

Research Paper



Curiosity meets the Gediz Vallis Ridge: Remnants of a late-stage debris-flow dominated fan in Gale Crater

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ABSTRACT

Recent studies using HiRISE and CTX imagery show that fan-shaped deposits are widespread especially across the walls of craters on Mars and likely formed during the Hesperian and Amazonian periods. Remote imagery and topographic analyses suggest that they may have been produced by both fluvial and debris-flow processes and are generally distinct from deltaic deposits such as those in Jezero. *Curiosity* rover-based observations of the deeply eroded deposits of the Gediz Vallis Ridge (GVR), lying at the base of Aeolis Mons (Mt. Sharp), now provide our first cross-sectional observation of a steep, central peak-based Martian fan, revealing episodic deposition by debris flows, and reworking by fluvial processes, and likely fluvial sheetfloods. The fan remnants today rest on top of the Stimson formation, a lithified aeolian sandstone that mantles the Greenheugh pediment. Above the Stimson is a discontinuous, fine-grained ($\leq$ gravel size), Basal unit with decimeter-scale bedding that is concordant with the top of the Stimson. The unit reaches a maximum observable thickness of 15 m and is unconformably overlapped by stratified gravel and boulder deposits. Discontinuous layering within these deposits ranges in thickness from 0.25 to 3 m and exhibits a mean slope of 7% oriented parallel to the ridge axis, which is much gentler than the slope of the basal contact. At progressively higher elevations, five depositional packages are exposed. Each package is composed of two facies whose frequency and lateral extent differ: 1) a light-toned, boulder-rich coarse facies, typically matrix-supported, with scattered erosionally resistant dark-toned gravel and boulders, and 2) a dark-toned gravel and boulder facies, typically clast-supported. Both facies have median grain sizes of about 180 mm and host 1–3 m boulders scattered throughout. The light-toned strata are interpreted as debris flow deposits. The dark-toned strata are interpreted to be fluvial, probably derived from surface runoff that concentrated dark clasts from the light-toned unit into channels and sheetflood deposits while

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disaggregating the weaker light-toned sediment during transport. The gentle slopes of these deposits indicate that the depocenter back-stepped up the steep underlying pediment surface, and probably formed the core of fan that may have extended as much as 10 km downslope, likely in the late Hesperian or early Amazonian. Most of the fan may have been composed of the light-toned sediment highly erodible by wind, leading to a near complete removal of the fan deposit after the cessation of fan construction. Stratigraphic differences in the five packages suggest significant changes in the magnitude and frequency of surface runoff and sediment supply likely reflecting variable climate conditions. Reconstructed fan sediment volume estimates indicate that at least many tens of meters (and likely much more) of precipitation-generating runoff from Mt. Sharp were needed to build the fan, and this involved thousands of runoff events. This variability supports proposals based on orbiter data that Martian fans are characterized by mixtures of debris flow and fluvial deposits and that fan development was episodic, rather than a single pulse. The GVR fan records one of the last significant hydrogeomorphic phases in Gale and likely the youngest observed in situ by *Curiosity*.

1. Introduction

The discovery of alluvial fans on Mars (Moore and Howard, 2005) and the subsequent studies that have now documented a global distribution across Mars (e.g., Wilson et al., 2021; Mondro et al., 2023; Morgan et al., 2022) have made alluvial fan investigations central to understanding the post-Noachian climate history of Mars. The transport of sediment downslope onto alluvial fans requires liquid water, either sufficient to carry sediment by rivers or sheetfloods, or to mobilize massive poorly sorted mixtures of sediment as mud (fine grained) or debris (coarse grained) flows (Blair and McPherson, 1994). The sediment comprising the fans must also be mobilized from a source area (catchment), which calls for understanding what climate-influenced physical and chemical processes comminute source bedrock, releasing transportable detritus that feeds the fan. Alluvial fans are accumulations of shed debris, with some Martian fans reaching over 1000 km² in area and 850 m in thickness (Morgan et al., 2014), and, as such, require liquid water-driven activities to act over potentially significant time and space. Water quantity, sediment production, and fan accretion all provide guides to the nature and duration of past sufficiently wet events on Mars. There is also an erosional record etched into the fans of Mars creating canyons and inverted topography that implies times of sufficient atmosphere density and winds to drive abrasion and entrainment of deposited sediments, most likely under atmospheric conditions different from current Mars.

With the arrival of the Mars Orbiter Camera (MOC), in 1997 (Malin et al., 1992; Malin and Edgett, 2001) which enabled the early discovery of fans, and then the launch in 2005 of the Mars Reconnaissance Orbiter (MRO) with onboard Context Camera (CTX; Malin et al., 2007; Dickson et al., 2024), which provides now near global coverage at 6 m resolution, and the High Resolution Imaging Science Experiment (HiRISE; McEwen et al., 2007, 2023), which acquires data at 25 cm pixel-scale over selected areas, a rapidly increasing catalogue of site investigations and regional to global comparisons have been reported. In 2022 and 2023 alone there have been several global compilations (Morgan et al., 2022; Mondro et al., 2023; Telbot and Goudge, 2022) and, among many other local studies, two independent in-depth investigations of the region north of Hellas basin (Anderson et al., 2023; Zhang et al., 2023) that provide insight about the fan timing and construction. The surveys reported for global compilations collectively raise the issue of distinguishing alluvial fans from fan-deltas, deltas, or other fan-shaped features of uncertain origin. Both Morgan et al. (2022) and Mondro et al. (2023) start with the 1049 fan-shaped landforms reported by Wilson et al. (2021) and these studies add 452 and 333 new fan features, respectively. Different criteria were then used to distinguish these features, and different terms were used for mapping, affecting their relative abundances. The new features frequently differ in locations and type assignment. There is general agreement, however, that alluvial fans are found across Mars with Morgan et al. (2022) reporting them found to 40 degrees south, while Mondro et al. (2023) report alluvial fans south to 78 degrees. Both surveys, while acknowledging the occurrence of deltas, were cautious about declaring features as deltas on their maps. Telbot

and Goudge (2022) used a slope break across exposed stratigraphy to indicate the occurrence of a delta and in the examination of 14 fan features concluded 11 did not meet the criteria for a delta. They emphasized the importance of significant post-deposition wind erosion making it impossible to evaluate the delta hypothesis for these features from existing orbital data.

It is the timing of alluvial fan construction that spotlights their significance in documenting the climate history of Mars. Morgan et al. (2022) and Anderson et al. (2023) summarized studies that show that most of the fans post-date the Late Noachian to early Hesperian Periods of extensive valley network development across Mars. The relatively small surface areas of alluvial fans compared to those required for accurate crater-counting statistics can make conventional crater counting to estimate ages subject to large uncertainty, especially given the eroded state of most surfaces (e.g., Palucis et al., 2020), but can yield useful results if fan areas are combined and there is supporting information indicating it is valid to do so (e.g., stratigraphic relationships and relative scale of preserved features. For example, activity on fans in Margaritifer Terra (Grant and Wilson, 2011, 2012) and even within Gale (Grant et al., 2014; Grant and Wilson, 2019; Grotzinger et al., 2015) likely extended into the Amazonian. In addition, early work by Williams and Malin (2008), documented fans in the Mojave crater whose impact age could be constrained from crater counts of the large circum-Chryse outflow channels it struck. Tanaka (1997) reported these outflow channels to be Late Hesperian or Amazonian. Holo et al. (2021) developed a statistical model that used the global geologic map reported by Tanaka et al. (2014), a global database of Martian impact craters (Robbins and Hynek, 2012a, 2012b), and the Wilson et al. (2021) global database of alluvial fans to estimate the timing and duration of alluvial fan formation. Their modeling also suggests that fan formation persisted well into the Amazonian. Morgan et al. (2022) added additional observations on impact crater ages that host alluvial fans across Mars that supports Hesperian and Amazonian ages of alluvial fans. These studies suggest that fan construction was not a singular event, but rather occurred at various times and places across Mars during an extended period of time of the Hesperian and Amazonian.

Given the widespread occurrence of alluvial fans well after the vast majority of valley network development period (e.g., Carr, 1995, 1996, 2006; Fassett and Head III, 2008; Alemanno et al., 2018), emphasis has been placed on deciphering the processes of fan construction and the duration of fan construction to provide further constraints on the nature of Martian climate that generated these features. Morgan et al. (2022) reported a tendency of crater-wall fans to be located on north, south and east facing slopes and to be located in the vicinity of local highest topography of the crater rim, but at lower elevations than the earlier valley networks. They proposed such locational attributes suggest a snow melt runoff mechanism to wet up and drive sediment transport.

In this paper we examine a 3-km-long ridge of coarse sediment downslope of Gediz Vallis along the lower flanks of Mt. Sharp in Gale Crater and propose that it likely records the remnants of a fan. We first review terrestrial and Martian empirical analyses of fan morphology and processes of formation that will help guide our analyses and place the

work in a broader context. We then describe the site and summarize previous work. This is followed by a description of a Basal sandy unit followed by an analysis of five coarse bedded distinct stratigraphic units that lie in the Gediz Vallis Ridge and likely are the remnant of a once extensive fan. The sedimentology indicates a dominance of debris flow deposits with interbeds of fluvial lenses and sheetflood deposits. The upper four units back-stepped up a steep slope as the fan aggraded. We complete our analysis with a discussion of the implications for both wet periods to build the fan and an extended dry period of wind much more erosive than present that drove the massive erosional removal of the fan.

1.1. Fan morphometric analysis and process inference

Early terrestrial studies on Earth suggested that the relationship between catchment area and fan area, as well as fan slope and fan area, may provide evidence of process controls of fan evolution (e.g., Bull, 1964, 1968; Denny, 1967) and many subsequent papers have reported power law dependencies (see review in Woor et al. (2023). Morgan et al. (2022) performed such analyses on their compilation of previous work and new observations across Mars and compared observed relationships with data from terrestrial studies. They proposed, based on comparison with terrestrial studies, that Mars fans were steeper than fans on Earth and suggested that this could be due to the reduced gravity of Mars.

Fig. 1 compares the Morgan et al. (2022) Mars compilation with a

collection of terrestrial data used by Morgan et al. (2022), but also a much larger data set, some of which post-date the Morgan et al. (2022) publication. Instead of comparing the different data by showing just regression lines through compilations (e.g., Morgan et al., 2022; Woor et al., 2023), the individual data points are shown here for selected data sources to show the wide variability of data. Added to the Morgan et al. (2022) collection are the data from the Hellas region (Anderson et al., 2023). The Earth data sources are derived from: 1) selected data from the Palucis et al. (2022) in which the field measurements were made to document dominant transport processes (debris flow, mixed, and fluvial), 2) Bull (1964) who documented fans generated by a mixture of mudflows and fluvial transport in the eastern coastal mountains of California, 3) Woor et al. (2023) who recorded fan morphology on the eastern side of the Hajar mountains in southeastern Arabia in which fluvial processes likely dominated (RK in Fig. 1), 4) Tomczyk (2021) high Arctic debris flow and fluvial fans, 5) Saito and Oguchi (2005) alluvial fans of Japan, Taiwan, and the Philippines (with four specific named example fans), nearly all fluvial dominated, and 6) debris flow and fluvial dominated fans in the Death Valley area reported by Blair (1999). Two data points from the Atacama desert are included (Morgan et al., 2014) where pebbly mudflows generated overbank deposits and, in places, left paired boulder lines that tracked the width of the flows. Morgan et al. (2014) compared this site with observations they made in the Saheki crater (a representative data point is shown).

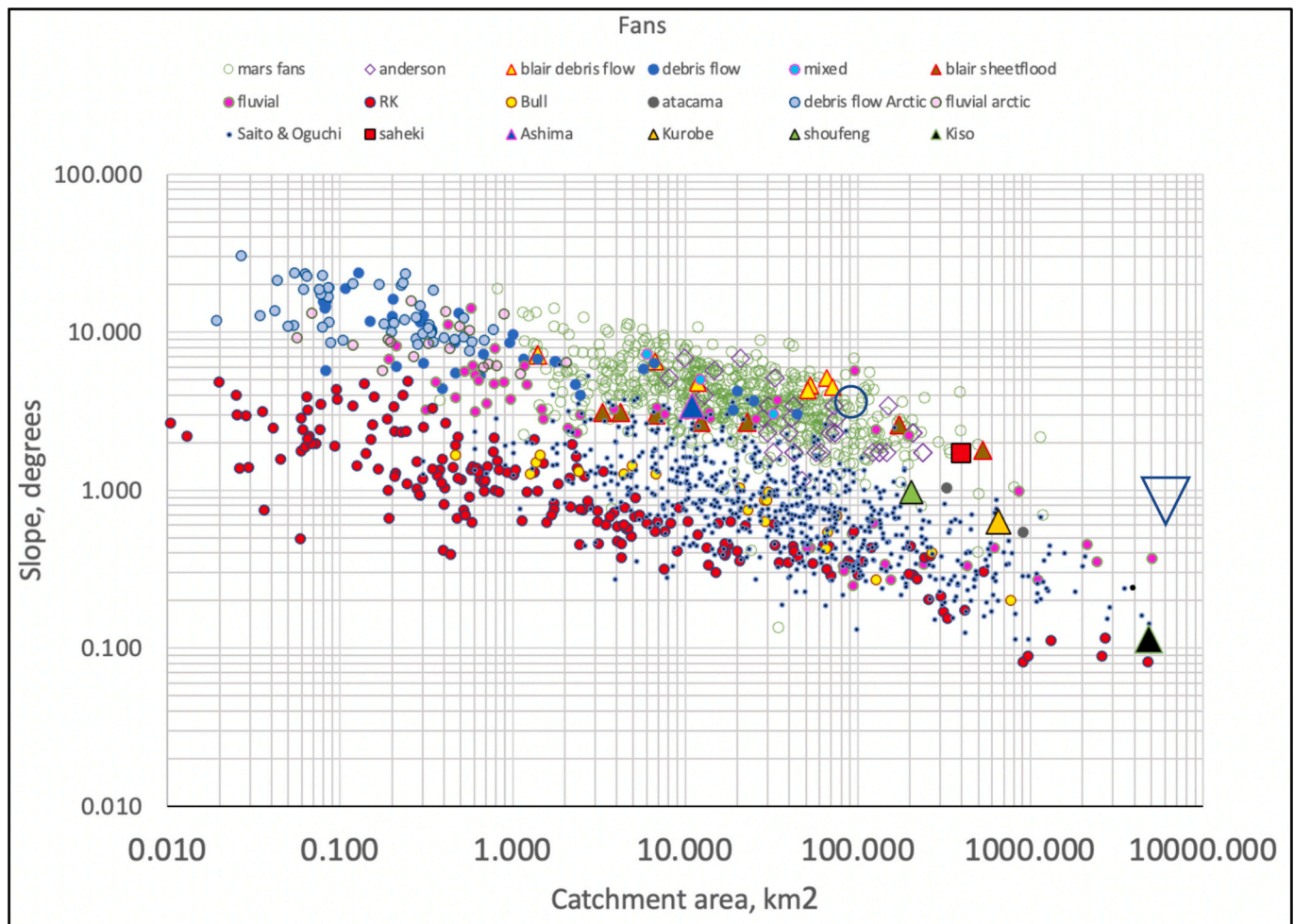


Fig. 1. Plot of fan slope vs. Catchment area for Earth and Mars. Data compiled for terrestrial fans are from: Blair debris flow (Blair and McPherson, 2009), debris flow, mixed, blair sheetflood (Blair and McPherson, 2009), fluvial (Dade and Verdeyen, 2007), RK (Woor et al. (2023) – Arabian fans), Bull (Bull, 1964), Atacama (Haug et al., 2010), Debris flow arctic (Palucis et al., 2022), Fluvial arctic (Palucis et al., 2022), Saito & Oguchi (Saito and Oguchi, 2005), Ashima (Saito and Oguchi, 2005), Kurobe (Tebakari et al., 2013), Shoufeng (Saito and Oguchi, 2005), Kiso (Ohata et al., 2023). Martian fan data compiled from: Mars fans (Morgan et al., 2022; Mondro et al., 2023), Anderson (Anderson et al., 2023), Saheki (Morgan et al., 2014). The large inverted triangle indicates the Jezero delta (Mangold et al., 2021).

Because of the typical concave-up profile of bedrock river channels (e.g., Sklar and Dietrich, 1998), the larger the drainage, the gentler the slope. This leads to the central tendency for mean slopes of fans to decline with increasing source area (Stock, 2013). Also, the larger the catchment area the greater the likelihood that the main channel that leads to the fan apex will decline to a slope below that which transmits debris flows and consequently, fans from large catchments will be likely dominated by fluvial processes with gentler slopes.

The Mars data mostly cluster in a well-defined power law relationship, but with a broad range of fan slope for a given catchment area. Only two fans smaller than 1 km² were included in the Morgan et al. (2022) plot. In the Hellas region, Anderson et al. (2023) compared estimated fan volumes with catchment eroded volumes and found commonly that the fan volume was smaller than the missing volume in the source catchment. They propose this is evidence of significant post-depositional wind erosion. They also found in the two largest fans (in the Sahel crater) that the fan volume exceeds the missing catchment area volume; the mechanism for this was not clear. The Anderson et al. (2023) data nonetheless plot in the Mars data field of Morgan et al. (2022) (Fig. 1). They classified the fans by the presence of chutes, inverted channels, or degraded and used morphology of channel features to infer discharge, interpreting these fans as being produced by fluvial processes.

Many terrestrial fans generated by debris flows, fluvial processes and mixtures of the two processes plot within the Mars data field. This is particularly clear in the Blair (1999) data in which process-dominance classification was based on extensive field documentation of stratigraphy. The Atacama fans overlap with the largest Mars fan, whereas the Saito and Oguchi (2005) data overlap the lower slopes of the high density of Mars data. Some Mars fans are as gentle as the least steep fans in the terrestrial compilation (i.e., RK and Bull). What is distinctive about the Mars fans is not that they are exceptionally steep, or necessarily dominated by mass flow processes (e.g., Anderson et al., 2023) but rather the range of fan slopes for a given catchment area on Mars is generally much lower than for a broad sampling of terrestrial fans. As Fig. 1 shows, fan systems on Earth can be both much gentler and much steeper than that found on Mars. The range of conditions under which fans form on Earth are much wider. Terrestrial tectonism, subsidence, widely varying lithology and weathered state, precipitation intensity and magnitude, topographic setting, and other controls may lead to much gentler slopes than found on Mars fans for the same catchment area. de Haas et al. (2024) show a plot of fan slope versus drainage area from a global compilation focused on debris flows, but much of their data area not reported so could not be incorporated into this plot. Their data also show many sites where fan slopes for a given drainage area are considerably steeper than that reported for Mars.

