Stratigraphy and paleohydrology of delta channel deposits, Jezero crater, Mars

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Abstract

The Jezero crater open-basin lake contains two well-exposed fluvial sedimentary deposits formed early in martian history. Here, we examine the geometry and architecture of the Jezero western delta fluvial stratigraphy using high-resolution orbital images and digital elevation models (DEMs). The goal of this analysis is to reconstruct the evolution of the delta and associated shoreline position. The delta outcrop contains three distinct classes of fluvial stratigraphy that we interpret, from oldest to youngest, as: (1) point bar strata deposited by repeated flood events in meandering channels; (2) inverted channel-filling deposits formed by avulsive distributary channels; and (3) a valley that incises the deposit. We use DEMs to quantify the geometry of the channel deposits and estimate flow depths of ~7 m for the meandering channels and ~2 m for the avulsive distributary channels. Using these estimates, we employ a novel approach for assessing paleohydrology of the formative channels in relative terms. This analysis indicates that the shift from meandering to avulsive distributary channels was associated with an approximately four-fold decrease in the water to sediment discharge ratio. We use observations of the fluvial stratigraphy and channel paleohydrology to propose a model for the evolution of the Jezero western delta. The delta stratigraphy records lake level rise and shoreline transgression associated with approximately continuous filling of the basin, followed by outlet breaching, and eventual erosion of the delta. Our results imply a martian surface environment during the period of delta formation that supplied sufficient surface runoff to fill the Jezero basin without major drops in lake level, but also with discrete flooding events at non-orbital (e.g., annual to decadal) timescales.

1. Introduction

Sedimentary rocks provide insight into past environmental conditions on planetary surfaces by recording the conditions of sediment transport and deposition by moving fluids (e.g., wind, water). Mars has an extensive sedimentary rock record (Malin and Edgett, 2000; Grotzinger and Milliken, 2012), which includes fluvial deposits formed early in martian history when liquid water availability and surface runoff were highest (Howard et al., 2005; Irwin et al., 2005a; Fassett and Head, 2008, 2011; Hynek et al., 2010; Mangold et al., 2012; Goudge et al., 2016). This period of peak fluvial activity, often referred to as the valley network-forming era, saw the incision of the largest-scale, most integrated valley systems on Mars (Pieri, 1980; Howard et al., 2005; Irwin et al., 2005a; Fassett and Head, 2008, 2011; Hynek et al., 2010). Other probable fluvial valleys formed later, but are generally more localized in extent (e.g., Fassett and Head, 2008; Hynek et al., 2010; Hobley et al., 2014; Wilson et al., 2016). The main valley network-forming era occurred ~3.5 Ga, ending in the Late Noachian or Early Hesperian (Howard et al., 2005; Irwin et al., 2005a; Fassett and Head, 2008; Hynek et al., 2010; Mangold et al., 2012).

Much of our understanding of this early period of fluvial activity comes from the geomorphic characteristics of incised valley network systems, which are readily quantified from orbital remote sensing data (e.g., Pieri, 1980; Howard et al., 2005; Irwin et al., 2005a,b; Jaumann et al., 2005; Ansan and Mangold, 2013; Penido et al., 2013). Studies of fluvial erosion on early Mars are complemented by the information preserved in the contemporaneous depositional record. However, only recently have orbital image and topographic data sets existed with sufficiently high resolution (meter to sub-meter pixel scale) to study the details of the martian sedimentary rock record (e.g., Malin and Edgett, 2000; Grotzinger and Milliken, 2012). These data enable quantitative studies of the geometry and architecture of martian sedimentary outcrop, which...
can be related to the formative conditions of the deposit (e.g., Lewis and Aharonson, 2006; Burr et al., 2009, 2010; Metz et al., 2009; Ansan et al., 2011; DiBiase et al., 2013; Stack et al., 2013; Williams et al., 2013a; Cardenas et al., 2017a; Goudge et al., 2017).

Jezero crater contains some of the most spectacular outcrops of fluvial sedimentary rock observed from orbit on Mars (Fassett and Head, 2005). This ~45 km diameter impact crater hosted a hydrologically open lake fed by inlet valleys to the north and west and drained by an outlet valley to the east (Fig. 1). At the mouths of the two inlet valleys are remnants of previously more extensive fluvial sedimentary deposits (Fassett and Head, 2005; Schon et al., 2012). The age of the Jezero crater paleolake is best constrained by work from Fassett and Head (2008), who derived cessation ages for 30 martian valley networks, including the Jezero inlet/outlet valley system. Their results suggest incision of the Jezero valley system ended at approximately 3.8 ± 0.1 Ga (in the Neukum isochron system), an age with one-sigma error bars overlapping the Noachian–Hesperian boundary. This is consistent with the system being active during the main phase of valley network activity (Fassett and Head, 2008). Therefore, it is likely that Jezero basin fluvial deposits provide a record of sediment transport during this main era of surface runoff and valley network activity on early Mars.

Of the two deposits, the western outcrop (Fig. 1B) is better preserved and is one of the most convincing examples of a martian river delta based on geologic context, geomorphology, mineralogy, and stratigraphic architecture (Fassett and Head, 2005; Ehlmann et al., 2008; Schon et al., 2012; Goudge et al., 2017). The western delta is heavily exhumed and exposes sedimentary structures interpreted as inverted channel-fills and point bar deposits (Fassett and Head, 2005; Schon et al., 2012). The inverted channel-filling deposits are stratigraphically above the point bar strata (Schon et al., 2012), which points to a shift in the style of channel sedimentation during building of the Jezero western delta (JWD).

The stratigraphic characteristics of fluvial sedimentary deposits are controlled by the discharge of water and sediment that built them. These deposits therefore encode the paleohydrology and evolution of the formative system (e.g., Allen, 1965; Mill, 1985; Bridge and Mackey, 1993; Mohrig et al., 2000). While the distinct classes of JWD channel deposits have been noted by Schon et al. (2012), the stratigraphy and geometry of these features have yet to be fully explored. The goal of this contribution is to reconstruct the evolution of the JWD and associated shoreline position using the geometry and architecture of the fluvial stratigraphy. Our results show that the JWD was a dynamic fluvoir-lacustrine system, with a deposit that records major shoreline transgression associated with the filling of the basin.

2. Data

The primary data used are stereo-derived digital elevation models (DEMs) and orthorectified images from the Mars Reconnaissance Orbiter Context Camera (CTX; Malin et al., 2007) and High Resolution Imaging Science Experiment (HiRISE; McEwen et al., 2007). DEMs were produced using the open-source NASA Ames Stereo Pipeline (Brockton and Edwards, 2008; Moratto et al., 2010; Shan et al., 2016). One CTX DEM was produced from stereo-pair images D14_032794_1989-D15_033216_1989. During processing, the stereo-derived point cloud was tied to overlapping elevation values from Mars Orbiter Laser Altimeter (MOLA) point shot data (Smith et al., 2001) to reduce errors in regional topography (Beyer et al., 2014). A DEM was produced from this corrected point cloud at a resolution of 18 m/pixel, and image D14_032794_1989 was orthorectified and projected at a resolution of ~6 m/pixel. Three HiRISE DEMs were produced from stereo-pair images ESP_036618_1985-ESP_037119_1985, ESP_037396_1985-ESP_042315_1985, and ESP_036618_1985-ESP_037119_1985 overlain on a mosaic of CTX images D14_032794_1989 and P04_002664_1988, and HiRISE images ESP_036618_1985, ESP_037396_1985 and PSP_003798_1985.

