Interior channels in Martian valleys: Constraints on fluvial erosion by measurements of the Mars Express High Resolution Stereo Camera

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[1] In High Resolution Stereo Camera (HRSC) images of the Mars Express Mission a 130 km long interior channel is identified within a 400 km long valley network system located in the Lybia Montes. Ages of the valley floor and the surroundings as derived from crater counts define a period of ~350 Myrs during which the valley might have been formed. Based on HRSC stereo measurements the discharge of the interior channel is estimated at ~4800 m^3/s, corresponding to a runoff production rate of ~1 cm/day. Mass balances indicate erosion rates of a few cm/year implying the erosion activity in the valley to a few thousand years for continuous flow, or one or more orders of magnitude longer time spans for more intermittent flows. Therefore, during the Hesperian, relatively brief but recurring episodes of erosion intervals are more likely than sustained flow. **Citation:** Jaumann, R., et al. (2005), Interior channels in Martian valleys: Constraints on fluvial erosion by measurements of the Mars Express High Resolution Stereo Camera, *Geophys. Res. Lett.*, *32*, L16203, doi:10.1029/2005GL023415.

1. Introduction

[2] Martian valley networks [*Pieri*, 1980b] or "runoff channels" [*Sharp and Malin*, 1975] have been cited as the best evidence that Mars maintained flow of liquid water across the surface. A valley cut by a river usually shows internal structures such as interior channels, terraces and benches. Although those features are extremely rare in Martian valleys [*Malin and Edgett*, 2000; *Carr and Malin*, 2000], short segments of interior channels have been identified in Mars Orbiter Camera (MOC) images [*Malin and Edgett*, 2000; *Irwin et al.*, 2004] and Mars Odyssey Thermal Emission Imaging System (THEMIS) images [*Mangold et al.*, 2004; *Irwin et al.*, 2005]. Interior channels indicate surface flow and constrain the amount of water (discharge) involved in surface runoff. As valley networks are thought to be formed by runoff in immature drainage basins, as well as by paleolake overflows and possible subsurface outflows, the timescales of the erosion are important to constrain the valley formation process [*Moore et al.*, 2003].

[3] A 130 km long and about 500 m wide interior channel has been discovered in Mars Express high-resolution imagery. The Mars Express High Resolution Stereo Camera (HRSC) is a multiple line scanner capable of resolving, at 250 km periapsis height, features as small as 10 m/pixel in stereo and color and by simultaneously covering large surface areas [*Neukum et al.*, 2004].

2. Observations and Geologic Settings

[4] The observed valley network and the interior channel are located between 1.4°N to 3.5°N and 81.6°E to 82.5°E in the western part of Libya Montes (Figure 1). The Libya Montes are thought to be eroded remnants of the Isidis basin rim that consist of Noachian highland materials [*Greeley and Guest*, 1987]. The oldest unit (Nm) forms rugged mountainous surfaces of massif materials, which are partly remnants of former impact crater rims, heavily modified and locally dissected by drainage patterns [*Crumpler and Tanaka*, 2003]. Within the Noachian terrain, younger Hesperian plain units (Hd_1) are exposed. These plains are characterized by valley systems and their poorly dissected interFluves [*Crumpler and Tanaka*, 2003]. Within Hd_1 a 400 km long valley network extends from south to north and includes major western and eastern branches that converge downstream. The valley floor is covered by smooth material (Hd_2). Within these smooth materials an interior channel (Figure 2) winds northward, showing a sinuous pattern. The interior channel incised the unit Hd_2 and developed terraces. Further downstream the channel cuts into the floor of a narrow valley and can be identified as an interior channel. Although at some places the valley rims are obscured by subsequent modifications due to mass wasting, the interior channel can be traced for about 130 km. Further north the interior channel crosscuts the ejecta of an impact crater of about 3 km diameter. The channel winds very closely around the crater rim and is not buried indicating its younger age compared to the crater.

2.1. Morphometry

[5] The HRSC high resolution stereo data allow the determination of the depth and width of the interior channel. Height measurements were carried out based on a high resolution DTM (horizontal resolution 50 m) in three different reaches (upstream, midstream and downstream) of the interior channel (Figures 2 and 3). DTM heights are stored with a resolution of 1 m. The expected mean point accuracy from analysis of a number of high resolution

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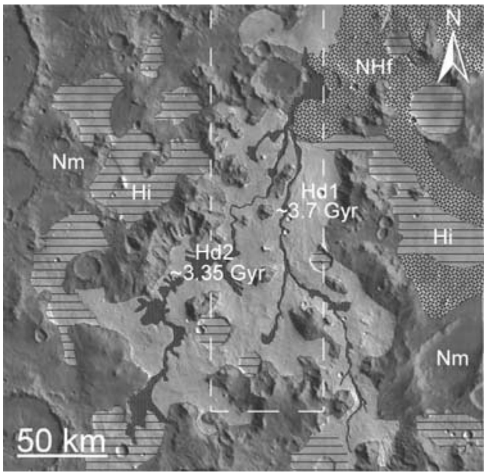

Figure 1. Geological map of the valley network in the Lybia Montes based on HRSC and MDIM2. The units (Nm, NHf, Hd_1, Hd_2, Hi) represent: Noachian massif material, Noachian slope material, Hesperian dissected plains, Hesperian intermontane plains [*Crumpler and Tanaka*, 2003]. The dashed area displays the position of HRSC-orbit 47 in Figure 2.