Plots of terrestrial fan slope versus fan catchment area have been used to infer influences of lithology (grain size and sediment supply rate), climate, tectonics, and process dominance (fluvial (both channelized and sheetfloods), debris flows, mudflows, and hyper-concentrated flows) (e.g., review in Bowman (2019) and Stock (2013)). Blair (1999) proposed that adjacent Death Valley fans differed in slope and dominant transport process due to lithologic differences. Bull (1964) concluded that the California fans in a semi-arid climate were built by fluvial processes and mudflows, and these mudflows had exceptionally high clay content, averaging 31%. Bull observed that the source basin lithology produces limited amounts of both gravel and silt. This absence of coarse debris and the ability of mudflows to travel to gentle gradients may have led to low slopes. Woor et al., 2023 note inverted sinuous channels on the RK fans and propose that the low gradients of these fans may be due to a dominance of fluvial processes during Quaternary wet events driven by Indian Ocean Summer Monsoon incursion (in an otherwise arid to hyperarid environment). The fans compiled by Saito and Oguchi (2005) developed in humid environments (1000 to 4000 mm annual precipitation). These data lie mostly above the Bull and RK data and below the mixture of Mars and terrestrial data

that include the Death Valley fans. Four fans described by Sato and Oguchi are shown in Fig. 1. The blue triangle is the Ashima River in Japan that is braided and mantled with large boulders (delivered by debris flows), the yellow triangle is the Shoufeng River in Taiwan, which transports pebbles across the fan, the green triangle is the Kurobe River in Japan, which is braided across the fan and the maximum grain size declines from 400 mm to 100 mm downslope (Oguchi, 1997), and the Kiso River in Japan, which is braided with a modal grains size of 64 mm just before the toe. These observations indicate that the low fan slopes for a given catchment area reported in Saito and Oguchi can occur with braided gravel bedded channels. Collectively, these terrestrial data show that fan slope for a given catchment area is not a simple function of process (debris flow versus fluvial), or grain size (gravel rich or poor), or climate (wet versus arid), or lithology, although all these factors matter for specific locations.

In their review of terrestrial fans, Blair and McPherson (1994) conclude that debris flows tend to deposit on fans at slopes of 5 to 15 degrees while sheetfloods occur on slopes of 2 to 8 degrees. Subsequently, Blair (2003) noted mudflow deposits (low in coarse debris) contributed to building the Cucomongo fan in Death Valley, which has a mean slope of 1.8 degrees. Whipple and Dunne (1992) in their study of fans in the Owens Range of California observed that high concentration flows tended to deposit above slopes of 4.5 degrees while low concentration flows could extend to 1 to 2 degrees. Stock and Dietrich (2003, 2006) in their study of canyon channels (upslope of the fan apex) report that debris flows rarely travel below slopes of 1 to 3 degrees.

These observations do indicate coarse debris flows and mudflows differ in the depositional slopes they build on fans, and tend to be steeper than fluvial, but the slope overlaps are significant, making an interpretation of process dominance based on slope alone uncertain. Some have elected to plot fan slope against a measure of the catchment source area mean slope (the Melton index) (e.g., Marchi et al., 1995). While correlations are found, process distinction still falls largely along threshold slope differences between debris flow and fluvial deposits. Many have noted that fans may shift from debris flow dominance near the apex of the fan to fluvial towards the distal end (e.g., Saito and Oguchi, 2005; Bowman, 2019).

Morgan et al. (2022) argued that debris flows should cease motion where the boundary shear stress falls below the yield strength of the flow (e.g., Whipple and Dunne, 1992). Because boundary shear stress can be estimated as the product of fluid density, gravity, flow depth and surface slope, it follows that the surface slope of the deposit at cessation of motion depends on the yield strength and inversely with gravity. Hence, lower gravity on Mars should lead to steeper depositional slopes. Theory and experiments, however, show that mobilization of flows is strongly influenced by the shear-induced dispersive pressures and the permeability of the supporting fluid matrix (e.g., Pierson, 1981; Costa, 1984; Iverson et al., 1997). Sustained mobilizing pore pressures are also strongly influenced by the grains size distribution of the coarse fraction (e.g., Kaitna et al., 2016; Chen et al., 2023). Under such controls, gravity may cancel out the various force balances, or it may lead to Mars flows transporting onto lower slopes if the sustained dispersive pressures occur because of reduced settling velocity of large clasts due to reduced gravity. The role of gravity in three of the most widely used debris flow models (e.g., Barnhart et al., 2021, FLO2D, RAMMS, and D-Claw has not been explored, and only D-Claw attempts a full accounting of the effects of dispersive pressures.

Remote sensing observations have been used to estimate debris flow and fluvial activity and duration on Martian fans based on surface features. For example, Williams et al. (2011) concluded that a debris flow occurred on a fan based on “rugged texture” and the presence of boulders. Morgan et al. (2014) examined the strongly wind scoured fans in the Sahel crater, and proposed that extensive erosion of stratified light-tone material occurred because it was composed of erodible mudflow deposits and the emergent flat topped intervening ridges recorded gravel transporting fluvial systems. They observed coarse boulders in places on

the ridges and estimated that the sizes decreased down fan, which they argue suggests a fluvial origin. Inverted channels on fans have been used to identify fluvial activity and their widths have been used to estimate water discharge (e.g., Anderson et al., 2023; Hayden et al., 2019; Morgan and Craddock, 2019; Palucis et al., 2014). Remote sensing of the larger fans on Mars indicates they are not the product of a single short lived construction event. Using superposition and crater preservation, Anderson et al. (2023), for example, argue that the fans in the Hellas region were episodically active from the Hesperian into the Amazonian periods, indicating multiple transient climatic events. This interpretation is consistent in general with the proposal by Kite et al. (2017) based on observed craters interbedded in fans that large alluvial fans may have built over 10s of millions of years.

On terrestrial fans, the primary indicator of the agent of transport and deposition is the surface expression of features (e.g., levees, bars, and channels) and the grain sizes, sorting and stratigraphy revealed in exposures (Blair and McPherson, 1994). Debris flows are characterized by poorly sorted, matrix supported clasts and a general lack of internal stratification locally. The clasts can reach meters in size with elongated particles that may be randomly oriented. The mud component can be abundant but may be well less than 10% of the mass and the clay size content can be zero (e.g., Iverson et al., 1997). Mudflows show similar patterns, but the frequency and size of clasts are smaller. Fluvial deposits display grain to grain contact, sorting, elongated particles lying parallel to the bedding, and fine to large scale stratification. Such criteria were used to categorize terrestrial fans in Fig. 1.

Grain size, sorting, and fine-scale stratigraphy, however, cannot be observed from orbital data on Mars. Williams et al. (2013) reported in Gale Crater the first rover-based evidence of fluvial transported sediment on Mars. While close to the toe of the Peace Vallis fan, the fluvial deposits were not directly part of the current extent of the fan. Mangold et al. (2021) report grain size of a gravel unit in the Jezero delta and use grain size and channel dimension to estimate stream discharge. Here we provide a detailed sedimentological analysis of the Gediz Vallis ridge deposits that record a much more extensive fan deposit.

2. The Gediz Vallis ridge: regional context and previous work

Across the northwest margin of Mt. Sharp in Gale crater, the strongly layered strata of the lower mound are dissected by a series of wind-etched valleys. One of these valleys, Gediz Vallis, comprises the outlet of a 125 km² drainage basin (Fig. 2). Downslope, emanating from the steep-walled valley rests a narrow (3 km long, 0.1–0.5 km wide) highly eroded ridge known as the Gediz Vallis Ridge (GVR) (Fig. 3; S1 and S2 in Supplemental Information S3). In contrast to the sizable upslope drainage basin, the volume of sediment encompassed by the modern Gediz Vallis Ridge is only ~0.018 km³ (Palucis et al., 2016). The GVR, consisting of a thick (up to 70 m) succession of layered sediments, lies stratigraphically above the Greenheugh pediment (e.g. Bryk et al., 2019; Figs. 2 and 3) and represents the erosional core of a once more extensive depositional fan (Anderson and Bell III, 2010; Palucis et al., 2016; Bryk et al., 2019).

All previous studies agree that the highly eroded deposit represents the remnants of an alluvial fan at the mouth of Gediz Vallis, but the depositional process that built the fan, as well as the former extent, have been debated since before *Curiosity* landed on Mars. Several studies have suggested a fluvial origin for the fan deposit – pointing to a sinuous channel-like feature in the base of Gediz Vallis that connects Gediz Vallis, to the Gediz Vallis Ridge (e.g., Thomson et al., 2008; Anderson and Bell III, 2010). Anderson and Bell III (2010) also suggested the fan once completely covered only the Greenheugh pediment and is responsible for the shape of the pediment. A closer inspection suggests that the shape of the pediment escarpment was set primarily by wind-eroded scarp retreat and the geometry of Vera Rubin Ridge (Fig. 2) and that the fan deposited extended across the ridge (e.g., Fraeman et al., 2020). Palucis et al. (2016), using HiRISE data, separated the GVR into 4 distinct deposits based on topographic, tonal, and textural differences. Here, these deposits (from lowest to highest) are termed informally the Tongue, Claw, Nose, and Transition to the upper GVR (Figs. 4 and 5). They hypothesized that the 4 deposits comprised 4 conformably stacked sedimentary packages that recorded a back-

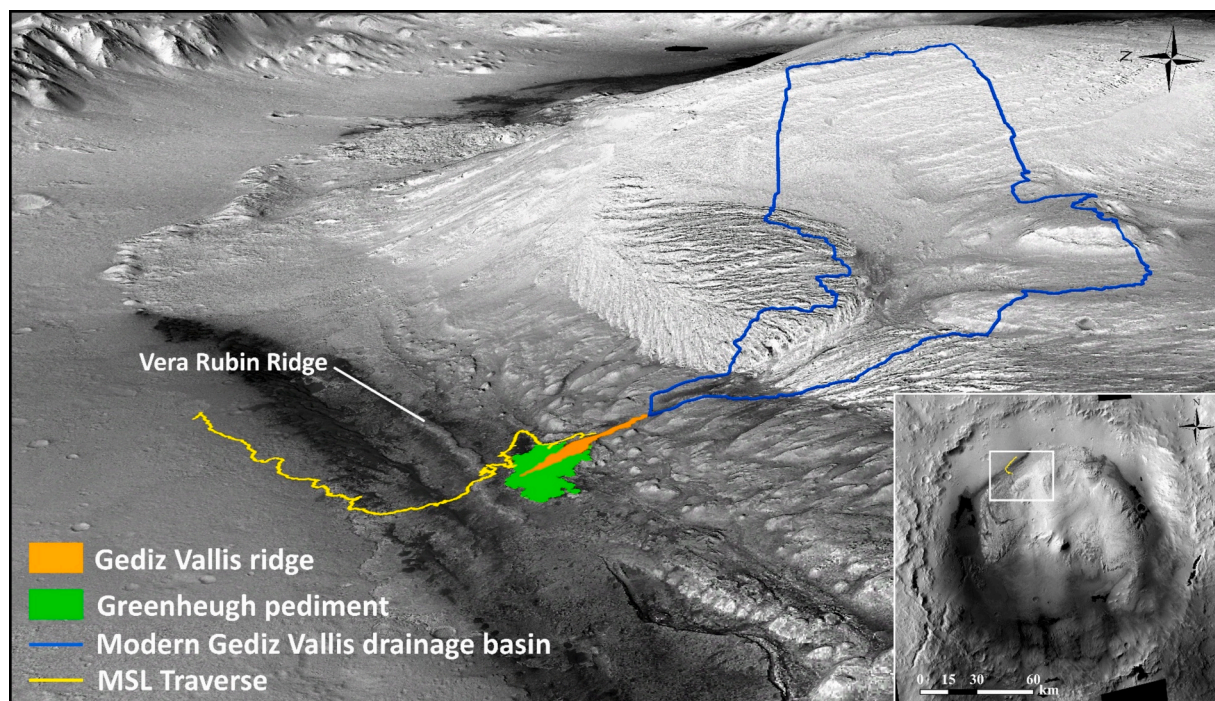


Fig. 2. 3D view facing West showing the NW margin of Mt. Sharp including inset map of Gale Crater made from Context Camera (CTX) imagery on board the Mars Reconnaissance Orbiter (MRO) (Malin et al., 2007). The Gediz Vallis Ridge (orange), the Greenheugh pediment (green), the MSL traverse (yellow) and the modern Gediz Vallis catchment (blue) are highlighted. The inset shows all of Gale crater and indicates (white box) the approximate focus region shown in 3D. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

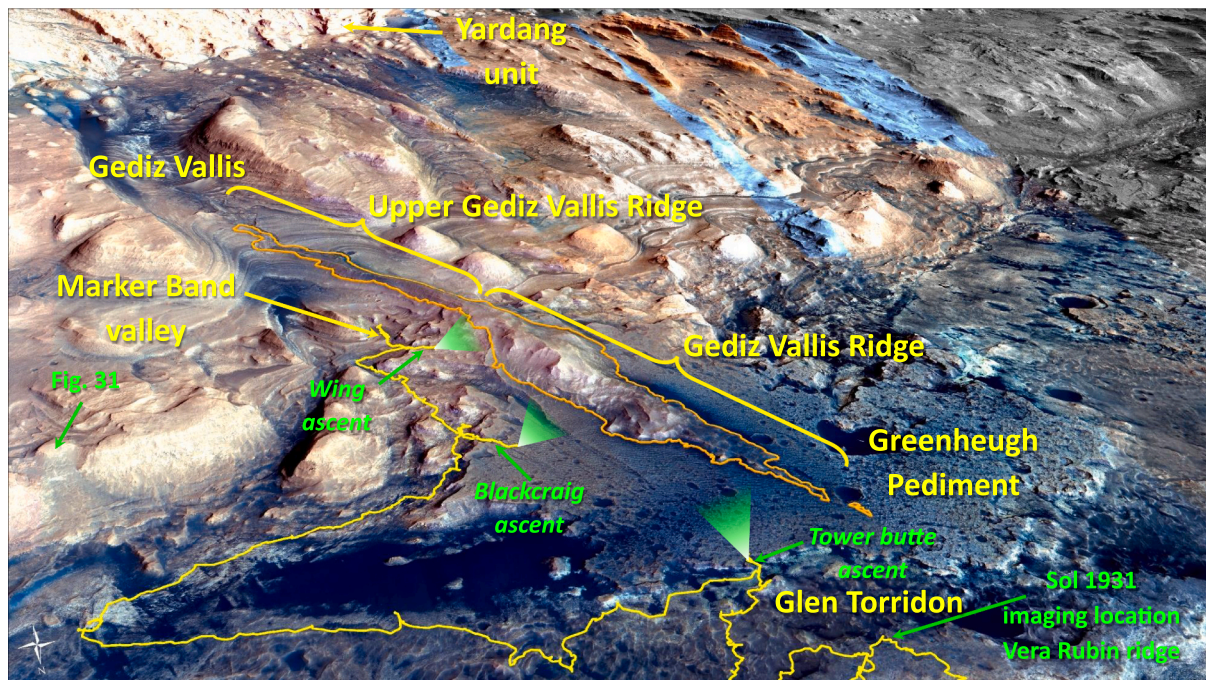


Fig. 3. 3D view of the Mt. Sharp footslope looking towards the south. The MSL traverse (yellow) and the Gediz Vallis Ridge (Orange outline) is also included. Note the braces showing the extents of the upper and lower ridge. Green labels indicate dedicated Gediz Vallis Ridge imaging locations. HiRISE image credit: NASA/JPL-Caltech/University of Arizona. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

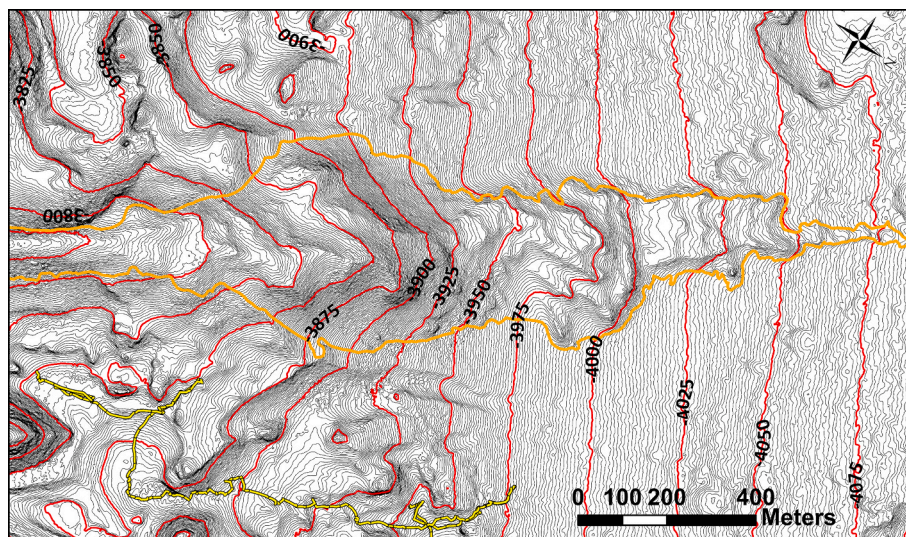


Fig. 4. Topographic map (1- and 25-m contours) of the Gediz Vallis Ridge area (outlined in orange) made from HiRISE stereo imagery. Segments of the MSL traverse are shown in yellow. HiRISE image credit: NASA/JPL-Caltech/University of Arizona. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

stepping aggrading system driven by successive hydrogeomorphic events. Specifically, they suggested that the Claw and the Transition to the upper GVR may have been influenced by the presence of large ephemeral lakes and therefore may contain deltaic deposits. Later, using a combination of HiRISE and Rover-based data, Bryk et al. (2019) qualitatively demonstrated that the GVR stratigraphy is primarily constructed of layered boulder-rich deposits, which are dipping significantly more gently than the surface on which they were deposited (the Greenheugh pediment).

