The volume of water required to carve the martian valley networks: Improved constraints using updated methods

Eliott N. Rosenberg\textsuperscript{a,b}, Ashley M. Palumbo\textsuperscript{a}, James P. Cassanelli\textsuperscript{a}, James W. Head\textsuperscript{b,x}, David K. Weiss\textsuperscript{a}

\textsuperscript{a} Department of Earth, Environmental and Planetary Sciences, Brown University, Providence, RI 02912, USA
\textsuperscript{b} Department of Physics, Cornell University, Ithaca, NY 14850, USA

\textbf{ARTICLE INFO}

\textbf{Keywords:} Martian climate
Martian valley networks
Fluvial erosion
Noachian
Martian geomorphology

\textbf{ABSTRACT}

The martian valley networks are a key piece of evidence for the presence of liquid water on early Mars, and understanding their formation conditions can provide valuable insight into the nature of the early climate. Previous studies have used various methods to estimate the volume of water required to carve the valley networks, with results ranging from 3–5000 m Global Equivalent Layer (GEL). In comparison, other workers have found that the surface/near-surface water inventory was likely to have been \(\sim\) 24 m GEL at the Noachian-Hesperian boundary. Thus, 3 m GEL may be consistent with recycling in a cold and icy Late Noachian-Early Hesperian climate, while 5000 m GEL may require continuous warm and wet conditions. In this study, we use updated methods and datasets to better constrain the necessary volume of water, finding a conservative lower limit of 640 m GEL. Based on valley network formation timescales, we find that our results do not preclude a cold and icy Late Noachian-Early Hesperian climate. Thus, this updated estimate of the minimum volume of water required to carve the valley networks is consistent with both formation in a cold and icy and warm and wet climate.

1. Introduction

Late Noachian surfaces on Mars contain widespread and abundant valley networks (VNs) (Fassett and Head, 2008; Carr, 1995; Hynek et al., 2010), fluvial features that were formed by flowing liquid water. These VNs demonstrate that the climate of Mars was different in the Late Noachian than it is today, but the nature of the early climate of Mars is a matter of dispute. The VNs are often cited as evidence that the climate of Mars was once warm and Earth-like, with mean annual temperature (MAT) \(\geq\) 273 K, regular rainfall, and a vertically integrated hydrological system (Luo et al., 2017; Craddock and Howard, 2002). However, recent 3-dimensional climate models predict a cold and icy early Mars (Wordsworth et al., 2013, 2015; Forget et al., 2013), in which water is preferentially deposited in the highlands as snow and ice and MAT is \(\sim\) 225 K, well below the melting point of water (Wordsworth et al., 2015; Head and Marchant, 2014). In this model, punctuated events, such as volcanic eruptions (e.g., Halevy and Head, 2014), impact events (e.g., Segura et al., 2008; Palumbo and Head, 2017), melting during the warmest hours of the summer season (e.g., Palumbo et al., 2018a), or the introduction of greenhouse gases into a transient reducing atmosphere (e.g., Wordsworth et al., 2017), have been proposed as mechanisms to melt the surface ice (Fastook and Head, 2015), leading to surface runoff and formation of the VNs.

In this work, we test the plausibility of the proposed cold and icy early Mars climate scenario by estimating the volume of water required to carve the VNs and determining whether the volume of water could be accounted for by transient/punctuated ice melting and fluvial activity, as proposed to be the formation mechanism of these features in a cold and icy climate, or whether the estimated volume of water requires a sustained vertically integrated hydrological cycle, continuous warm and wet/arid climate, and (at least) seasonal rainfall (e.g., Craddock and Howard, 2002; Ramirez, 2017; Luo et al., 2017; Ramirez and Craddock, 2018).

Rosenberg and Head (2015) produced a preliminary estimate of the water volume required to carve the VNs. First, Rosenberg and Head (2015) utilized the valley network distribution data compiled by Williams and Phillips (2001) (derived from MOLA data, horizontal resolution 460 m/pixel) and the volumes of eight large VNs (Hoke et al., 2011) to estimate the volume of sediment removed to carve the VNs. Next, they used an empirically derived fluid:sediment flux ratio and the total volume of sediment removed from the VNs to estimate the minimum volume of water required to carve the VNs. They estimated...
the required volume of water to be a global equivalent layer (GEL) of 3–100 m. This volume of water does not preclude formation through transient fluvial activity in a cold and icy climate scenario. In a more recent analysis, Luo et al. (2017) implemented methods based on those of Rosenberg and Head (2015) and performed a global evaluation of the volume of sediment removed from all martian VNs in order to determine the volume of water required to carve all of the VNs. After applying several fluid:sediment volume ratios, to convert the sediment volume to a water volume, Luo et al. (2017) reported that an ∼5000 m GEL volume of water was required to have carved the VNs. Because this volume of water could not reasonably have been introduced to the surface through transient fluvial activity in a predominantly cold and icy climate when considering reasonable water inventory constraints (Carr and Head, 2015), Luo et al. (2017) claim that this updated water volume estimate is evidence for a warm and wet early Mars and that this volume of water also requires the presence of a Noachian ocean in the northern lowlands.

Here, we revisit the estimates of Rosenberg and Head (2015) and Luo et al. (2017), but implement both a more inclusive estimate of the VN cavity volume and an improved estimate of the fluid:sediment flux ratio based on a terrestrial dataset that is more applicable to the martian VNs. By taking these steps, we provide improved estimates of the minimum amount of water required to carve the VNs.

