The Pace of Fluvial Meanders on Mars and Implications for the Western Delta Deposits of Jezero Crater

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Abstract Although there is little doubt that rivers once flowed on Mars' surface, how sustained and frequent their flows were remains enigmatic. Understanding the hydrology of early Mars, nonetheless, is a prerequisite to resolving the planet's climate history and the astrobiological potential of various ancient putative ecosystems. In 2021, NASA's Perseverance rover will attempt to land near ancient fluviodeltaic deposits in Jezero crater. Deltas offer enhanced organic-matter burial and preservation on Earth but translating this notion to early Martian environments remains speculative in the absence of information on flow intermittency and sedimentation rates. Here we develop a new model to infer the lateral migration rate of Martian river meanders, which, combined with orbiter-based observations of the fluviodeltaic deposits at Jezero crater, allows us to determine a minimum timescale for the formation of its delta. We then independently constrain the total duration of delta formation, including dry spells. Our best estimates suggest that delta formation spanned ~19–37 years over a total duration of ~380,000 years, i.e., that rivers flowed for a minimum ~1 sol/15–30 Martian years and conceivably more frequently, but uncertainties on total duration are large. Despite a possibly arid climate, predicted sedimentation rates are high, suggesting a rapid burial of putative organics in distal deposits. Altogether, our results support Jezero crater's potential as a prime target to look for ancient Martian life and acquire samples to return to Earth. Any discrepancies between our predictions of the deposits' grain-to-bedform-scale architecture and future rover observations will shed critical light onto Mars' early surface environments.

Plain Language Summary Rivers once flowed on Mars, but how often, and for how long? Answering these questions will increase our understanding of Mars' habitability at a time when life was already evolving on Earth. NASA's Perseverance rover will land by the remnants of an ancient river delta in Jezero crater. Here we develop a new model to calculate the pace of shifting Martian rivers, which, when applied to orbital observations of the Jezero delta, allows us to determine a minimum duration for delta formation. Combined with an independent estimate for the total duration of delta formation (including dry spells), our results suggest that the delta took a few decades to form over a total timespan of, most likely, hundreds of thousands of years. This result suggests that Mars was likely arid at the time, with rivers flowing for at least 1 Martian day every 15–30 Martian years, and possibly more often. Nonetheless, we predict that sediments would have been buried quickly in the delta, favoring the long-term preservation of possible organic matter. Altogether, our results confirm that Jezero crater is a prime location to understand Mars' early climate, look for traces of ancient Martian life, and return samples from for further analysis on Earth.

1. Introduction

A diverse suite of geomorphic and sedimentary evidence at the surface of Mars unambiguously demonstrates that the planet had an active sedimentary cycle involving liquid water in its early history (e.g., McLennan et al., 2019). Open lake basins fed by valley networks (e.g., Fassett & Head, 2008a) and delta deposits (e.g., Di Achille & Hynek, 2010; Fassett & Head, 2005), in particular, demonstrate a clear source-to-sink parallel between terrestrial and Martian fluvial systems. However, given a lower solar flux and no plausible, efficient, and relatively stable greenhouse effect on Mars, this geologic evidence has been a challenge to explain for most state-of-the-art climate models (e.g., Wordsworth, 2016). A recent 1-D climate model that incorporates ice-albedo feedback, cirrus-cloud warming, and updated CO₂-H₂O and CO₂-CH₄ collision-induced absorption parametrization suggests that early Mars might have been arid but warm, a scenario under which surface runoff would have resulted from rainfall (Ramirez, 2017; Ramirez & Craddock, 2018). However, 3-D...
models that incorporate all of the above effects cannot sustain a warm atmosphere on geologically relevant timescales (e.g., Wordsworth et al., 2015), leaving transient warming of the early Martian atmosphere as a possible solution to reconcile model predictions with geological evidence (e.g., Halevy & Head, 2014; Wordsworth et al., 2017). In such a scenario, the surface of early Mars would have been otherwise cold and arid, with intermittent episodes of active surface hydrologic cycling driven by either precipitation or ice melting (e.g., Bishop et al., 2018; Kite, 2019; Palumbo et al., 2018). The climate of early Mars thus remains a major unsolved mystery in planetary science (Wordsworth et al., 2018)—one that bears critical implications for the planet’s habitability at a time when life was most likely already evolving on Earth (e.g., Ohtomo et al., 2014; Tashiro et al., 2017). Solving the Martian climate puzzle will require not only the integration of orbiter-based data sets with climate models, but also the acquisition of more in situ data by rovers, as well as the return of Martian rock samples to Earth for detailed analyses (e.g., Ehlmann et al., 2016).

With compelling evidence for surface flows (e.g., Fassett & Head, 2005; Goudge et al., 2017, 2018; Schon et al., 2012), a rich mineralogic diversity (Goudge et al., 2015; Horgan et al., 2020; Salvatore et al., 2018; Tarnas et al., 2019), and the promise of astrobiologically relevant samples (International Mars Sample Return Objectives and Samples Team (iMOST), 2019; McMahon et al., 2018), the western delta deposit of Jezero crater (hereafter referred to as the Jezero delta; Figure 1) was chosen as the landing site for NASA’s Perseverance rover, which is expected to launch in 2020 and serves as the first phase of a Mars sample-return mission (Grant et al., 2018; iMOST, 2019). Although the Jezero delta is thought to have formed when a river flowed into a crater lake sometime around the Late Noachian/Early Hesperian boundary (e.g., Schon et al., 2012), its age remains uncertain, and the duration of surface flows that formed the delta is poorly constrained. Whereas crater counting can help with dating individual surfaces (e.g., Hartmann & Neukum, 2001), it takes a very specific and fortuitous set of circumstances, as well as favorable stratigraphic relationships, to obtain the age or duration of geomorphic events from crater-retention ages alone (e.g., Fassett & Head, 2005; Kite et al., 2013; Palucis et al., 2020). In contrast, landed spacecraft may directly determine the absolute and exposure ages of some stratigraphic units or events (e.g., Farley et al., 2014; Martin et al., 2017), but very few sites have been investigated in situ on Mars to date. Finally, both orbiter-based and in situ observations of the sedimentary record may help to assess the duration of specific geomorphic and sedimentary events (e.g., Buhler et al., 2014; Grotzinger et al., 2015; Kleinhans, 2005), provided that one has robust mechanistic constraints on the rates of geological processes.

By combining an empirical relationship between the width and discharge of terrestrial channels that was further empirically calibrated for Mars-like conditions by Irwin et al. (2005), Fassett and Head (2005) estimated that it would have taken ~10–20 years to fill Jezero crater and breach its rim in a single, continuous event; however, this technique does not take into account possible water loss by infiltration or evaporation, and it cannot determine whether one or multiple episodes of surface runoff occurred. In order to account for the intermittency of formative flow events, Fassett and Head (2005) assumed an Earth-like hydrograph and multiplied their lower-bound duration estimate by 100, bringing their total estimated duration of surface runoff to thousands of years. Schon et al. (2012) argued that, assuming lateral migration rates that are representative of terrestrial vegetated rivers, individual scroll-bar sets at Jezero crater would have required 20–40 years to form. These estimates, however, assume an Earth-like hydrograph as well as ignore the effects of overbank vegetation and gravity. Further combining Earth-like lacustrine sedimentation rates with an inferred crater-fill depth, Schon et al. (2012) estimated that the Jezero lake accumulated sediments over $10^5$–$10^6$ years in total. However, in addition to assuming Earth-like rates of sedimentation, these inferences ignore the possible deposition (whether pre-, syn-, or post-delta formation) of other materials such as eolian sediments, airfall, or even lava flows (e.g., the dark-toned mafic floor unit; Shahrazad et al., 2019) that likely contributed to the total crater-infill thickness. Our ability to place the rocks of Jezero crater in proper temporal and environmental contexts will ultimately be key to interpreting any returned samples. Specifically, the delta deposits offer a unique opportunity to assess the timing and duration of wet climate intervals on Mars.