The rover-based field study presented here confirms the existence and distinction of the 4 depositional units proposed by Palucis et al. (2016) and evaluates, using rover-based imagery, the facies,

stratigraphic architecture, and likely depositional processes of each. In addition to the 4 packages defined by Palucis et al. (2016), we include two additional packages: a Basal unit defined stratigraphically and observed by *Curiosity* on the NE side of the ridge, and a transition zone between the Claw and the Nose (Fig. 6). The remainder of this manuscript will be separated according to each of these features: *Basal unit*, *Tongue*, *Claw*, *Claw-Nose transition*, *Nose*, *Transition to upper GVR (uGVR)*. We also include a section devoted to the initial conditions prior to GVR deposition. For each feature, HiRISE orbital data and rover-based imagery collected by *Curiosity* will be presented including graphic logs and grain size data, followed by a brief process-interpretation. Finally, we include a reconstruction of the likely extent of the fan and demonstrate

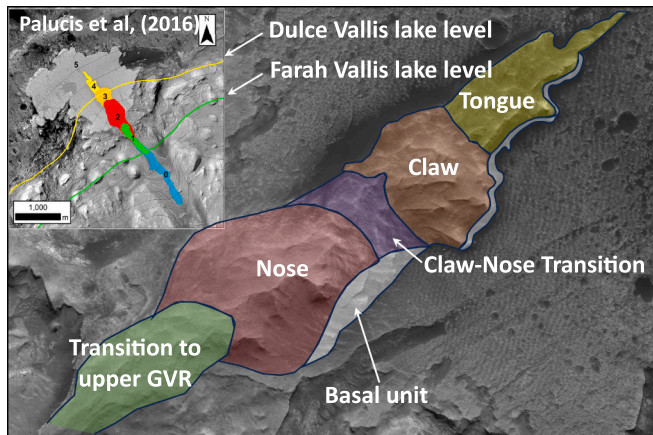


Fig. 5. Annotated HiRISE image of the Gediz Vallis Ridge showing 5 of the 6 distinct units (Tongue, Claw, Claw-Nose transition, Nose, and Transition to upper GVR) analyzed in this study. The inset shows the similar unit definitions (units 1–4) of Palucis et al. (2016) as well as two of the three lake highstands they identified. HiRISE image credit: NASA/ JPL-Caltech/University of Arizona.

that the fan could have been built through a combination of back-stepping fluvial and debris flow processes in the absence of baselevel control.

3. Field investigation of the Gediz Vallis ridge

Our Investigations of the Gediz Vallis Ridge relied primarily on imagery generated by Curiosity's Mastcam 34- and 100-mm lens cameras (Malin et al., 2017; Bell III et al., 2017), long-distance imaging using the ChemCam long distance Remote Micro-Imager (ChemCam LD RMI) (Maurice et al., 2012; Wiens et al., 2012), and Curiosity's navigation cameras (Navcam) (Maki et al., 2012).

Orbital mapping was completed in ArcGIS using a HiRISE (McEwen et al., 2007) MSL Gale Merged Orthophoto Mosaic 25 cm/pixel base map

created by Calef and Parker (2016). Elevation data were derived from HiRISE stereo pairs calibrated to Mars Global Surveyor (MGS) Mars Orbiter Laser Altimeter (MOLA) data, by Parker and Calef III (2016). Topographic profiles, bedding orientation measurements and fan reconstructions were completed in MATLAB. Figures showing all of Gale crater (Fig. 1) were created using ArcGIS from data collected by the Context Camera (CTX) on board the Mars Reconnaissance Orbiter (MRO) (Malin et al., 2007). HiRISE-based bedding dip calculations (Section 3.3) were made in ArcGIS. Grain size analysis was completed using both Mastcam and ChemCam LD RMI mosaics. Individual grains were counted by hand using ImageJ (see Supplemental Information S1 and S3 for pixel/distance calibrations).

3.1. Depositional surface and initial conditions

3.1.1. Observations

The exposed pediment surface subjacent to the GVR is capped by the Sinson formation, a lithified aeolian deposit (e.g., Banham et al., 2022, 2024). The top surface of the Sinson near the GVR assumes a washboard-like morphology with 10–12 m wavelength low amplitude (<1 m) ridges and swales, all of which are covered by a rubbly decimeter-scale surface texture (Fig. 7a; Bryk et al., 2025). Sand bordering the GVR deposit obscures the basal contact with the Sinson. The washboard pattern is likely related to the lithified dunes of the Sinson formation (Bryk et al., 2025). Continuation of this washboard surface, with similar character on either side of the GVR suggests that GVR was deposited on top of the Sinson formation. Mastcam imagery at the base of the Tongue shows a sharp contact between the overlying boulder GVR deposit and the Sinson (Fig. 7b).

Fig. 8a shows the GVR on which a profile A-A' has been traced along the basal contact and shown in Fig. 8b. In the middle of the profile at an elevation of ~ -3800 m there is a clear slope break (yellow star in figure) which also roughly corresponds to the most southerly exposure of the Sinson formation at the Greenheugh pediment. This slope break is also found to the west and east adjacent to the GV ridge and is the lowest of the major retreating escarpments that characterize the NW

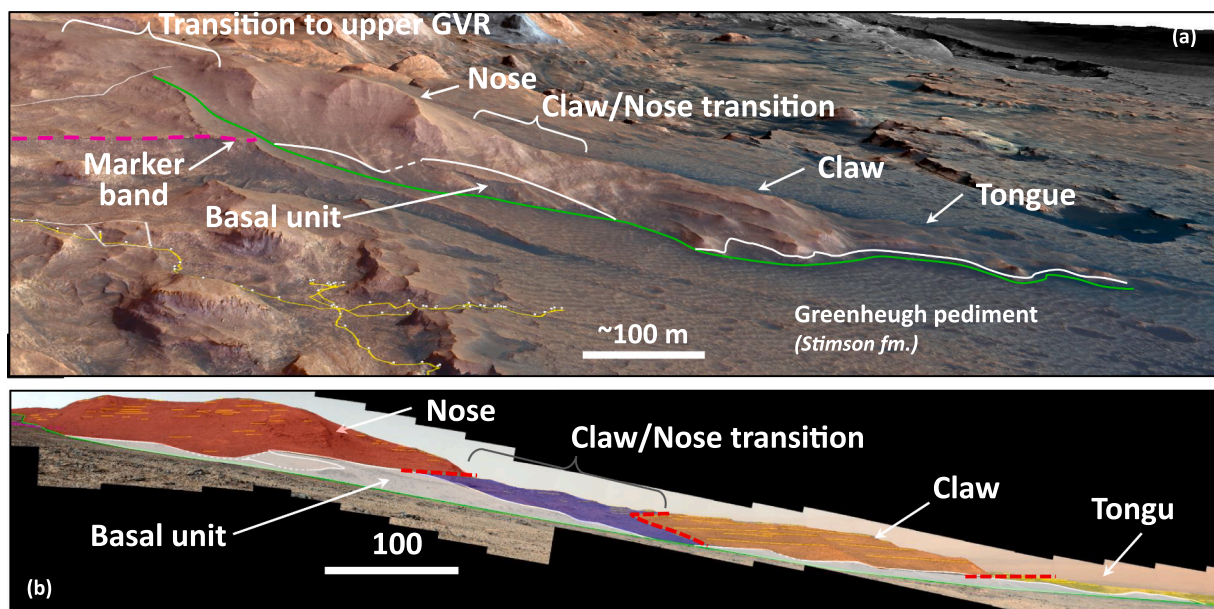


Fig. 6. Annotated cross-sectional views of the Gediz Vallis Ridge. a) 3D HiRISE image of the GVR showing the 6 units (Basal unit, Tongue, Claw, Claw-Nose transition, Nose, and Transition to upper GVR) analyzed in this study. The green line indicates the base of the section above the Greenheugh pediment. b) Mastcam mosaic showing the same unit definitions from (a). Yellow lines denote approximate bedding planes and red dashed lines indicate approximate unit boundaries. Note the transition to the uGVR package is not visible from this angle. HiRISE image credit: NASA/ JPL-Caltech/MSSS. See Table S1 in Supplemental Information S3 for image sequence identifiers. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

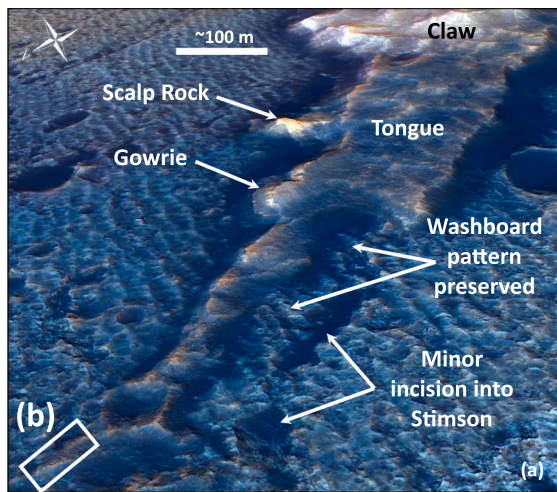


Fig. 7. Distal Gediz Vallis Ridge deposits. a) 3D HiRISE view facing SE showing the distalmost GVR deposits (Tongue) resting atop the Greenheugh pediment including locations of key outcrops Gowrie and Scalp rock. b, Mastcam view referenced to white box in (a). Note the red line delineating the Stimson fm. from the overlying Gediz Vallis Ridge boulders. HiRISE image credit: NASA/JPL-Caltech/Mt. Sharp. Mastcam image credit: NASA/JPL-Caltech/MSSS. See Table S1 in Supplemental Information S1 for image sequence identifiers. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

margin of Mt. Sharp (Bryk et al., 2025) Upslope from this point, the profile corresponds to the contact between upper GVR sediments and the underlying Mt. Sharp stratigraphy, which serves as a reasonable approximation of the valley slope at the time of deposition. The average gradient along the valley floor is 6.4%. The modern entrance to Gediz Vallis lies roughly 1 km up slope of the slope break. Downslope of the yellow star (Fig. 8), the pediment slopes declines, diminishing gradually from >30% near the slope-break to ~10% at the NW edge of the pediment.

3.1.2. Interpretation

Collectively, these observations support the hypothesis that the Gediz Vallis Ridge lies stratigraphically above the Stimson formation, and therefore post-dates it. The Stimson formation has previously been interpreted to be a depositional interlude in the overall denudation of Mt. Sharp (Watkins et al., 2022; Bryk et al., 2025). This stratigraphic relationship suggests that the Gediz Vallis Ridge deposits represent a second, later depositional phase that interrupted the overall denudation of lower Mt. Sharp. It is likely that the GVR deposits are among the youngest depositional features in Gale crater.

The GVR sediments upslope of the yellow star in Fig. 8, known as the upper GVR, are inverted – rising high above the modern valley floor on either side, and extending more than 1 km beyond the modern outlet of Gediz Vallis (Fig. 3). It is likely that these sediments, like the deposits further up slope in Gediz Vallis, were deposited along the valley floor, and only later became inverted developing a channel-like planform morphology after extensive wind erosion of the surrounding bedrock.

The distinctive feature of the GVR topographic setting is that the main feeder system for the fan did not progressively decline towards the Gale crater floor, but instead spilled across an erosionally steepened surface downslope in which arriving debris must have initially accelerated and possibly eroded into the surface. As discussed below, this generally has important consequences for how the resulting fan developed.

3.2. The basal unit

3.2.1. Observations

The Basal unit overlies the Stimson formation across much of the Gediz Vallis. The unit thickness varies greatly from 3 m near Gowrie to 15 m thick upslope (Fig. 6; S3 in Supplemental Information S3), but in few places is absent. The unit is cliff forming, and sheds angular meter-scale boulders down slope (Fig. 9a-d). These boulders are lighter in colour and more angular than boulders in the dark-toned facies or blocks shed from the Stimson.

The Basal unit has decimeter scale bedding and lacks coarse (> gravel size) sediment (Fig. 9b, c). The bedding orientation within the Basal unit is in general concordant with the local pediment slope and dip of the underlying Stimson formation (Fig. 9a). The closest *Curiosity* was able to get to the Basal unit (~420 m), which yielded a highest ChemCam LD RMI resolution of 8.4 mm/pixel (see section S1.4). In Fig. 9c and d, the vertical outcrop face is ~2.3 m thick and exhibits bedding with an apparent dip of 19–22 degrees (~34–40%), 8–10 degrees steeper than the overall 11–12 degree (~20%) dip of the unit in this location. Some areas of the outcrop exhibit convex, concave, and possibly wavy bedding traces. However, it was not possible to identify clear stratal terminations and/or obvious fine-scale (decimeter) cross-stratification (Fig. 9c, d). This area is the only location where the apparent bedding dip within the Basal unit is not parallel to the depositional surface, and the only location in the entire Gediz Vallis Ridge where the apparent dip of the beds is greater than the local slope of the pediment.

The Basal unit contact with the Stimson formation was not observed due to sand cover. In contrast, the top surface is clearly onlapped by the overlying boulder-rich strata (yellow traces in Fig. 9e).

3.2.2. Interpretation

While the absolute grain size of the Basal unit is unknown, a lack of identifiable grains at 8.4 mm/pixel (Fig. S34 in Supplemental Information S3) suggests the grain size of the unit is equal to or smaller than fine gravel. The convex/concave bedding traces (Fig. 9d), although not diagnostic, are reminiscent of fine-scale ripple and dune cross-stratification. The steep apparent dip in Fig. 9c, d compared to the overall dip of the unit more convincingly indicates the presence of larger (1–3 m scale) cross-stratification. While both meter- and decimeter-scale cross-stratification can occur among a wide range of grain sizes (e.g., Ashley, 1990), the observations in Fig. 9 together point to a likely average grain size in the sand size fraction for the Basal unit.

The most likely interpretation is that the Basal unit was formed by an aeolian erg similar to or coincident with the Stimson sandstone. Initial observations of the Stimson at the Greenheugh pediment from Vera Rubin Ridge (see Fig. 3 for location) revealed multi-directional meter-scale cross-stratification at 450–500 m distance within the outcrop, much thinner than the Basal unit in Fig. 9. However, the pediment front outcrops offered more three-dimensionality than the Basal unit of the GVR, and it was not until the rover got much closer that finer (decimeter-scale) cross-stratification could be identified (Banham et al., 2022). Also like the Stimson formation at the Greenheugh pediment, an aeolian origin could explain the concordance between the slope of the pediment and the apparent dips of the Basal unit. Finally, it is common for aeolian dunes to generate low-angle cross-stratification when aggregation rates are low and/or when the dunes are large – as would be expected from a 1–3 m foreset thicknesses. The large planar foresets of the Ladder interval within the Stimson at the Greenheugh pediment (see Banham et al., 2022) could be a reasonable analog for the possible cross-stratification shown in Fig. 9.

The top surface of the Basal unit is onlapped by the overlying boulder-rich stratigraphy (Fig. 9e). This, coupled with the wide variability in thickness of the Basal unit, suggests that this surface is an erosional unconformity preserved by the overlying boulder-rich strata. No angular truncation of the Basal unit internal stratification was exposed. However, if the deposit is aeolian in origin and the top surface

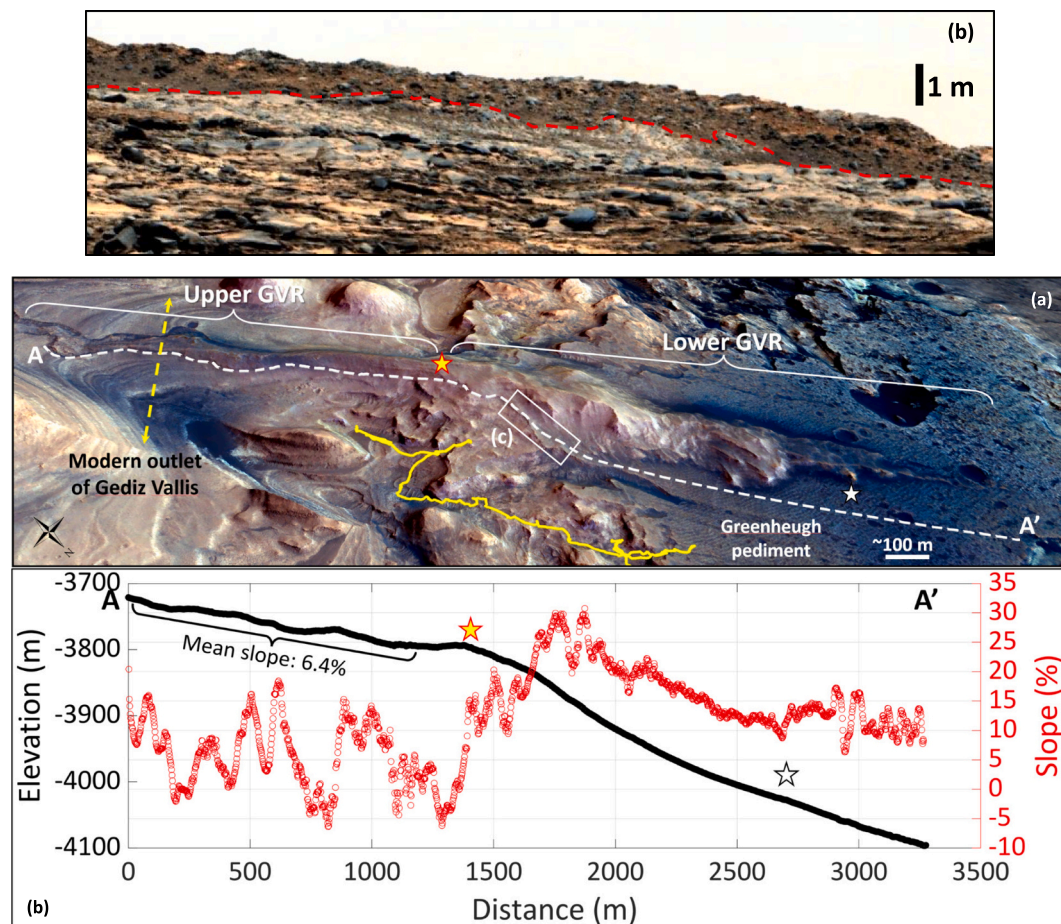


Fig. 8. 3D HiRISE view and topographic long profile of the upper and lower Gediz Vallis Ridge. a) 3D HiRISE view showing the full Gediz Vallis Ridge. The MSL Traverse is shown in yellow. The white dashed line indicates the topographic profile A-A' extracted from HiRISE topography. b) Topographic and slope profiles A-A' taken from (a). Yellow and white star locations reference regions of relatively abrupt slope changes. Red dots record slope values at 1 m elevation intervals smoothed with 5 m spacing. High variability upslope records the variable erosional boundary and the effects of partial coverage of the contact by colluvial debris on the steep border slopes. As shown in the profile the amplitude of variation is small and the mean slope of 6.4% is well defined. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

was not an erosional feature, one would expect to find a level of concordance between the internal stratigraphy and the surface morphology of the deposit. This was not observed anywhere along the Basal unit at the GVR. It is therefore likely that the Basal unit was once a roughly tabular unit like the Stimson formation is at the Greenheugh pediment, and that it was subsequently eroded.

The process of erosion that created the preserved non-tabular form of the Basal unit (e.g., Fig. 6 and S3 in Supplemental Information S3) may have been wind erosion prior to the GVR sediment delivery or driven by the arriving sediment-laden flows. While the unit exhibits 15 m changes in thickness across 10s to 100 s of meters laterally, the top surface of the unit is smooth at the 1–10 m scale (e.g. Fig. 9a). Coupled with the cliff-forming and boulder-shedding properties of the unit, it is likely the deposit was lithified prior to erosion of the top surface. The process of lithification remains unknown, but likely involved at least one liquid water event and subsequent dry period at a time after the deposition, lithification, and significant erosion of the Stimson – all of which occurring prior to initial of boulder deposition in the Gediz Vallis Ridge.