HRSC DTMs is about 90% of the nadir image resolution [*Gwinner et al.*, 2005]. Thus we assume a mean height accuracy of 10 m in the case of orbit 47. In order to increase the accuracy of the depth measurements, we combine cross sections and point measurements along the floor and the terraces of the interior channel (Figure 3 and Table 1). The width was measured using both the map projected ortho-images and cross-sectional profiles. The measured channel depths are minimum values because mass wasting and aeolian processes may have filled the channel after formation and for sampling reasons. The depth of the channel (Figure 4) is in the upper reach ~27 ± 10 m, in the middle reach ~35 ± 10 m and in the lower reach ~55 ± 10 m, indicating an increasing incision of the channel bed downstream. The interior channel widths in all areas are ~450 ± 50 m, independently of whether they are measured in the image data or in the profiles. The interior channel lays

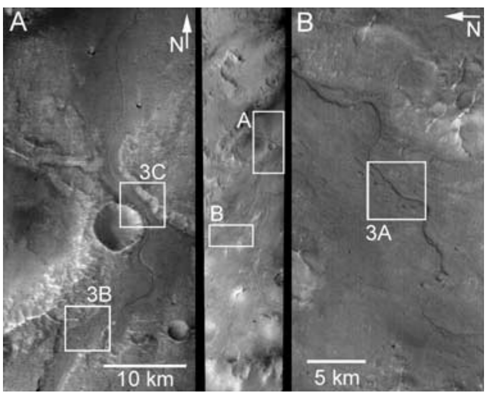

Figure 2. Part of HRSC-orbit 47 (middle) with A (downstream subframe) and B (upstream subframe). Frames of 3A-C show the locations in Figure 3.

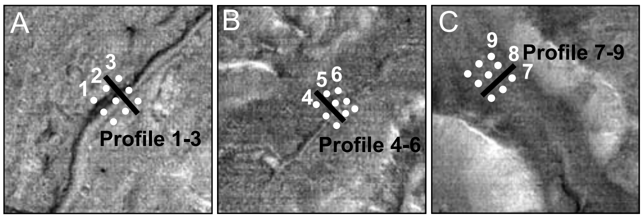

Figure 3. Position of the measurements in the HRSC image data (12.5 m/pixel) subframes of the upper, middle and lower part of the interior channel regions (see Figure 2). All images are 5 km wide and north is at top. The numbers show the position of cross section measurements (see Table 1).

within a 280 ± 10 m deep and 3500 ± 50 m wide valley (Figure 5).

2.2. Discharge

[6] The derived morphometric parameters were used to calculate the hydraulic parameters of the interior channel. From these parameters we estimated maximum discharges at different locations as shown in Figure 3 by using a modified Manning's equation for steady, uniform flow taking into account the lower gravitational acceleration on Mars [*Manning*, 1891; *Komar*, 1979; *Wilson et al.*, 2004]:

$$Q = AU = A\left(SR^{4/3}\right)^{1/2}/n_{mars}$$

where A is the flow cross-sectional area, U is the flow velocity, S is the local slope, R is the hydraulic radius and n_{mars} is the Manning roughness coefficient for Mars. For the slope S we used the measured gradient of 0.01 of the interior channel as derived from the topographic data. The Manning roughness coefficient for Mars is constrained by a dimensionless constant K, the bed roughness scale r and the acceleration due to gravity. With $n_{mars} = r^{1/6}\, g^{-1/2}\, K^{-1}$ and using $K = 6.01$, $g = 3.74$ m/s^2 for Mars and $r = 0.064$ m *Wilson et al.* [2004] derived an optimum value for n_{mars} equal to 0.0545 s m$^{-1/3}$. The peak rates for a bankfull discharge of the interior channel range from (22 ± 14) × 10^4 m^3/s upstream to (68 ± 27) × 10^4 m^3/s downstream. The channel could not have contained more water than at bankfull and so this is an limiting condition for the total amount of water running through the valley system.