3. Observations of fluvial stratigraphy

We mapped the fluvial stratigraphy exposed on the upper portion of the JWD (Fig. 2) using HiRISE DEMs and orthorectified images in Esri’s ArcMap Geographic Information System (GIS). Based on distinct texture, tone, topographic expression, and stratigraphic position, we identify three classes of fluvial stratigraphy: curvilinear strata, straight ridges, and an incised valley.
that occupy the majority of the JWD surface (Fig. 2). The ridges have a surface slope that trends towards the basin center, are typically continuous for several hundred meters to a few kilometers in the dip direction, and a few hundred meters in the strike direction. Ridges are typically on the order of a few to tens of meters in height above the surrounding terrain (Fig. 5). Individual ridges outcrop as multiple, approximately planar surfaces separated by distinct topographic breaks or steps (Fig. 6A, red arrows). Ridge surface exposures are rough and commonly covered in meter-scale boulders, with no other evident internal structure or layering (Fig. 5C,F).

3.3. Valley

The final structure observed at the JWD is not depositional, but a valley eroded through a portion of the delta and channel deposits (Figs. 2, 7). The valley has sharp walls (Fig. 7) and has primarily eroded the straight ridge class of fluvial stratigraphy of the upper JWD deposit. There are also isolated outcrops of straight ridge materials within the valley (Fig. 2). The valley is several hundred meters wide and a few kilometers long; however, the precise length of the valley is not easily measured, as the upstream portion connects to the Jezero western inlet valley. A CTX DEM profile along the valley floor has an average slope of ~0.014, although there is significant topographic variability (Fig. 7B).

4. Interpretations of fluvial stratigraphy

4.1. Point bar strata

We interpret the curvilinear strata (Figs. 3 and 4) as exhumed lateral accretion strata of fluvial barform deposits (e.g., Allen, 1965; Miall, 1985). As individual strata are continuous for many tens to hundreds of meters and occur in large, coherent groups, or accretional packages (Figs. 3 and 4), we conclude these strata were primarily deposited by fixed point bars, as opposed to free mid-channel bars (Allen, 1965; Edwards et al., 1983; Miall, 1985; Willis, 1993). This interpretation is consistent with previous work (Schon et al., 2012).

Point bar lateral accretion strata are built through the systematic accumulation of sediment at the inner banks of channel bends during floods, and in association with lateral migration of the channel (e.g., Allen, 1965; Edwards et al., 1983; Miall, 1985; Ikeda, 1989; Willis, 1993). Given that neighboring strata within a single accretional package have conformable relationships with no evidence of inter-stratal unconformities (Figs. 3 and 4A), we interpret these structures as a record of lateral migration of an individual point bar (Fig. 4B, white arrow). However, distinct accretional packages do show clear evidence of cross-cutting relationships (Fig. 3, orange arrows), indicating a complex record of stacking of multiple lateral accretion packages at the deposit scale.

To quantify the geometry of the lateral accretion strata, we use the three-dimensional exposure of each mapped layer to model the associated original point bar surface. For this modeling, we chose to use an inverted, elliptic paraboloid, which is the simplest geometric surface that can capture the curved, rounded topography of a point bar surface (Fig. 8). Model surfaces are described by the general equation:

\[ Z = ax^2 + by^2 + c \]  

(1)

where \( a, b, \) and \( c \) are shape parameters unique to each fit. All model surfaces have the following geometric characteristics: an elliptical cross-section in the \( X-Y \) plane with a freely varying aspect ratio, parabolic cross-sections in the \( X-Z \) and \( Y-Z \) planes, and opening in the negative \( Z \) direction.

3.1. Curvilinear strata

The first class of fluvial stratigraphy is alternating light- and dark-toned, curvilinear strata (Fig. 3; Schon et al., 2012). These strata occupy the lowest stratigraphic position on the JWD, primarily exposed in erosional windows through overlying stratigraphy (Fig. 2). Individual strata can be mapped over lengths of many tens to a few hundred meters. The apparent exposed thickness of individual strata is variable, ranging from <1 m to ~5 m; however, it is likely that stratification also exists below the image resolution. A defining characteristic of the curvilinear strata is that they occur in coherent groups, typically composed of many dozen individual strata that are conformable from one stratum to the next (Figs. 3 and 4). These groups of strata do not have a clear topographic expression (Fig. 4B). Neighboring groups often show complex cross-cutting relationships (Fig. 3, orange arrows), indicating they provide a time-integrated record of sediment accumulation.

3.2. Straight ridges

The second class of fluvial stratigraphy is relatively straight ridge deposits (Fig. 5; Fassett and Head, 2005; Schon et al., 2012).
Fig. 3. Example alternating light- and dark-toned curvilinear strata. North is up in all images. (A) and (C) Portion of HiRISE image ESP_037396_1985. Orange arrows indicate unconformities between discrete accretional packages. (B) and (D) Same as (A) and (C), but with all mapped strata shown in black, and strata robustly modeled by paraboloid surfaces shown in red. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article).

To fit model point bar surfaces, we first manually mapped exposed curvilinear strata that are laterally continuous over >20 m. Next, three-dimensional location information (i.e., X, Y, and Z) was extracted along each mapped stratum from the underlying HiRISE DEM at 1 m intervals in a local gnomonic projection, which minimizes distortion around the local tangent point (18.5°N, 77.5°E). We then used this set of three-dimensional points to calculate the best-fit paraboloid surface, defined by Eq. (1), using the MATLAB nonlinear least-squares solver, which uses the trust-region-reflective algorithm (Coleman and Li, 1994, 1996). The result of this algorithm is unique a, b, and c parameter values that define the best-fit paraboloid surface for each mapped point bar stratum (e.g., Fig. 8A).

We imposed three constraints on the model fit based on a priori assumptions about the shape of the point bar surface. First, we require that the fit has values for a and b that are <0, meaning the surface is elliptic (i.e., elliptical in a cross-section taken in the X-Y plane) and opens downwards. Second, we limited the X/Y position of the paraboloid apex to ± 500 m from the mean X/Y value of the mapped point bar stratum. Finally, we limited the Z position of the
Fig. 4. (A) Example large accretional package of point bar strata showing evidence for significant lateral migration. Note that the strata exposed across the length of profile B–B’ can be traced to a common point (orange arrow), indicating deposition of this accretional package was associated with the lateral migration of a single point bar. Mosaic of CTX image P04_002664_1988 and HiRISE image ESP_037396_1985. (B) Same as (A), but showing surface elevation of the point bar accretional package exposure. Interpreted migration direction of the formative channel is indicated by white arrow. HiRISE-derived DEM from stereo-pair ESP_037396_1985-ESP_042315_1985 overlain on a mosaic of CTX image P04_002664_1988 and HiRISE image ESP_037396_1985. (C) Profile B–B’ along point bar migration direction shown in part (B). Elevation data are extracted from HiRISE-derived DEM from stereo-pair ESP_037396_1985-ESP_042315_1985. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article).
Fig. 5. Example straight ridge deposits. North is up in all images. (A) and (D) HiRISE-derived DEM from stereo-pair ESP_037396_1985-ESP_042315_1985 overlain on a mosaic of CTX image P04_002664_1988 and HiRISE image ESP_037396_1985. Labeled black boxes indicate locations of parts (C) and (F). (B) and (E) Same as (A) and (D), but with mapped surfaces shown in blue and estimated downslope directions shown by white arrows. (C) and (F) Surface of straight ridge deposits, showing lack of internal structure and meter-scale boulders. Portion of HiRISE image ESP_037396_1985. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article).
paraboloid apex to between 0 and 30 m from the maximum elevation (Z position) of the mapped point bar stratum. The paraboloid was allowed to rotate freely around the local Z-axis (i.e., vertical axis through the apex point) during fitting.

