Supplementary Information for “Valley Networks and the Record of Wet-Based Glaciation on Ancient Mars”

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Supplementary text

S1 Note of the comparative geometry of Terrestrial and Martian ice sheets.

For the sake of isolating the effect of gravity on glacial drainage and glacial sliding velocity, the manuscript assumes two identical ice sheets on Earth and Mars. This idealized scenario is, of course, an assumption and overlooks differences that go beyond the dynamic regime studied here. Here we discuss the most significant aspects where ice sheets

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on Earth and Mars may differ, which are mass balance, thermophysics, and ice surface topography.

**S1.1 Earth and Mars ice sheet geometry.**

An ice sheet has to transport the mass accumulated upstream to its ablation area, so that in equilibrium, ice must transport mass fast enough to balance the gain and loss. Ice flow velocity, mass balance rates, and rheology thus determine the equilibrium surface shape of a glacier or ice sheet. Ice is a viscous fluid with a non-linear rheology given by Glen’s law:

\[ \dot{\varepsilon} = A \tau^n. \]  

(1)

Considering a mass of ice resting on a slope, the shear stress is given by \( \tau = \rho_i g H S \), where \( S = dH/dx \). \( A \) is the temperature dependent softness parameter, \( n \) is Glen’s exponent \( (n \sim 3) \), and \( \dot{\varepsilon} \) is the strain rate. Integrating Glen’s law considering null lateral gradients, we find the Shallow Ice Approximation equation:

\[ u(z) = \frac{2A}{n + 1} \left( \rho_i g \frac{\partial H}{\partial x} \right)^n \left[ H^{n+1} - z^{n+1} \right] + u_s, \]

(2)

where \( H \) is the surface of the ice sheet, \( z \) is the elevation over the bed, and \( u_s \) is the basal sliding velocity. To be in steady state, the ice velocity flux given by \( u(z) \) has to balance the mass balance flux. This balance sets the shape of an ice sheet (Cuffey & Paterson, 2010), which in absence of basal sliding is given by:

\[ H^{2+2/n} = \frac{2(n + 2)^{1/n}}{\rho_i g} \left( \frac{\dot{b}}{2A} \right)^{1/n} x^{1+1/n}, \]

(3)

with \( \dot{b} \) the mass balance rate (m/s) integrated assumed independent of position or elevation. A comparison between the surface elevation of a terrestrial and a Martian ice
sheet using equation 1 is given in figure S1. Because of the dependence on gravity, the equilibrium shape of an ice sheet on Mars is thicker and shorter if we assume mass balance is homogeneous and equivalent between Earth and Mars. Slopes should be steeper, particularly at the ice margin.

S1.2 Mass balance.

Mass balance, however, is never homogeneous and independent on elevation. The patterns of accumulation and mass loss or ablation depend strongly on the temperature and humidity of the atmosphere, both depending on the elevation of the ice surface, atmospheric lapse, surface temperatures and water partial pressures, etc. Any of these variables differ between Earth and Mars, so that the true equilibrium shape of an ice sheet can only be found coupling a climate model with an ice sheet model until it reaches steady-state. Several studies (Fastook et al., 2008; Fastook & Head, 2015; Wordsworth et al., 2013) have attempted local and global reconstructions of the Martian glacial mass balance. The largest Earth-Mars differences are in the ablation patterns, in particular the relative contribution to melt vs. sublimation and calving (mass loss at the margin). Temperature on Earth and Mars would decrease roughly linearly with elevation (the so-called lapse rate), but in the case of Mars this decrease would be a factor \( g_E/g_M \sim 3 \) lower. Lower Martian surface pressures would then contribute to a stronger presence of sublimation than on Earth, whereas steeper margin slopes likely enhance ice mass wasting events at the ice margin (calving).

If the mass balance on early Mars was similar to that occurring on Earth nowadays, then a scenario with thicker ice bodies and steeper marginal slopes such as that shown...
in figure S1 would develop. If instead early Martian water vapor pressure was lower and sublimation was more important than on Earth, then a mass balance with a double Equilibrium Line Altitude (ELA, where accumulation equals ablation) emerges (figure 5, (Fastook et al., 2008)), as sublimation dominates the mass balance at high elevations. This second scenario seems likely because of the lower water budget and surface pressure values expected for early Mars. In this second case, ice sheets would tend to become much flatter at higher elevations, with a thickness controlled by the second ELA where accumulation equals sublimation rates (figure S2).