Here we utilize the stratigraphy of the Jezero delta (Goudge et al., 2018) and combine it with a new empirical relationship between the lateral migration rate of single-thread rivers and channel width
(Ielpi & Lapôtre, 2020) to infer the duration of surface-runoff events that deposited the sediments. We combine the empirical relationship of Ielpi and Lapôtre (2020), corrected for climate and scaled to Mars gravity, with orbiter-based measurements of river-deposit stratigraphy to estimate sediment aggradation rates and determine the minimum duration of delta formation (i.e., the time it would have taken to form the Jezero delta in a single, continuous sediment-transport event). We then independently estimate the total timespan over which the Jezero delta formed, including dry spells, using the technique of Buhler et al. (2014). Altogether, minimum and total durations of delta formation inform us about the intermittency of delta-forming hydrologic conditions at Jezero crater. Finally, we combine sediment volumes with estimated delta-formation duration to assess sediment fluxes, and we use the latter to make predictions of the bedform-scale to grain-scale architecture of the fluviodeltaic deposits. Our results have direct implications for the exploration of the Jezero delta by the Perseverance rover and its quest for biogenic signatures; in turn, future in situ stratigraphic observations will allow us to refine our estimates of delta-formation duration and intermittency. Furthermore, comparisons between our model predictions and detailed sedimentological data will serve as direct tests for different climate scenarios for early Mars.

Figure 1. (a) Context map of Jezero crater and catchment of the western Jezero delta. The feeder valley to the Jezero delta is highlighted in blue. Cross-sections V1–V6 along the feeder valley are shown in Figure S2b. (b) Close-up view of the western Jezero delta deposits. The base map for both panels is the global CTX mosaic of Dickson et al. (2018).
2. Lateral Migration Rate of Single-Thread Rivers on Earth and Mars

2.1. Lateral Migration of Terrestrial Single-Thread Rivers

Alluvial meanders arise when channel curvature, flow inertia, and scour lead to the development of a lateral pressure gradient across channel bends (Bathurst et al., 1977). Combined with the downstream nonrotational flow component, cross-stream flow induced by lateral pressure gradients causes the formation of a helical flow that erodes the channel's outer bank and deposits sediments along its inner bank (Bluck, 1971; Dietrich et al., 1979; Jackson, 1975, 1976). This balance of erosion and deposition leads to the lateral migration of the channel at a rate that generally increases with channel size and bend curvature (e.g., Braudrick et al., 2009; Hickin & Nanson, 1984; Howard & Knutson, 1984; Sylvester et al., 2019). Following the concepts of Partheniades (1965) and Partheniades and Paaswell (1970) for the fluvial erosion of cohesive soils, quantitative models of meander lateral migration typically relate the channel lateral migration rate to the near-bank excess flow velocity and bank erodibility (e.g., Hasegawa, 1977; Howard, 1992; Howard & Knutson, 1984; Ikeda et al., 1981; Johannesson & Parker, 1989). Bank erodibility is controlled by a complex suite of factors, including bank height (e.g., Matsubara & Howard, 2014), bank-material properties (e.g., Grabowski et al., 2011), stratigraphy (e.g., Motta et al., 2014), and vegetation (e.g., Polvi et al., 2014). Therefore, the empirical calibration of migration rates is typically performed on a case-by-case basis (e.g., Matsubara & Howard, 2014). While vegetation plays a critical role in bank stabilization (e.g., Braudrick et al., 2009; Davies & Gibling, 2010; Parker et al., 2011; Smith, 1976), recent field and modeling studies demonstrated the stability of single-thread rivers in unvegetated or sparsely vegetated basins (Ielpi, 2019; Ielpi & Lapôtre, 2019a, 2019b, 2020; Lapôtre et al., 2019; Li et al., 2015; Matsubara et al., 2015; Santos et al., 2017, 2019). Because their morphodynamics are unaffected by plant life, modern unvegetated single-thread rivers may offer particularly relevant analogs for ancient Martian single-thread rivers.

Through a global compilation of channel migration rates spanning all but polar climate zones (but including some large Arctic rivers), Ielpi and Lapôtre (2020) showed that, for a given channel width, unvegetated meanders migrate about an order of magnitude faster than their vegetated counterparts (Figure 2a). Specifically, the lateral migration rates of vegetated and unvegetated rivers ($M_{E,v}$ and $M_{E,uv}$, respectively; in m/year), averaged over several years (i.e., including periods of inactivity), are functionally correlated with channel width, $w$ (in m), through distinct power laws of similar exponents (Ielpi & Lapôtre, 2020), which we herein approximate as

$$M_{E,v} \approx (0.023 \pm 0.001)w^{0.85}, \quad (1a)$$

$$M_{E,uv} \approx (0.133 \pm 0.006)w^{0.85}. \quad (1b)$$

In equations (1a) and (1b), the exponent is forced to be 0.85 for both vegetated and unvegetated systems, and the range in prefactors is taken as the 90% confidence interval as estimated from the data of Ielpi and Lapôtre (2020). These correlations reflect that excess velocity is the primary control on bank erosion, and that width is a good proxy for flow discharge (and thus velocity and sediment flux; e.g., Prata de Moraes Frasson et al., 2019). Scatter in the data likely reflects the effect of channel curvature (e.g., Sylvester et al., 2019) and variable bank materials.

Because equations (1a) and (1b) reflect channel lateral migration rates integrated over timespans that are longer than individual migration events, they incorporate both periods of activity and inactivity, the
frequency of which is primarily controlled by basin physiography and climate. In order to estimate the lateral migration rate of Martian rivers without forcing an Earth-like basin physiography and climate regime, one needs to first correct the data set for migration intermittency and estimate an “instantaneous” migration rate that solely reflects the balance of bank stabilizing and destabilizing forces at times of active migration. To do so, we utilize the concept of migration intermittency; following the premise that most geomorphic work is done when flow discharge is bankfull, i.e., when the channel is filled to the brim with water (e.g., Blom et al., 2017; Wolman & Miller, 1960), we assume that lateral migration only occurs when bankfull conditions are reached or exceeded. We define a bankfull intermittency factor, \( I_{bf} \), as the yearly fraction of time spent under discharges exceeding bankfull conditions, such that “instantaneous” lateral migration rates, \( M_{bf, inst} \), can be estimated as

\[
M_{bf, inst} \approx \frac{M_{bf}}{I_{bf}}
\]

Estimating \( I_{bf} \) requires knowledge of river hydrographs, which are largely unavailable today for unvegetated meandering streams. However, the discharge record of one such stream—the Amargosa River draining into Death Valley, California—is available through one gauging station that recorded near-continuous data since 1962 at Tecopa, California. In the absence of a more inclusive and geographically varied database, the discharge record of the Amargosa River is herein considered as overall representative of other arid rivers in the unvegetated-river compilation of Ielpi and Lapôtre (2020). Bankfull discharge for the Tecopa gauging site was estimated from field observations as described in Ielpi and Lapôtre (2020).