Model fits were rejected if they did not satisfy all of the following criteria: (1) root-mean-square error (RMSE) <0.5 m; (2) major radius of curvature at the mean elevation of the mapped point bar stratum <1000 m; (3) apex vertical offset >1 m from the upper and lower bound; and (4) sufficient vertical and lateral variation of the mapped point bar stratum to provide a robust fit to a three-dimensional surface. We enforce this final condition by adapting the criteria outlined by Lewis et al. (2008), namely by specifying that a principal component analysis of the points yield a variance explained by the first principal component of <99.5% (to exclude fits where the variation in the point positions is insufficient to constrain the fit in more than one dimension).

Of the 924 point bar strata mapped across the delta deposit, 257 have fits that satisfy the criteria outlined above (Table SM1). We estimated the height of each point bar by measuring the vertical distance from the best-fit model surface apex to the minimum elevation of the mapped exposure of the point bar stratum (Fig. 8B). This provides a lower-bound estimate, as it is unclear how much of the point bar stratum remains unexposed below the surface. Estimated heights have an approximately normal distribution, with a mean of ~20 m and a standard deviation of ~6.8 m (Fig. 9A). One of the most prominent accretional packages of JWD point bar strata shows a record of continuous lateral migration across >500 m horizontally, and is exposed over a vertical distance of >25 m (Fig. 4). This vertical exposure provides an independent lower bound on point bar height, and agrees well with the values from our fitting technique.

Point bar height records the bankfull depth of the associated river at the bend apex (Allen, 1965; Edwards et al., 1983; Miall, 1985, 1993; Willis, 1993). This flow depth is greater than the average bankfull depth for the channel because of differences in channel cross-sectional shape between bends and the relatively straight reaches that connect bends. Bridge and Mackey (1993) propose that channel depths in bends tend to be greater than the mean channel depth by a factor of approximately three. Applying this correction to the set of estimated point bar heights yields an estimate for mean formative flow depth of ~6.7 m, with a standard deviation of ~2.3 m, for the meandering paleo-channels connected to the point bar strata (Fig. 9A).

The limited areal extent of exposed strata from the lower JWD does not allow for direct measurement of channel widths from the point bar deposits. We were therefore forced to generate estimates of channel width using only the model point bar surfaces for each mapped stratum. To do this we calculated the major radius of curvature of the point bar surface from the cross-sectional ellipse in the X-Y plane at the mean elevation of the mapped exposure of the point bar stratum (Fig. 8C). These calculated values show a positively skewed distribution with a mean of ~140 m and a standard deviation of ~130 m (Fig. 9B).

The radius of curvature for river bends, measured at the channel centerline, is typically a factor of ~2~3 larger than the mean channel width (Leopold and Wolman, 1960; Williams, 1986). However, point bar curvature is always somewhat tighter than overall bend curvature. To convert between the channel centerline radius of curvature (Rc) and the radius of curvature we estimate from the model point bar surfaces (Rpb), we use a simple geometric argument. If we assume Rpb represents the radius of curvature at the inner bank and the channel is symmetric with width B, then Rc = Rpb ± \( \frac{1}{2}B \). Since Rc = [2~3]B (Leopold and Wolman, 1960; Williams, 1986), we can rearrange to get Rpb = [1.5~2.5]B. Applying this correction using the central value of two gives a mean channel width of ~70 m, with a standard deviation of ~65 m, for the JWD meandering channels (Fig. 9B).

4.2. Inverted channel-filling deposits

We interpret the straight ridges as erosionally inverted channel-filling deposits (e.g., Pain et al., 2007; Williams et al., 2011; DiBiase et al., 2013; Nuse, 2015; Cardenas et al., 2017b) based on: (1) their geometric characteristics (i.e., they slope towards the basin center and are many times longer in the dip direction than in the strike direction; Fig. 5); (2) the geologic setting (i.e., as a part of the JWD deposit); and (3) the observation that ridges outcrop at multiple stratigraphic levels (Fig. 6), consistent with large-scale internal stratification separating discrete channel bodies. This interpretation is consistent with previous work on the JWD (Fassett and Head, 2005; Schon et al., 2012). We take the dearth of internal accretion strata within the channel-filling deposits (Fig. 5C,F) as evidence for a lack of significant lateral migration of the channels in the upper JWD.

Field studies of similar erosionally inverted fluvial deposits on Earth show that single ridges are composed of multiple, vertically stacked channel-filling deposits, or channel bodies (Mohrig et al., 2000; Williams et al., 2011; Nuse, 2015; Cardenas et al., 2017b). Surface exposures of different channel bodies are commonly
Fig. 7. Valley incised into the delta deposit. (A) Overview of the downstream portion of the mapped valley (black outline). Mosaic of CTX-derived DEM from stereo-pair D14_032794_1989-D15_033216_1989, and HiRISE-derived DEMs from stereo-pairs ESP_037396_1985-ESP_042315_1985 and ESP_036618_1985-ESP_037119_1985 overlain on a mosaic of CTX image D14_032794_1989, and HiRISE images ESP_037396_1985 and ESP_036618_1985. North is up. (B) Profile A–A’ along the valley center (see Fig. 1B for profile location) and estimated valley erosional relief, which is a lower limit on the incision depth. Error bars for erosional relief are calculated standard deviation on measured values. Profile elevation data are extracted from CTX-derived DEM from stereo-pair D14_032794_1989-D15_033216_1989. (C) and (D) Profiles C–C’ and D–D’ across the valley. Profile locations shown in part (A). Elevation data are extracted from HiRISE-derived DEM from stereo-pairs ESP_036618_1985-ESP_037119_1985 (C) and ESP_037396_1985-ESP_042315_1985 (D).
Fig. 8. (A) Example model surface fit to a mapped point bar stratum (red line in Fig. 4A) for estimation of the original point bar geometry. Thick black line is three-dimensional location information along the mapped point bar stratum extracted from HiRISE-derived DEM from stereo-pair ESP_037396_1985-ESP_042315_1985. Colored surface is the best-fit paraboloid to those data. Grey surfaces indicate planes from which the point bar height (dark grey) and major radius of curvature (light grey) were extracted. Point bar height plane is defined by the Z axis and the elliptic paraboloid major radius, which can rotate freely (here rotation is 7° south of east). Major radius of curvature plane is defined by the X and Y axes, and intersects the Z axis at the mean elevation of the mapped point bar exposure. (B) Parabolic cross-section of the best-fit model surface shown in part (A). Point bar height is estimated as the vertical distance between the parabola apex (yellow circle) and the lowest elevation of the mapped point bar exposure (red circle). (C) Elliptical cross-section (white) of the best-fit model surface shown in part (A) at the mean elevation of the mapped point bar exposure. Point bar radius of curvature is estimated from the major radius of the ellipse (white arrows). Mapped point bar exposure (black) and colored model surface shown for reference. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article).