Table 1. Measured Morphometric Parameters

Cross Section[a]	Height 1[b] m	Height 2[b] m	Height 3[b] m	Depth m	Width m
1	−583	−616	−597	26 ± 10	425 ± 50
2	−586	−610	−592	21 ± 10	425 ± 50
3	−579	−617	−589	33 ± 10	425 ± 50
P 1–3	−592	−619	−589	29 ± 10	440 ± 50
4	−1510	−1555	−1538	31 ± 10	440 ± 50
5	−1527	−1575	−1537	43 ± 10	440 ± 50
6	−1531	−1562	−1538	28 ± 10	440 ± 50
P 4–6	−1542	−1575	−1541	34 ± 10	615 ± 50
7	−1789	−1837	−1785	50 ± 10	420 ± 50
8	−1831	−1890	−1722	64 ± 10	420 ± 50
9	−1796	−1852	−1791	59 ± 10	420 ± 50
P 7–9	−1844	−1890	−1832	52 ± 10	450 ± 50

[a]Cross section numberings correspond to point and profile (P) measurements shown in Figures 3 and 4.
[b]Height 1, 2 and 3 correspond to measurements of the left terraces, the interior channel floors and the right terraces, respectively.

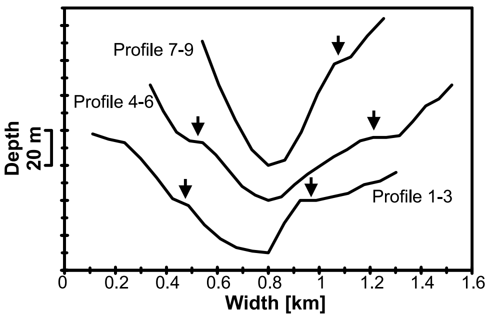

Figure 4. Cross section profiles of the interior channels (see Figure 3). Arrows show the terraces of the interior channel.

However, the low width to depth ratio, the entrenched characteristics of the channel, and the lack of overflow features as well as the immature development of the total network suggest that the dominant discharge never filled the interior channel to the top of its banks. The width to depth ratio is 14. This is about an order of magnitude less than what we would expect for bankfull discharge in terrestrial alluvial rivers, and Martian channels were assumed to have relatively high width to depth ratios in the conservative discharge estimates of *Irwin et al.* [2005]. As the interior channel is very old and still preserved over a long distance, we also have to assume resistant banks. By applying the width to depth ratio to an entrenched and confined channel, which might never have been filled to the top of its banks, the active flow depth of the dominant discharge might have been reduced by one order of magnitude to no more than 1/10 of bankfull. This translates to a flow depth of about 3 m. 10% bankfull yields a discharge of $\sim$4800 m^3/s. This is comparable within a factor 2 with other discharge estimates for interior channels of the same width [*Irwin et al.*, 2004, 2005]. Unknown factors such as varying cohesion of channel banks or width and depth variabilities due to postfluvial modifications, can explain discharge discrepancies of this order between different interior channels. A discharge of $\sim$4800 m^3/s would fit with a production rate ($P = Q/A$) of $\sim$1cm/day based on the measured drainage area (A) of 43000 km^2, which is consistent with estimates for other Martian valley systems [*Irwin et al.*, 2004, 2005].

2.3. Ages and Chronology

[7] The age of the valley system was determined by crater counts on the valley floor containing the interior channel and the surrounding terrains, in which the valley eroded. The crater retention ages for the surrounding area (Hd_1) are $N(1\ km) = 5.6 \times 10^{-3}$ and for the valley floor (Hd_2) are $N(1\ km) = 2 \times 10^{-3}$. To derive the absolute model ages from the crater frequency distributions, we utilized the Martian impact cratering model of *Hartmann and Neukum* [2001] and polynomial coefficients of *Ivanov* [2001]. The valley was formed in Hesperian times with absolute model ages for the valley floor including the channel of $\sim$3.35 ± 0.05 Gyr and for the surrounding terrains of $\sim$3.7 ± 0.05 Gyr (Figure 6). Therefore, the maximum valley formation time amounts to $\sim$350 ± 100 Myr.

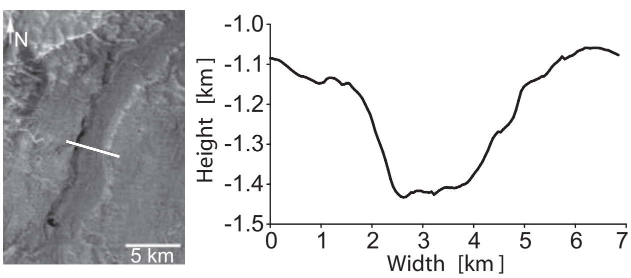

Figure 5. Cross section profile of the northern valley segment.