The data collected at the Tecopa gauging station include peak annual discharge and mean daily discharge, which allow us to estimate lower and upper bounds on \( I_{bf} \). A lower bound is estimated under the assumption that channels migrate for a full day on days when peak daily discharge is at or greater than bankfull. To estimate peak daily discharge from mean daily and annual peak discharges, we first identify the day when annual peak discharge was reached in any given year, and then divide the mean daily discharge from that day by the peak annual discharge. This ratio serves as an indicator of the relationship between mean and peak daily discharges, and peak daily discharge is estimated by dividing mean daily data by this ratio for every day on record. A conservative upper bound, in contrast, is estimated under the assumption that channels only migrate when mean daily discharge is at or greater than bankfull discharge. Our estimated “instantaneous” lateral migration rate for terrestrial unvegetated rivers accounts for uncertainties related to the bankfull discharge at the Tecopa gauging station, the frequency of flood events exceeding bankfull conditions, and scatter in the functional relationship between \( M_{bf} \) and \( w \). In the following, our best estimate for channel lateral migration rate is calculated from the mean of our lower and upper bounds on \( I_{bf} \), but uncertainties are propagated throughout. A final step to determine the lateral migration rate of ancient Martian rivers is to scale our estimates for the “instantaneous” lateral migration rates of terrestrial channels to Martian gravity.

### 2.2. Single-Thread Rivers on Mars: Scaling for Martian Gravity

Martian sinuous ridges have long been interpreted as remnants of meandering channel belts (e.g., Burr et al., 2006, 2010; Cardenas et al., 2018; DiBiase et al., 2013; Kite et al., 2015; Williams, Irwin, et al., 2013). Although adjoining floodplain materials appear to have been eroded, as commonly observed in inverted fluvial landscapes on Earth (e.g., Hayden et al., 2019), river flows were most likely confined by banks that contained fine-grained materials (e.g., Lapôtre et al., 2019). Clay minerals have been detected globally on Mars by orbiting spectrometers (e.g., Ehlmann & Edwards, 2014; Mustard et al., 2008; Poulet et al., 2005), including within delta deposits (e.g., Ehlmann et al., 2008; Goudge et al., 2015). Thus, there is little doubt that clays played a significant role in stabilizing riverbanks where they are associated with delta deposits (e.g., DiBiase et al., 2013; Goudge et al., 2017, 2018), and we herein utilize modern terrestrial unvegetated rivers, which form within mud-rich endorheic basins today, as analogs for ancient Martian single-thread rivers.

Following the model of Hasegawa (1977) and Ikeda et al. (1981), we consider that bank migration rate is proportional to the near-bank excess velocity, which can also be rewritten in terms of a near-bank shear stress, \( \tau_b \) (e.g., Howard, 1992), as
where $E$ is bank erodibility and $\tau$ is the average boundary shear stress. The near-bank shear stress ($\tau_b$) and the average boundary shear stress ($\tau$) are both expected to vary linearly with gravity ($g$), such that the effect of $g$ cancels out when dividing the near-bank excess shear stress by the average boundary shear stress. However, bank erodibility ($E$) may depend on $g$. Specifically, $E$ is expected to be controlled by the rate-limiting erosional process (Howard, 1992)—mainly, sediment transport of readily available loose sediment (transport-limited scenario), hydrodynamic erosion (e.g., Ternat et al., 2008), and mass wasting processes (e.g., bank failure). In all three scenarios, we expect bank erodibility to be controlled by sediment removal by water (mass wasting is virtually “instantaneous,” such that even in that scenario, bank migration is controlled by the removal of slumped materials when banks are actively migrating). Thus, we posit that

$$E \propto \sqrt{g},$$

as is the case in all three rate-limiting scenarios (e.g., Meyer-Peter & Muller, 1948; Ternat et al., 2008). Finally, channel width is not expected to exert any major control on $E$, such that the 0.85 exponent in equations (1a) and (1b) should remain unchanged under Martian conditions. Thus, the “instantaneous” lateral migration rate of Martian channels, $Mr_{\text{M,inst}}$, is estimated through

$$Mr_{\text{M,inst}} = \sqrt{\frac{g_M}{g_E}} Mr_{E,\text{inst}},$$

where $g_M \approx 3.71 \text{ m/s}^2$ and $g_E \approx 9.81 \text{ m/s}^2$ are Mars’ and Earth’s surface accelerations of gravity, respectively. Following this procedure, our best estimate for the “instantaneous” lateral migration rate of Martian channels is given by

$$Mr_{\text{M,inst}} \approx 9.81^{+10.28}_{-3.15} \times 10^{0.85}.$$

Figure 2b shows our best estimate along with our absolute minimum and maximum constraints, including a full propagation of uncertainty. For example, our best estimate predicts that a 45-m wide meandering channel on Mars bounded by mud-rich banks would have migrated at an “instantaneous” rate of ~250 m/year (where a year here refers to the orbital period of Earth).

### 3. Timescales of Hydrological Activity at Jezero Crater

With a model for lateral migration rate of Martian channels on hand, we next utilize the stratigraphy of the Jezero delta to estimate the timescales associated with surface-flow events during delta formation.

#### 3.1. Minimum Duration for the Formation of the Western Jezero Delta

A minimum estimate for the duration of delta formation (or the cumulative duration of delta-forming sediment-transport events), $T_c$, is obtained from the total thickness of channel deposits, $z_{\text{dep}}$, and the bed aggradation rate, $\dot{A}r$, as

$$T_c = \frac{z_{\text{dep}}}{\dot{A}r \varepsilon},$$

where $\varepsilon = \frac{z_{\text{comp}}}{z_{\text{comp}}}$ is a stratigraphic-completeness factor and $z_{\text{comp}}$ is the thickness of a putative equivalent stratigraphically complete deposit. The stratigraphic-completeness factor accounts for the occurrence of more localized or short-lived events that are less prone to be represented in the stratigraphic record (e.g., Durkin et al., 2017; Sadler, 1981). A stratigraphic completeness of 1 implies that all deposits get preserved. Typically, meander belts on Earth display $\varepsilon \approx \frac{1}{4}$ to $\frac{1}{2}$, and factors decreasing stratigraphic completeness include downstream translation of meanders, bar rotation, higher sinuosity, and increased incidence of avulsion and cutoff (Durkin et al., 2017). To calculate $T_c$, we first need to estimate $\dot{A}r$ from a simple set of measurable parameters.
Taking advantage of the physical control of aggradation rate on the channel avulsion timescale (e.g., Ganti et al., 2016; Jerolmack & Mohrig, 2007), the channel-avulsion timescale, \( T_a \), may be expressed as

\[
T_a = \frac{h}{A_p}
\]

(8a)

where \( h \) is the average channel depth and \( h^* \) is a dimensionless constant that describes the fraction of \( h \) that needs to be filled with sediments to trigger an avulsion. Equation (8a) may then be rewritten to solve for aggradation rate,

\[
A_p = \frac{h^* h}{T_a}
\]

(8b)

The avulsion timescale, in turn, may be estimated from the width of channel-body deposits, \( W_d \) (Jerolmack & Mohrig, 2007; Mohrig et al., 2000). Sediments accumulate along meanders’ inner banks in the form of point bars as the channel migrates laterally (Figures 3a–3c) until avulsion (Figure 3d) or cutoff occurs. Because of the moderate sinuosity of the channels at Jezero crater, avulsions likely occurred more frequently than cutoffs, such that avulsion was likely the dominant channel-body-width-limiting process. The stratigraphic analysis of Goudge et al. (2018) determined that the median channel-body width atop the Jezero delta is ~67% larger than the median channel width (~75.5 vs. ~45.3 m, respectively), indicating that lateral migration typically did occur between avulsions. Thus, we assume that the width of channel bodies, \( W_d \), is determined by the distance the channel migrated laterally over the avulsion timescale, i.e.,

\[
W_d = T_a M_{M, \text{inst}}
\]