We therefore interpret the distinct stratigraphic surfaces of the JWD that are separated by topographic breaks (Mohrig et al., 2000; Cardenas et al., 2017b) as exposures of the top of individual channel bodies of composite inverted fluvial ridges. It is important to note that the vertical section of channel-filling deposit bound by two stratigraphic surfaces represents at least one, but potentially multiple, channel bodies, as not every channel body erodes as an observable stratigraphic surface during topographic inversion (Cardenas et al., 2017b).

To quantify the geometry of the inverted channel-filling deposits, we mapped the surface exposure of each individual channel
body (e.g., Fig. 5), and modeled this as a plane (e.g., Fig. 6B). Since the mapped surface is a modern erosional surface, and not necessarily a stratigraphic surface, the results of this approach provide only an approximation of the geometry of the true stratigraphic surface associated with the channel-filling deposit. However, erosional surfaces of inverted channel-filling deposits on Earth can accurately record the channel bed topography at the time of abandonment (Cardenas et al., 2017b). Additionally, this is the most reproducible approach given the available orbital remote sensing data. We expand below on the inherent sources of uncertainty for each of the different geometric properties estimated for the inverted channel-filling deposits of the JWD.

To define the model stratigraphic surfaces, three-dimensional location information was extracted from a grid of points spaced 1 m apart across the mapped channel body tops, again in a local gnomonic projection. From these points, the best-fit plane was modeled using a linear least squares method. Fits were rejected if they did not satisfy all of the following criteria: (1) RMSE <0.5 m, (2) slope <0.05, and (3) dip direction approximately parallel to the large-scale transport direction for the JWD (i.e., basinward).

Of the 606 channel body tops mapped across the delta deposit, fits for 159 satisfy these criteria (Table SM2). We estimated channel bed slope values directly from each model surface, assuming the stratigraphic surface provides a reasonable approximation of the channel bed topography (Cardenas et al., 2017b). The topographic breaks separating exposures of discrete channel bodies have scarp sets that face basinward (Fig. 6A, red arrows), which indicates the erosional surface of the JWD has a steeper slope than the channel body stratigraphy. Therefore, the estimated slope values are likely to be an upper limit on the channel bed slope. Estimated slopes have an approximately normal distribution with a mean of ~0.027, and a standard deviation of ~0.012 (Fig. 10A).

We estimated channel body width as the absolute width of the mapped channel body top exposure perpendicular to the downslope direction. This provides an approximation of the true width of the formative channel, and may be either an over- or underestimated. The formative channel width would be overestimated in the case where the stratigraphic surface extends across multiple channel bodies mapped as one surface. The formative channel width would be underestimated if it has been significantly modified by erosion or if it is partially obscured by a different channel body stacked upon it (e.g., Fig. 6). Estimated widths have a slightly positively skewed distribution with a mean of ~84 m and a standard deviation of ~45 m (Fig. 10B).

Where channel-filling deposits display clear stacking patterns, neighboring model surfaces define the top and base of a channel body (e.g., Fig. 6). This allowed us to estimate the thickness of 55 channel bodies by calculating the vertical distance between the two adjacent bounding model surfaces (Fig. 6B). This measurement was taken at the geometric center of the basal model stratigraphic surface. The estimated channel body thickness may be overestimated if the two stratigraphic surfaces bound multiple channel bodies, or if erosion has lowered the basal stratigraphic surface more than the upper stratigraphic surface. The estimated thickness may also be underestimated if erosion has lowered the upper stratigraphic surface more than the basal stratigraphic surface. Estimated channel body thicknesses have a positively skewed distribution with a mean of ~1.7 m and a standard deviation of ~1.4 m (Fig. 10C). Field studies of the thickness of alluvial channel-filling deposits indicate that channel body thickness provides an appropriate proxy for the formative channel depth (Mohrig et al., 2000).

4.3. Fluvial valley

The laterally continuous incision and rough grading of the JWD valley with the Jezero western inlet valley (Figs. 2, 7) suggest this valley is fluvial in origin. We interpret this feature as an erosional valley, as opposed to a river channel, as it is significantly larger in scale (both width and depth) than the inverted channel-filling deposits into which it incises (Fig. 7).

To quantify the valley geometry, we extracted across-stream profiles from the HiRISE DEMs every 500 m, starting just downstream of where the valley crosses completely into the basin (Figs. 2, 7). From these profiles, we manually identified the valley rim and measured the rim-to-rim width and the erosional relief (difference between the minimum valley elevation and the rim elevation, both averaged across 15 m). The modern erosional relief is taken as a lower limit on the incision depth of the valley given the amount of exhumation the JWD deposit has experienced (Fassett and Head, 2005; Schon et al., 2012).

The valley has a mean width of ~600 m with a standard deviation of ~200 m. The valley has a mean erosional relief of ~30 m with a standard deviation of ~11 m. Valley erosional relief shows significant scatter, although there is a slight trend of decreasing values downstream (linear fit with $R^2 = 0.38$; Fig. 7B); however, it is difficult to obtain a robust trend given the amount of deposit erosion and exhumation.
5. Styles of channelization preserved in the stratigraphic record

Our interpretations of the three classes of JWD fluvial stratigraphy demonstrate a record of three contrasting styles of channelized flow and sedimentation/erosion. Additionally, these three classes of stratigraphy have distinct horizontal and vertical positions across the deposit (Fig. 2), indicating they record fluvial activity separated in time and/or space during the evolution of the JWD. Below we discuss some of the likely characteristics of these periods/zones of channelization as inferred from the stratigraphic record.

5.1. Meandering channels

The first style of channelization is recorded as point bar strata formed by laterally migrating channels (Figs. 3 and 4). When interpreting the characteristics of the formative meandering channels, it is important to consider both the geometric properties estimated from the point bar strata, as well as the JWD coastal environment. As coastal rivers near the shoreline, they enter the backwater zone, where channel transport is influenced by the standing water in the receiving basin. This influence suppresses lateral migration of channel bends and point bar growth (Hudson and Kesel, 2000; Nittroer et al., 2012; Fernandes et al., 2016). The JWD point bar strata record lateral migration over distances equivalent to several channel widths (Figs. 4, 9B). This is consistent with higher lateral migration rates in channels outside of the backwater zone, but it might also be an expression of significant lengths of time captured by the slow lateral migration of channels within the backwater zone.