The modern aerial coverage of the Basal unit does not extend beyond the margins of the modern Gediz Vallis Ridge. Nevertheless, our analysis this unit suggests that it once may have extended laterally far beyond its current outcrop exposure, which is limited to where it was recently covered by GVR deposits.

3.3. Stratigraphy above the basal unit

Above the Basal unit there are 5 packages of stacked sedimentary deposits that are characterized by boulder-rich depositional facies and bedding orientations significantly less steep than the surface on which they were deposited. Differential wind erosion of these stacked sedimentary packages has generated distinct geomorphic expression, which we exploit as a means of referencing the individual packages. In stratigraphic order these 5 packages (Units 1–5) are informally named: the Tongue, Claw, Claw-Nose transition, Nose, and Transition to the upper GVR (Fig. 6a). Fig. 10a shows a map of the Gediz Vallis Ridge including 43 bedding attitude measurements compiled from HiRISE orthoimagery and topography. Each of the blue regions was mapped as the surface expression of a bedding plane. Fig. 10b shows a histogram of bedding dips across all 43 locations. The mean dip of the beds is 4° (7%) and the median is 3.9° (6.8%). Compiled dip directions across the entire GVR are shown in the form of a polar histogram in Fig. 10c. The average dip direction among all GVR measured data is 315 degrees – roughly parallel to the axis of the ridge. Dip direction diagrams split out among the lower 4 units (Tongue, Claw, Claw-Nose transition, and Nose) including mean dips are also shown in Fig. S4 in Supplemental Information S3. The average bedding dip is lowest in the Tongue (Mean dip: 5.3%) and highest in the Claw-Nose transition (Mean dip: 10%) with the Claw and Nose having mean dips of 7.3% and 7.0% respectively (Fig. S4 in Supplemental Information S3). The Tongue is also characterized by the

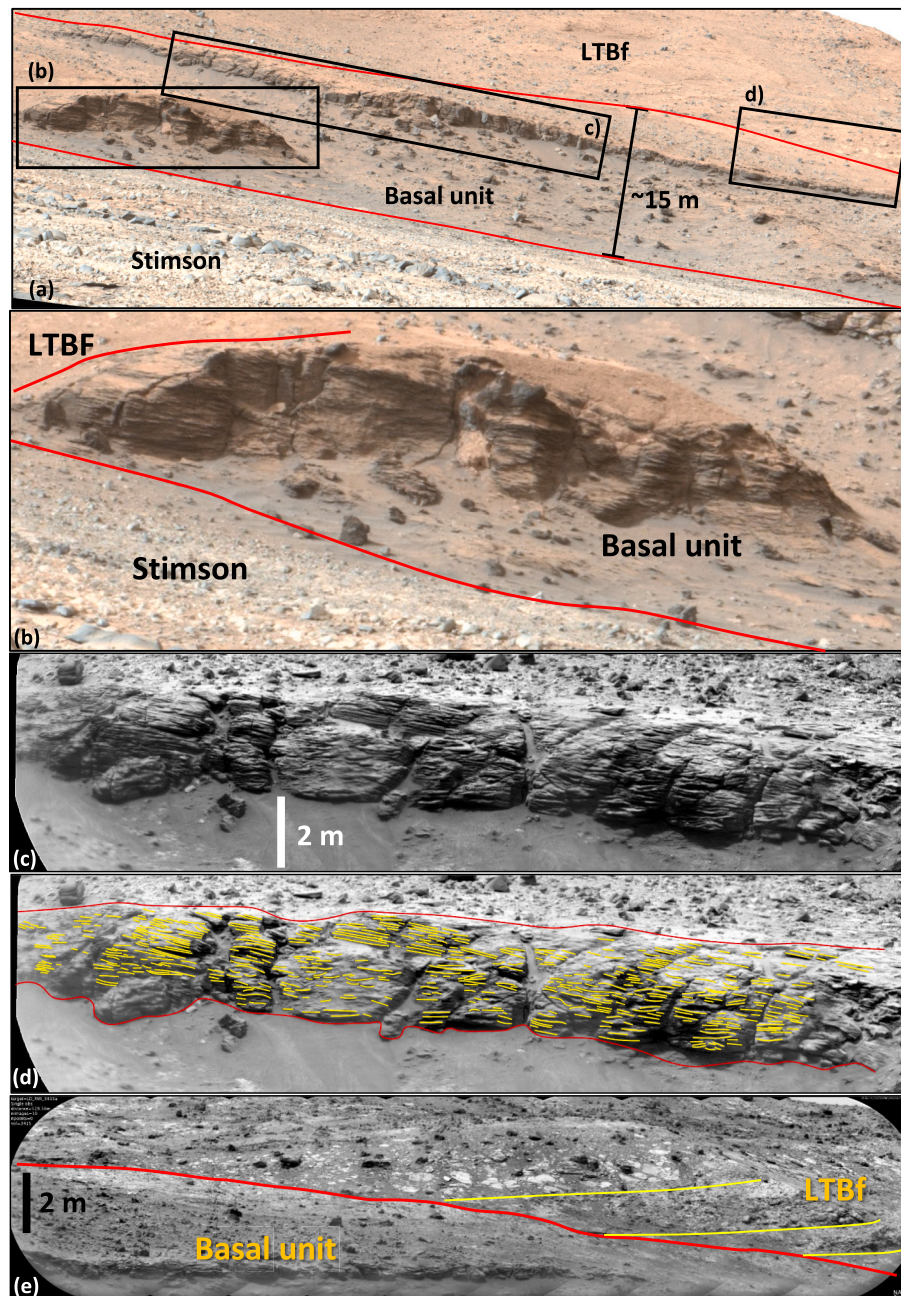


Fig. 9. Mastcam (colour) and Chemcam LD_RMI (black and white) mosaics of the Basal unit. a) close up Mastcam view showing the outcrop-forming Basal unit with facies boundaries (red lines) between the LTBf (above) and Stimson fm. (below). c) Close up view of the cliff-forming Basal unit including facies boundaries (red lines) between the LTBf (above) and Stimson fm. (below). c) close up unannotated and d) annotated versions of the Basal unit. Line annotations with background image removed showing possible low angle 1–3 m scale cross-stratification. e) Chemcam LD RMI data showing the boundary between the Basal unit and the overlying (onlapping) LTBf (Light-Toned Boulder facies) of the Claw-Nose transition. Mastcam image credit: NASA/JPL-Caltech/MSSS. Chemcam image credit: CNES/LANL/IRAP/IAS. See Table S1 in Supplemental Information S1 for image sequence identifiers. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

tightest distribution of dip directions (Fig. S4a in Supplemental Information S3). In contrast, the 3 units above the Tongue all have a wider range of dip directions (Fig. S4b–d in Supplemental Information S3). In all locations above the Basal unit, the bedding orientation was significantly gentler than the dip of the pediment (Fig. 10a). There were no locations where any of the upper stratigraphy could be seen down-lapping onto either the Basal unit or the pediment. The only strata that appeared parallel to, or steeper than, the pediment was within the Basal unit. The only discordant stratal relationship identifiable between the Basal unit and the overlying 5 packages is the onlapping relationship shown in Fig. 9e. These bedding orientations are the primary evidence

suggesting the fan was once significantly larger than it is now (see Sections 4 and 6 for further discussion).

The five units above the Basal unit comprise the majority of the modern Gediz Vallis Ridge and all five are comprised of two distinct interbedded facies: a facies dominated by light-toned clasts, termed the Light-Toned Boulder facies (LTBf), and a much less abundant facies dominated by distinctly dark-toned clasts, termed the Dark-Toned Boulder facies (DTBf) (Fig. 11; supplemental section S2 for facies descriptions). We have chosen to restrict facies definitions to these two (LTBf and DTBf), primarily due to the difficulty in defining sedimentary facies and sub-facies from hundreds of meters distance.

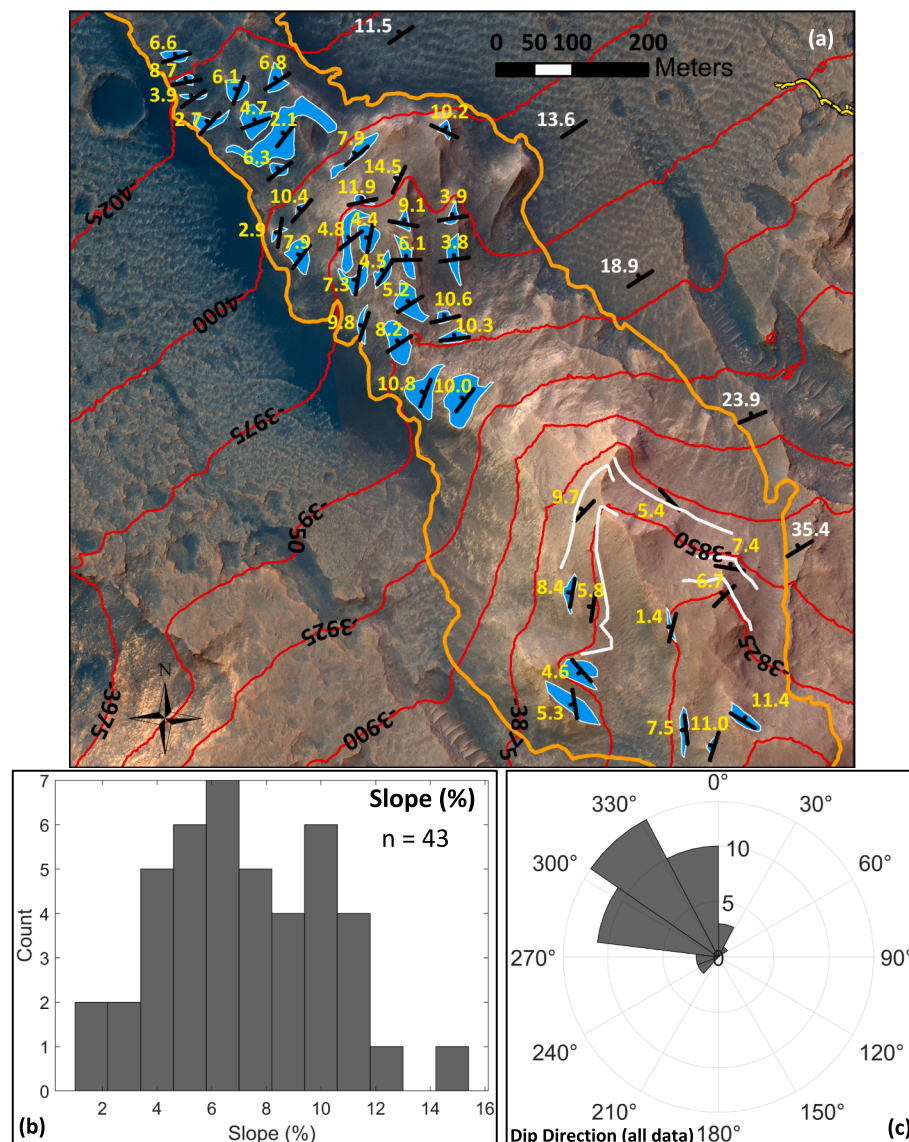


Fig. 10. Bedding orientations measured from HiRISE-based topography. a) Reference map made from HiRISE orthoimagery showing the GVR with 25 m topography overlain. Bedding (blue polygons) and bedding orientations are shown with GVR dips (yellow) and Greenheugh pediment dips (white) reported as percent grade. b) Histogram of bedding dips extracted from measurements shown in (a). c) Polar histogram (rose diagram) of dip direction data shown in (a). HiRISE image credit: NASA/ JPL-Caltech/University of Arizona. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

3.4. Unit 1 - the tongue

3.4.1. Observations

The current distal extent of the GVR atop the Greenheugh pediment, known as the Tongue, comprises a narrow (<200 m wide) 500 m long deposit with a maximum thickness of 14 m (Fig. 12 and S5 in Supplemental Information S3). In Fig. 12a, the A-A' cross-section shows the ridge exhibiting a relatively 'flat' cross sectional profile while the more distal (downslope) B-B' profile indicates the east and west margins rising as topographic highs relative to the rest of the cross-section. The distal most (northwest) deposits rest directly atop the Stimson formation (see Fig. 7), while boulders in the more proximal Tongue overlie the Basal unit. The HiRISE data (Fig. 12a) suggest that the interior surface of the Tongue contains abundant dark-toned boulders, while the margins of this remnant feature appear spectrally brighter and contain fewer boulders. Mastcam data (Fig. S6 in Supplemental Information S3) support the orbital observations, suggesting that the NE margin of the Tongue (including key outcrops Gowrie and Scalp Rock) is significantly less boulder-rich than the interior.

Gowrie is the most distal occurrence of the Basal unit observed via *Curiosity* (Fig. 13a; S6 in Supplemental Information S3). At Gowrie, the Basal unit is 3 m thick and is directly overlain by the light-toned boulder facies (LTBf). Here, the LTBf is matrix-supported with sparse individual boulders that are difficult to identify. There is some evidence for sub-horizontal bedding visible near the top of the outcrop (Fig. 13a and S7 in Supplemental Information S3). The top of the section at Gowrie consists of a thin (<2 m thick) cap of the clast-supported dark-toned boulder facies (DTBf). The base of the DTBf at this location is undulatory and appears to erosively cut into the LTBf below. The weak stratification in the LTBf may be truncated by the dark-toned facies above (Fig. 13a and Fig. S7 in Supplemental Information S3).

Approximately 100 m upslope from Gowrie, Fig. 13b shows another key outcrop (Scalp Rock) with overlain facies boundary interpretations. Here the Basal unit is slightly thinner than at Gowrie and the DTBf is absent at the top of the outcrop. Above the Basal unit, the outcrop is entirely made of LTBf, with sparse but obvious isolated dark- and light-toned clasts. The outcrop lies adjacent but connected to the interior of the Tongue where, behind Scalp Rock in Fig. 13b, extensive dark-toned

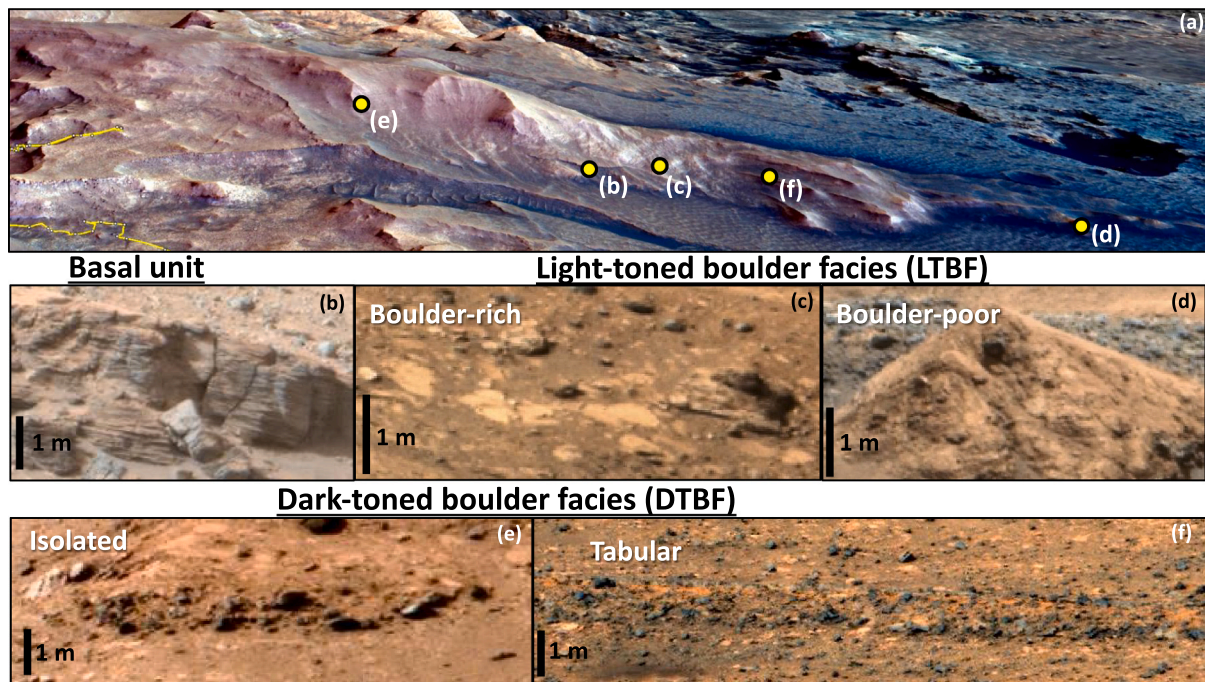


Fig. 11. GVR facies summary figure. a) 3D HiRISE view of the GVR showing locations for facies examples (b-f). b) Basal unit. Note the Basal unit is monolithologic, so is considered a single facies. c) Boulder-rich light-toned boulder facies (LTBF). d) Boulder-poor light-toned boulder facies (LTBF). e) Isolated dark-toned boulder facies (DTBF) f) Tabular dark-toned boulder facies (DTBF). Mastcam image credit: NASA/JPL-Caltech/MSSS. See Table S1 in Supplemental Information S3 for image sequence identifiers.

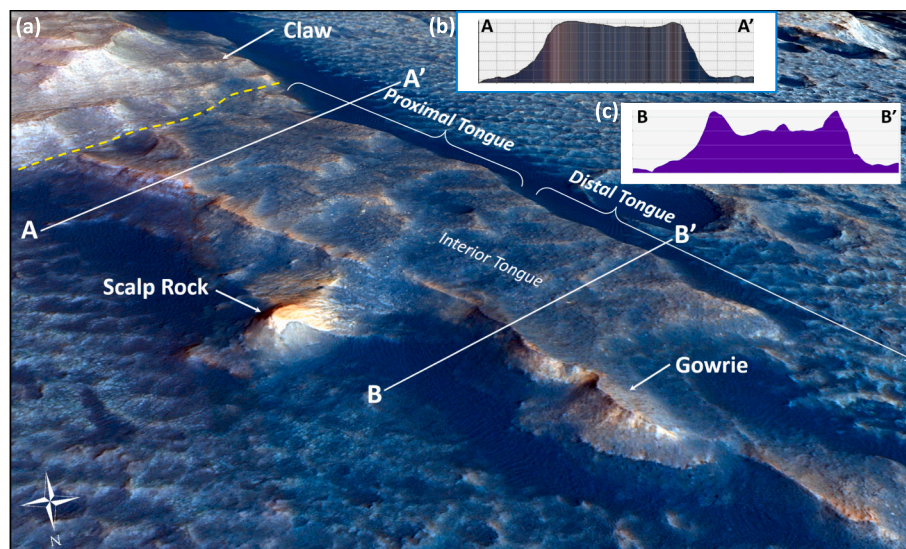


Fig. 12. HiRISE view of the Tongue. a) 3D view facing South showing key outcrops (Scalp rock and Gowrie), boundary definitions for the proximal and distal Tongue, and cross-section reference locations A-A' and B-B'. b) Cross-sectional topographic profile A-A'. c) Cross-sectional topographic profile B-B'. Note total stratal thickness are less than 15 m in the Tongue. HiRISE image credit: NASA/JPL-Caltech/University of Arizona.

boulder deposits can be seen mantling the surface of the interior Tongue.