S1.3 Temperature and rheology.

Ice temperature, basal temperature, and surface temperature all play important roles in determining the size of an ice sheet in equilibrium. Temperature in particular has a large effect on ice effective viscosity through the softness parameter $A(T)$: warmer ice sheets will flow faster, and therefore will have equilibrium forms that are thinner and span greater lengths.

Mars and Earth are differently affected by temperature changes: whereas a cooling temperature produces thicker ice sheet equilibrium surfaces in both cases for a given length span, the difference in equilibrium elevation increases non-linearly, as shown in figure 1.

The temperature at the ice sheet surface, as discussed above, is one of the controls of the water exchange with the atmosphere and the distribution of accumulation and ablation zones. If melting either existed at the surface or confined under a thin layer of ice (Kaufmann et al., 2006), water will travel to the ice margin as either surface supraglacial
channels or percolating to the interior through moulins or crevasses. Once in the ice sheet, water may flow in channels embedded in the ice (englacial channels) or accumulate at the base of the ice. Most of the water originating at the surface of terrestrial ice sheets, in occasions more than 90%, is delivered through moulins or crevasses to the base (Cuffey & Paterson, 2010; Smith et al., 2015).

The temperature at the base of the ice controls whether the ice sheet is in a cold-based (frozen to the ground) or wet-based regime (water exists at the base, also called a ‘warm-based’ or temperate regime). Basal temperature is set by the heat transfer through the ice and by the geothermal heat flow, predicted to be in the range 45-65 mW/m² for early Mars (Fastook & Head, 2015; Solomon et al., 2005). Areas of high stress (such as steep slopes or around obstacles) may produce internal deformation leading to frictional heating. Supercooled water is also commonly present in liquid state under ice sheets below its pressure melting temperature (Cuffey & Paterson, 2010).

### S1.4 Basal meltwater sources and rates.

Basal water production rates are controlled by the release of basal heat, which involves contributions from geothermal heat flux, frictional heating due to glacial sliding, heat conduction through the ice layer, and meltwater sourced from surface melt (from left to right, terms in the right hand side in equation S3), as detailed in chapter 4 in (Cuffey & Paterson, 2010),

\[
\dot{b} = \frac{G_M}{\rho L} + \frac{u_s \tau_b}{\rho L} + \frac{\kappa_i \Delta T}{\rho LH} + \dot{b}_s
\]  

The following discussion refers specifically to the Dorsa Argentea southern circumpolar ice sheet (DASCIS). Whereas the precise values of the geothermal heat flow for early
Mars are poorly constrained, numbers are typically considered to be on the range of $G_M = 45-65 \text{ mW/m}^2$ for the DASCIS (Fastook et al., 2012). The presence of subglacial volcanoes could locally enhance these values. Whereas the meltwater production rates by geothermal heat flux likely did not differ greatly from terrestrial values, the slower sliding rates discussed in the manuscript would greatly decrease meltwater production by frictional melt. Similarly, the low surface temperatures (-50°C to -75°C) predicted for the southern circumpolar region in the Noachian-Hesperian transition (circa 3.5 Byr ago, (Fastook et al., 2012)) would result in a net loss of basal heat by conduction. The last source of meltwater in equation S3, coming from surface melt, could be locally important given the active cratering rate expected for early Mars. Taking the following parameters as reference:

An order of magnitude estimation for the background meltwater production rate $\dot{b}$ following equation S3, in m/s, yields:

$$\dot{b} \sim 10^{-10} + 10^{-15} - 10^{-10} + \dot{b}_s$$ (5)