We use the updated VN cavity volume from Luo et al. (2017) which employed a progressive black top hat (PBTH) transformation method to calculate the cavity volume of the VNs based on an earlier automated VN mapping (Luo and Stepinski, 2009) and on the VN mapping by Hynek et al. (2010). This new cavity volume includes all VNs, not just the eight measured by Hoke et al. (2011), and thus represents an improvement over the cavity volume used by Rosenberg and Head (2015). Additionally, we use an empirically derived distribution of fluid:sediment flux ratios (median 3900; displayed in Fig. 3), rigorous error analysis, and we correct errors in previous analyses. For comparison, Luo et al. (2017) used a fixed fluid:sediment flux ratio of ∼4000.

We then apply our updated methods to the VNs studied by Hoke et al. (2011) to estimate the volume of water required to form each of the eight VNs included in that study as well as provide updated timescales of formation. Next, we apply our updated water volume estimates and test whether the VNs could have formed in a cold and icy climate through several mechanisms, and assess in more detail a specific formation mechanism: seasonal melting during the warmest hours of the summer season (e.g., Palumbo et al., 2018a). We conclude with a discussion of the implications of our results for the nature and evolution of the early martian climate.

2. Methods and discussion

2.1. Sediment volume

The first step in estimating the volume of water that was required to erode the VNs is to estimate the volume of sediment that was removed from the VN cavity. The volume of sediment removed is related to the total present-day cavity volume (ignoring subsequent infilling) by

\[ V_i = V_f (1 - \lambda) \]

Let \( V_i \) denote the cavity volume, \( V_f \) denote the sediment volume, and \( \lambda = \frac{V_e}{V_i} \) denote the fraction of the total cavity volume that was pore space prior to erosion, i.e., the porosity. It follows that the sediment volume is related to the cavity volume by

\[ V_i = V_f (1 - \lambda) \]

The porosity of martian regolith is estimated to range between ∼0.2 – 0.4 (Kleinhans, 2005; Clifford, 1993). It is important to note that Luo et al. (2017) erroneously calculated sediment volume as

\[ V_i = V_f (1 - \lambda) \]

which does not represent the desired physical quantity and overestimates the sediment volume by a factor of \((1 - \lambda)^{-1} \), which is equivalent to overestimation by a factor of 1.6 – 2.8 for our preferred range of \( \lambda \).

As previously mentioned, in our implementation of Eq. (2) we use the updated VN cavity volume from Luo et al. (2017). Using MOLA data and automated PBTH methods, Luo et al. (2017) measure the total cavity volume of the VNs to be \((1.74 \pm 0.8) \times 10^{14} \text{ m}^3\) based on the VN mapping of Luo and Stepinski (2009) (referred to here as “automated mapping”), or \((2.23 \pm 1.0) \times 10^{14} \text{ m}^3\) based on the combined VN mappings of Luo and Stepinski (2009) and Hynek et al. (2010) (referred to here as “combined mapping”). We perform calculations for both cavity volumes, as shown in Table 1, but use the automated mapping to state the main results of the paper in order to draw a more conservative, or lower limit, conclusion.

In the work of Luo et al. (2017), the authors scale the estimated cavity volumes to account for additional cavity volume that is typically observed in the higher-resolution HRSC DEMs, but is below the resolution of the data used in the automated mapping and combined mapping efforts. However, Luo et al. (2017) incorrectly scale an observed cavity volume of zero to a scaled cavity volume of \(5 \times 10^{13} \text{ m}^3\) (this is the y-intercept of the regression line in their Fig. 3). Thus, if the scaling were consistently applied to individual VNs, it would artificially inflate the cavity volume; there are many locations on the martian surface where VNs are not observed even in the highest resolution data. For this reason and to provide a lower bound on the VN cavity volume, we do not adopt the methods of Luo et al. (2017) to scale our results and directly implement the cavity volume estimates from the automated mapping and combined mapping efforts. If included, the scaling used by Luo et al. (2017) would increase our estimates by a factor of ∼1.3, the slope of the regression line in Fig. 3d of Luo et al. (2017) and the factor by which their volumes increased after scaling.

Substituting the VN cavity volume from Luo et al. (2017) into Eq. (2), the sediment volume derived from the automated mapping is

\[ V_i = \left(1.13 + 0.90 \frac{0.57}{-0.57}\right) \times 10^{14} \text{ m}^3, \]

where we have used a preferred porosity of 0.35, following Luo et al. (2017), but have also considered other values that span the range 0.2–0.4. The sediment volume resulting from the combined mapping is, by the same logic,

\[ V_i = \left(1.45 + 1.13 \frac{1.13}{-0.71}\right) \times 10^{14} \text{ m}^3. \]

To increase the accuracy of our estimate, we perform a more careful error analysis in Section 2.3, treating the cavity volume and the porosity as random variables and convolving them in order to obtain a 95% confidence interval.
2.2. Fluid:sediment flux ratio

The sediment volume that we estimated in Section 2.1 is directly proportional to the volume of water required to erode the VNs. Following Rosenberg and Head (2015), we estimate the fluid volume as

\[ V_f = \frac{Q_f}{Q_s} \]

where \( Q_f \) is the volumetric fluid flux and \( Q_s \) is the volumetric sediment flux. \( Q_s \) is the flux of bed-material load, i.e., the flux of sediment with grain sizes found in appreciable quantity on the VN bed (Brownlie, 1981). This approximation of \( Q_s \) excludes wash load, which consists of finer grains and is generally limited by supply rather than by the dynamics of the flow (e.g., Garcia, 2008).