(9a)

which may be rearranged to solve for \( T_a \) into

\[
T_a = \frac{W_d}{M_{M, \text{inst}}}
\]

(9b)

Combining equations (7)–(9b), we obtain

\[
T_c = \frac{z_{\text{dep}} W_d}{\epsilon h h M_{M, \text{inst}}}
\]

(10)

3.2. Flow Intermittency and Total Duration of Delta Formation

We define the intermittency of delta-forming sediment-transport events, \( I \), as

\[
I = \frac{T_i}{T_{\text{tot}}}
\]

(11)

where \( T_{\text{tot}} \) is the total duration of delta formation, including dry spells. Surface flows around the Noachian/Hesperian boundary, i.e., when the Jezero delta is thought to have formed, were inferred to have been intermittent from the geomorphic record (Buhler et al., 2014). For example, Buhler et al. (2014) estimated that fluvial activity only occurred ~0.01–0.1% of the time for \( 10^5 \)–\( 10^6 \) years near Paraná Valles, over 5,500 km away from Jezero crater. To constrain the episodically of delta-forming sediment-transport events at Jezero crater proper, we estimate the total delta-formation duration, \( T_{\text{tot}} \), as
\[ T_{\text{tot}} = \frac{V_d - V_v}{A_d E_{N/H}} \]  

(12)

where \( V_d \) is the total volume of the delta deposits, \( V_v \) is the volume of the paleoriver-valley that feeds the delta (blue lines in Figure 1a), \( A_d \) is the drainage area of the Jezero catchment as measured from the delta apex, and \( E_{N/H} \) is an independently determined Noachian/Hesperian denudation rate. This method, as proposed by Buhler et al. (2014), serves as an order-of-magnitude estimate. The difference between the delta and feeder valley volumes represents the total volume of rock that was eroded within the Jezero catchment, which divided by \( A_d \), provides an estimate of the catchment-averaged denudation thickness during delta formation. Further assuming that denudation on the catchment occurred at a rate of \( E_{N/H} \), \( T_{\text{tot}} \) may be estimated using equation (12). Estimates of \( V_d \) and \( V_v \) from modern topography and inferred stratigraphy are reported in supporting information Text S1 and Figures S1-S2.

4. Distal and Proximal Deposits of the Western Jezero Delta

Building on our estimate of delta-formation duration, we can determine the average total sediment flux (including both bed and suspended loads) to the Jezero crater lake. Combined with independent estimates of channel paleoslope, total sediment flux allows us to constrain the median grain size of channel and bank sediments, the bed configuration, bedload and suspended sediment fluxes, and suspended sediment concentration.

4.1. Total Sediment Flux

The total sediment flux fed to the delta per unit channel width, \( q_t \), may be estimated by dividing the total volume of deposited sediment by the width of the feeder channel, \( w_f \), and by the minimum formation duration, \( T_c \):

\[ q_t = \frac{Q_t}{w_f} = \frac{(1 - \phi)V_d}{w_f T_c}. \]  

(13)

where \( Q_t \) is the total sediment flux and \( \phi \) is an average porosity of the delta deposits, such that \((1 - \phi)V_d\) provides an estimate of the solid volume occupied by sediment grains. Assuming that sediments transported in washload settled in distal lacustrine deposits further downstream in the crater and are not volumetrically significant within the delta fan, \( q_t \) as estimated by equation (13) encompasses sediments that were transported in bedload and suspended load.

4.2. Grain Sizes and Bed Paleoslope

4.2.1. Grain Sizes of Bed Materials

According to Engelund and Hansen (1967), \( q_t \) can be estimated through

\[ q_t = 0.05 \sqrt[5]{\frac{R g d_{50}^3}{C_f \tau^*}}, \]  

(14)

where \( C_f \) is a dimensionless bed friction coefficient, \( R = R_s = \frac{\rho_s - \rho}{\rho} \) is the submerged specific gravity of the sediment (where \( \rho_s \) and \( \rho \) are the sediment and water densities, respectively), \( d_{50} \) is the median grain size in the channel (excluding washload), and \( \tau^* = \frac{\tau}{(\rho_s - \rho)g d_{50}} \) is the Shields stress. Equation (14) can be rearranged to express the average boundary shear stress, \( \tau \), as a function of the total sediment flux per unit width, \( q_t \), as

\[ \tau = \left[ \left( \frac{C_f q_t d_{50}}{0.05} \right)^2 \rho_s (\rho_s - \rho)g \right]^{\frac{1}{5}}. \]  

(15)

Concurrently, the boundary shear stress, \( \tau \), may be estimated under the assumption of steady uniform flow through \( \tau = \rho g h S \), where \( S \) is the channel paleoslope. Lapôtre et al. (2019) showed that the effective bank grain size, \( d_{\text{bank}} \), may be determined from \( S \), channel width, \( w \), and channel depth, \( h \). Thus, provided that channel paleoslope and bed friction coefficient can be determined independently (section 4.2.2 and Texts...
S2–S3), one may solve for an effective \( d_{\text{bank}} \) using the model of Lapôtre et al. (2019) (section 4.2.2), and for \( d_{50} \) by equating equation (15) to \( \rho ghS \) and rearranging into

\[
d_{50} = \frac{0.05}{q_{C_t} (d_{50}, \, h, \, S)} \sqrt{\frac{ghS^2}{R^2}}.
\]

Because \( C_t \) varies with \( d_{50} \) and bed configuration, equation (16) needs to be solved iteratively. To this end, we first constrain \( S \) to calculate \( \tau \) (section 4.2.2). From \( \tau \), we determine the bed configuration for a range of \( d_{50} \) values and estimate associated bedform dimensions (Text S3). For each \( d_{50} \) value, we then calculate \( C_t \) (Text S3) and find the value of \( d_{50} \) that satisfies equation (16). A similar calculation was performed using the generalized Engelund-Hansen equation of Ma et al. (2019) as a sensitivity analysis (Text S4). Results are virtually unchanged.

### 4.2.2. Channel Bed Paleoslope

Based on the premise that, in average over both banks, single-thread rivers adjust their width such that the shear stress imparted by the flow on the riverbanks is equal to the erosion threshold stress associated with the bank materials (e.g., Dunne & Jerolmack, 2018), and using a first-principle model for the cohesion of fine sediments (Ternat et al., 2008), Lapôtre et al. (2019) presented a model to relate the equilibrium bed slope of a river to \( d_{\text{bank}} \) for given channel width and depth. Here we apply the model of Lapôtre et al. (2019) to the Jezero channels, utilizing the channel width and depth measurements of Goudge et al. (2018). Because both \( S \) and \( d_{\text{bank}} \) are unknown and correlated, more independent constraints are required to narrow down the range of possible \( S \) and \( d_{\text{bank}} \) values. First, Goudge et al. (2018) estimated an upper bound on channel paleoslope by measuring the slope of best-fit planes to channel-body tops. These measurements represent conservative upper bounds on paleoslope because the erosional surface exposes discrete channel bodies and is thus steeper than the channel-body stratigraphy (Goudge et al., 2018). We use the 10th percentile of the measured slopes as an upper bound for \( S \). Second, it was empirically shown that terrestrial rivers universally follow a relationship between Shields stress, \( \tau^* = \frac{K_{\text{crit}}}{d_{50}} \), and particle Reynolds number, \( Re_p = \sqrt{\frac{gd_{50}^3}{v}} \) (where \( \nu \) is the kinematic viscosity of water) of the form \( \tau^* \propto Re_p^{1/2} \) (e.g., Trampush et al., 2014; Wilkerson & Parker, 2011). Although this relationship was largely derived from a compilation of vegetated rivers and utilizes bed grain size in the calculation of \( Re_p \), Lapôtre et al. (2019) suggested that a similar relationship may hold for single-thread rivers forming within mud-rich unvegetated banks when using bank-specific particle Reynolds number and Shields stress. In addition, Martian rivers are expected to have been steeper than terrestrial rivers of similar width because of the lower acceleration of gravity at the Martian surface (e.g., Konsoer et al., 2018; Lapôtre et al., 2019). Thus, we use the empirical relationship of Trampush et al. (2014) to estimate a conservative minimum bound on paleoslope.