Areal conduction and melt by geothermal heat flux would tend to cancel each other, whereas the contribution from basal frictional melt is much lower than on Earth given the slower sliding rates. With the lateral extent of the Dorsa Argentea southern circumpolar ice sheet predicted to be $L_{span} \sim 1200 \text{ km}$ (Head & Pratt, 2001; Fastook et al., 2012), the total steady discharge produced under the whole ice sheet at a given time is order $Q \sim 10 \text{ m}^3/\text{s}$ without accounting for peaks in basal meltwater input rates, given by surface melt inputs or volcanic sources.
It is likely that surface melt percolating to the base would dominate the glacial hydrology sources, further preventing areal sliding to develop, particularly given the higher cratering rate on early Mars. Thus, on non-volcanic, relatively flat terrains on early Mars, the basal melt sources would be dominated by surface melt produced in impacts, and delivered to the bed via moulins, and by basal melt in localized high heat flux regions ($G_{MV}$ in volcanic or hydrothermal centers). Surface temperatures are otherwise too low to produce significant constant melt in the DASCIS, and steady, significant areal production of basal meltwater is similarly inhibited. Thus,

$$\dot{b} \sim \frac{G_{MV}}{\rho L} + \dot{b}_s \quad (6)$$

The discharge rates achieved with these melt production rates scale with the production area. If we consider a cavity with a typical area of $\sim 1\text{ km}^2$, the discharge produced would be order $Q \sim 10^{-10}$ to $10^{-15}\text{ m}^3/\text{s}$, which is unreasonably small to open a cavity. The discharge produced accounting all background sources under the whole DASCIS (with a diameter of $\sim 1200\text{ km}$ (Fastook et al., 2012, e.g.,) is only of order $Q \sim 10\text{ m}^3/\text{s}$. Thus we assume $\dot{b} \sim \dot{b}_s$. Therefore, the discharge values presented in the manuscript are taken from the estimated discharge necessary to produce the eskers observed in the geological record of the DAF. The minimum constrained discharge is $50\text{ m}^3/\text{s}$ through an average esker, and the maximum is the peak discharge given (Butcher et al., 2017; Scanlon et al., 2018). These values are larger than the background rate of meltwater production, and thus highlight the need for localized meltwater input, either originating from the surface $\dot{b}_s$ or from basal sources (hydrothermal, volcanic, high salt concentrations, etc.).

**S1.5 Summary note.**
In summary, the equilibrium shape of a Martian ice sheet is likely shorter and thicker than a terrestrial one if the same temperature conditions are considered, with thickness capped by the elevation at which sublimation becomes dominant. The resulting body is likely flatter than a terrestrial ice sheet after ELA 2 (figure S2) and has a steeper ice margin (figures S1 and S3). Mass balance distribution and temperature conditions are certainly a key difference between Earth and Mars, and both play important roles determining the shape and thickness of an equilibrium ice sheet. Therefore determining the geometry of an ice sheet on Mars requires the full coupling of a climate model with an ice sheet evolution model, which is beyond the scope of this paper but has been done in previous studies (Fastook et al., 2008; Wordsworth et al., 2013; Fastook & Head, 2015).

In particular, we take the work by Fastook (Fastook et al., 2008; Fastook & Head, 2015) as a guide to justify that ice sheets of similar geometry and thickness to terrestrial ice sheets can result from the coupling of realistic climate and ice flow models for Mars.

S2 Details of the subglacial conduit evolution model.

We use the model presented by Schoof (Schoof, 2010), which synthesizes the models of channel opening and conduit closure (J. Nye, 1976; Röthlisberger, 1972; J. F. Nye, 1953) and the formulations of cavity hydrology established in the literature (Walder, 1986; Kamb, 1987, 1970). Below we summarize the main details of the subglacial hydrology evolution model presented in the following equation:

\[
\frac{\partial X_s}{\partial t} = c_1 Q \Psi + u_s h - c_2 N^n X_s
\]  

(7)

The rate of change of conduit cross-section with respect to time \(\partial X_s/\partial t\) depends on the rate of subglacial channel opening \(c_1 Q \Psi\), on the rate of cavity growth \(u_s h\), and on the
rate of channel creep closure \( c_2 N^n X_s \). Here \( Q \) is discharge, \( u_s \) is sliding velocity, \( h \) is the height of bed protrusions, \( N = P_i - P_w \) is effective pressure (ice overburden minus water pressure), \( n \) is the exponent of the ice flow law (Cuffey & Paterson, 2010) and \( c_1 \) and \( c_2 \) are constants (Schoof, 2010). \( \Psi \) is the hydraulic potential gradient along the conduit, which describes the pressure gradient due to the hydrostatic gravitational component as well as changes in the water pressure:

\[
\Psi = -\rho_w g \frac{\partial z_b}{\partial x} - \frac{\partial P_w}{\partial x} \quad (8)
\]

Below we discuss each individual term, using the term ‘conduit’ as in the manuscript to refer to any opening at the ice sheet bed where water can flow, either a channel or a cavity.