To solve for \( V_f \) using Eq. (3), we first write an expression for the fluid:sediment ratio, \( Q_f/Q_s \), in order to understand its parametric dependence. Then, once we establish its parametric dependence, we will test our understanding against empirical data and use these data for our error analysis. To do this, we follow the derivations from Garcia (2008), which we reproduce here for completeness. The flow velocity, \( u \), at a height \( z \) above the bed is given by Eq. (2-4) in Garcia (2008):

\[ u = \frac{1}{\kappa} \ln \left( \frac{z}{z_0} \right) \]

where \( \kappa \approx 0.41 \) is von Karman’s constant, \( z_0 \) is the bed roughness length (the distance above the bed at which the flow velocity is zero), and \( u_* = \sqrt{gHS} \) is the shear velocity, where \( H \) is the flow depth, \( g \) is the acceleration due to gravity, and \( S \) is the bed slope. Integrating the flow velocity over the height \( z \), we find that the volumetric fluid flux per unit width is

\[ q_f = uH \left( \ln \left( \frac{H}{z_0} \right) + 1 \right) \]

Equipped with an equation for fluid flux, we must next consider the sediment flux. This can be decomposed into the bed load sediment flux and the suspended sediment flux. First, we consider the suspended flux. In terms of the time-averaged volumetric sediment concentration, \( c(z) \), at a height \( z \), the suspended flux per unit width, \( q_{s, sus} \), is
\[ q_{\text{uw}} = \int_{0}^{H} u(z) \sigma(z) dz \]  

Integrating Eq. (6) leads to Eq. (2-219) in García (2008):

\[ q_{\text{uw}} = \frac{1}{k} c_{i} u_{w} H \left( \frac{1}{H} \left( 30 \ln \left( \frac{H}{\nu} \right) \right) \right) + j \]  

where \( J_{1} \) and \( J_{2} \) are integration constants described by García (2008), \( k_{c} \) is the composite roughness, and \( c_{i} \) is the time averaged concentration near the bed. García (2008) lists various relations that attempt to predict the near-bed concentration in their Table 2-6. Some of these relations give the near-bed concentration as a function of the Shields parameter, \( \tau^{*} \), where \( R \) is the submerged specific gravity, \( H \) is the flow depth, \( D \) is the minimum grain size, and \( S \) is the bed slope.

Next, we consider the bed load flux, the second part of the sediment flux. Again, García (2008) reviews the necessary derivations for estimating bed load flux (see their Section 2.6.4 for more detail), concluding that bed load flux can be represented as

\[ q_{b} = 23.2 D \sigma_{u_{w}} \approx 23.2 \frac{D}{H} q_{\text{uw}} \]  

where \( D \) is the representative particle diameter and the other variables are defined the same as in Eq. (7).

Eq. (8) implies that the bed load flux is suppressed in comparison to the suspended flux by a factor of the grain size divided by the flow depth. Therefore, because the ratio of grain size over flow depth is very small, we can assume that the bed load flux is negligible. Thus, we can approximate sediment flux as equal to \( q_{\text{uw}} \), which is defined in Eq. (7). It follows that the fluid:sediment ratio can then be written as

\[ \frac{q_{s}}{q_{b}} \approx \frac{\ln \left( \frac{u}{\nu} \right) + 1}{c_{i} \left( \frac{u}{\nu} \right) + j_{1}} \approx \frac{1}{G} \]  

where, in the last line, we have noted that, for \( H \gg z_{0} \), the ratio of the logarithmic factors approaches a constant. Recall that units of both \( q_{s} \) and \( q_{b} \) are flux per unit width.

To compare this analytical solution with empirical data, we now compare the values of \( c_{i} \) predicted by the various relations summarized in Table 2-6 of García (2008) to the measured fluid:sediment ratios in the Brownlie (1981) dataset (Fig. 1). Specifically, in Fig. 1, we find that the data from Brownlie (1981) fits well with the relation given by van Rijn (1984) (as cited in Table 2-6 of García 2008). The offset is likely to be due to the suppressed constant of proportionality in Eq. (9).

Next, we explore in detail the equation from van Rijn (1984) because it provides a good fit for the observational data (Fig. 1). For cases in which the bedform height is not known, the van Rijn (1984) equation becomes

\[ c_{i} = 1.57 \tau^{1/2} \left( \frac{D}{H} \right)^{3/2} \left( \frac{v^{2}}{D} \right)^{1/10} \]  

where \( T = \tau^{*} \) is called the transport stage parameter, \( R \) is the submerged specific gravity and \( v \) is the kinematic viscosity. \( \tau^{*} \) is the critical value of \( \tau^{*} \) for the initiation of sediment transport. To calculate \( \tau^{*} \) we use Eq. (2-59a) of García (2008). Note that \( T = \alpha - 1 \), where \( \alpha \) is the flow strength defined by Rosenberg and Head (2015).

The values for many parameters needed to evaluate Eq. (10) are known. The volume-weighted average of the slopes reported by Hoke et al. (2011) is 0.002. If, following Luo et al. (2017), we assume that the sediment density is 2900 kg/m³, then the submerged specific gravity is \( R = 1.9 \). The acceleration due to gravity on the surface of Mars is \( g = 3.711 \) m/s² and the kinematic viscosity of water at 0 °C is \( \nu = 1.79 \times 10^{-6} \) m²/s.

The remaining unknown parameters are the flow depth, \( H \), and the representative grain size, \( D \). These parameters are interdependent. A deeper flow (larger \( H \)) will pick up larger grains of sediment and thus leave a larger representative grain size in the bed (larger \( D \)). The condition for the initiation of suspension of sediment grains of a particular size is that their fall velocity equal the shear velocity. The fall velocity can be computed numerically using Eq. (2-46) of García (2008). By numerically solving the condition that the fall velocity equal the shear velocity, we can arrive at an estimate for minimum possible median grain size for non-suspended grains (García, 2008). Although the actual median grain size in the bed may be significantly larger than this minimum and, as such, using the minimum grain size may not be a very realistic assumption, approximating it as the minimum provides the best possible estimate without prior knowledge of the grain size distribution. Additionally, implementing the minimum grain size ensures that our results provide a lower bound on the total volume of water required to carve the VNs, which is consistent with our final result being a conservative estimate. Using this reasoning, Rosenberg and Head (2015) estimated the grain size and flow depth as \( (H, D) \approx (1 \) m, \( 1 \) mm) \( - (16 \) m, \( 6.2 \) mm).