Upslope of Scalp Rock, the stratigraphy of the NE margin of the Tongue is characterized by interbedded dark- and light-toned boulder facies each with thickness less than 1 m (Fig. 14). Overlain on the figure are blue, green, and red lines showing three stacked DTBF intervals. Fig. 14c shows that the DTBF beds are interbedded with thin LTBF deposits similar to the LTBF seen at Gowrie and Scalp Rock. This differs from the more distal areas of the Tongue (at and downslope of Scalp Rock; see Fig. 12a). In Fig. 14d the blue, green, and red lines correspond to the interpreted bedding shown in Fig. 14a-c. The lower/more distal

strata in this area (red line) exhibit a dip of 9.6% (5.4°). Up section, the dip of the beds becomes gradually more gentle with the middle (green) and highest (blue) measurable beds dipping at 3.9% (2.2°) and 2.7% (1.6°) respectively. These thin beds in some cases can be traced laterally (~140 m) across the entire Tongue. Importantly, all of the bedding measurements in the Tongue (Mean dip: 3.0°; 5.3%) are lower than the dip of the underlying depositional surface. There is no evidence of downlapping strata. Together this suggests that, if the contact was visible, these beds would exhibit an onlapping relationship with the depositional surface.

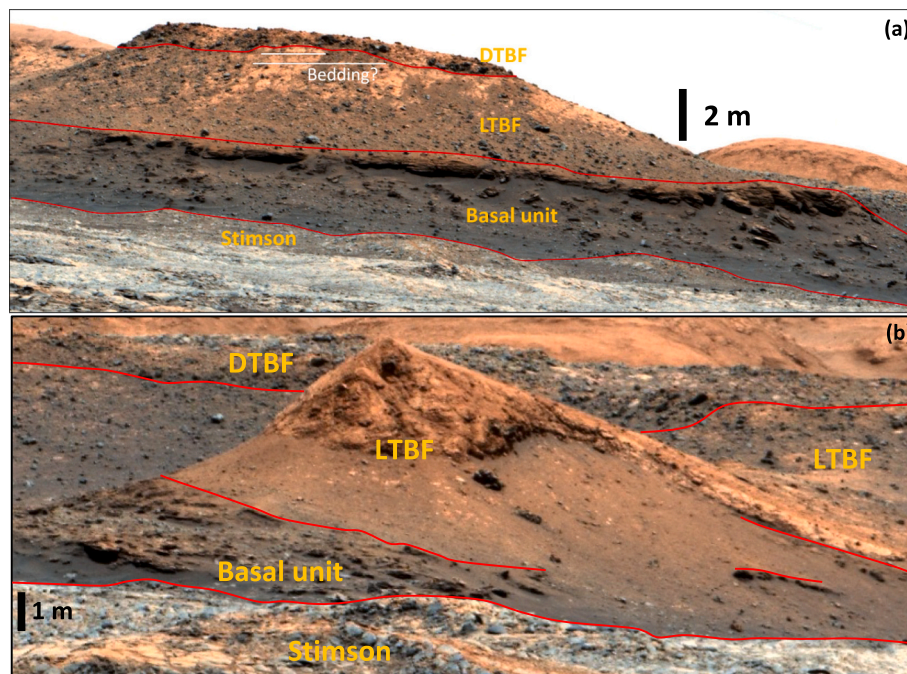


Fig. 13. Mastcam data from key outcrop locations in the Tongue. a) Gowrie location including possible bedding traces (white) in the boulder-poor Light-Toned Boulder facies (LTBF). b) Scalp rock location. Facies and unit names are shown in orange; facies and unit boundaries are overlain in red. Note the Basal unit is monolithic, so is considered a single facies. Mastcam image credit: NASA/JPL-Caltech/MSSS. See Table S1 in Supplemental Information S1 for image sequence identifiers. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Two graphic logs were created for the Tongue: one in the region of Gowrie, and one in the proximal Tongue near the Tongue-Claw contact (Fig. S9 in Supporting Information S3; summary comparison of all logs shown in Fig. 26). The columns demonstrate the significant difference in DTBF character between the distal and proximal Tongue. Complementing the columns in the Gowrie region is an interpreted facies map (Fig. S8 in Supplemental Information S3) that summarizes the stratal architecture of the distal Tongue. The outcrop contains both the dark- and light-toned boulder facies as well as thin (≤ 3 m thick) exposures of the Basal unit. Grain size in the Tongue was only measurable in the DTBF near the Gowrie outcrop where a significant sample of measurable clasts ($n = 520$) were observed (Fig. S10a, b in Supporting Information S3). The median grain size for the DTBF near Gowrie is 196 mm (coarse cobble; Fig. S10c in Supporting Information S3). The D_{16} was 108 mm (fine cobble) and the D_{84} was 339 mm (fine boulder) yielding a D_{84}/D_{50} ratio of 1.7. These sizes will be an overestimate because the pixel size at this 575 m distance of viewing is 11.5 mm (see section S1.4). The maximum clast size observed in this region was a 771 mm diameter dark-toned boulder and the mean sphericity (crudely estimated based on an aspect ratio between the longest and shortest axes) is 0.544. (see Table S5 in Supplemental Information S1 for compiled grain size statistics).

3.4.2. Interpretation

Unit 1 (The Tongue) can be divided into proximal and distal segments (Fig. 12a). The proximal Tongue exhibits interbedded LTBF and DTBF facies with individual beds less than 2 m thick (Fig. 14 and S9 in Supplemental Information S3). The cross-sectional (NE/SW) profile in this area is relatively flat-topped (Fig. 12b) and the topography appears ‘stepped’ – consistent with this stacked-facies association (Fig. 12 and S9 in Supplemental Information S3). The top surface of the Tongue is mantled with dark-toned boulders. These DTBF deposits form thin but laterally extensive (up to 140 m wide today) sheet deposits (Figs. 14 and 15), although they may have once been more extensive. The correspondence between the cross-sectional stratigraphy (Fig. 14a-c) and the HiRISE topography (Fig. 14d), suggest these DTBF sheets are primary depositional features. In contrast, the distal Tongue (downslope of Scalp

rock) is more degraded. Coherent sheets are absent. Gowrie and Scalp Rock lack the interbedded facies association observed upslope. Rather, the LTBF appears erosionally truncated by the DTBF, which caps the Gowrie outcrop (Fig. 13a; S7 in Supplemental Information S3).

The LTBF observed along the NW margin of the Tongue, with relatively low boulder concentrations are unsorted, and clasts range from rounded to angular with no clear evidence for mechanical segregation by shape or size. The clasts are matrix supported, although the origin of the ‘matrix’ (non-boulder) component of the LTBF could not be clearly identified at hundreds of meters distance. Together the most likely depositional process for the LTBF in the Tongue is through debris flow transport. In most areas, the LTBF appears massive (structureless). However, at Gowrie (Fig. 13a; S7 in Supplemental Information S3) there is evidence for weak layering within the ‘matrix’ component of the LTBF, such that some fluvial transport may have occurred.

Unlike the LTBF interbeds, the laterally extensive DTBF sheets of better sorted sediment in the proximal Tongue probably record fluvial and sheetflood transport. Segregation of dark and light-boulders is unlikely to occur during debris flow deposition. Thin (<2 m thick) wide planar sheet deposits are characteristic of fluvial braid plain and sheetflood deposition. Within individual DTBF sheets, subtle ridges and furrows observable in HiRISE (yellow lines in Fig. 14d) suggest that some of these laterally extensive ‘sheet-like’ DTBF beds may be comprised of amalgamated smaller individual events (e.g., levees, channels) with locally variable flow directions. Together, the stratigraphy of the proximal Tongue observed by Mastcam is consistent with multiple intervals of fluvial and sheetflood deposition interrupting phases of LTBF debris flow deposition.

Down slope of Scalp Rock in the distal Tongue, the successive stacked DTBF and LTBF intervals are absent, nor are there obvious sheet-like DTBF features in the interior. In this region, all of the facies boundaries above the pediment are interpreted to be unconformable and/or disconformable (i.e. where erosion was likely parallel to the bedding orientation) (Fig. S8b in Supplemental Information S3). Here, the margins of the Tongue are topographically higher than the interior (see Fig. 12 A-A' profile). The stratigraphically lower distal Tongue may not

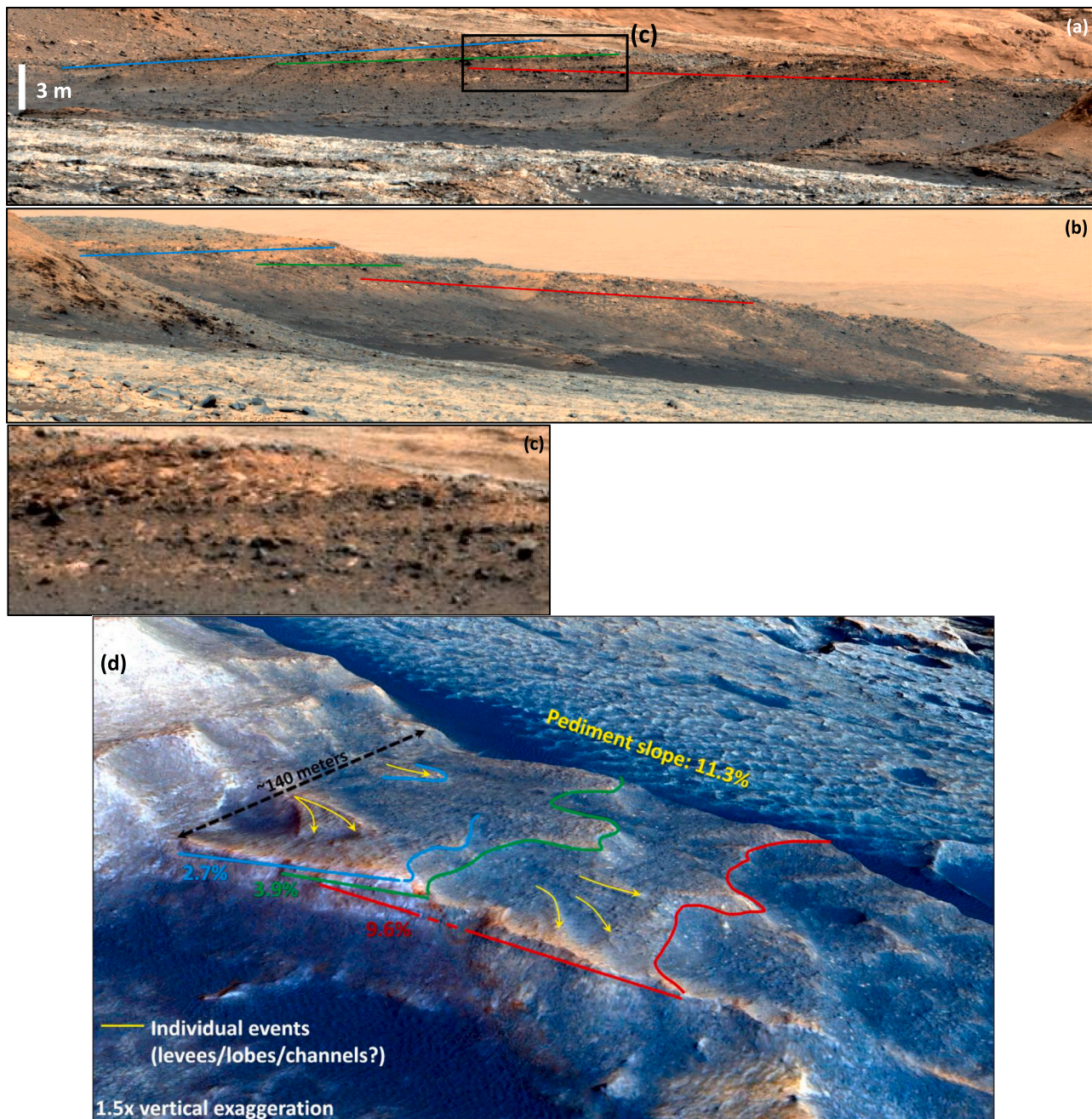


Fig. 14. Mastcam data from the proximal tongue shot from a) the Tower butte location and b) the blackcraig ascent area (see Fig. 3 for dedicated imaging locations). Bedding traces in red, green, and blue are overlain. c) Inset closeup unannotated view from (a) showing interbedded LTBf and DTBf sheets. d) 3D HiRISE view of the proximal Tongue with overlain red, green, and blue bedding traces (DTBf sheets) that correspond to those in (a) and (b). HiRISE image credit: NASA/JPL-Caltech/University of Arizona. Mastcam image credit: NASA/JPL-Caltech/MSSS. See Table S1 in in Supplemental Information S3 for image sequence identifiers. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

have been driven by cyclic deposition.

Another possibility is that there was more lateral variability within a given stratigraphic interval in the distal Tongue than there was up slope. There is a disparity between the stratigraphy of Scalp Rock and the adjacent deposits to the SW. The Scalp Rock outcrop is roughly 9 m tall (Fig. 13b), the top of which sits at -4027 m elevation. By contrast, within the adjacent deposits (behind Scalp Rock in Fig. 13b) dark boulder facies can be observed at lower elevations than the top of Scalp Rock. Essentially, the lateral equivalent of the top of Scalp rock is DTBf. It is difficult to tell if these DTBf deposits are in-place, but if so, this would indicate a lateral change in facies at similar stratigraphic horizon.

A third possibility is that the distal Tongue unit experienced

differential erosion into the equivalent DTBf sheet flood unit, leaving a rubble lag. This would require Scalp Rock to be more resistant to wind erosion, which given the prominent LTDF facies in it seems unlikely.

A fourth possibility explaining the valley-like topography in the distal Tongue is that the LTBf outer margins represent older debris flow deposits which were once incised into and subsequently backfilled by successive debris flows and sheetflood deposition. Evidence for back-stepping can also be seen in the proximal Tongue, where bedding dip systematically decreases upsection (Fig. 14). Possible confinement from older debris flow deposits (e.g., Scalp Rock and Gowrie) could also help explain why the Tongue is so narrow compared to most of the rest of the GVR.

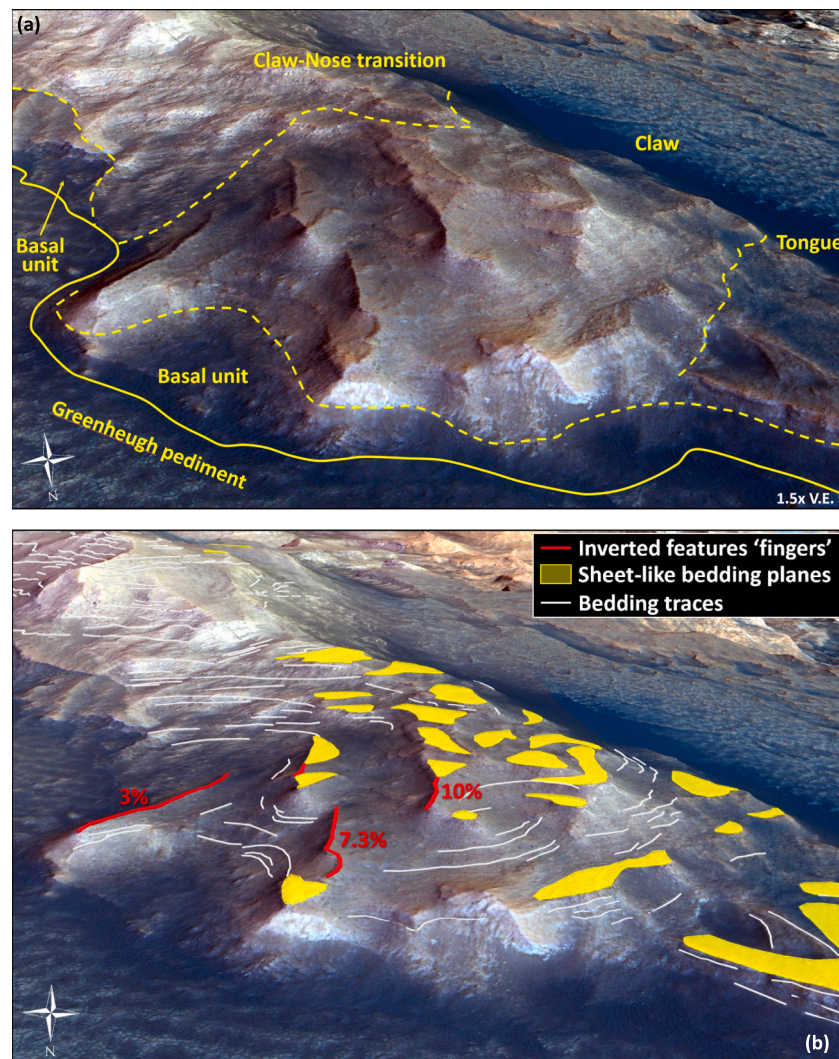


Fig. 15. 3D HiRISE view facing SSW focusing on the Claw. a) 3D HiRISE view with 1.5× vertical exaggeration showing unit boundaries in yellow. b) Similar HiRISE view (unexaggerated) showing bedding traces (white), DTBF sheets (yellow) and inverted ridges (red). HiRISE image credit: NASA/JPL-Caltech/University of Arizona. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Importantly, the systematic decrease in bedding dip in the Tongue suggests back-stepping debris flow (LTBF) deposition with interbedded fluvial and sheetflood DTBF deposits regardless of whether or not these beds were confined. This interpretation is supported by the lack of downlapping stratal geometries and the fact that all measured bedding dips are gentler than the depositional surface.

3.5. Unit 2 - the claw

3.5.1. Observations

Unit 2 (The Claw) is the most distinct feature on the Gediz Vallis Ridge and rests both upslope of- and stratigraphically above- the Tongue. Named for its 'claw-like' planform morphology which wind erosion has etched out the underlying variable stratigraphy, the 30 m thick unit shares an abrupt NW boundary escarpment at the upslope end of the Tongue, expands laterally to 370 m at its widest, and then gradually gives way upslope to a chaotic and rubbly area known as the Claw-Nose transition (Fig. 5; 6; 15; S10c in Supporting Information S3).

