wide polygons on Mars is not well constrained, a formation time of up to \( \sim 10^3 \) years might be a reasonable first-order estimate. Episodes of mantle deposition and erosion, if tied to obliquity or other orbital parameters (e.g., Mellon and Jakosky, 1995), may occur on a time frame \( \sim 10^5 \) years—
which is an order of magnitude larger than likely polygon formation times but is still geologically youthful. Thus given that depositional/erosional and polygon-forming processes operate on distinct time scales, both sets of processes may be accommodated on a given surface (though perhaps sequentially rather than simultaneously). The presence of thermal contraction cracks also suggests an additional potential mechanism for increasing the surface rock abundance: frost-heave driven stone uplift (e.g., Washburn, 1997), though such a process has yet to be fully modeled under martian conditions.

4.5. Net erosion rate

Given that the pedestal craters measured in this study may have occurred before or after the mantle deposit attained its maximum thickness, we use the pedestal crater with the greatest relief to constrain the erosion rate. Such an approach still represents a minimum thickness and therefore minimum erosion rate. The average net erosion rate, \( r \), is expressed by the thickness of removed material \( (h) \) over the duration of erosion \( (t) \). One of the difficulties unresolved by this simple relationship are that, although resurfacing has continued until the present, the erosion was unlikely to act continuously. Depending on whether the onset of erosion occurred in the Late Hesperian, Early Amazonian, or Middle Amazonian, the inferred net erosion rates range from \( 6 \times 10^{-8} \) to \( 10^{-7} \) m/yr [using the estimated absolute chronologies of Hartmann and Neukum (2001)]. The net erosion rate at the Viking 2 Lander site determined in this study (Fig. 9) is similar to a previously determined estimate for the region (Arvidson et al., 1979). This rate is 2 orders of magnitude larger than the inferred rate at the VL1 site and 3 orders of magnitude larger than that inferred at the Pathfinder site (Golombek et al., 2001).
The high erosion rates at VL2 are consistent with the inferred erosional history of other unconformable, volatile-rich sedimentary deposits (e.g., Schultz and Lutz, 1988; Grizzaffi and Schultz, 1989; Grant and Schultz, 1990). Similar erosion rates have also been estimated for the south polar layered terrain (Bleacher et al., 2003).

5. Summary

The VL2 landscape has undergone massive eolian deflation with at least 102 m of friable material largely stripped away as evidenced by erosional remnants such as inverted secondary and pedestal craters. The significant depletion of small craters (<100 m in diameter) suggests that the resurfacing process is currently active, and the loss of craters 100–500 m in diameter suggests erosion of a mantle layer at least 100 m thick. These results are consistent with previous observations of the presence of a high-latitude eolian mantling deposit that has been partially stripped by erosion. Extending the crater count down to small crater diameters was necessary to assess the recent resurfacing history of this mantle deposit. The emplacement of the mantle deposit, indicated by the \( N(2) \) crater density of differentially eroded craters, occurred prior to the end of the Late Hesperian period. This age constraint is confirmed by the measured crater density of a superposing feature, namely Mie crater ejecta, since the mantle layer had to be emplaced prior to the Mie impact to account for the observed topographic inversion of Mie crater rays.

Furthermore, the presence of nested or superposed pedestal craters such as Canberra, preservation of fine crater structure on armored surfaces, and general absence of fluvial scour features all imply that both the mantle deposition and removal mechanisms were dominantly eolian. The dominance of eolian processes in this region likely precludes any post-Late Hesperian oceanic, lacustrine, or glacial activity in the northern plains. However, the results of this study place no constraints upon a hypothesized Early Noachian ocean (Clifford and Parker, 2001), the deposits of which would be buried beneath the current surface. Thermal inertia values derived from THEMIS indicate that the VL2 landed in an intracrater area that is similar to the surrounding northern plains. The thermal signature of the smaller pedestal superposed on Canberra indicates surfaces with coarser and/or more indurated particles, including perhaps impact melt breccias, can be derived from impacts into fine-grained sedimentary targets. Such a process has been documented for impacts into terrestrial loess (Schultz et al., 2004, 2006).

The micro-structure observed with the Viking 2 Lander reveals a rock-rich landscape. Rock abundances at VL2 are consistent with the addition of impact-emplaced material from a population of craters that have been eroded, and some rocks are possibly impact-derived. These observations are consistent with macro-scale processes inferred from orbit, namely the evidence of deep deflation from relict landforms and elevated thermal inertia values indicating an accumulation of coarse and/or indurated material. The present surface is an erosional lag deposit, and this explains the relatively high thermal inertia values observed by TES and THEMIS despite the presence of an active dust mantle in this region.

The VL2 site likely typifies martian landscapes that are dominated by partially stripped eolian mantle deposits. In particular, the VL2 site may be an appropriate analog for the upcoming Mars Phoenix mission (Smith, 2006) that will land at a high northern latitude site (65°–72° N). Given that many of the processes that have shaped the VL2 site also operate at higher latitudes, a similar landscape might be expected at the Phoenix site.

References