Finally, substituting these parameters into Eq. (10) leads to \( c_{i} \in (0.0694, 0.125) \). As shown graphically in Fig. 3, we focus our attention on the data points from Brownlie (1981), which, when substituted into Eq. (10), give values of \( c_{i} \) in this preferred range. For the observational data points satisfying this condition, the 25th and 75th percentiles for the fluid:sediment flux ratio are 1.5 \( \times 10^{3} \) and 4.2 \( \times 10^{3} \), respectively. For the laboratory data points, the same percentiles are 2.0 \( \times 10^{3} \) and 5.4 \( \times 10^{3} \) respectively. The 5th percentile fluid:sediment flux ratios are 589 for the observational data and 941 for the experimental data.

### 2.3. Valley network water volume estimates

Armed with updated estimated volumes of removed sediment (Section 2.1) and fluid:sediment flux ratios (Section 2.2), our goal is to place a conservative lower bound on the volume of water that flowed through the VNs. We are not currently in a position to place an upper bound on the water volume because we assume the smallest possible bed sediment grain size in our analysis. Further, large cumulative volumes of water could have trickled through the VNs weakly enough not to pick up very much sediment; we can only estimate the amount of water responsible for moving sediment and eroding the VN cavity.

For our first attempt at such a lower bound, we treat the porosity, grain size, flow depth, slope, and sediment density as fixed assumptions, which may be further constrained in the future as more observational and in situ measurements are taken. We treat the cavity volumes reported in Luo et al. (2017) as normally distributed with a standard deviation equal to half the reported error. If the sediment volumes are distributed according to a probability density function (PDF) \( f(v) \) and the fluid:sediment flux ratios are distributed according to a cumulative distribution function (CDF) \( G(r) \), then the CDF of the water volume \( \tilde{G}(v) \) is

\[ \tilde{G}(v) = \int_{-\infty}^{v} ds f(s) G(v/s) \]  

Using Eq. (11), we find the 5th percentile water volumes. The results are shown in Table 1. They range from 5.36 \( \times 10^{16} \) m³ (370 m GEL), using field data from Brownlie (1981) and the automated VN mapping of with a porosity of 0.4, to 13.79 \( \times 10^{16} \) m³ (952 m GEL), using laboratory data from Brownlie (1981) and the combined VN mapping with a porosity of 0.2.

Now, to report a final result, we (1) focus on the experimental (laboratory) dataset because sediment loads are more accurately measured in laboratory experiments than in rivers due to external influences in rivers, including vegetation, and because it is easier to measure sediment concentration in a controlled flume experiment; and (2) use the automated mapping from Luo et al. (2017) because it is a more conservative estimate for VN cavity volume. Treating the porosity as a...
normally distributed random variable with a mean of 0.3 and a standard deviation of 0.05 leads to a 5th percentile water volume of $9.3 \times 10^{16} \text{ m}^3$ (640 m GEL) (see Fig. 4), as does treating the porosity as a uniformly distributed random variable between 0.2 and 0.4. Therefore, we make a conservative estimate that at least $9.3 \times 10^{16} \text{ m}^3$ of water (640 m GEL) flowed through the VNs over the course of their formation. For the analysis done here, this result represents a 95% confidence value. While many assumptions and values with significant uncertainty were used in this analysis, we have consistently chosen to use conservative estimates for these assumptions and uncertain values. Thus, the results presented here represent minimum estimates and we conclude that a minimum of 640 m GEL water was likely to have been required to carve the VNs.

3. Application to individual VNs

The techniques used here can also be applied to individual VNs to estimate the volume of water required to form each of them. To illustrate this, we estimate the volume of water required to form each of the eight VNs studied by Hoke et al. (2011). Similar to the analysis described above, to do this requires estimates of sediment volume removed and fluid:sediment flux ratio. Table 3 of Hoke et al. (2011) lists the cavity volumes, channel widths, and slopes for eight VNs. These data were extracted from MOLA ($\sim 463 \text{ m/pixel}$) and HRSC ($\sim 75 \text{ m/pixel}$) datasets. First, we multiply the cavity volumes by $(1 - \lambda)$ to estimate the sediment volume as in Eq. (2). Note that Hoke et al., (2011) use $V_c$ to refer to what we refer to as the cavity volume, not the sediment volume, and erroneously divide the cavity volume by $(1 - \lambda)$ to obtain the sediment volume (cf. their Eq. (14)). We note that the methodology and equation that was implemented by Hoke et al. (2011) was originally introduced by Kleinhans (2005) and that this error in the described relationship between cavity volume and porosity has been inherited by multiple works since then (e.g., Barnhart et al., 2009; Jaumann et al., 2010; Matsubara et al., 2013; Luo et al., 2017), making it an important error to highlight here.

Next, in order to estimate the fluid:sediment flux ratio, we need to estimate the grain size and flow depth for each of the VNs. For each VN, we can find a relationship between the minimum bed grain size and the flow depth, similar to that displayed in Fig. 2 but different for each VN because of their different slopes. For each VN, we let the flow depth vary over the range explored by Hoke et al. (2011) and find the corresponding range of minimum bed grain sizes. We then substitute the morphometric parameters into Eq. (10) to estimate the near-bed concentration, $c_b$. Finally, we use the data from Brownlie (1981) to find the distribution of fluid:sediment ratios corresponding to the preferred range of $c_b$.