**S2.1 Conduit closure.**

A tunnel of cylindrical symmetry and semi-circular cross-section established at the bottom of an ice sheet will close driven by the difference between the ice cryostatic pressure and the water pressure inside of the channel, given by \( N \). Ice flows inwards to close the channel at a rate given by its rheology, channel cross-section, and driving stress (J. F. Nye, 1953; Schoof, 2010):

\[
\frac{\partial X_{sc}}{\partial t} = 2A n^{-n} N^n X_s \quad (9)
\]

Where \( A, n \) establish the flow rheology and effective viscosity of ice through Glen’s law (Cuffey & Paterson, 2010).

**S2.2 Channel growth.**

The mechanics of channel evolution were initially described by Hans Röthlisberger (Röthlisberger, 1972) and John Nye (J. Nye, 1976; J. F. Nye, 1953), and are estab-
lished by a competition between the rate of wall melt and the rate of ice creep closure. The inwards creep of ice is thus balanced by wall melt due to the turbulent dissipation of heat in the conduit. Assuming negligible changes in kinetic energy, conservation of energy establishes that the power done by the turbulent flux of water driven by the hydraulic gradient $Q\Psi$ dissipates into heat.

If this heat is instantly delivered to the ice wall and turned into latent heat, the rate of turbulent wall melt is:

$$\frac{\partial X_{sm}}{\partial t} = \frac{Q\Psi}{\rho_i L}. \quad (10)$$

The evolution of a subglacial channel is thus given by the balance between opening and closure rates:

$$\frac{\partial X_{sch}}{\partial t} = \frac{Q\Psi}{\rho_i L} - 2An^{-n}N^n X_s. \quad (11)$$

We can now identify the constants $c_1$ and $c_2$ in equation S1:

$$c_1 = \frac{1}{\rho_i L} \quad \text{and} \quad c_2 = 2An^{-n}. \quad (12)$$

S2.3 Cavity opening.

Water drainage through the action of cavities is much more complex, and takes the form of a continuum, macroporous problem. A model presented by Walder (Walder, 1986) and followed by Schoof (Schoof, 2010) in the formulation of equation S1 captures the basic dynamics of cavity growth and closure that appear in more complex models. A cartoon of this simplified model (fig S5) shows a block of ice sliding over a protrusion of size $h$ with a sliding velocity $u_s$. The size of the cavity increases as sliding velocity or protrusion height increase, and decreases when ice flows inside the gap.

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The linear dependence of opening rates on sliding velocity, and of closing rates on the $n-th$ power of the effective pressure assumes implicitly that cavity aspect ratio is order 1: height and length are similar. This assumption is reasonable when sliding velocity and/or protrusion height $h$ is small, so that cavities are not too shallow (see cartoon in figure S5, left panels). When this is not true, a non-linear dependence on effective pressure and protrusion size emerge, which need to be considered in the closure term (Schoof, 2010; Kamb, 1987). Because the model proposed by Schoof (Schoof, 2010) better describes cavity opening rates at small sliding velocity rates, we expect this physical representation of cavity opening to better describe Martian ice sliding than terrestrial ice sliding.

S3 Timescales of conduit closure.