The Claw, like the Tongue, is comprised of stacked sedimentary strata with bedding attitudes significantly less steep than the slope of the Greenheugh pediment (Fig. 10a; 16b; S4b in Supplemental Information S3). In addition to the sheet-like features (yellow regions in Fig. 15b) and bedding traces (white lines in Fig. 15b) The Claw exhibits several

characteristic NE-SW trending 'finger-like' ridges appearing to fan slightly away from the overall NW-SE axis of the GVR (red lines in Fig. 15b), which suggest some underlying stratigraphic control.

Above this initial LTBF unit, bedding planes (yellow regions in Fig. 15b) are comprised of thin (<2 m thick) DTBF facies that interrupt the dominant LTBF stratigraphy with increasing frequency higher in the section (Fig. 17d). In most areas, these beds are discontinuous, rarely extending laterally over a few 10s of meters (Fig. 17d). Some of the bedding traces (white lines in Fig. 15b) grade laterally into thin DTBF deposits suggesting corresponding lateral changes in depositional process over 10s of meters. Surface expression of these DTBF bedding planes is 'sheet like', similar to those found in the proximal Tongue (see Fig. 15b and S12e in Supplemental Information S3). These 'sheets' create a 'step-like' topography at the top of the section comprised of stacked resistant DTBF deposits with weaker LTBF interbeds.

In contrast to the relatively common DTBF sheets observed in the upper Claw, the ridges contain narrow ridge-capping DTBF boulder deposits that are generally steeper than the surrounding bedding (Fig. 15b). Outcrops of the ridges show that the capping DTBF is likely erosively cut into the underlying LTBF deposits (Fig. 17a-c). Boulder long-axes within the DTBF facies (Fig. 17d) show a consistent downslope (north-verging) imbrication. The dominant LTBF is periodically interrupted by the DTBF and the irregular DTBF basal contact is consistent

with erosion into the LTbf (Fig. 17d). The facies map also shows the DTbf deposits are only traceable upslope over 10s of meters (maximum ~100 m). The variable thickness of the Basal unit is not due to cover or view angle, but rather records areas of local erosional removal of the unit by the subsequent LTgf processes. A graphic log (Fig. S13 in Supplemental Information S3, summary in Fig. 26) depicts both the increase in DTbf prevalence near the top of the section and the erosional bases common to the facies.

At two adjacent ridge capping locations (Fig. S14b, c in Supplemental Information S3) grain size analysis was done on the underlying LTbf and capping DTbf facies (Fig. S14b, c in Supplemental Information S3). Grain size analysis was combined for the same facies at the successive ridges. Median resolvable grain size of the LTbf combined layers is 159 mm, while the DTbf layers is 207 mm. At a LTbf bluff (Fig. S14b, e

in Supplemental Information S3) median grain size was 186 mm, whereas the combined grain size of two DTbf sheet layers was 175 mm. The D_{84}/D_{50} ratio measured among both facies was similar to that measured in the Tongue, with mean values between 1.5 and 1.9. The maximum clast size measured in LTbf deposits was 745 mm. The largest dark-toned boulder in the measured area was 656 mm (Fig. S14f in Supplemental Information S3; see Table S5 in Supplemental Information S1 for compiled grain size statistics).

3.5.2. Interpretation

Despite the contrast in morphology between of the Tongue and Claw (including the abrupt boundary separating the two features), the stratigraphy of the two features is relatively similar. Like the Tongue, the Claw is comprised dominantly of LTbf deposits that are poorly sorted,

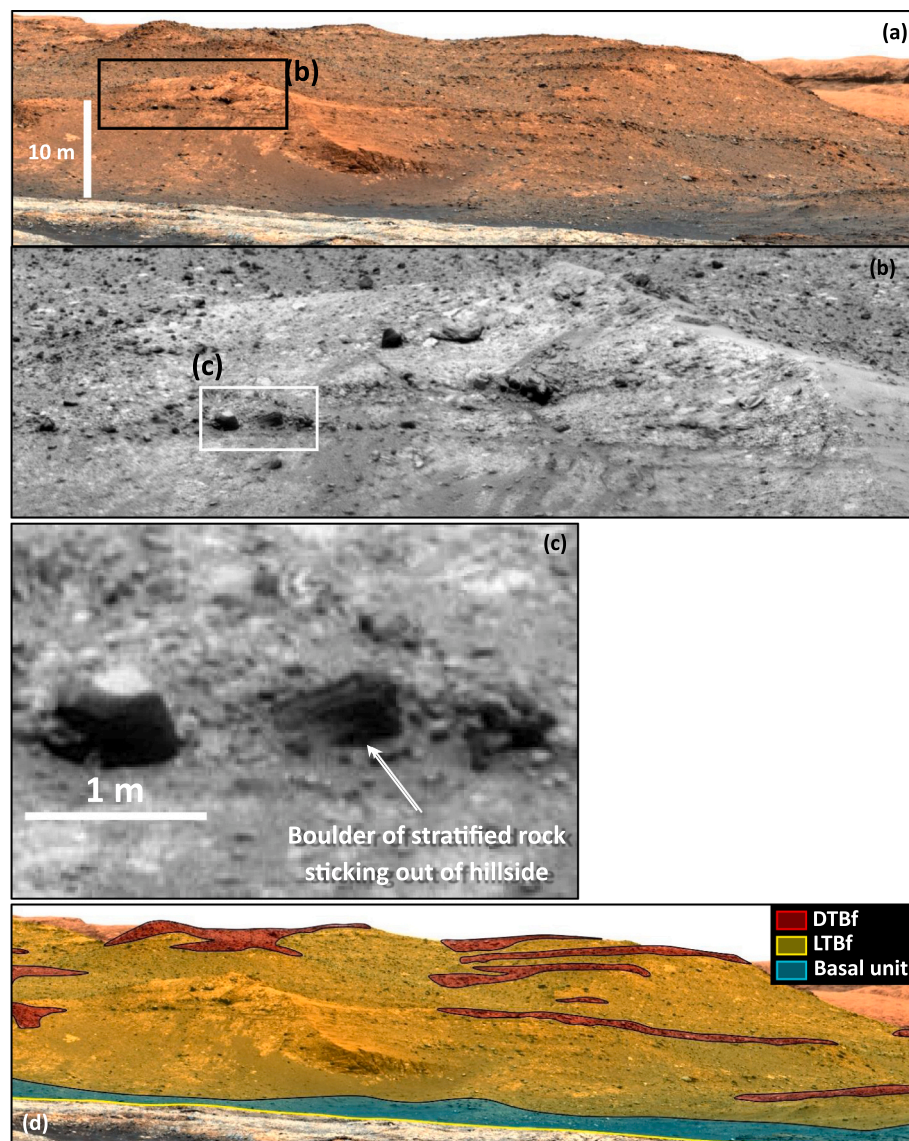


Fig. 16. Key Mastcam and Chemcam LD_RMI data of the Claw taken from Tower butte (Fig. 3). a) Wide view referencing LD_RMI locations shown in (b, c). b) LD_RMI view of bedded LTbf deposits. c) LD_RMI data referenced to (b) showing stratified boulder sticking out of the rockface. d) Facies map overlain on the Mastcam from (a) showing discontinuous Dark Toned Boulder facies (DTbf) sheets surrounded by the dominant Light Toned Boulder facies (LTbf). Mastcam image credit: NASA/JPL-Caltech/MSSS. ChemCam image credit: CNES/LANL/IRAP/IAS. See Table S1 in Supplemental Information S1 for sequence identifiers. Figure 17 shows Mastcam views of The Claw taken from the Tower butte pediment ascent (see Fig. 3 for reference) including key outcrop locations. Stratigraphically below the boulder-rich Claw, the Basal unit directly overlies the Stimson formation and is relatively thin – varying in thickness from 0.5 to 5 m thick (Fig. 17a, d; S12 in Supplemental Information S3). Above the Basal unit the lower Claw is comprised of LTbf with variable boulder concentrations that are generally greater than those found in the LTbf of the Tongue (Fig. S12 in Supplemental Information S3). Within this LTbf, clearly stratified horizons contain occasional dark-toned boulders that themselves show bedding typically of recycled sedimentary rocks (Fig. 17b, c).

appear to be matrix-supported, and have large boulders (up to 1.5 m), all typical of debris flow deposits. This is exemplified by the bedded LTBF of the lower Claw with laterally variable concentrations of chaotically oriented matrix-supported blocks (Fig. 16b, c). There are both light- (dominant) and dark- (rare) toned boulders in the LTBF. Also similar to the Tongue are the sheet-like DTBF deposits that interbed the dominant LTBF deposits, whose characteristics suggest sheetflood processes. In the Claw we see possible imbrication of clasts in the DTBF (Fig. 17d), the clear segregation of dark and light-toned boulders that define the two facies, and the commonly sheet-like form of the DTBF outcrops at the top of the section, all point to a fluvial origin.

The Claw exhibits several characteristics that distinguish it from the Tongue. The Claw is wider and lacks any features suggesting confinement. Rather, the Claw even in its modern eroded form exhibits a distinct fan-like morphology, exemplified by the ‘finger-like’ ridges capped by narrow linear DTBF deposits (Fig. 15b; 17), which deflect away from the ridge axis to the NNE to more than twice the width of the Tongue. This form of DTBF deposition is the most significant difference between the stratigraphy of the Tongue and the Claw. The average dip of the beds in the Claw is also steeper than those in the Tongue (7.3%; 4.2°

in the Claw vs. 5.3%; 3.0° in the Tongue) and the dip directions of the Claw are also more laterally variable (see Fig. S4d-e in Supplemental Information S3). The stratigraphic relationships indicate that the Claw was constructed as the depocenter backstepped across the Tongue deposits and advanced up the pediment.

The ‘finger-like’ ridges fanning away from the ridge axis were first noted by Palucis et al. (2016) who proposed that they were inverted channels and noted a correspondence between an inferred lake level and these features, suggesting an origin as a delta. We have found no evidence of foresets supporting a delta hypothesis. But the severe erosion of the Claw may not have preserved such stratigraphy. Are the ridges, then, inverted channels? Tracing some of these ridges upslope reveals a connection between the sheet-like (Fig. 15b) and the finger-like (Fig. 15b) DTBF deposits. In other cases, there appears to be no lateral or upslope sheet-like equivalent to the DTBF-capped ridges. Ridgeline slopes range from 3% to 10% but are not profoundly different than other nearby bedding measurements (Fig. 10a; 15b). The modern ‘finger-like’ ridges that give the Claw its name are clearly erosional features formed through aeolian erosion of laterally adjacent deposits. But why do they occur in the Claw and not elsewhere in the GVR? The simplest

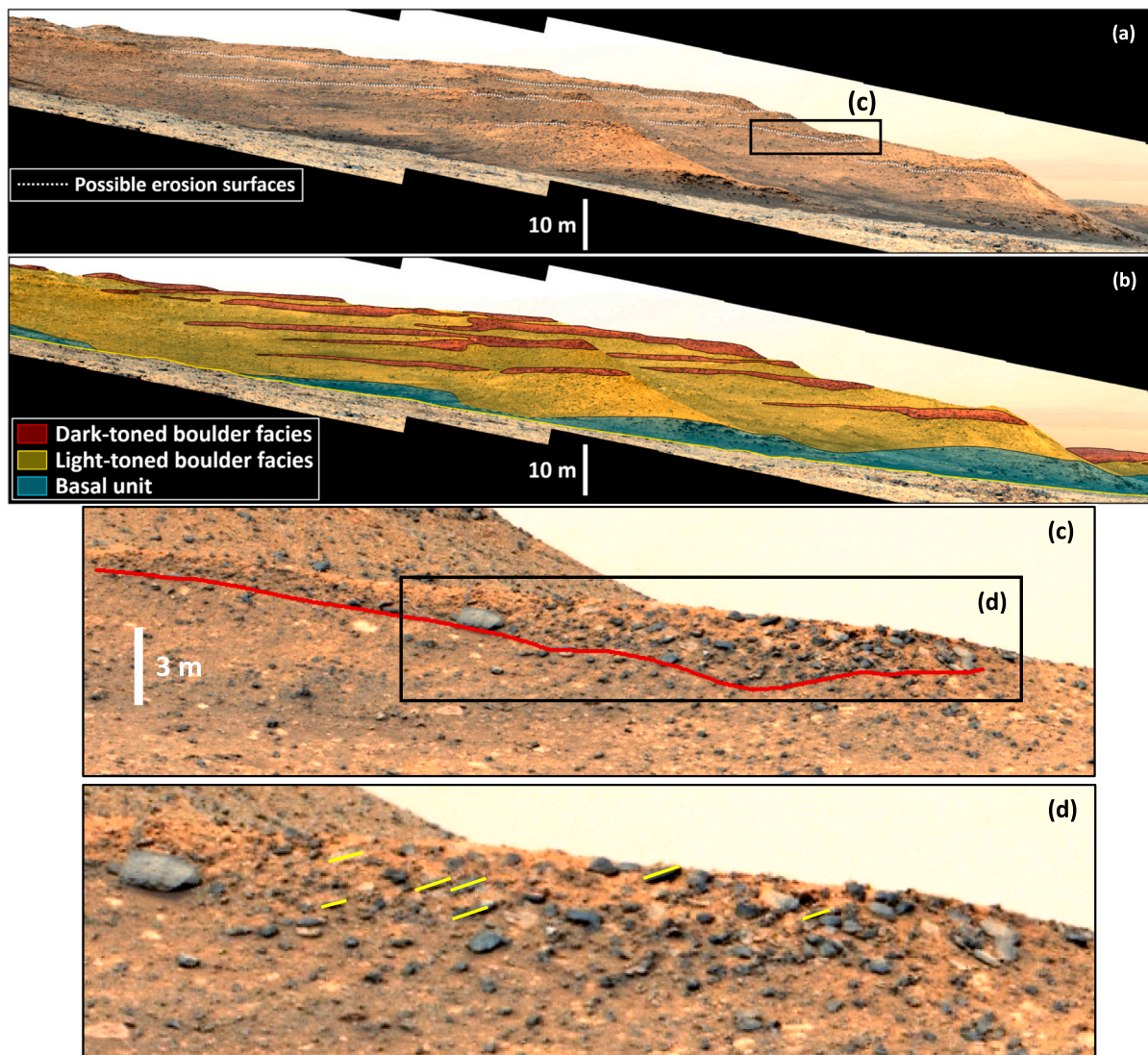


Fig. 17. Key Mastcam observations of the Claw taken from the Blackcraig ascent location (see Fig. 3). a) Wide Mastcam view of the Claw shot roughly parallel to the average strike of the bedding. b) Facies map overlain on the Mastcam mosaic shown in (a). c) Zoomed-in view of an inverted ridge with the basal DTBF contact shown downcutting into the underlying LTBF deposits. d) Possible imbricated clasts (yellow lines) from the same location shown in (c). Mastcam image credit: NASA/JPL-Caltech/MSSS. See Table S1 in Supplemental Information S1 for image sequence identifiers. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

explanation for these ridges is that they are the topographic expression of localized wind-eroded interbedded LTBF and DTBF sheet (not channel) deposits. They share an average long-axis orientation with other wind-etched features in the region. These include the pediment washboard (Bryk et al., 2025) and the Periodic Bedrock Ridge (PBR) topography in Glen Torridon (Stack et al., 2022). The fan like form of the Claw outline may be due to wind erosion, dominantly from the North, that etched into and around the sheet-like dark-toned bedding that frequents the upper Claw exposure.

Alternatively, fluvial channels could have formed during the initial construction phase of the Claw. DTBF deposition would have occurred sometimes through sheetflood processes and at other times as channelized flow with minor incision into the recently deposited LTBF debris flow facies. Fluvial transport of the overall $D_{50} = 190$ mm would require flow depths roughly 40 cm deep (or twice the median grain size) (Lamb et al., 2008). If shallow surface channels on the fan preferentially retained erosionally resistant dark boulders, they could later become inverted deposits through wind erosion.

3.6. Unit 3 – the claw-nose transition

3.6.1. Observations

Stratigraphic mapping confirms a transitional boundary between the Claw and the upslope Nose defined here as the “Claw-Nose Transition” (Unit 3; Figs. 5, 6; 9c; 18a). This relatively thin (25 m thick) package, sandwiched stratigraphically between the upper Claw and Lower Nose stratigraphic packages, hosts several anomalous features not found elsewhere in the GVR. Directly upslope of the Claw, the Basal unit pinches out completely such that the stratigraphy of the Claw-Nose transition is resting directly atop the Stimson formation. This location

also corresponds to the upslope extent of the ridges as well as and DTBF sheet facies of the Claw. The package is characterized by massive to weakly bedded LTBF stratigraphy containing a very coarse-grained chaotic poorly sorted mixture of dark- and light-toned boulders (Fig. 19b, c; S15 in Supplemental Information S3). The dominant facies in the transition zone is LTBF, with discontinuous and poorly exposed ~1 m thick DTBF beds near the top of the section. Some sheet-like features on the SW side of the GVR are present (see Fig. 10a; 15; 18b). Poorly exposed DTBF on the NE side of the ridge (Fig. 18b) appear as more isolated than those in the Claw. The LTBF facies include several light-toned boulders that exceed 3 m in diameter (Fig. 18b; S15 in Supplemental Information S3). These are the largest light-toned clasts observed anywhere in the GVR. The average dip of the beds in the Claw-Nose transition (Mean: 10%; 5.7°) is also among the steepest dips measured anywhere in the ridge (Fig. 10a; S4 in Supplemental Information S3), although still nearly one-half the slope of the underlying pediment surface.