Table 2 shows the 5th percentile water volumes required to form these VNs. Note that in this estimate we have fixed the porosity at $\lambda = 0.3$, the sediment density at 2900 kg/m$^3$, and treated the Hoke et al. (2011) cavity volumes as exact since the uncertainties were not provided.

4. Discussion and implications

4.1. Minimum timescale of formation

The methods used here do not by themselves constrain the intermittency of erosional activity, defined as the fraction of the total time over which the VNs formed in which erosion occurred (e.g., Hoke et al., 2011). However, we can constrain the minimum time required to carve the VNs by assuming bankfull flow. We are not suggesting that the VNs experienced bankfull flow during their entire formation; we are using this assumption to provide a minimum time estimate for VN formation. Specifically, we can estimate the instantaneous fluid flux through the VNs using Eqs. (5) and (6) of Hoke et al. (2011) and utilizing the flow depths, slopes, and channel widths that they report in their Table 3. Fixing the porosity at 0.3, the sediment density at 2900 kg/m$^3$, and neglecting the uncertainties of the cavity volume and channel width leads to the formation timescales shown in Table 3. We estimate (1) minimum timescales for formation by assuming continuous fluvial activity capable of erosion and (2) intermittent timescales where we assume that fluvial activity capable of erosion occurred $5\%$ of the time. The minimum timescales estimated here (Table 3) are shorter than those found by Hoke et al. (2011) because (1) they used a median grain size of 10 cm, whereas, for each VN, we assume the smallest grain size that can exist on the bed without going into suspended flow; (2) they assumed a sediment density of 3400 kg/m$^3$, whereas we assume a sediment density of 2900 kg/m$^3$, and (3) their Eq. (14) contains an error and should read (in our notation) $T = V_c(1 - \lambda)/Q_c$, where $T$ is the formation timescale and our $V_c$ is their $V_c$. This error causes their timescales to be a factor of $(1 - \lambda)^{-2} \approx 2$ too long, for $\lambda \approx 0.3$.

4.2. Climatic implications: Revisiting the ambient climate scenario

With an improved understanding of the volume of water that is required to carve the VNs, we can now revisit the question: Does the formation of the martian VNs require a long-lived warm and wet climate, characterized by temperatures $> 273 \text{ K}$ and abundant rainfall, or could the VNs have formed through transient or punctuated periods of heating, melting of surface ice and runoff in a predominantly cold and icy climate, characterized by mean annual temperatures $< 273 \text{ K}$ and all water trapped as ice in the southern highlands? In other words, although the necessary volume of water, ~640 m GEL, can likely be reconciled by an active vertically integrated hydrological cycle and abundant pluvial activity in a warm and wet climate, can this amount of liquid water alternatively be sourced by melting of ice in a cold and icy climate? The predicted distribution of ice in a cold and icy climate is spatially well-correlated with the distribution of valley networks and lakes, implying that melting of surface ice and runoff could offer a plausible explanation for the formation of these features (Head and Marchant, 2014). Multiple mechanisms for transient or punctuated heating in a cold and icy climate have been proposed, including impact cratering-induced heating (e.g., Segura et al., 2008), a transient reducing greenhouse-rich atmosphere (Wordsworth et al., 2017), volcanism-induced heating (e.g., Halevy and Head, 2014; Kerber et al., 2015), and seasonal temperature variations and summertime melting (e.g., Palumbo et al., 2018a). Previous work by Fastook and Head (2015)
Fig. 3. Left: Scatter plot comparing the empirical values of fluid:sediment flux ratio reported by Brownlie (1981) to the values of $1/c_b$ predicted by Eq. (10), a version of the van Rijn (1984) equation. The vertical lines indicate the boundaries of the preferred region of $1/c_b$. Right: Normalized histogram of the fluid:sediment flux ratio for the data points between the two vertical lines in the scatter plot.

Fig. 4. We use a CDF (blue curve) to predict the probability (y-axis) that the VN could have been formed with a water volume less than a given amount (x-axis). In our calculations, we assume porosity of 0.3 and use the laboratory data (instead of the field data) compiled by Brownlie (1981) that satisfies $0.0694 < c_b < 0.125$, the preferred range of $c_b$.

Table 2
5th percentile water volumes required to form the VNs studied by Hoke et al. (2011), assuming a porosity of 0.3 and neglecting the uncertainties of the cavity volumes, which Hoke et al. (2011) did not provide.

<table>
<thead>
<tr>
<th>Valley Network</th>
<th>5th percentile GEL (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>12° S, 12° E Evros</td>
<td>38.28</td>
</tr>
<tr>
<td>7° S, 3° E</td>
<td>2.20</td>
</tr>
<tr>
<td>3° S, 5° E</td>
<td>1.09</td>
</tr>
<tr>
<td>0° N, 23° E</td>
<td>7.17</td>
</tr>
<tr>
<td>2° N, 34° E Naktong west</td>
<td>1.82</td>
</tr>
<tr>
<td>2° N, 34° E Naktong east</td>
<td>31.06</td>
</tr>
<tr>
<td>12° N, 43° E</td>
<td>6.42</td>
</tr>
<tr>
<td>6° S, 45° E</td>
<td>8.15</td>
</tr>
</tbody>
</table>

Table 3
Formation timescales for the VNs studied by Hoke et al. (2011) derived using the methods of Section 4.1. “Shallower flow depth” and “deeper flow depth” refer to the two flow depths assumed by Hoke et al. (2011) for each VN. Intermittent timescales assume that the erosion occurred 5% of the time. Timescales are in Earth years.