Conduit closure rate is given by the negative of conduit opening rate:

$$\frac{\partial X_s}{\partial t} = -c_1 Q \Psi - u_s h + c_2 N^n X_s.$$  \hspace{1cm} (13)

The time span required for the closure of a conduit (channel or cavity) with an initial cross-section of $X_{so}$ down to a final cross-section $X_{sf}$ is:

$$\Delta t = \int_{X_s}^{X_{sf}} \frac{dX_s}{-c_1 Q \Psi - u_s h + c_2 N^n X_s}.$$  \hspace{1cm} (14)

Although the integral is not analytic, here we are only interested in seeing whether subglacial channels close between one melt cycle and the next on Mars, i.e., whether the Martian subglacial drainage system is stable. To do so, we can start by taking the fastest rate at which conduits may close, which is obtained setting $N \sim P_i$, ($P_w \ll P_i$) and of course implies $c_1 Q \Psi \sim 0$ and $u_s h \ll c_2 N^n X_s$. However, $Q \sim 0$ indicates a complete lack of flow inside the channel. To keep a functional channel, the critical discharge must be
achieved $Q = Q_c, \Psi = \Psi_c$. This critical discharge will approach zero if and only if $u_s \to 0$. 

$$\Delta t = \frac{2 \ln \left( c_2 P_i^n X_s^{7/2} - c_1/c_3 Q_c^3 \right)}{7c_2 P_i^n},$$

which evaluated between $X_{so}$ and $X_{sf}$ yields

$$\Delta t = \frac{2}{7c_2 P_i^n} \ln \left( \frac{c_2 P_i^n X_{sf}^{7/2} - c_1/c_3 Q_c^3}{c_2 P_i^n X_{so}^{7/2} - c_1/c_3 Q_c^3} \right).$$

This expression calculates the time needed to shut down a subglacial channel on Mars or Earth, assuming a steady critical discharge. The time required for the shutdown of the subglacial drainage system depends on ice thickness and cryostatic pressure, ice softness through temperature and likely presence of dust, discharge, initial, and final cross-section. Figure S6 shows the comparison of Mars and Earth shutdown times, including the duration of the winter season, resulting from this equation.

If the drainage falls below critical, the closure rate of a subglacial conduit is set by the initial cross-section and effective pressure. In this case, the rate of ice creep closure in a conduit is given by the work by Nye (J. F. Nye, 1953; J. Nye, 1976):

$$\frac{\partial X_s}{\partial t} = c_2 N^n X_s.$$  \hspace{1cm} (17)

The timescale that can be derived from this rate is, taking effective pressure to be ice overburden,

$$\Delta t = \int_{X_{so}}^{X_{sf}} \frac{dX_s}{c_2 N^n X_s} = \frac{1}{c_2 (\rho g H)^{\eta}} \ln \left( \frac{X_{sf}}{X_{so}} \right).$$

An interesting hysteresis may emerge here, in which terrestrial channels close in the winter due to a higher critical discharge and a faster closure rate, whereas Martian channel may remain open even throughout the longer winter season. This hysteresis, which depends on the the minimum and peak discharges (nominally, winter and summer discharges),
the duration of the freezing and melting season, and the effective pressure, is depicted in figure S7 below.

S4 Model sensitivity and Martian parameter space.

In the manuscript, we interrogate the feedback between ice sliding velocity and the subglacial drainage system and compare the results for Earth and Mars. The Martian parameter space is ill defined in comparison with the terrestrial, in particular because any wet-based glaciation likely occurred hundreds to thousands of million years ago. To test the robustness of our results because of a range of possible mass balance scenarios, we consider a variation of ice thickness and ice surface slope. Variability in temperatures capture the range of Martian surface temperatures and geothermal heat flow scenarios. Finally, meltwater availability and basal input are captured through variations in the discharge and hydraulic gradient. Changes in ice thickness and hydraulic gradient inform the variability in ice effective pressure, in turn.

We consider the effect of varying conditions (including ice thickness, water availability, ice temperature, and topography) on our sliding velocity results. We take as a benchmark the combination $Q=100 \text{ m}^3/\text{s}$, $T=270 \text{ K}$, $S_i=0.002$, $H=1500 \text{ m}$, $X_s = 10 \text{ m}^2$, and explore the following ranges, for Earth and Mars:

- Ice thickness: $H = 500 - 2000 \text{ m}$ (figure S8).
- Ice surface slope: $S_i = 0.001 - 0.007$ (figure S9).
- Ice temperature $T = 233 - 273 \text{ K}$ (figure S10).
- Meltwater discharge $Q = 10 - 1000 \text{ m}^3/\text{s}$ (figure S11).