The Basal unit vertically thickens across this transition unit from zero meters to its GVR-wide maximum thickness of 15 m across only 100 m distance upslope (Fig. 8; 11a; 19a, c). In this region, the coarse weakly bedded LTBF facies clearly onlaps the Basal unit (Fig. 9e; S16 in Supplemental Information S3). This is the only location where *Curiosity* observed an obvious discordance between the Basal unit and the overlying boulder-rich stratigraphy. This area also provided one of the best (and most dust-free) opportunities to observe the relative difference in competence between the light and dark-toned boulders (Fig. 19c). The deposit is dominated by light-toned boulders that have been wind eroded flush with the outcrop. Interspersed among these are embedded dark-toned boulders, protruding from the outcrop. Two graphic logs for the Claw-Nose transition (Fig. S17 in Supplemental Information S3, see

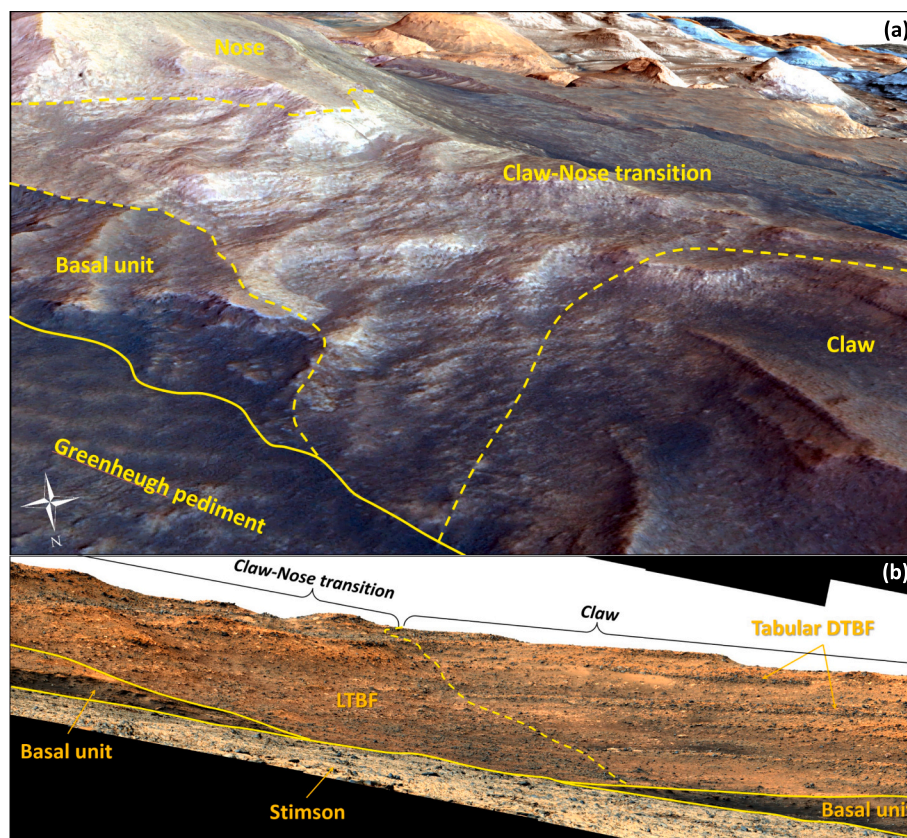


Fig. 18. HiRISE and Mastcam data showing the Claw-Nose transition a) 3D HiRISE view facing SSW focusing on the Claw-Nose transition. Unit boundaries are shown in yellow. b) Mastcam with annotations showing unit boundaries (yellow) and facies names (orange). HiRISE image credit: NASA/JPL-Caltech/University of Arizona. Mastcam image credit: NASA/JPL-Caltech/MSSS. See Table S1 in Supplemental Information S1 for image sequence identifiers. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

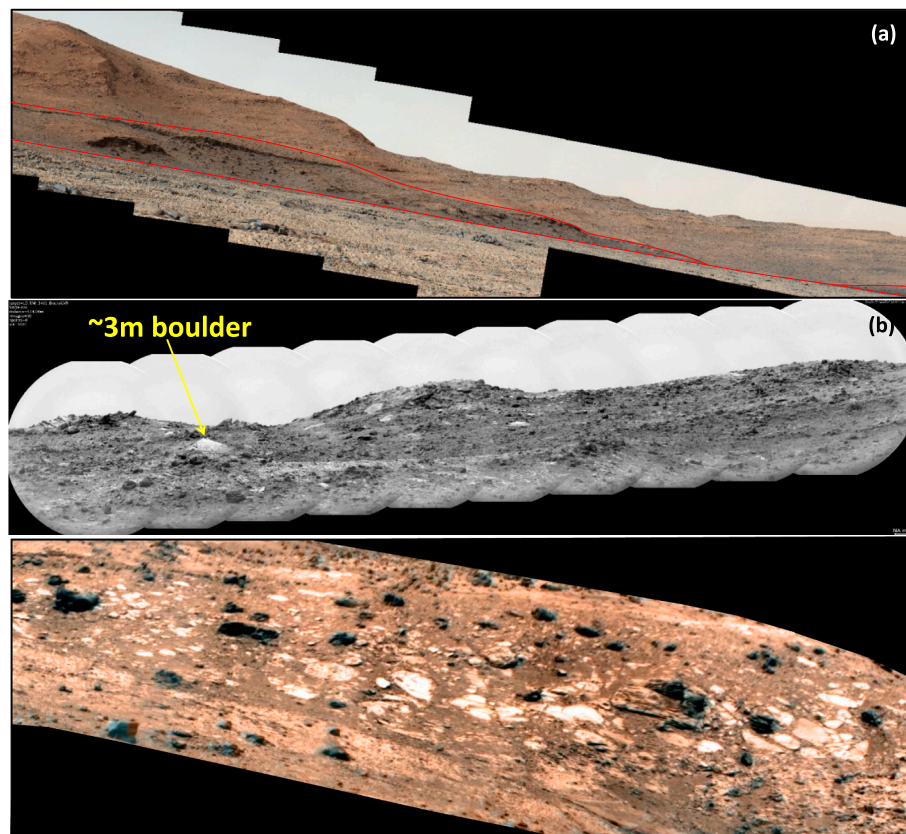


Fig. 19. Key Mastcam and Chemcam LD_RMI observations in the Claw-Nose transition taken from the Blackcraig ascent location (Fig. 3). a) Mastcam mosaic of the Claw-Nose transition showing reference locations for b) ChemCam LD_RMI data showing coarse LTBf facies including occasional 3 m light-toned boulders. c) Merged Mastcam RGB colour data and higher-resolution ChemCam LD_RMI data of the LTBf onlapping the Basal unit in the Claw-Nose transition. The figure shows the clear difference in outcrop expression between the resistant dark-toned boulders and the weaker light-toned boulders, which have been planed off flush with the hillside. Mastcam image credit: NASA/JPL-Caltech/MSSS. ChemCam image credit: CNES/LANL/IRAP/IAS. See Table S1 in Supplemental Information S1 for image sequence identifiers.

summary in Fig. 26), also demonstrate the coarse grain sizes and stark difference in thickness of the Basal unit. Resolvable median grain size of the outcrop is 336 mm, much coarser than elsewhere in the GVR, with a higher D84/D50 ratio and higher aspect ratio (minimum/maximum length) of 0.5 (Fig. S18 in Supplemental Information S3; Table S5 in Supplemental Information S1 for compiled grain size statistics).

The boundary between the Claw and the Claw-Nose Transition dips to the NW (Fig. 18b). This is the only location in the GVR where down-dip facies changes are visible along a single stratigraphic horizon. In Fig. 18b, the coarse and chaotic transitional strata give way downslope to the finer-grained and more laterally continuous interbedded LTBf/DTBf facies.

3.6.2. Interpretation

Like the Claw and Tongue below, the facies interpretation for both LTBf and DTBf deposits remains consistent. LTBf facies likely indicate boulder-rich debris flow depositional processes, whereas the poorly exposed DTBf facies in the region, although difficult to interpret, likely indicate fluvial winnowing and transport as was found downslope.

Although minimal (<10 m) stratigraphy is exposed, the lowermost strata in the Claw-Nose transition can be traced down-dip to the uppermost Claw stratigraphy (Fig. 15a and 18b). This suggests down-dip facies changes from the chaotic coarse-grained Claw-Nose transition to the more organized and finer-grained LTBf/DTBf interbedded upper Claw deposits. The boundary itself dips steeply to the NW cross-cutting the depositional layering suggesting a retrogradational stacking pattern. This would indicate the system backstepped as the fan aggraded.

Increases in resolvable grain size and bedding dip also suggests a

change in process up-section from episodic debris flow deposition punctuated by fluvial intervals (found in the Claw) to a period of more dominant, coarse debris flow transport (in the Claw-Nose Transition). This is further supported by the absence of observable DTBf sheet facies and points to variable hydrogeomorphic activity in the Gediz Vallis catchment.

3.7. Unit 4 - the nose

3.7.1. Observations

The Nose makes up the bulk of the Gediz Vallis Ridge and contains the greatest thickness of preserved stratigraphy rising over 70 m vertically above the Greenheugh pediment (Fig. 20 and S19 in Supplemental Information S3). The base of the section conformably overlies the Claw-Nose transition. No facies change was observed between the upper Claw-Nose transition and Nose units. The lower and middle Nose deposits are dominated by boulder-rich LTBf facies which appear similar to those observed down section in the Claw and Claw-Nose transition (Fig. 21). Mixed in with the light-toned matrix supported boulders are protruding resistant dark-toned boulders. (Fig. 21b, c; S20 in Supplemental Information S3). The thickness of the Nose and steepness of its NE flank allowed for rough estimation of LTBf bedding thickness. Individual beds range from roughly 1 to 3 m thick and in general, cannot be traced laterally more than 100 m. Internal stratification within individual beds is massive and matrix supported like other LTBf observations down section.

In contrast to the similarity among LTBf in the Nose and the strata downslope, the DTBf deposits in the nose are significantly less common

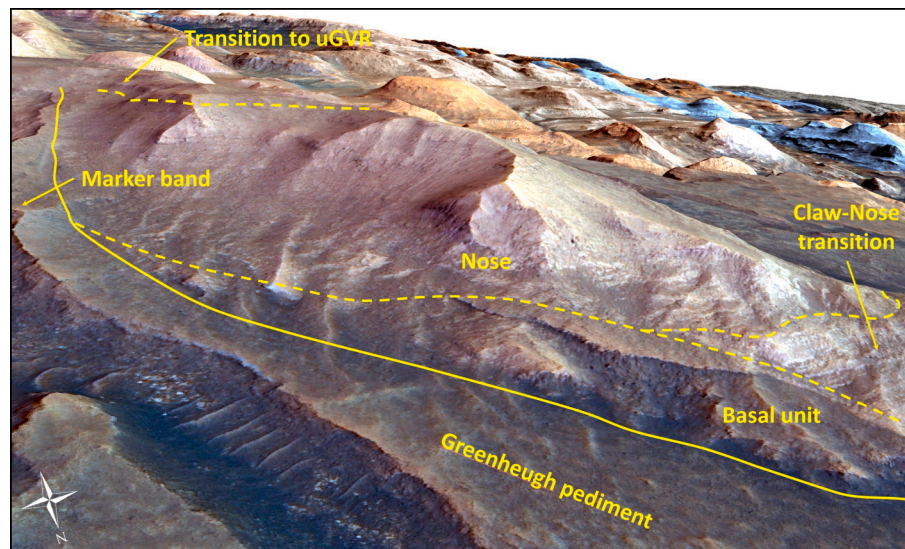


Fig. 20. 3D HiRISE view facing SSW focusing on the Nose. Unit boundaries are shown in yellow. HiRISE image credit: NASA/JPL-Caltech/University of Arizona. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

and occur as isolated collections of outcropping dark boulders and thin boulder-rich layers of limited extent (Fig. 21c; S21; S22 in Supplemental Information S3). Fig. 21c shows a lenticular (channel-like) DTBf deposit with a concave-up basal contact. Isolated clast-supported DTBf layers range from 1 to 2 m thick and extend to a maximum of 13 m laterally. Sheet-like DTBf deposits were absent on the NE side of the GVR and may not occur anywhere in the Nose.

Angular, poorly sorted and matrix-supported character of the LTBF facies dominate the stratigraphy of the Nose sedimentary package (Fig. 21b, d). In some areas, weak but consistent bedding can be seen – common to the LTBF facies (Fig. S23c and d in Supplemental Information S3). This area clearly shows the dramatic difference in bedding orientation between the steeply dipping Greenheugh pediment and the much more gently dipping overlying boulder facies (Fig. 21a; 22; S23c in Supplemental Information S3). In this region of the GVR, the underlying Greenheugh pediment steepens significantly to >30% grade (16.7°) above a distinct stratigraphic unit labeled the Amapari Marker band (Fig. 6a; Weitz et al., 2022; Thompson et al., 2024; Kite et al., 2025). Despite this extreme slope, the internal stratigraphy of the GVR in the Nose is (on average) dipping more gently (mean dip: 4.0°; 7%) than either the beds in the Claw (mean dip: 4.2°; 7.3%) or the Claw-Nose transition (mean dip: 5.7°; 10%) (see Fig. 10 and S4 in Supplemental Information S3). The overwhelming dominance of LTBF in the Nose compared to elsewhere downslope is demonstrated in the facies map and graphic logs shown in Fig. 22 and S24 in Supplemental Information S3 respectively (and see summary in Fig. 26). The facies map also shows the obvious absence of sheet-like DTBf deposits.

Roughly at the midpoint between the upper and lower boundaries of the Nose, the Amapari Marker Band, a laterally extensive prominent dark-toned unit within the Mt. Sharp group (Milliken et al., 2010; Kite et al., 2013; Weitz et al., 2022; Thompson et al., 2024) intersects the Greenheugh pediment (Fig. 6; 20; S19a in Supplemental Information S3). The Amapari Marker Band is cut by the basal Siccar Point group unconformity (Watkins et al., 2022; Bryk et al., 2025), but unlike the otherwise smooth pedimentation surface, the band was resistant to planation and forms a slight bump in the topography of the pediment and capping Stimson. The Marker Band does not, however, influence the form of the GVR or outcrop across it. Upslope of the southern extent of the Stimson formation, the GVR Basal unit pinches out (Fig. 20). Stratigraphically above the Basal unit, the beds remain dipping consistently more gently than the depositional surface indicating an onlapping of the surface by the boulder-rich GVR deposits. There is no evidence of

downlapping. The onlapping observed in the Nose is even more extreme than in the lower units (Tongue, Claw, Claw-Nose trans.).

The Nose provided some of the largest exposures for grain size analysis anywhere in the ridge. Extensive outcrop allowed grain size data to be collected for both the dark- and light-toned facies, resulting in the largest grain size dataset ($n = 8433$ clasts). Fig. S25 in Supplemental Information S3 shows measurement locations for Nose grain size data including distributions for all data collected in the Nose. For two DTBf dominated exposures the median resolvable grain size for DTBf regions was 149 mm (Fig. S25c in Supplemental Information S3) and 212 mm (Fig. S25d in Supplemental Information S3), while for three LTBF dominated exposures regions median grain sizes were 158 mm (Fig. S25c in Supplemental Information S3), 99 mm (Fig. S25d in Supplemental Information S3), and 97 mm (Fig. S25f in Supplemental Information S3). Maximum clast sizes in the measurement area were similar among the LTBF (1457 mm) and DTBf (1449 mm) facies. The D_{84}/D_{50} ratio for the Nose varied between 2 and 4.6, but did not show any trends by facies and was 3 on average, higher than that measured in any of the units downslope. Mean approximate sphericity for LTBF was 0.635 and for DTBf was 0.643 similar to those found in the Claw, but greater than both the Tongue and the Claw-Nose transition (Fig. S25g in Supplemental Information S3; see also Table S5 in Supplemental Information S1 for compiled grain size statistics).

3.7.2. Interpretation

The stratigraphy of The Nose is dominated by poorly sorted, matrix-supported, angular to sub-rounded, boulder-rich stratigraphy, interpreted as debris flow deposits of the LTBF, as is visible in Fig. 21 and S23 in Supplemental Information S3. The lenticular DTBf facies of the Nose are interpreted to have formed through fluvial transport and reworking of the LTBF and local deposition. In some cases (e.g., Fig. 21c; S22 in Supplemental Information S3), the deposits may record channel infilled bodies, with concave (likely erosive) lower- and flat upper-contacts. Elsewhere, they appear more as collection of clasts, which can occur in fan surfaces where shallow fluvial scour mobilizes embedded clasts which then collectively hydraulically interfere and form local patches (Fig. S27 in Supplemental Information S3). In most cases it is not possible to identify whether the DTBf lenses represent true channel deposits, or if they indicate preserved barforms common to gravel bed rivers. The range in thicknesses observed among the lenticular DTBf is consistent with both possible interpretations. The difficulty in identifying true sedimentary architectural elements (e.g., channels, bars,

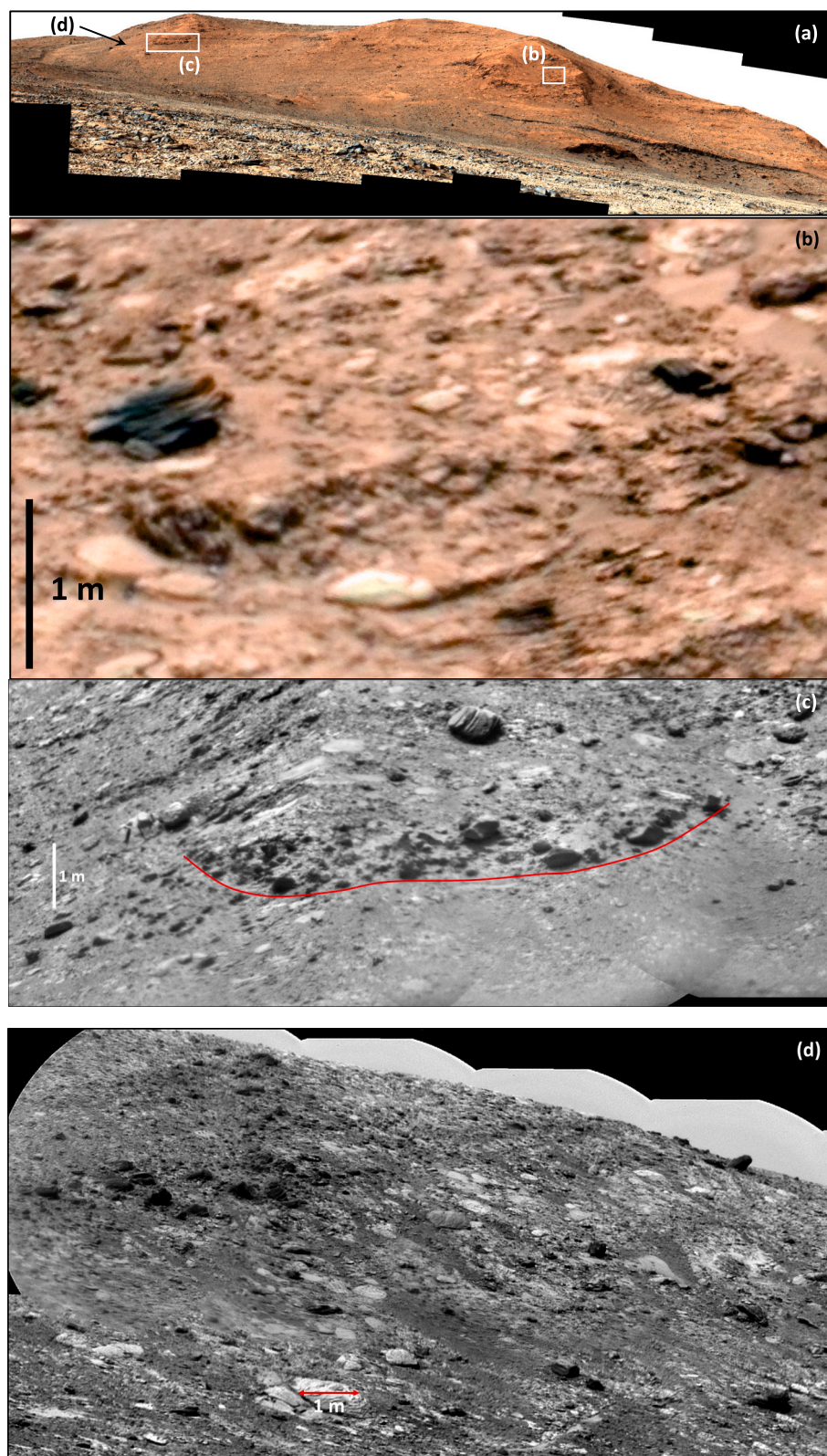


Fig. 21. Key Mastcam and Chemcam LD_RMI observations in the Nose taken prior to the Blackcraig ascent (see Fig. 3). a) Wide Mastcam view referencing Mastcam and ChemCam LD_RMI locations shown in (b, c). b) Merged Mastcam and ChemCam LD_RMI data showing the difference in outcrop expression between the resistant dark-toned boulders and the weaker light-toned boulders, which have been planed off flush with the hillside. c) ChemCam LD_RMI mosaic showing an isolated channel-like lenticular DTBf deposit. d) ChemCam LD_RMI image of the Nose showing chatotic matrix-supported light- and dark-toned boulders. Mastcam image credit: NASA/JPL-Caltech/MSSS. ChemCam image credit: CNES/LANL/IRAP/IAS. See Table S1 in Supplemental Information S1 for image sequence identifiers.