<table>
<thead>
<tr>
<th>Valley Network</th>
<th>Continuous timescale of formation (Earth years)</th>
<th>Intermittent (×20) Timescale (Earth years)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Evros</td>
<td>Continuous timescale of formation (Earth years)</td>
<td>intermittent (×20) timescale (Earth years)</td>
</tr>
<tr>
<td>12° S, 12° E Evros</td>
<td>Continuous timescale of formation (Earth years)</td>
<td>intermittent (×20) timescale (Earth years)</td>
</tr>
<tr>
<td>7° S, 3° E</td>
<td>7.21 × 10^3</td>
<td>3.10 × 10^4</td>
</tr>
<tr>
<td>3° S, 5° E</td>
<td>1.73 × 10^2</td>
<td>3.10 × 10^3</td>
</tr>
<tr>
<td>2° N, 34° E Naktong west</td>
<td>1.85 × 10^3</td>
<td>3.70 × 10^4</td>
</tr>
<tr>
<td>2° N, 34° E Naktong east</td>
<td>3.91 × 10^3</td>
<td>7.83 × 10^4</td>
</tr>
<tr>
<td>12° N, 43° E</td>
<td>1.14 × 10^3</td>
<td>6.08 × 10^3</td>
</tr>
<tr>
<td>6° S, 45° E</td>
<td>6.68 × 10^2</td>
<td>5.80 × 10^3</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Valley Network</th>
<th>Continuous timescale of formation (Earth years)</th>
<th>Intermittent (×20) Timescale (Earth years)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Evros</td>
<td>Continuous timescale of formation (Earth years)</td>
<td>intermittent (×20) timescale (Earth years)</td>
</tr>
<tr>
<td>12° S, 12° E Evros</td>
<td>Continuous timescale of formation (Earth years)</td>
<td>intermittent (×20) timescale (Earth years)</td>
</tr>
<tr>
<td>7° S, 3° E</td>
<td>4.26 × 10^3</td>
<td>5.50 × 10^2</td>
</tr>
<tr>
<td>3° S, 5° E</td>
<td>7.99</td>
<td>5.26 × 10^3</td>
</tr>
<tr>
<td>0° N, 23° E</td>
<td>8.02 × 10^3</td>
<td>7.46 × 10^2</td>
</tr>
<tr>
<td>2° N, 34° E Naktong west</td>
<td>3.01 × 10^3</td>
<td>1.68 × 10^2</td>
</tr>
<tr>
<td>2° N, 34° E Naktong east</td>
<td>5.92 × 10^3</td>
<td>4.50 × 10^3</td>
</tr>
<tr>
<td>12° N, 43° E</td>
<td>5.00 × 10^3</td>
<td>2.66 × 10^2</td>
</tr>
<tr>
<td>6° S, 45° E</td>
<td>8.39 × 10^3</td>
<td>7.29 × 10^2</td>
</tr>
</tbody>
</table>
implemented an ice sheet evolution model, approximated global temperature maps, and a positive degree day analysis to determine the amount of meltwater that can be produced through transient and punctuated heating in a cold and icy climate, finding that a +18 K heating event (global MAT ~ 243 K) would only have to persist for 2000 years to produce sufficient meltwater to fill the open-basin lakes. Thus, it is reasonable to assume that some or all of these transient and punctuated heating mechanisms could have introduced significant volumes of liquid water to the surface. To make a first order approximation regarding whether or not these mechanisms can account for formation of the VNs in a cold and icy climate, we can compare the volume of water produced through these mechanisms with the volume of water required to carve the VNs.

Impact cratering-induced heating was likely to be an important mechanism following the formation of basin-scale impact events (Segura et al., 2008; Palumbo and Head, 2017). However, the basins formed in the early- to mid-Noachian, not in the LN-EH, when the majority of the VNs formed (e.g., Fassett and Head, 2011). For this reason, we do not consider the role of impact cratering-induced heating in further detail.

The role of a transient reducing greenhouse atmosphere has recently been proposed (Wordsworth et al., 2017), however the magnitude of heating and amount of meltwater produced is dependent upon the concentration of the specific atmospheric greenhouse gases, including H₂ and CH₄. The likely atmospheric concentrations of these gases in the LN-EH must be better constrained before the amount of meltwater produced by this mechanism can be assessed in more detail.

Although volcanism was an active and abundant process in the LN-EH, the duration of volcanism-induced heating was likely to have been short due to the rapid conversion of sulfur-based gases into aerosols, which are atmospheric cooling agents. The short duration of volcanism-induced heating, possibly only one season to one year (e.g., Kerber et al., 2015), paired with the fact that reasonable concentrations of sulfur-based gases cannot increase global MAT > 273 K (e.g., Mischna et al., 2013), has commonly led researchers to disregard this mechanism when considering meltwater production and the formation of the VNs. However, recent work has revisited this idea again, finding that significant melting could occur during the summer season (Palumbo et al., 2018b).

Additionally, for reasonable sulfur-based greenhouse gas concentrations, hundreds to thousands of volcanic eruptions would be required to produce the necessary volume of meltwater for formation of the VNs (Palumbo et al., 2018b), which is not unreasonable given the coincident timing of the formation of the volcanic Hesperian-riddged plains (Head et al., 2002).

Finally, Palumbo et al. (2018a) found that significant volumes of meltwater can be produced through summertime melting. Palumbo et al. (2018a) utilized 3D GCM simulations, positive degree day calculations which use temperature as a proxy to estimate the amount of ice melted, the volume of water required to carve the VNs (considering the results from Rosenberg and Head, 2015; 3–100 m GEL), and estimates of the necessary duration of fluvial activity to form specific VNs ranging from ~10⁵–10⁷ years (Hoke et al., 2011), to estimate whether the VNs could have formed through this seasonal melting mechanism. Specifically, Palumbo et al. (2018a) compared the volume of meltwater produced in one year to the total volume required to carve the VNs to determine whether the VNs could have formed through this mechanism within the predicted duration of VN formation.