Here we propose the simpler model that the point bar strata formed outside of the backwater zone, with a shoreline located farther downstream towards the basin center, and at a lower elevation. This is consistent with the stratigraphic position of the point bar deposits tens of meters below the minimum hydrologically open lake level defined by the outlet breach floor (Fig. 2; Fassett and Head, 2005), and require that these strata were deposited before the basin overflowed. This model is also consistent with isolated outcrops of potential delta sediment identified farther towards the basin center (Schon et al., 2012).

The length of the backwater zone for a coastal river is scaled by the ratio of channel depth to channel bed slope (Paola and Mohrig, 1996; Jerolmack, 2005). We have no constraint on the slope of the JWD meandering channels; however, using the slope estimated from the inverted channel-filling deposits as an order of magnitude proxy (Fig. 10A), the backwater length scale for the meandering channels should be on the order of a few hundred meters to a few kilometers. We interpret the JWD point bar strata as recording sedimentation during an initial phase of basin filling, when the shoreline was in excess of several hundred meters basinward from these preserved channel deposits.

5.2. Avulsive distributary channels

The JWD inverted channel-filling deposits (Figs. 5 and 6) overlying the point bar strata record delta growth associated with
relatively straight channels showing little evidence for lateral mobility via systematic erosion along one bank and sedimentation along the other. Transport directions estimated from the inverted channel-filling deposits show apparent mid-system bifurcations, consistent with the occurrence of a distributary network of channels (Fig. 11). Additionally, the large spread in transport directions emanating from the erosional inlet valley (Figs. 1 and 11) is characteristic of a deltaic avulsion point (Jerolmack, 2009). It is also possible that this nodal avulsion point is controlled by the transition from a confined bedrock channel to an unconfined alluvial channel (e.g., Ganti et al., 2014); however, we suggest this is unlikely as the point where the Jezero western inlet would have become unconfined (i.e., where it crosses the crater rim into the basin; Fig. 2, green star) is >2 km upstream from the first mapped occurrence of inverted channel-filling deposits. The stacking of multiple channel bodies (Fig. 6) is also consistent with a record of numerous avulsions, and further suggests the routing of new channel paths during avulsions was influenced by subtle topographic signatures left by abandoned, under-filled channels (e.g., Heller and Paola, 1996; Mohrig et al., 2000; Edmonds et al., 2016).

Why was the JWD system so avulsive? The channel avulsion process is a consequence of gravitational potential energy as a channel becomes increasingly perched, or supereroded, above the surrounding terrain via highest sedimentation rates always occurring in the vicinity of the channel (Bryant et al., 1995; Heller and Paola, 1996; Mohrig et al., 2000). Avulsions characterize both sub-

aerial and subaqueous channelized systems. Fortunately, the character of deposits tied to both environments is expected to be different and these differences can be evaluated using preserved stratigraphy of the JWD.

Given the low excess density of flows that form subaqueous channels compared to flow in subaerial channels, Imran et al. (1998) argued that the potential energy available for levee erosion from overbank flows, and thus channel avulsion, should be significantly higher in subaerial channels. Everything else being equal, this should result in relatively thicker deposits associated with any time-integrated channel in the subaqueous environment compared to the subaerial environment (Mohrig et al., 2000). The stratigraphic consequence of this difference on Earth has recently been quantified by Jobe et al. (2016). Their data set shows channel-belt width-to-thickness ratios that are less than 30 in the subaqueous environment and greater than 30 in the subaerial environment (see Fig. 2G of Jobe et al., 2016). The measured value here of 49 (mean channel deposit width/mean channel deposit thickness; Figs. 10B and C) indicates the JWD inverted channel-filling deposits are most consistent with subaerial formation.

We interpret the lack of lateral accretion associated with the JWD inverted channel-filling deposits as evidence for these channel bodies being deposited near or within the backwater zone, a short (< ~1 km) distance upstream from the shoreline. Under this scenario, the backward stepping arrangement of the inverted channel bodies (Fig. 10D, orange ellipse) is consistent with significant shoreline transgression as the lake level rose through time.

Lake level rise also drives increased bed aggradation (e.g., Posamentier and Vail, 1988; Posamentier et al., 1988; Jerolmack, 2009). Experimental studies have shown a positive correlation between channel avulsion frequency and overall sedimentation rate (Bryant et al., 1995; Ashworth et al., 2004, 2007). Therefore, we conclude that bed aggradation associated with lake level rise was a likely driver for the density of channel avulsions preserved in these strata. It is also possible that the density of channel avulsions records slow bed aggradation rates over a significant period of time; however, we suggest this is less likely given the independent evidence for lake level rise preserved by the inverted channel-filling deposits of the JWD (Fig. 10D).

The stratigraphically highest channel bodies are above, or within a few channel depths of, the lake breach level (Fig. 10D, black ellipse). During filling of the Jezero basin, the lake level would have reached a maximum elevation ∼200–300 m above the present breach level prior to overflow and outlet valley incision (Fassett and Head, 2005; Fassett and Goudge, 2017). Therefore, the position of these upper channel bodies is consistent with formation either during ongoing shoreline transgression during basin filling, or during shoreline regression after the basin breached and had a fixed lake level near the breach floor elevation. Given the lack of major unconformity separating these two ‘groups’ of channel-filling deposits, and the continued backward stepping geometry of the stratigraphically highest channel bodies (Fig. 10D), we conclude that the former scenario is most likely.

5.3. Valley incision

The final stage of channelization recorded by the JWD is the incision of the deposit by a younger fluvial valley (Fig. 7). The valley has relatively sharp walls (Fig. 7), which indicates that either the flows responsible for incision occurred following sufficient lithification for the deposit to act as rock, not unconsolidated sediment, or the valley topography has been sharpened by later aeolian erosion (e.g., Perkins et al., 2015).

Incision of the valley required either a lower lake level or a decrease in the relative sediment to water discharge in the fluvial system. Knickpoint formation, beginning toward the
downstream end of the fluvial system, is expected with lake level fall (Paola et al., 1992; Heller et al., 2001; Sun et al., 2002). Incomplete regrading of the fluvial channel by any resultant knickpoint would yield a valley longitudinal profile with a sharp break in slope bounded upstream and downstream by two reaches with comparable bed slopes (Howard and Kerby, 1983; Stock and Montgomery, 1999; Crosby and Whipple, 2006). The valley profile does not show this nearly diagnostic signature of lake level fall, and is instead roughly graded, although there is significant short-wavelength variability (Fig. 7B).

The valley has a slight decrease in erosional relief in the downstream direction (Fig. 7B), which is the expected pattern of erosion driven by an upstream decrease in the ratio of sediment discharge to water discharge (Paola et al., 1992; Sun et al., 2002). The erosional pattern of the valley is thus most consistent with formation due to a drop in the sediment load of the system, though the possibility of a knickpoint having propagated upstream through the entire JWD deposit and into the inlet valley cannot be ruled out. Therefore, we conclude that the data are insufficient to conclusively reconstruct the forcing behind final valley incision.