### 4.3. Sediment Loads

Flows with high sediment concentrations occur during high discharge events when a large volume of fine sediment is available for entrainment. This phenomenon is amplified when water flow has long recurrence intervals, allowing fine sediment to accumulate in the landscape for prolonged timespans between transport episodes (e.g., Cannon et al., 2010; Eaton, 1935; Wells, 1987). Thus, suspended-sediment concentration may provide another clue to understand the hydrologic regime under which the Jezero delta formed, with relatively high suspended-sediment concentration potentially hinting at highly intermittent flows separated by prolonged periods of sediment accumulation in the landscape. The equations used to calculate sediment fluxes are presented in Text S4.

### 5. Parameter Values and Results

#### 5.1. "Instantaneous" Lateral Migration Rate of River Channels Over the Jezero Delta

Based on detailed mapping of exposed fluvial deposits and geometric considerations, Goudge et al. (2018) estimated the width of paleochannels (\( w \)) at the Jezero delta. This technique, notably, provides a more reliable estimate of paleochannel width than measurements of inverted-channel-body widths (as the latter represent amalgamated and subsequently eroded channel belts rather than pristine individual channels; Hayden et al., 2019). The data of Goudge et al. (2018) constrain the median paleochannel width atop the
Jezero delta to be ~45.3 m, with a 10th and 90th percentiles of 18.7 and 148.1 m, respectively. Combining these measured channel widths with our estimates for the lateral migration rate of Martian channels, we infer that the single-threaded streams atop the Jezero delta migrated laterally at “instantaneous” rates of 80–1,406 m/year (or ~0.2–4 m/sol, where a sol is a Martian day), with a best estimate of ~250 m/year (or ~70 cm/sol; Figure 2b). Furthermore, Goudge et al. (2018) determined that the median $W_0$~75.5 m (with a 10th and 90th percentiles of ~38 and ~135 m, respectively), suggesting that avulsions typically occurred 3–4 times per year of active delta formation (equation (9b)). We emphasize that these rates only reflect times of active migration, and that net lateral migration rates, including dry spells, would have been slower. We estimate flow intermittency and net migration rates in section 5.3.

5.2. Minimum Duration of Delta Formation

Based on detailed measurements of point-bar height, Goudge et al. (2018) inferred that the median $h$~6.7 m (with a 10th and 90th percentiles of ~3.4 and ~9.4 m, respectively). Their inferred stratigraphy suggests that $z_{dep}$ ~45 m, including the thickness of inverted-channel bodies and underlying point-bar strata. Recent mapping over the Jezero delta reveals that the point-bar strata alone are more likely as thick as ~60 m (Sangwan & Gupta, 2019), such that $z_{dep}$ could be as thick as ~90 m. We thus use $z_{dep} = 45 – 90$ m, with a best estimate of ~90 m. The flume experiments of Ganti et al. (2016) revealed that $h^*$ is tightly distributed around a median value of ~0.22 (with a 10th and 90th percentile of ~0.15 and ~0.46, respectively). Finally, we illustrate the effect of stratigraphic completeness by discussing the results for $\varepsilon = 1$ (complete section) and $\varepsilon = 0.5 \pm 0.17$ (Earth-like preservation; Durkin et al., 2017). Using these values as inputs to equation (10), our best estimate for the minimum formation duration of the Jezer delta is $T_c \approx 19^{+14}_{-3}$ years if the deposit is stratigraphically complete, or $T_c \approx 37^{+46}_{-30}$ years for an Earth-like completeness.

5.3. Total Delta Formation Duration and Intermittency

From our paleotopographic reconstruction (Text S1), we find $V_d = 22.5^{+3.9}_{-3.9}$ km$^3$ and $V_v = 13.0^{+3.6}_{-3.6}$ km$^3$. As expected, the reconstructed delta volume is greater than the volume of modern deposits exposed above the crater floor (~10 km$^3$) since we accounted for deposits that are possibly buried under the crater-floor unit as well as denudation and backwasting of the delta top and toe; similarly, our reconstructed inlet valley volume is smaller than its present value (~55–60 km$^3$; Fassett & Head, 2005), as we accounted for widening through erosion and possible infilling over the billions of years since abandonment. Used as inputs to equation (12), our inferred volumes yield a best estimate for the total delta-formation duration of $T_{tot} \approx 380,000$ years. Propagating our uncertainty from our paleotopographic reconstructions, we estimate a minimum and maximum delta-formation durations of $T_{tot} \approx 2,200$ years and $T_{tot} \approx 3,800,000$ years, respectively. This large range in total durations reflects our attempt at considering rather extreme minimum and maximum bounds on several parameters to acknowledge sources of uncertainty, but our results do favor a total formation time of several hundreds of thousands of years. Using our estimates of $T_c$ and $T_{tot}$ in equation (11), we find an intermittency of delta-forming flows at Jezero crater of $I \approx (1.2 \times 10^{-4} – 1.3)%$ with a best estimate of $I \approx 0.005%$ if the deposits are stratigraphically complete, or $I \approx (1.8 \times 10^{-4} – 3.8)%$ with a best estimate of $I \approx 0.01%$ for an Earth-like stratigraphic completeness (Figure 4). For reference, the estimated intermittencies correspond to a recurrence interval of ~1 sol/30 MY (~9 sols/MY to ~1 sol/1,250 MY, where MY denotes a Martian year) for a stratigraphically complete delta, or of ~1 sol/15 MY (~25 sols/MY to ~1 sol/830 MY) for an Earth-like stratigraphic completeness (Figure 4b). These intermittencies only reflect the frequency of delta-forming bankfull flows, whereas lower-discharge baseflows not conducive to significant geomorphic work likely were more frequent.