We consider ~10⁵–10⁷ years (Hoke et al., 2011) to be a reasonable estimate for the lower limit of the cumulative duration of VN formation. One reason that this must be a lower limit for the cumulative duration of VN formation is that, in order for all VNs to have formed with ~10⁵–10⁷ years and for the formation of a single VN to take ~10⁵–10⁷ years, all of the VNs must have formed contemporaneously; in fact, cumulative VN formation likely took much longer. Hoke and Hynek (2009) found that fluvial activity at many VNs terminated between ~3.6–3.8 Ga (~2 × 10⁸ years). This range in timing of the end of fluvial activity at different VN systems implies that all systems were not active at once and, more specifically, that fluvial activity likely persisted on Mars for at least 200 million years, although fluvial activity may not have persisted at one location for that long. Additional reasons that we are only capable of estimating a lower limit for the cumulative duration of VN formation through geomorphologic analyses (such as those by Hoke et al. (2011) and Hoke and Hynek (2009)) include: (1) significant temporal and spatial intermittency of fluvial activity likely persisted during the period of VN formation, and (2) significant amounts of lower magnitude fluvial activity that was not very erosive may have occurred in the LN-EH and not left a signature on the surface.

In this work, we have revisited the duration of fluvial activity that is required to form the eight large VNs that were previously analyzed by Hoke et al. (2011) by implementing our improved analyses and conservative lower-bound estimates for relatively unconstrained parameters (Table 3). We find formation timescales that are shorter than those predicted by Hoke et al. (2011) and that the maximum amount of time that it took to form one of these eight large VNs is ~10⁴ years. Thus, we determine that the absolute minimum duration of cumulative VN formation is ~10⁴ years; this duration requires that the VNs all formed contemporaneously, which we reiterate is unlikely.

Next, we will utilize the results from Palumbo et al. (2018a) to determine whether 640 m GEL of meltwater can be produced within ~10⁵–10⁷ years through seasonal melting in a cold and icy climate. We perform our calculations using the timescales from Hoke et al. (2011), but note that the reader can easily make comparisons to the timescales predicted by the work of Hoke and Hynek (2009) (a minimum of ~200 million years) and the results of this study (an absolute minimum of ~10⁴ years), as well.

4.2.1. Revisiting the seasonal melting hypothesis

We now revisit the seasonal melting hypothesis, implementing our updated estimate for the minimum volume of water required to carve the VNs. Palumbo et al. (2018a) found that with a 1 bar CO₂ atmosphere, 25° obliquity, and a circular orbit, 1.26 × 10⁶ m GEL of meltwater would be produced annually. Therefore, based on our calculations, it would have taken at least 507 million years to produce 640 m GEL of meltwater and form the VNs by this process alone. Our estimates regarding meltwater production through seasonal melting requires a duration longer than the estimated amount of time required to carve individual VNs, 10⁵–10⁷ years. Although it is possible that fluvial activity persisted for much longer than the estimates produced by analyses of eight specific VNs (Hoke et al., 2011), this simple calculation suggests that it may not be likely that the VNs formed through seasonal melting under the spin-axis/orbital conditions that are specified above.

If the eccentricity is increased to 0.17 (upper limit of statistically probable eccentricity values for the past 5 billion years with ~20% probability of having occurred; Laskar et al., 2004) and the other conditions are kept the same, then 1.97 × 10⁻⁴ m GEL of meltwater is produced for the condition of perihelion at southern hemispheric summer (Palumbo et al., 2018a), implying that it would have taken at least 6.95 million years to form the VNs by this process alone. However, the precession cycle for longitude of perihelion is ~50,000 years. Thus, making the simple assumptions that these perihelion conditions would exist for ~50% of the period of VN formation and that minimal ice melting occurs in regions where VNs exist when under the conditions of the reverse perihelion scenario (e.g., Palumbo et al., 2018a), we can infer that VN formation through this process would take ~13.9 million years. This estimate is consistent with the upper limit of required duration of fluvial activity to form the specific VNs analyzed by Hoke et al. (2011) (10⁵–10⁷ years), implying that the formation of the VNs can be explained through this seasonal melting scenario when considering some additional intermittency in fluvial activity.

Alternatively, with a circular orbit, additional greenhouse warming (global MAT ~ 243 K), and the other spin-axis/orbital conditions kept the same, 1.45 × 10⁻⁴ m GEL of meltwater would have been produced
annually, implying that it would have taken at least 4.4 million years to form the VNs by this process alone. Again, this time estimate is consistent with the required duration of fluvial activity to form the specific VNs analyzed by Hoke et al. (2011).

In conclusion, the volume of water required to carve the VNs could be explained by seasonal melting in a predominantly cold and icy Late Noachian climate under the conditions of either a slightly more eccentric orbit or a small amount of additional greenhouse warming (e.g., MAT \( \sim 243 \) K). However, the volume of water is not the only important factor to consider when constraining whether or not a mechanism can explain the formation of the VNs – other important factors include runoff rates and predicted geomorphology.

4.2.2. Consideration of runoff rates and geomorphology

Although the water volume can be reconciled by transient and punctuated heating, ice melting, and surface runoff in a predominantly cold and icy climate, the question remains: Are the runoff rates required to form the VNs and the geomorphology of the VNs consistent with formation through these processes? Or, alternatively, are the characteristic runoff rates and the geomorphology of the VNs more consistent with rainfall and runoff in a warm and wet climate?