6. Estimating paleohydrology

When estimating paleohydrology from fluvial stratigraphy, the most important system-scale constraints are channel geometry (depth, $H$, width, $B$, and slope of the bed, $S$) and the size of transported sediment, often characterized by the median grain diameter, $D_{50}$ (Leopold and Maddock, 1953; Parker et al., 2007; Wilkerson and Parker, 2011). We have estimates for all three geometric characteristics for the avulsive distributary channels (Fig. 10), and two ($H$ and $B$) for the meandering channels (Fig. 9). However, it is not possible to directly estimate $D_{50}$ of the system using only remote sensing data.

Appropriate $D_{50}$ values for martian fluvial systems have been estimated by various methods, including in situ observations of the regolith (Wilson et al., 2004; Kleinhans, 2005) or fluvial deposits (Williams et al., 2013b), assignment of empirical values (Jerolmack et al., 2004; Metz et al., 2009), and the assumption that equilibrium channel slopes are controlled by the critical Shields stress of the transported sediment (DiBiase et al., 2013). Estimates of $D_{50}$ from these methods range over three orders of magnitude, from ~0.3 to 100 mm (Jerolmack et al., 2004; Wilson et al., 2004; Kleinhans, 2005; Metz et al., 2009; DiBiase et al., 2013; Williams et al., 2013b). Clearly, constraints on grain size are a major uncertainty when estimating paleohydrology from martian fluvial deposits (Kleinhans, 2005).

Therefore, we opt to take a novel, conservative approach by assessing the paleohydrology of these channel deposits in a relative sense, instead of estimating absolute values for an underconstrained problem. This is of particular use for the JWD when comparing the two types of channel deposits (meandering versus avulsive), which represent a change in delta sedimentation from either autogenic (internal) or allogetic (external) forcing(s). The stratigraphically lower meandering channel deposits record a period of flow that is a factor of ~4 deeper and a factor of ~1.2 narrower than the period of flow recorded by the avulsive distributary channels (Figs. 9 and 10).

To quantitatively estimate what this means in terms of the paleohydrology for the system, we use the widely applied theory that the hydraulic geometry of alluvial channels is defined by the bankfull discharge of water ($Q_w$) and sediment ($Q_s$; e.g., Leopold and Maddock, 1953; Parker et al., 2007; Wilkerson and Parker, 2011). To develop a relationship between $Q_w$ and $Q_s$ and the cross-sectional geometry of the river channel at bankfull flow, we start with relationships that describe: (1) the hydraulic conditions of the fluid motion; (2) the relationship between fluid movement and sediment transport; and (3) a Shields criterion for sediment motion.

The hydraulic conditions of fluid motion are described using a momentum balance equation for open-channel, quasi-normal flow:

$$\tau_b = \rho g HS = \frac{C_f \rho Q_s^2}{B^2 H^2}$$  \hspace{1cm} (2)

where $\tau_b$ is the shear stress applied to the bed by the moving fluid, $\rho$ is the density of the fluid, $g$ is the gravitational acceleration, and $C_f$ is an empirical drag coefficient (Dietrich and Whiting, 1989; Parker et al., 2007; Wilkerson and Parker, 2011). This equation balances the momentum carried by the moving fluid with the drag on the fluid at the base of the channel, assuming there is no fluid acceleration (in time or space).

Sediment transport is described by the physically scaled equation developed by Meyer-Peter and Muller (1948), which can be written in the most general form as:

$$Q_s = BD_{50} \sqrt{gD_{50} \alpha (\psi \tau_{f_{\text{form}}} - \tau_{c_{\text{cr}}})^\eta}$$  \hspace{1cm} (3)

where $R$ is the submerged specific gravity of the sediment, $\tau_{c_{\text{cr}}}$ is the non-dimensional critical shear stress for incipient sediment motion, and $\eta$, $\alpha$, and $\psi$ are empirically-derived coefficients (e.g., Meyer-Peter and Muller, 1948; Fernandez Luque and van Beek, 1976; Wong and Park, 2006).

Finally, the non-dimensional formative shear stress, or the formative Shields number, for bankfull flow is defined as:

$$\tau_{f_{\text{form}}} = \frac{HS}{RD_{50}}$$  \hspace{1cm} (4)

Combining Eqs. (2)–(4), we can derive the following three relationships for the bankfull hydraulic geometry of alluvial channels (see Appendix A for additional detail):

$$B = \frac{1}{D_{50} \sqrt{RgD_{50} \alpha (\psi \tau_{f_{\text{form}}} - \tau_{c_{\text{cr}}})^\eta} Q_w}$$  \hspace{1cm} (5)

$$H = \frac{D_{50} \alpha (\psi \tau_{f_{\text{form}}} - \tau_{c_{\text{cr}}})^\eta \sqrt{\tau_{c_{\text{cr}}}} Q_w}{\sqrt{\tau_{f_{\text{form}}}} Q_s}$$  \hspace{1cm} (6)

$$S = \frac{R (\tau_{f_{\text{form}}})^{3/2} Q_s}{\sqrt{\tau_{c_{\text{cr}}}} (\psi \tau_{f_{\text{form}}} - \tau_{c_{\text{cr}}})^\eta Q_w}$$  \hspace{1cm} (7)

This gives three equations for bankfull hydraulic geometry, as well as three primary unknowns ($Q_w$, $Q_s$, and $D_{50}$). We are fully aware that this is a potentially solvable system of equations; however, these equations depend on a number of coefficients ($\tau_{c_{\text{cr}}}$, $\eta$, $C_f$, $\alpha$, and $\psi$) that have been empirically derived from alluvial channels on Earth, which vary as a function of $Q_w$, $Q_s$, and $D_{50}$ (Meyer-Peter and Muller, 1948; Fernandez Luque and van Beek, 1976; Dietrich and Whiting, 1989; Wong and Park, 2006; Parker et al., 2007; Wilkerson and Parker, 2011). While a robust null hypothesis suggests that the physics controlling sediment transport in alluvial channels is the same on Earth and Mars, applying these equations to remotely studied martian deposits is beyond the scope where these empirical coefficients were derived, and would lead to values that are subject to significant uncertainty.

For our approach of looking at relative paleohydrology, we can make the simplifying assumption that both the empirical coefficients and non-geometric, system-specific characteristics (e.g., $R$, $g$, $D_{50}$) remain approximately constant between the flows recorded by the meandering versus avulsive distributary channel deposits.
Therefore, we can reduce Eqs. (5)–(7) to three proportional relationships:

\[ B \propto Q_s \]  
\[ H \propto \frac{Q_{uw}}{Q_s} \]  
\[ S \propto \frac{Q_s}{Q_{uw}} \]  

The estimated channel widths indicate a mean value of \( Q_s \) associated with the avulsive distributary channels that is a factor of \( \sim 1.2 \) larger than for the meandering channels; however, the range in estimated width values between the two types of channels shows significant overlap (Figs. 9B, 10B). Due to this overlap and the uncertainty in the effect of exhumation and channel stacking on the reliability of inverted channel-filling deposit width measurements, we focus our attention on estimates for channel flow depth, which show considerably less overlap (Figs. 9A, 10C).