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Figure S1. Comparative equilibrium surface topography of an ice sheet spanning 500 km on Earth (blue) and Mars (red) considering an equal and spatially homogeneous mass balance of $a=5 \text{ mm/yr}$ and all ablation occurring due to calving.
Figure S2. Mass balance for a characteristic terrestrial ice sheet (left side) and a Martian ice sheet (right side). Upper panels show a cartoon with the Equilibrium Line Altitudes (ELA) highlighted with an arrow. Blue stars show snow accumulation, blue arrows represent melt, and the smoky symbols represent sublimation. Dashed black line represents ice flux. Bottom panels are adapted from figure 5 in (Fastook et al., 2008) and represent models for a terrestrial (left) and Martian (right) ice sheet, and include the effects of precipitation, melting, and sublimation as extracted from a climate model adapted to the Martian conditions (Fastook et al., 2008)
**Figure S3.** Effect of varying ice temperature on the equilibrium shape of Martian (red) and Terrestrial (blue) ice sheet surfaces, from 273 K (left) to 243 K (right).

**Table S1.** Reference parameters (orders of magnitude)

| variable | description                        | value  | units   |
|----------|------------------------------------|--------|---------|
| $G_M$    | geothermal heat flow               | 0.05   | W/m²    |
| $\rho$   | ice density                        | 917    | kg/m³   |
| $L$      | latent heat of ice fusion          | 334000 | J/kg    |
| $u_s$    | sliding velocity                   | $3 \times 10^{-11}$ | m/s     |
| $H$      | ice thickness                      | 2000   | m       |
| $\tau_b$ | driving stress                     | 7000   | Pa      |
| $\Delta T$ | base to surface temperature difference | -50   | K       |
| $\kappa_i$ | thermal conductivity of ice        | 2.3    | W/mK    |
Figure S4. Cartoon showing the dynamics of opening and closure of a subglacial channel.

![Diagram of subglacial channel dynamics](image)

- **Ice creep closure**
- **Turbulent heat dissipation**
- **Wall melt**

**Cavity growth**

**Cavity closure**

Figure S5. Cartoon showing the dynamics of opening and closure of a subglacial cavity. Sliding velocity ($u_s$) increases with the length of the arrow. Ice creep closure is represented with downward pointing black arrows.
Figure S6. Shutdown timescale of Earth (blue line) and Mars (red line) subglacial channels showing the duration of the winter season in terrestrial months (taken to be 3 months for Earth and 12.8 months for Mars). Vertical axis are the duration a channel is open carrying the critical discharge in Earth months, and horizontal axis indicate the ratio of final to initial cross-section. The projection assumes 1500 m thick ice, an ice temperature of 265 K, and a critical discharge of $Q_c = 0.003 \text{ m}^3/\text{s}$. 
Figure S7. Drainage evolution throughout a year cycle of the subglacial drainage on Earth (upper row) and Mars (bottom row), from the late summer season into the early summer season of the following year. Note that a Martian year (687 terrestrial days) lasts more than a terrestrial year (365 terrestrial days).
Figure S8. Sensitivity of critical discharge (left panels) and sliding velocity (right panels) to a range of ice sheet thickness $H = 500$–$2000$ m. Earth curves are shown in blue, and Mars curves in red. Cv and Ch arrows mark the onset of cavities and channels, respectively.
Figure S9. Sensitivity of critical discharge (left panels) and sliding velocity (right panels) to a range of ice surface slopes $S_i = 0.001 - 0.007$. Earth curves are shown in blue, and Mars curves in red. Cv and Ch arrows mark the onset of cavities and channels, respectively.
Figure S10. Sensitivity of critical discharge (left panels) and sliding velocity (right panels) to a range of ice temperatures $T = 233–273$ K. Earth curves are shown in blue, and Mars curves in red. Cv and Ch arrows mark the onset of cavities and channels, respectively.
Figure S11. Sensitivity of critical discharge (left panels) and sliding velocity (right panels) to a range of basal water discharges $Q = 10^{-10} - 1000$ m$^3$/s. Earth curves are shown in blue, and Mars curves in red. Cv and Ch arrows mark the onset of cavities and channels, respectively.