Our best estimates of intermittency suggest net channel lateral migration rates of ~1.2–2.5 cm/year depending on stratigraphic completeness, although uncertainties are high and net lateral migration rates could potentially have been as fast as 53 m/year. Our best estimates thus suggest that the channels atop the Jezero delta migrated ~25–50 times slower than modern vegetated rivers of similar width on Earth when averaged over the total duration of their formation, although uncertainty is large and permits faster than terrestrial net lateral migration rates.
5.4. Sediments

To estimate the grain size of fluvial sediments, we first need to estimate the volume of sediments that constitutes our total reconstructed delta volume (equation (13)). The reconstructed delta volume accounts for the effect of denudation but is not corrected for sediment compaction. In addition, a fraction of the pore space in the compacted delta volume was likely filled by diagenetic cements like those observed in the fluvialacustrine deposits of Gale crater (e.g., Grotzinger et al., 2015; Siebach et al., 2017). In equation (13), $\phi$ needs to reflect the porosity of a compacted delta without pore-filling cements. We use $\phi = 20 – 70\%$ (with a best estimate of $\phi = 45\%$; consistent with clay-to-sand-sized sediments under 0–1 km of overburden; e.g., Hantschel & Kauerauf, 2009; Sclater & Christie, 1980). Finally, we estimate all sediment fluxes at the apex of the delta within the feeder channel. Because the feeder valley width, $W_v$, only provides an upper bound on the width of the feeder channel, $W_f$, we here explore two scenarios: one where $W_f$ is equal to the maximum channel width estimated by Goudge et al. (2018) (i.e., $W_f \approx 412$ m), and one where $W_f$ is equal to their 90th percentile (i.e., $W_f \approx 148$ m), both consistent with our estimate of $W_v \approx 350 \pm 150$ m.

Next, to constrain bed shear stresses, we determine the paleoslope within the inlet channel (section 4.2.1). Our best estimate of paleoslope is shown in Figure S3. The suspension threshold provides a conservative lower-bound on paleoslope; however, the suspension threshold for bank materials does not provide a strong constraint on paleoslope. Altogether, we estimate a typical channel paleoslope atop the Jezero delta in the

![Figure 4](image-url)

**Figure 4.** (a) Duration of delta formation as a function of formation intermittency, with inferred total formation duration highlighted in gray, illustrated here under the assumption of a stratigraphically complete delta ($\varepsilon = 1$). (b) Best estimates of total formation and intermittency for a complete delta stratigraphy (circles and pink shaded area) as well as for an Earth-like stratigraphic completeness (squares and blue shaded area).
Figure 5a shows our mechanistic predictions for $\tau$ as a function of $d_{50}$ (equation (15)). Combined with our independent best estimate on channel paleoslope (gray shaded area in Figure 5a), we predict a median bed-material (including both bedload and suspended load) grain size in the silt to medium sand range, with a best estimate of $\approx 100–175$ $\mu$m (very fine to fine sand) depending on stratigraphic completeness. Using our best estimate for the median bed-material grain size as a lower bound for the median grain size of the bedload, inferred bed stresses place the channels atop the Jezero delta in the upper flow regime (Figure 5b) with best estimate configurations yielding Froude subcritical flows within the feeder channel (Text S2–S3).

Calculations of sediment concentration are also conducted for sediments within the main feeder channel. Thus, the flow depth used in the calculation must reflect conditions within the feeder channel, which was inherently deeper than the distributary channels located further downstream; to this end, we consider two cases: one where flow depth is taken as the maximum flow depth measured by Goudge et al. (2018) (i.e., $h \approx 14.8$ m), and one where it is taken as their 90th percentile (i.e., $h \approx 9.3$ m). Combining equations (14), (16), (S11), and (S12), we estimate a sediment concentration of $\approx 2–10$% by volume, or, for basaltic sediment, $\approx 50–300$ g/L under both stratigraphic-completeness scenarios.

### 6. Discussion

#### 6.1. Uncertainties and Implications of Putative Discrepancies Between Predictions and Future Observations

Our estimate of minimum formation time mainly relies on our scaling relationship between lateral migration rate and channel width, and on the assumption that the width of the generated channel body scales linearly with the avulsion timescale. Even if our gravity correction was incorrect, our estimate of the lateral migration rate would most likely remain within our reported uncertainty as long as banks were not stabilized by some other agent (e.g., ice) directly at the time of lateral migration. For example, if “instantaneous” lateral migration rate scaled linearly with gravity, lateral migration rates on Mars would be $\approx 1.6$ times smaller than we estimated, which would translate into a best estimate for the minimum formation time of $\approx 12$ years instead of $\approx 19$ years for a stratigraphically complete deposit (i.e., well within reported uncertainty). Another potential source of error comes from $W_d$; $T_c$ scales linearly with $W_d$ (equation (10)), such that if channel deposits were significantly eroded and narrowed, reported values of $T_c$ would be underestimated. We based our calculations on the full distribution of measured $W_d$ such that our results should not be biased by a single section that has been particularly eroded. Furthermore, channel sinuosity imparts a secondary control on lateral migration rate. Because the Jezero channels have relatively low sinuosity, their lateral migration rates may have been on the lower end of those estimated through equation (6). Thus, our estimates of $T_c$ should be regarded as minimum bounds.

One major unknown, however, lies in the possibility that riverbanks were frozen (e.g., Bishop et al., 2018; Cassanelli & Head, 2019), as the presence of ground ice in riverbanks, channel hydraulic geometry, and lateral migration rate are interlinked through complex feedbacks (e.g., Zheng et al., 2019). However, the nature of this relationship is not known on Mars, and it likely would have been a strong function of temperature. For example, if banks were permanently frozen, ground ice would have likely...
strenthened the banks, perhaps slowing down lateral migration if flow stresses were relatively low, requiring a longer minimum duration for the formation of the Jezero delta. However, if formative discharge was associated with thawing of ground ice (i.e., with the development of a talik as observed in permafrost-floodplain rivers on Earth), lateral migration rates may have been faster than those we predict from equation (6) (with, possibly, increased sediment input in the channel), which would further reduce \( T_c \). As future rover observations either corroborate or contradict different aspects of our predictions, arising discrepancies will help refine our assumptions. For example, rover-based observations of bedforms in channel deposits will better inform us about bed stresses, and combined with channel width estimates, may help constrain bank strength, and determine whether riverbanks were permeated with ground ice.

Finally, estimates of \( T_{tot} \) rely on delta and feeder-valley volumes, which are highly uncertain. An overestimate of the delta volume, an underestimate of the feeder-valley volume, or underestimates of either or both the Noachian/Hesperian and Hesperian/Amazonian denudation rates in our paleotopographic reconstruction would result in shorter total formation times, and thus, in a hydrologic regime involving more frequent runoff events (Figure 4b). Detailed stratigraphic work on the ground combined with in situ characterization of sediment composition may reveal erosional unconformities that cannot be resolved from orbiter data, as well as potential changes in sediment source (e.g., Siebach et al., 2017) that would hint at possibly prolonged dry spells. Combined with quantitative models such as those herein presented, the analysis of carefully chosen samples at Jezero crater may ultimately place a definitive constraint on the total duration and climate setting of fluvio-deltaic deposition.

### 6.2. Implications for the Hydrology and Climate of Early Mars

Regardless of flow intermittency and stratigraphic completeness, we find that the formation of the Jezero delta only required active surface hydrology for \( T_c \approx 5 \text{–} 85 \) years, cumulatively. Various channel-bend migration patterns that tend to decrease stratigraphic completeness (such as downstream translation and bar rotation) are observed atop the Jezero delta (Sangwan & Gupta, 2019); we thus favor our scenario for an incomplete delta stratigraphy, from which we infer \( T_c \approx 37^{+46}_{-30} \) years. This minimum formation timescale is consistently much greater than the time it would have taken to fill in the paleo-lake to its mapped upper-stand level using modern topography (e.g., Fassett & Head, 2005; Schon et al., 2012) under the range of flow discharges we estimate in the feeder channel (<1.8 year). This first constraint implies that climate models for early Mars are not required to produce surface temperatures above freezing for geologically long timescales in order to explain the formation of the Jezero delta. Even on Earth, formative bankfull conditions are only reached <~1% of the time (or about once every 1–2 years) regardless of aridity (e.g., Wolman & Leopold, 1960), and the frequency of flow events could have been even lower on Mars. Thus, our minimum timescale for the formation of the Jezero delta neither speaks to the total duration of delta formation nor to the episodicity of runoff events. Our attempt at determining the total formation duration, \( T_{tot} \), does not provide a strong constraint on intermittency. Our best estimate indicates that \( T_{tot} \approx 380,000 \) years (or an intermittency of ~1 sol/15–30 MY), but it could be as short as \( T_c \) if, e.g., our paleotopographic reconstruction underestimated \( V_d – V_c \), and as long as ~3.8 million years based on our propagated uncertainty. Although highly uncertain, these estimates are overall consistent with inferred depositional timescales for the Eberswalde delta (e.g., Irwin et al., 2015) and fluvio-lacustrine deposits at Gale crater (e.g., Grotzinger et al., 2015). It is important to note that our calculated intermittency only reflects delta-forming events such that more frequent baseflows without significant sediment transport remain possible.