Channel flow depth estimates indicate a ratio of water to sediment discharge (\( Q_{uw}/Q_s \)) that is a factor of approximately four larger for the meandering channels compared to the avulsive distributary channels. This provides an important constraint for understanding the change in channelization recorded by the JWD: the shift from meandering to avulsive distributary channels was associated with an approximately four-fold decrease in \( Q_{uw}/Q_s \). This change may be due to an allogenic forcing, where water discharge decreased due to changes in regional climate and/or sediment discharge increased due to enhanced landscape erosion, or it could be due to an autogenic forcing as the JWD system dynamically evolved.

7. Discussion

7.1. Evolution of the Jezero western delta

To understand the evolution of the JWD, we reconstruct a stratigraphic framework that is internally consistent with the constraints provided by both the fluvial stratigraphy and the channel deposit paleohydrology (Fig. 12). The stratigraphy and geometry of the JWD channel deposits record two styles of channelization and associated sedimentation: (1) fluvial deposits of deeper meandering channels that formed some distance upstream from the paleolake shoreline (Figs. 3 and 4); and (2) coastal deposits of shallower avulsive distributary channels that formed proximal to the paleolake shoreline (Figs. 5 and 6). The paleohydrology of these two classes of channel deposit indicates that the shift from meandering to avulsive distributary channels had a corresponding approximately four-fold decrease in \( Q_{uw}/Q_s \). The youngest episode of fluvial activity recorded by the JWD is a valley incised into the deposit.

We conclude that the JWD deposit primarily records a rising lake level and shoreline transgression as the Jezero basin filled with water. During the initial filling of the basin, the shoreline remained towards the basin center and meandering channels developed near the inlet breach (i.e., the position of the preserved JWD deposit). Goudge et al. (2017) identified subaqueous foreset and bottomset strata within the JWD stratigraphically below the most basinward meandering channels. This vertical stratigraphic succession of bottomset to foreset to topset/fluvial deposits indicates that sedimentation rates during this initial phase of delta growth were sufficient to keep pace with the rising lake level and cause delta progradation and shoreline regression (Fig. 12A).

As the lake level continued to rise, and assuming there were no major changes in the input sediment load, the accommodation space beyond the delta front became too large for the sediment supply to fill, leading to shoreline transgression (Fig. 12B). This evolution of minor shoreline progradation followed by major transgression under constant sediment supply and lake level rise has been well studied in laboratory experiments, and is termed shoreline autoretreat (Muto and Steel, 1992; Muto, 2001). Autoretreat gives the distinct signature of downstream coastal channel deposits overlying upstream fluvial channel deposits (Figs. 2, 12; Muto and Steel, 1992; Muto, 2001).

As autoretreat progressed, the shoreline approached the inlet breach of the basin, forming the JWD avulsive distributary channels with backward stepping elevations (Fig. 10D, orange ellipse). During this process, the coastal distributary channels of the JWD system underwent bed aggradation from steady lake level rise (Posamentier and Vail, 1988; Posamentier et al., 1988; Jerolmack, 2009), driving significant channel avulsion (Fig. 11; Bryant et al., 1995; Ashworth et al., 2004, 2007).
We interpret the factor of approximately four change in $Q_m/Q_s$ between the JWD meandering and avulsive distributary channels as a direct consequence of their relative position to the shoreline and development of a distributary channel network (Fig. 11). As coastal distributary channels partition water and sediment, the $Q_m/Q_s$ ratio decreases downstream as increasing amounts of water leaves the channels onto the floodplain or delta islands, while most of the sediment remains in the channels (e.g., Hiatt and Passalacqua, 2015; Shaw et al., 2016). Meandering rivers are uncommon in distributary settings with multiple active channels (e.g., Jerolmack and Mohrig, 2007), consistent with single thread JWD meandering channels feeding a more distal shoreline.

It is possible that the two classes of channel deposits and the associated change in $Q_m/Q_s$ ratio recorded by the JWD was driven by an allogenic forcing, such as changing climate and/or landscape-scale erosion and sediment delivery. However, we favor the much more straightforward model that these changes record the dynamic evolution of a fluvio-deltaic system responding to the filling of the Jezero basin, a process that is expected a priori based on the presence of an incised outlet valley (Fassett and Head, 2005).

After continued autoretrat, the lake level in the Jezero basin overtopped the crater rim and the outlet valley incised the rim to an elevation of approximately $-2395$ m. Following the outlet forming flood and associated lake level drop, the outlet valley would have held lake level relatively fixed (Fassett and Head, 2005). Assuming ongoing input of water and sediment, the fixed lake level ended shoreline autoretrat and caused delta progradation and shoreline regression (Fig. 12C; Posamentier and Vail, 1988; Posamentier et al., 1988; Jerolmack, 2009). This phase of basin evolution may potentially be recorded by the stratigraphically highest inverted channel-filling deposits, which have an elevation near the outlet valley breach level (Fig. 10D, black ellipse). However, we suggest this is unlikely given the lack of a major unconformity separating the upper and lower inverted channel-filling deposits, and the continuous backward stepping geometry of the deposits (Fig. 10D). Instead, we hypothesize that the majority of fluvial stratigraphy that post-dates the outlet breaching of the Jezero basin has been eroded, as would be expected for the highest stratigraphy of a heavily exhumed deposit (Fassett and Head, 2005; Schon et al., 2012).

Following the formation of the JWD, the deposit was incised by a fluvial valley (Fig. 7). The cause of valley incision is poorly constrained based on available data, and may have been driven by a decrease in the input sediment to water discharge ratio or a falling lake level. Finally, the deposit was eroded to its present form, likely due to an extended period of aeolian erosion (Fassett and Head, 2005; Schon et al., 2012), also leaving behind isolated outcrops of delta sediment (Fig. 12D). Our stratigraphic model suggests that these isolated outcrops are likely to have an offshore, lacustrine origin.

To provide a first order estimate of the amount of JWD sediment eroded by aeolian activity, we calculated the area bound by the delta erosional front and a circle centered on the deposit that reaches the delta deposit remnant outcrops ($\sim 50$ km$^2$; Fig. 13A). We then measured the height of the remnant outcrop ($\sim 53$ m; Fig. 13B), yielding an estimated eroded volume of $\sim 2.7$ km$^3$. In comparison, Fassett and Head (2005) estimate a total volume of $\sim 5$ km$^3$ for both the western and northern Jezero fluvial sedimentary deposits, indicating the loss of a significant fraction of JWD sediment (Fassett and Head, 2005; Schon et al., 2012).

7.2. Implications for the Jezero western delta’s formative climate

The JWD point bar deposits consist of coherent packages composed of dozens of individual lateral accretion strata (Figs. 3 and 4A). Each stratum is interpreted to preserve one, or a small number of, individual flood events separated by periods of minimal-to-non-deposition (e.g., Allen, 1965; Edwards et al., 1983; Willis, 1993). The amount of time separating individual flood events is unconstrained; however, we hypothesize that timescales were likely on the order of years to decades as opposed to millennia given the lack of unconformities between individual strata (Figs. 3 and 4A). Therefore, the JWD fluvial stratigraphy is inconsistent with a single episode of sustained peak runoff, as has been proposed in models of rapid martian delta formation (e.g., Kleinhans, 2005; Kraal et al., 2008; Kleinhans et al., 2010).