Hoke et al. (2011) performed geomorphological analyses to estimate runoff rates required to carve the VNs, finding that runoff rates of mm/day to cm/day are required. When considering surface runoff in a cold and icy climate, we can assume that meltwater would tend to undergo surface runoff (in contrast to infiltration and runoff in a warm and wet climate) because the substrate is assumed to be frozen (permafrost) due to the extremely low temperatures (e.g., Cassanelli and Head, 2018). Thus, higher snowmelt rates produce the highest runoff rates (e.g., Scanlon et al., 2016). Palumbo et al. (2018a) found that snowmelt rates were higher in the relatively higher eccentricity simulation at \( 0.6 \) mm/day in the summer season, implying that runoff rates in the relatively higher eccentricity simulation are also the higher than runoff rates in the circular orbit simulations. Scanlon et al. (2016) found that for simulations with relatively higher mean annual temperature (close to or greater than 273K), snowmelt runoff rates become too high to explain the VNs. Thus, runoff rates in colder simulations, such as those studied by Palumbo et al. (2018a), are more likely to be consistent with the runoff rates required to carve the VNs. Furthermore, short-lived heating and melting (1) from volcanism-induced heating could produce a range of runoff rates depending on the magnitude of the volcanic eruption and associated degree of warming from sulfur-based gases, and (2) from a transient reducing greenhouse-rich atmosphere could produce a range of runoff rates depending on the concentration of different gases in the atmosphere, such as H\(_2\) and CH\(_4\).

In a warm and wet climate, it is not expected that all rainfall will lead to surface runoff because some rainfall will inevitably infiltrate into the unfrrozen substrate. Future modeling efforts that predict rainfall distributions and rates are required to better assess whether or not rainfall in a warm and wet climate can produce runoff rates comparable to those required to carve the valley networks (mm/day to cm/day; Hoke et al., 2011).

The geomorphology, such as hydrologic complexity, of the VNs are also important characteristics to consider when contemplating whether or not the VNs could have formed through snowmelt. In an effort to better understand whether the geomorphic characteristics of the VNs can be explained by snowmelt, recent work by Quintel et al. (2018) has analyzed the climate controls on the seasonal fluvial activity and geomorphic characteristics of the Onyx River in the McMurdo Dry Valleys. The Onyx River and associated streams are sourced by top-down melting of existing cold-based glaciers and are highly active during the warmest part of the southern summer season (e.g., Atkins and Dickson, 2007; Dickson et al., 2017; Head and Marchant, 2014; Shaw and Healy, 1980); observations of the current fluvial activity and erosion at the Onyx River offers insight into the formation and evolution of rivers through melting of snow and ice and runoff in a hyper-arid and hypothermal environment. When analyzed at similar spatial resolution and accounting for erosion of smaller streams that may have once been associated with the martian VNs but have since been eroded away, Strahler stream order analyses of the Onyx River and martian VNs suggest that the Onyx River is equal to or more developed than \( \sim 90\% \) of the martian VNs (Quintel et al., 2018). This result implies that the characteristics and degree of development of the majority of the VNs does not preclude formation through seasonal, transient, or punctuated snowmelt and runoff in a cold and icy climate.

4.3. Relation to global water inventory

The volume of water required to carve the VNs is not simply related to the global inventory at any moment in time (e.g., Carr and Head, 2015). Water can cycle multiple times through the VNs via climatic recycling (i.e., evaporation or sublimation followed by re-melting of surface ice and snowfall or rainfall in the source region to replenish the water source; e.g., see Head and Marchant (2014), for terrestrial and martian examples). Water can also remain trapped as ice, as groundwater, or in distant geographic locations, effectively removing it from the integrated hydrologic system. Therefore, significant caution must be used if attempting to estimate the global inventory of water from the cumulative volume of water required to carve the VNs. In other words, our results do not require that 640 m GEL of water existed on the surface of Mars at any one time. In fact, the volume of water on the martian surface at any one time could have been much less due to atmospheric recycling.

4.4. Conclusions

We have implemented updated and more accurate methods and datasets to better constrain the volume of water required to carve the martian VNs. The minimum water volume found here (640 m GEL) is larger than that found by Rosenberg and Head (2015) (3–100 m GEL) but smaller than that found by Luo et al. (2017) (5000 m GEL). Further, we find that this volume of water is consistent with either a warm and wet or a cold and icy climate scenario. In a warm and wet scenario, sufficient liquid water can be introduced to the surface through rainfall and runoff. Alternatively, in a cold and icy scenario, we find that it is possible to produce enough seasonal meltwater to form the VNs within a geologically plausible timescale and further speculate that punctuated volcanism-induced heating or a transient reducing greenhouse-rich atmosphere could also produce sufficient meltwater (c.f. Section 4.1). Our conclusion that the presence of the VNs does not preclude a cold and icy climate is counter to that of Luo et al. (2017), who found that a warm and wet climate was necessary to form the VNs and that the presence of a Noachian ocean was required in the northern lowlands.

Acknowledgments

We thank Rebecca Williams for sharing the original data used in the Williams and Phillips (2001) analysis. We gratefully acknowledge financial support from NASA for participation in the Mars Express High Resolution Stereo Camera (HRSC) Team (JPL 1488322) to JWH, and for support by NASA Headquarters under the NASA Earth and Space Science Fellowship Program to AMP – Grant 90NSSC17K0487. This project began with support from the Brown University Undergraduate Teaching and Research Awards (UTRA) program to ENR and JWH, which is gratefully acknowledged.

Supplementary materials

Supplementary material associated with this article can be found, in the online version, at doi:10.1016/j.icarus.2018.07.017.
References

Atkins, C., Dickinson, W., 2007. Landscape modification by meltwater channels at the margin of cold-glaciated, Dry Valleys, Antarctica. Boreas, 36, 47-55.