3. Discussion

[8] The erosion of unit Hd_1 down to Hd_2 is most prominently exposed in the northern part of the valley system, where Hd_1 is cut by a narrow and deep segment of the valley (Figure 5). Taking into account the $\sim$350 ± 100 Myr time difference between Hd_1 and Hd_2 and the valley depth of 280 ± 10 m, the average erosion rate for this valley segment is 0.8 ± 0.3 μm/a, which is very low. However, it is unlikely that such low erosion rates would result from fluvial processes in a long-lived Earthlike climate and the valley might have been eroded during a shorter period. In order to investigate this question we need to know how much material was eroded and transported in the valley. The erosion rates of the valley can be constrained by using the erosion power of the interior channel. Within a distance of about 100 km (Figures 3a–3c) the depth of the interior channel increases by a factor of 2 while the width remains almost constant. However, is a 10% bankfull discharge of the interior channel sufficient to erode and transport enough material to explain the dimensions of the valley? The frictional shear velocity $u = (\tau/\rho)^{1/2}$ of the interior channel where τ is the stress between the flowing water and the bottom and ρ is the water density constrains the flow conditions: For near uniform flow τ can be determined from the flow depth h and the channel slope S through the relationship $\tau = \rho g h S$ where g is the acceleration of gravity (3.74 m/s^2 for Mars). For a 10% bankfull discharge the frictional shear velocity is 33.5 cm/s. According to the relationship between frictional shear velocity and transport load on Mars as developed by *Komar* [1980] the threshold grain diameter in the interior channel is $\sim$40 cm, the cutoff diameter between bed load and suspension is

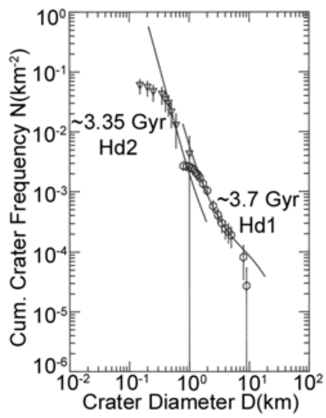

Figure 6. Crater count results of the valley floor (Hd_2) (triangles) and surrounding areas (Hd_1) (circles).

~8 mm and the grain diameter at onset of the wash load is ~0.3 mm. Thus according to the frictional shear velocity, a flow depth of ~3 m in the interior channel is capable of eroding and transporting a significant amount of sediment.

[9] The area of the valley network (unit Hd_2) is 1850 km^2 and the average depth of the valley is assumed to be 250 m. Thus, the maximum amount of excavated material is 462.5 km^3. With the density of basalt of ~2.9 g/cm^3 this translates to 1.34×10^{12} tons of material. In order to constrain the time needed to erode the valley network, the sediment discharge of the interior channel has to be known. Typical transport loads in terrestrial rivers are in the order of 0.03 kg/m^3 to a few kg/m^3 but may reach in semiarid areas extremely high concentrations of about 40% sediment by weight [*Leopold et al.*, 1964; *Beverage and Culbertson*, 1964; *Nordin and Beverage*, 1965]. Such high concentrations are mostly dependent on sediment availability, which might have been substantial under Martian semiarid conditions. Martian rivers would transport sediment more efficiently than terrestrial rivers per unit discharge owing to the low settling velocity and low critical velocity for suspension and could reach sediment concentrations as high as 60–70% by weight [*Pieri*, 1980a; *Komar*, 1980]. Such concentrations would also contribute to the flow power, enhancing the channel's ability to erode [*Bagnold*, 1962; *Komar*, 1980]. Based on these considerations a sediment load of about 5 kg/m^3 seems to be reasonable. Although this sediment concentration is a rough estimate, it can be used to constrain the duration of the valley formation based on an order-of-magnitude calculation. Together with a discharge of ~4800 m^3/s, sediment loads of $\sim 2 \times 10^6$ tons were transported per day. The erosion of the valley would then take 1800 years for continuous flow, or one or more orders of magnitude longer time spans for more intermittent flows, resulting in erosion rates of a few cm/year. Owing to this result, sustained flow over millions of years is unlikely. The valley formation might have occurred either during a period of less than $\sim 10^5$ years in the late Hesperian or during a number of repeated flow events, which ended 3.35 billion years ago.

4. Conclusions

[10] The observation of an interior channel in HRSC stereo data demonstrates a fluvial environment and channelized flow within the corresponding valley. The valley formed during the Hesperian Period. Fluvial erosion rates, discharges and sediment transport indicate a relatively short valley formation time rather than sustained flow during the Hesperian. The valley might have formed during a period of wet climate that lasted over about 10^5 years during the Late Hesperian, or it may have been cut by multiple short-term flooding events within a longer time span. Episodic floods might have been triggered by short periods of intense precipitation or by sudden release of large groundwater volumes. However, we cannot exclude changes of alternating dry and wet climate causing intervals of sustained flow within the valley system for a few thousands years each during Hesperian time.

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