Apart from the late stage valley incising the deposit, we observe no major erosional unconformities or alternation of channel deposit types moving upsection through the stratigraphy (e.g., point bar strata that overlie inverted channel-filling deposits that overlie point bar strata). This is inconsistent with multiple lake level rise/fall cycles during the formation of the JWD. While we acknowledge the limitations of remote sensing data for identifying subtle unconformities or upsection changes in stratigraphy, we suggest it is unlikely that drops in lake level sufficiently large to impart a delta-wide signal in the stratigraphic record (e.g., a laterally continuous erosional unconformity) would be missed in our HiRISE-scale analysis. Therefore, we conclude that our
results indicate a relatively continuous filling of the Jezero basin to the point of outlet breaching with no significant drops in lake level. This conclusion is consistent with the results of Fassett and Head (2005) and Schon et al. (2012), who both argued for prolonged fluvial activity associated with the formation of the Jezero crater paleolake and delta deposits.

8. Conclusions

The Jezero crater western delta is a well-exposed example of fluvial stratigraphy formed during the era of valley network activity on Mars. We mapped two classes of channel deposits across the delta – point bar strata deposited by single thread meandering channels, and inverted channel bodies deposited by avulsive distributary channels. Using a novel approach for constraining relative paleohydrology, we conclude that the meandering channels had a water to sediment discharge ratio ($Q_w/Q_s$) that was a factor of approximately four larger than the avulsive distributary channels.

Using the stratigraphy, geometry, and paleohydrology of the mapped channel deposits as constraints, we developed an internally consistent stratigraphic framework and model for delta evolution (Fig. 12). We conclude that the growth of the Jezero western delta was dominated by an overall transgression in shoreline position through time as the basin filled with water to the point of overtopping. Subsequently, the deposit was incised by a younger valley, due to either a drop in the input sediment supply or a drop in the lake level, and eroded to its present form by aeolian activity.

The stratigraphic record of the Jezero western delta is best interpreted as requiring a formative environment with persistent surface runoff, such that the basin filled to overtopping without major drops in lake level. Hydrologic fluctuations that did occur are consistent with flooding at annual to decadal timescales, recorded by individual point bar lateral accretion strata. We emphasize the fact that the western delta deposit in the Jezero basin is unlikely to exemplify all processes occurring everywhere on Mars during the major period of valley network formation. Instead, our stratigraphic and paleohydraulic reconstruction provides one piece in the larger, complicated puzzle for understanding the early martian climate and hydrologic cycle.

Acknowledgments

We express our gratitude to Jay Dickson for helpful discussions on stereo image processing, and to the Ames Stereo Pipeline developers for their efforts. We thank two anonymous reviewers for fair and insightful reviews that helped improve the quality of the manuscript. Jeff Johnson is thanked for editorial handling. T.A.G. acknowledges financial support from the Jackson School of Geosciences Distinguished Postdoctoral Fellowship.

Appendix A. Derivation of bankfull hydraulic geometry relationships

Here we present the derivation of the three relationships for bankfull hydraulic geometry given by Eqs. (5)–(7), starting from Eqs. (2) to (4). First, we will start with the relationship for width ($B$). Rearranging Eq. (3) gives:

$B = \frac{1}{D_{50} \sqrt{R_g D_{50} \alpha (\phi \tau_{form}^{\star} - \tau_{cr}^{\star})}} Q_s$ \hspace{1cm} (A.1)

i.e., Eq. (5).

Next, we will focus on the relationship for depth ($H$). We start by rearranging Eq. (3) to get unity:

$1 = \frac{1}{B D_{50} \sqrt{R_g D_{50} \alpha (\phi \tau_{form}^{\star} - \tau_{cr}^{\star})}} Q_s$ \hspace{1cm} (A.2)

We then rearrange Eq. (4) to isolate $H$:

$H = \frac{\tau_{form}^{\star} R D_{50}}{S}$ \hspace{1cm} (A.3)

Then divide Eq. (A.3) by Eq. (A.2):

$\frac{H}{S} \sqrt{\frac{R_g D_{50} \alpha (\phi \tau_{form}^{\star} - \tau_{cr}^{\star})}{Q_s}} = \frac{\tau_{form}^{\star}}{BD_{50} \sqrt{R_g D_{50} \alpha (\phi \tau_{form}^{\star} - \tau_{cr}^{\star})}}$ \hspace{1cm} (A.4)

Now rearrange Eq. (2) to get unity, simplify, and take the square root:

$1 = \frac{\sqrt{C_f} Q_w}{B \sqrt{H^2 g S}}$ \hspace{1cm} (A.5)

Rearrange Eq. (4) to get unity, then raise everything to the power of 3/2:

$1 = \frac{\sqrt{C_f} Q_w}{B \sqrt{H^2 g S}} \sqrt{\frac{\tau_{form}^{\star}}{3 R D_{50}^{3/2}}}$ \hspace{1cm} (A.6)

Now, multiply Eq. (A.4) by Eqs. (A.5) and (A.6):

$H = \frac{\tau_{form}^{\star} R D_{50}}{S} \sqrt{R_g D_{50} \alpha (\phi \tau_{form}^{\star} - \tau_{cr}^{\star})} \sqrt{\frac{C_f Q_w}{Q_s}} \frac{\sqrt{S_f}}{S_f}$ \hspace{1cm} (A.7)

which simplifies to Eq. (6):

$H = \frac{D_{50} \alpha (\phi \tau_{form}^{\star} - \tau_{cr}^{\star})^{3/2} \sqrt{C_f} Q_w}{\sqrt{\tau_{form}^{\star}} Q_s}$ \hspace{1cm} (A.8)

Finally, we will derive the relationship for slope ($S$). Start by rearranging Eq. (4) to isolate $S$:

$S = \frac{\tau_{form}^{\star} R D_{50}}{H}$ \hspace{1cm} (A.9)

Multiply Eq. (A.9) by Eq. (A.2):

$S = \frac{\tau_{form}^{\star} R D_{50}}{H} \frac{1}{B D_{50} \sqrt{R_g D_{50} \alpha (\phi \tau_{form}^{\star} - \tau_{cr}^{\star})}} Q_s$ \hspace{1cm} (A.10)

Substitute Eq. (A.9) into Eq. (A.5), and simplify:

$1 = \frac{\sqrt{C_f} Q_w}{B H \sqrt{\tau_{form}^{\star} R D_{50}^{3/2}}}$ \hspace{1cm} (A.11)

Now divide Eq. (A.10) by Eq. (A.11):

$S = \frac{\tau_{form}^{\star} R D_{50}}{H} \frac{1}{B D_{50} \sqrt{R_g D_{50} \alpha (\phi \tau_{form}^{\star} - \tau_{cr}^{\star})}} Q_s \frac{B H \sqrt{\tau_{form}^{\star} R D_{50}^{3/2}}}{\sqrt{C_f} Q_w}$ \hspace{1cm} (A.12)

which simplifies to Eq. (7):

$S = \frac{R \left(\tau_{form}^{\star}\right)^{3/2}}{\sqrt{C_f} (\phi \tau_{form}^{\star} - \tau_{cr}^{\star})^{3/2}} \frac{Q_s}{Q_w}$ \hspace{1cm} (A.13)

Supplementary materials


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