Estimated sediment concentrations are modest, such that they do not require the accumulation and rapid release of large volumes of fine sediments into the channels. Our results thus indicate that either (i) flow events were relatively frequent (which we cannot rule out but is not supported by our best estimate of delta-formation intermittency) or (ii) minimal denudation occurred within the catchment during dry spells. An alternative but less likely scenario would be for flow events to have been rare but prolonged, such that the average suspended-sediment concentration over the duration of flow events was moderate. On these grounds, we favor a scenario under which little denudation occurred in the catchment during dry spells, also consistent with the absence of major erosional unconformities or upsection changes in channel-deposit types as noted by Goudge et al. (2018).
Finally, it is worth noting that our best estimate for the total formation of the Jezero delta is broadly consistent with the results of Wordsworth et al. (2017), in which they find that the emission of a percent-level pulse of CH₄ in the atmosphere would maintain a mixed CO₂-CH₄-H₂ atmospheric composition over a timescale of ~10⁵ years. Such a reduced atmospheric composition would raise the annual mean surface temperature by tens of degrees, allowing for temperatures above 273 K in some scenarios. Thus, our best estimate suggests that only one to a few pulses of percent-level CH₄ emissions would have been required to maintain an overall reducing atmosphere during the formation of the Jezero delta, and that surface temperature at Jezero crater would need to have exceeded 273 K for a minimum of ~0.005–0.01% of that duration.

6.3. Comparison Between Previous Ground Observations of Martian Sedimentary Deposits and Predictions

Prior to Curiosity's landing at Gale crater, sedimentary rocks observed by rovers mainly consisted of successions formed in eolian environments, including dune and interdune deposits (e.g., Grotzinger et al., 2005). In contrast, Curiosity observed a diverse suite of alluvial and fluvial (e.g., Edgar et al., 2018; Williams, Grotzinger, et al., 2013), lacustrine (e.g., Grotzinger et al., 2015), and eolian (e.g., Banham et al., 2018) deposits. The fluvial-lacustrine deposits observed by Curiosity are well-cemented (e.g., Grotzinger et al., 2015; Siebach et al., 2017), and contain a broad diversity of bedform stratification types, including planar beds, ripples, and dune cross-stratification (e.g., Edgar et al., 2018; Grotzinger et al., 2015), as well as variably dipping low-angle stratification that was interpreted as a possible signature of antidunes or cyclic steps formed by plunging river plumes (Stack et al., 2019). Grain sizes between fine and very coarse sand were observed in fluvial sandstones of Gale crater (e.g., Edgar et al., 2018), although cementation, dustiness, low contrast, and a pixel resolution of ~10 to a few tens of microns per pixel with the Mars Hand Lens Imager (MAHLI) make it difficult to resolve and measure very fine sand and finer grains (e.g., Siebach et al., 2017). Our lower-bound estimates of grain size in channel deposits of Jezero crater are overall similar to those observed in the fluvial deposits of Gale crater. Even though our best estimate indicates that the beds of Jezero rivers may have been dominated by upper-flow regimes (also qualitatively consistent with the analysis of Konsoer et al., 2018), these inferences are based on average properties of the fluvial deposits, and we certainly do not expect all channel deposits to consist of planar to low-angle stratification. Rather, spatial and temporal variability in bed stresses most likely led to a variety of subcritical bedforms. Given the overall low preservation of supercritical bed configurations, subcritical bedforms could even dominate the channel deposits, unless the terminations of wet episodes at Jezero crater were marked by a rapid decrease in water discharge (e.g., Fielding, 2006). Furthermore, errors in our estimates of the total sediment flux, e.g., through an overestimated delta volume or underestimated inlet channel width, would result in the prediction of a coarser bed, placing the Jezero rivers closer to the dune stability field (Figure 4).

6.4. Implications for NASA's 2020 Investigation of the Western Jezero Delta

As a Noachian-Hesperian clay-rich fluvial-lacustrine silicilastic system, the Jezero delta deposits represent a promising astrobiological target (McMahon et al., 2018), provided that putative Martian life colonized surface or near-subsurface environments. Contingent that putative life survived possibly prolonged dry spells and that organics were protected from galactic cosmic rays by a thicker atmosphere throughout the formation of the Jezero delta (Fox et al., 2019; McMahon et al., 2018), our findings corroborate the hypothesis that the distal deposits of Jezero may have preserved evidence of past Martian life. Our estimated “instantaneous” lateral migration rates for the Jezero rivers imply an “instantaneous” bed aggradation rate of $Ar \approx 5^{+0.5}_{-0.3} \text{ m/year}$, or $Ar \approx 1.5^{+0.8}_{-0.5} \text{ cm/sol}$. Although those inferred rates only strictly apply to the channel deposits, our predicted average bedload flux is <1% of predicted suspended-sediment flux, such that burial rate of putative organics by fines in the distal part of the delta was likely very rapid. Such a rapid burial would have promoted accelerated compaction, creating a closed chemical environment with low permeability that inhibits diffusion and promotes rapid mineralization (e.g., Farmer & Des Marais, 1999; McMahon et al., 2018). Furthermore, clays like those detected from orbit in distal deposits (Ehlmann et al., 2008) were shown to absorb organics into their mineral structure (e.g., Farmer & Des Marais, 1999). Altogether, mineralogy and our inferred rapid burial strongly support the preservation potential of the Jezero deposits, making them a prime target to
search for traces of putative ancient Martian organisms. Given the presented relationships between channel width, migration rates, aggradation rates, and frequency of avulsions, rapid burial was more likely to occur along the inner depositional banks or at the frontal, distal terminations of wider channels that were abandoned ensuing avulsion, providing a clear strategy for sampling of the Jezero fluvio-deltaic deposits.

7. Conclusions

NASA’s Perseverance rover will be exploring fluvio-deltaic deposits at Jezero crater to search for traces of putative ancient life. Although deltas preserve organic matter well on Earth, the hydrology of early Mars is poorly understood, and in particular, the duration and frequency of surface flows remain elusive such that it is unclear whether Martian deltas would also be good preservers of organic compounds. Here we first propose a new relationship between the lateral migration rate of Martian single-thread rivers and channel width. This model is based on a recent global compilation of unvegetated single-thread rivers on Earth (Ielpi & Lapôtre, 2020) and is scaled to account for Martian gravity. Combined with detailed measurements of the dimensions of fluvial deposits atop the Jezero delta, our new model allows us to estimate an integrated duration of delta-forming sediment-transport events, providing a lower bound for the duration of surface flows. We further estimate the total duration spanned by delta-forming events by comparing the reconstructed volumes of delta sediments and the delta-feeding valley, following the technique of Buhler et al. (2014). Depending on stratigraphic completeness, we find that surface flows were required for a minimum of ~5–85 years, with best estimates of ~19–37 years. In contrast, we find that delta formation spanned a total of ~380,000 years, although uncertainty is large and allows for delta formation over ~2,200–3.8 × 10^6 years. Combined together, our best estimates of these timescales suggest a delta-formation intermittency of 0.005–0.01%, with rivers flowing a minimum of 1 sol every 15–30 Martian years. Our estimates of intermittency only reflect delta-forming events, such that more frequent baseflows without significant sediment transport remain possible. In comparison, the frequency of bankfull flows, which perform most geomorphic work, is almost universally of once every 1 to 2 years on Earth, regardless of aridity.

Our estimates of delta-formation duration can further be used to infer sediment fluxes and concentration and to make predictions of the grain-to-bedform-scale architecture of the deposits that will be explored by the Perseverance rover. We predict that the median grain size of bedload and suspended sediments in the feeder channel is in the very-fine to fine-sand range. Although our results suggest that channel-bed deposits may display plane beds deposited in upper flow regime, spatial and temporal variability in bed stresses also likely led to the formation of a variety of subcritical bedforms such as ripples and dunes. Suspended sediment concentration is inferred to have been relatively high (~2–10% by volume), yet below what is required for hyperconcentration. We find that, averaged across channels and over delta formation, riverbeds aggraded by ~1–3 cm/sol, and that bedload represented <1% of the total sediment flux. In summary, our results hint at punctuated surface flows over relatively long timescales, which induced rapid sedimentation of fines in the distal portions of the delta, and in turn, rapid burial of putative organic matter. Our results thus quantitatively confirm that, if life ever arose near Jezero crater, the delta deposits near the landing site of NASA’s Perseverance rover have a high chance to have preserved biosignatures.

Data Availability Statement

All data used in the paper were published prior to our study, including the compilation of lateral migration rates (Ielpi & Lapôtre, 2020), Jezero delta stratigraphy (Goudge et al., 2015, 2018; Shahrzad et al., 2019), Mars’ denudation rates (Golombek et al., 2006, 2014; Golombek & Bridges, 2000; Sweeney et al., 2018), estimates of stratigraphic completeness on Earth (Durkin et al., 2017), experimental river-avulsion data (Ganti et al., 2016), and imagery and topography (Dickson et al., 2018; Fergason et al., 2018). All of these published data are cited throughout where appropriate in the text and can be accessed directly through cited references.
## Appendix A: List of Notations Used in the Present Study

| Symbol   | Variable                                                                 |
|----------|--------------------------------------------------------------------------|
| $A_d$    | Drainage area of the Jezero catchment above the delta apex               |
| $Ar$     | Aggradation rate of the riverbed                                          |
| $C(z)$, $\mathcal{U}$ | Suspended-sediment concentration by volume at a given height above the bed and averaged over the water column, respectively (Text S4) |
| $C_f$    | Friction coefficient of the riverbed                                     |
| $d_{50}, d_{90}$ | 50th and 90th percentiles (Text S2) of the grain-size distribution of bed-sediments (bedload and suspended load) |
| $d_{\text{bank}}$ | Effective grain size of bank-forming sediments                           |
| $E$      | Erodibility of riverbanks                                                |
| $E_{\text{NIA}}, E_{\text{HHA}}$ | Noachian/Hesperian and Hesperian-to-Amazonian average denudation rates |
| $F_{\text{r}}$ | Froude number of the flow (Text S3)                                       |
| $g$, $g_E$, $g_M$ | Acceleration of gravity, with subscripts “E” and “M” indicating Earth and Mars, respectively |
| $h$      | Bankfull flow depth                                                      |
| $h^*$     | Fraction of bankfull flow depth required to fill with sediments to trigger a channel avulsion |
| $h_t$    | Height of the delta toe above the delta base (Figure S1b)                |
| $H_d$    | Height of the delta apex above the delta base (Figure S1b)               |
| $H_p$    | Present-day depth of the feeder valley (Figure S2)                       |
| $I$      | Intermittency of delta-forming sediment-transport events                  |
| $I_{\text{bf}}$ | Intermittency of bankfull flows                                           |
| $k_s$    | Characteristic roughness height of the riverbed                          |
| $l$      | Horizontal distance between the delta apex and the most proximal delta materials along the delta base (Figure S1b) |
| $L$      | Distal extent of delta deposits (Figure S1)                             |
| $M_r$, $M_{r_E}$, $M_{r_{\text{inst}}}$, $M_{r_{E_{\text{inst}}}}$, $M_{r_{E_{\text{inst,nv}}}}$, $M_{r_{E_{\text{nv}}}}$, $M_{r_{M_{\text{inst}}}}$ | Channel lateral migration rate, where subscripts “E” and “M” denote Earth and Mars, respectively, “nv” and “v” denote unvegetated and vegetated, respectively, and “inst” denotes “instantaneous” |
| $q_b$, $q_s$, $q_t$ | Bedload, suspended-sediment, and total sediment fluxes per unit feeder-channel width, respectively |
| $Q_t$    | Total sediment flux in the feeder channel                                |
| $R$      | Specific submerged density of the sediments                               |
| $Re_p$   | Particle Reynolds number                                                  |
| $S$      | Streamwise slope of the riverbed                                          |
| $T_a$    | Avulsion timescale                                                        |
| $T_c$    | Integrated duration of delta-forming flow events                          |
| $T_{\text{tot}}$ | Total timespan of delta formation, including dry spells          |
| $u(z)$, $\mathcal{U}$ | Flow velocity at a given height above the bed and averaged over the water column, respectively |
| $V_{d}$, $V_e$ | Reconstructed volume of the delta deposits and feeder valley, respectively |
| $w$, $w_T$ | Channel width, where subscript “$v$” denotes the channel within the feeder valley near the delta apex |
| $W_d$    | Width of channel-belt deposits                                            |
| $W_e$    | Width of the feeder valley                                                |
| $x$, $x_s$, $x_{e,i}$ | Streamwise coordinates along the feeder valley bed, where subscripts “$s$” and “$e$” denote start and end points, respectively, and “$i$” denotes the $i$th reach of the feeder valley (Text S1) |
| $z$      | Elevation above the riverbed                                              |
| $z_{\text{comp}}$, $z_{\text{dep}}$ | Thickness of the fluvial section (inverted channel bodies and point-bar strata), where “comp” and “dep” denote a putative stratigraphically complete and the actual deposits, respectively |
| $z_{\text{denud}}$ | Reconstructed height of eroded materials atop the delta (Figure S1) |
| $\alpha$ | Shape factor related to the asymmetry of bedforms (Text S3)               |
| $\Delta$ | Bedform height (Text S3)                                                  |
| $\epsilon$ | Stratiographic-completeness factor                                        |
| $\theta$ | Angle subtended by the delta in map view (Figure S1b)                    |
| $\kappa$ | Von Kármán constant (Text S2)                                            |
| $\lambda$ | Bedform wavelength (Text S3)                                              |
| $\nu$   | Kinematic viscosity of water                                              |
| $\rho$, $\rho_s$ | Water and sediment density, respectively                                  |
| $\tau_b$, $\tau$ | Near-bank and average boundary shear stresses, respectively                |
| $\tau_r$, $\tau_v$ | Shields stress and critical Shields stress, respectively                 |
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