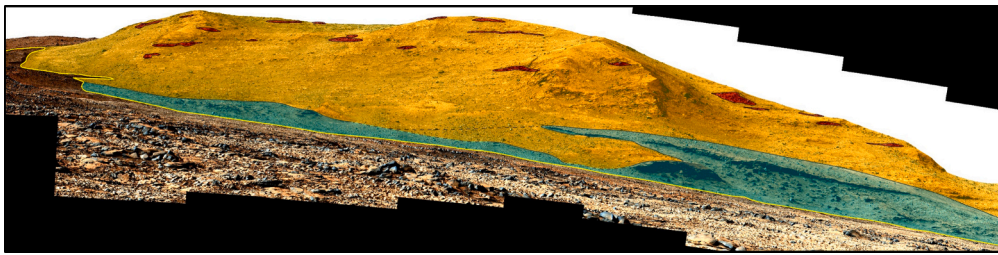


Fig. 22. Facies map overlain on mastcam data (Fig. 21a) for the Nose. Image data were collected from the Blackcraig pediment ascent area. Mastcam image credit: NASA/JPL-Caltech/MSSS. See Table S1 in Supplemental Information S1 for image sequence identifiers.

levees) is compounded by the likelihood that transport directions were radial – as is common in fan environments and supported by the radial spread in measured bedding orientations (see Fig. 10 and S4 in Supplemental Information S3). It is therefore nearly impossible from hundreds of meters distance to know for a given DTBf lens the angle between the transport direction and the outcrop face.

The median resolvable grain size of the DTBf was slightly larger than the that of the LTbf, but together they were in general similar to those measured in the Claw and the Tongue. However, the maximum clast sizes as well as the D_{84}/D_{50} ratio were both greater in the Nose. It is possible the material deposited on the Gediz Vallis fan higher in the stratigraphy may have come from closer to the canyon mouth, which would have resulted in lower transport distances and thus less mechanical breakdown.

The absence of downlapping stratal geometries, and consistent evidence for onlapping of boulder-rich facies onto the depositional surface suggests, like the deposits below, the Nose formed as the depocenter backstepped upslope. The Nose is significantly larger than any of the other deposits in the GVR. It is likely that the strata of the Nose and the (now missing) lateral and downslope correlative units represented the bulk of the Gediz Vallis fan volume prior to erosional removal.

The relatively abrupt loss of sheet-like DTBf deposits could indicate a change in the fluid discharge or sediment supply delivered from Gediz Vallis. It is also possible these changes do not indicate fundamental changes in external forcings, but rather indicate internal dynamics endemic to fan environments. However, due to the large size and thickness of the Nose stratal package, it seems more likely that this change actually reflects a temporal shift in hydrology within Gale.

3.8. Unit 5 – transition to upper GVR

3.8.1. Observations

The contact between the Nose and the Transition to uGVR is difficult to define. The LTbf is nearly identical between the two features and remains the dominant facies in the transition to the uGVR package. Near the top of Nose stratigraphy (Figs. 23 and 24) there is an increase in the prevalence of DTBf facies, whereas lower in the Nose, the DTBf is rare and forms isolated – sometimes lenticular pockets. This is denoted with a yellow dashed line in Figs. 23 and 24a, c. Here, not only does the DTBf increase in prevalence, but the grain size also increases, including several areas where 1–3 m dark-toned boulders are common. We place a transition between the Nose and the Transition to uGVR stratal package based on these criteria (Figs. 23 and 24; S26–28 in Supplemental Information S3).

Like the Nose below, DTBf units appear lenticular and clast-supported, but as the upslope extent of this transition unit is reached, the dark-tone facies extend laterally across greater distances (up to 30 m) and are on average slightly thicker than those of the Nose. LTbf thicknesses (1–3 m) were similar in the Nose and the Transition to the uGVR package. There was no indication of repeated sheet-like DTBf strata. The most glaring difference in the stratigraphy of the Nose and the Transition to uGVR package is the presence of large dark boulders (up to 3 m long) in many of the DTBf lenses. Fig. 24a and S27 in Supplemental Information S3 show the lower strata of the Transition to uGVR package, where DTBf lenses contain large individual boulders and may also exhibit erosive bases. Upslope, Fig. 24b shows an example location where the grain size in the DTBf is relatively uniform, and clast supported. Additional areas demonstrating this trend are shown in Fig. S28 in Supplemental Information S3. It is possible Fig. 24b indicates a single DTBf lens that downcuts several meters into the underlying LTbf stratigraphy. Directly overlying one of the smallest DTBf lens in this area

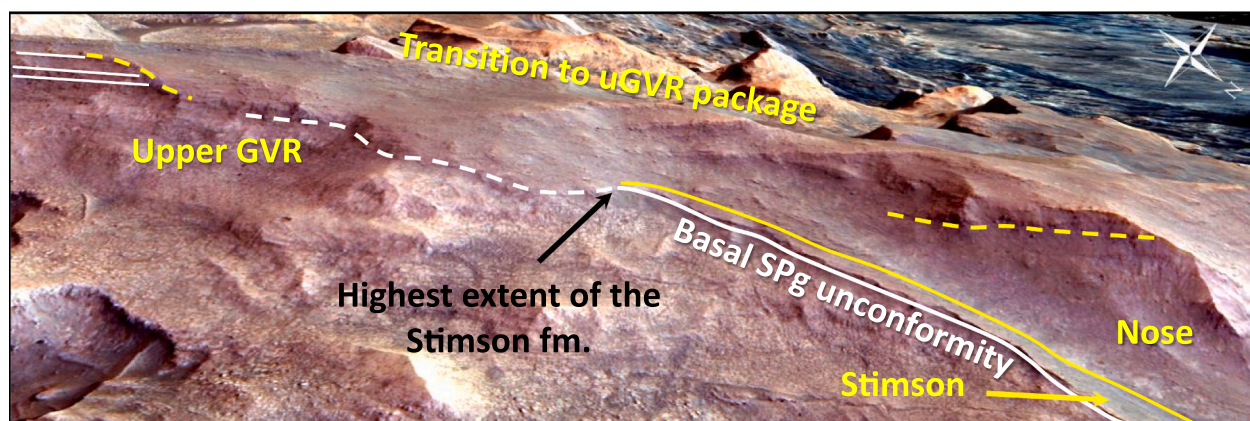


Fig. 23. 3D HiRISE view facing SW focusing on the Transition to the upper Gediz Vallis Ridge (uGVR) package. Unit names and boundaries are shown in yellow for the Transition to the uGVR package. The basal Siccar Point group (SPg) unconformity (white) and highest extent of the Stimson (black) are also included. HiRISE image credit: NASA/JPL-Caltech/University of Arizona. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

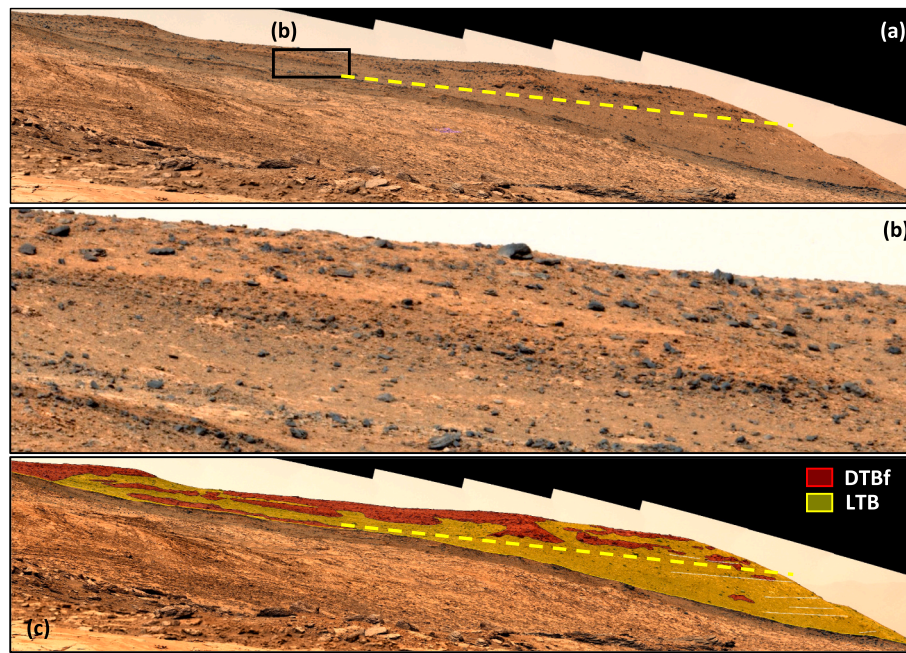


Fig. 24. Key outcrop locations for the Transition to uGVR package. a) b) Mastcam mosaic of the upper Nose and transition to uGVR package shot from the Wing ascent (Fig. 3). b) DTBf lens downcutting several meters into underlying LTBf deposits. c) Annotated facies map overlain on the Mastcam mosaic shown in (a). HiRISE image credit: NASA/JPL-Caltech/University of Arizona. Mastcam image credit: NASA/JPL-Caltech/MSSS. See Table S1 in Supplemental Information S1 for image sequence identifiers.

are several planar beds that could indicate finer grained deposition (see Fig. S28b in Supplemental Information S3). This is the only location in the Transition to uGVR where finer grained material could be inferred.

The bedding within the Transition to the uGVR package dips more gently than the underlying pediment suggesting an onlapping relationship with the depositional surface. There was no evidence for downlapping. The increase in frequency and thickness of the DTBf prevalence that marks the lower boundary of the transition can be clearly seen in the facies map developed for the Transition to uGVR package (Fig. 24c). In this region, the pediment extends as a smooth surface upslope, but the capping Stimson disappears roughly half-way through the Transition to uGVR stratigraphy (Fig. 23). This basal unconformity clearly continues upslope beyond this location across the underlying Mt. Sharp bedrock. Due to poor outcrop visibility, and a paucity of high resolution Mastcam or ChemCam LD RMI data. HiRISE imagery was used to define the upper boundary between the Transition to the uGVR package and the uGVR (Fig. 23). The upslope boundary of the Transition to uGVR appears to truncate the uGVR strata. This apparent discontinuity cuts down across 10 m of stratigraphy where the Transition to uGVR package layering appears to onlap the boundary. Unfortunately, we rely on HiRISE images here, as we were not able to obtain sufficiently high resolution Mastcam or any ChemCam LD RMI image of this critical transition. Complementing the interpreted HiRISE data and facies map (Figs. 23 and 24c) are two graphic logs (Fig. S29 in Supplemental Information S3 and summary Fig. 26) showing the abrupt increase in DTBf in that marks the lower contact of the Transition to uGVR package, as well as the enigmatic upper contact. There does not appear to be major facies change crossing the upper boundary of this package, despite the clear discontinuity observed in the HiRISE data.

3.8.2. Interpretation

The LTBf beds in the Transition to the uGVR are indistinguishable from those of the upper Nose and are interpreted to have formed through similar debris flow depositional processes. In the lower strata of the Transition to uGVR, where extreme grain sizes (>1 m diameter) occur in DTBf lenses (e.g., Fig. S27 in Supplemental Information S3), the DTBf facies may record coarse grain clusters collected in relatively shallow

water due to grain interactions during transport. Small bars may have developed. The more uniform grain sizes of the DTBf lenses suggests fluvial sorting. It is probable the DTBf lenses shown in Fig. 24b; S27a, c, and e in Supplemental Information S3 represent an obliquely sectioned channel body downcutting ~4 m into the underlying LTBf strata (see Fig. S27a in Supplemental Information S3). If this interpretation is correct, this location features the greatest incision of the DTBf into the underlying LTBf observed by *Curiosity*.

The lack of downlapping stratal geometries and the onlapping relationship with the Greenheugh pediment suggests, like the strata below, the transition to the uGVR was deposited as the depocenter backstepped upslope. There is no indication of a large hiatus in deposition between the Nose and the Transition to the uGVR package.

If the boundary between Transition to uGVR and the Nose is cross-cutting the uGVR strata (Fig. 23, yellow dashed line at left), it would indicate the Transition to uGVR package is younger than the stratigraphy in the uGVR upslope. In addition, this would indicate the first major depositional discontinuity observed in the GVR stratigraphy above the Basal unit. The top of this possible discontinuity rests at elevation −3780 m. This location corresponds to the slope break interpreted as a likely fan apex during depocenter upslope transition (yellow star in Fig. 8), as well as the elevation of the Farah Vallis lake level proposed by Palucis et al. (2016).

The Gediz Vallis fan – even at these high elevations – was likely significantly wider than it is now. The discontinuity (yellow dashed line on the left-hand side of Fig. 23) may therefore be a local feature in a large-scale fan structure and deposition coincident with uGVR construction. The correspondence with the Farah Vallis Lake level raises the possibility that the transition could be a shoreline. Finally, although unlikely, it is possible the discontinuity could be the upslope extension of the basal Siccar Point unconformity (Bryk et al., 2025), which would require the Stimson to post-date the uGVR deposition, which contradicts observed stratigraphic relationships downslope.

4. Stratal architecture

The steep walls showing gently sloping truncated stratigraphy

strongly indicates that the GVR deposits are a remnant of a likely much larger deposit. Given its location downslope from a relatively steep Gediz Vallis which drains (at least presently, a 127 km² drainage area rising >3 km in elevation above this deposit), and its occurrence on a wide planated surface, it seems reasonable to conclude, as others proposed based on satellite imagery, that the GVR deposits represent the eroded core of a once more extensive alluvial fan. Significant erosion has removed the majority of the fan, but in so doing, has revealed vertically more than 175 m of well exposed stratigraphy.

The extensive cross-sectional exposure, uncharacteristic of fans elsewhere on Mars, allowed us to construct facies maps (Fig. 24c, Fig. 25) of the entire GVR in multiple cross-sectional orientations as well as multiple stratigraphic logs that further summarize the stratigraphy of the remaining cross-sectioned fan (Fig. 26). Scaling these graphic logs and hanging them along with the unit boundaries on topography yields a reasonable cross-sectional stratigraphic model of the GVR (Fig. 27). As described above, all the contacts and stratigraphy indicate that the fan was built as a back-stepping aggrading depocenter that began kilometers downslope and progressively climbed the steep mountain front towards Gediz Vallis. This is unusual and deserves further explanation.

In a standard progradational setting, strata at the base of the fan in contact with the depositional surface (here the Basal unit or Stimson) would be expected to lie with bedding either parallel to or downlapping that surface. Instead, we see only onlapping of boulder-rich strata at the boundary. It is possible to produce this style of onlapping stratal geometry through standard progradation in fan environments if the deposit has been cross-sectioned obliquely or perpendicular to the transport direction. In the upper strata of the GVR, it is clear from the variable bedding dips (Fig. 10 and S4 in Supplemental Information S3) and cross-sectioned channel deposits (e.g., Fig. 21c; 24b; S22 and S27; S28 in Supplemental Information S3) that transport directions do vary across the ridge. But the overall axis of the outcrop is roughly parallel to the axis of the former fan and exposure is excellent. In the case of an unfavorably sectioned outcrop, (e.g. oblique to the bedding dip direction) the correlative strata that downlapped onto the Stimson would only be visible if the transport direction was nearly parallel to the outcrop exposure and the lowermost strata would furthermore appear parallel to the depositional surface. This was not observed anywhere in the Gediz Vallis Ridge, despite the extensive outcrop exposure that is on average perpendicular to the strike of the beds (parallel to the average

transport direction).

It is unlikely that downlapping or slope-parallel strata near the base of the GVR were obscured due to poor outcrop geometry. Rather, the complete absence of downlapping or slope-parallel stratal relationships anywhere in the ridge strongly suggests that, above the Basal unit, the fan was built through back-stepping of the depocenter upslope. The only direct evidence for backstepping is where the Claw and Claw-Nose transition interact (see Fig. 18 and S15; S16 in Supplemental Information S3). The boundary between the Claw and the Claw-Nose transition is marked by a facies change from extremely coarse light-toned debris flow deposits to interbedded LTBF and DTBF facies in the Claw downslope. The boundary dips downslope cross-cutting the NW dipping bedding planes and clearly demonstrates the down-dip facies change moves upslope as fan aggradation continued. Additional evidence for back-stepping can also be seen in the proximal Tongue, where some of the bedding dips appear to systematically decrease upsection (Fig. 14). The favorably oriented cross-sectional view of the ridge shown in Fig. 27 can be recast in the form of a Wheeler diagram (e.g., Wheeler, 1958; Dromart et al., 2021). Without adequate temporal constraints on fan formation, arbitrary time steps are assigned to the stratigraphy of the ridge and then the strata are expanded such that times of non-deposition and erosion are more easily viewed (Fig. 28). The diagram suggests that much of the Basal unit overlying the Stimson is likely missing due to extensive erosion of its top surface, although it is unclear if the unit ever extended upslope beyond the marker band. The diagram also demonstrates clearly that the vast majority of the former Gediz Vallis fan is missing. Above the Tongue, for example, based on a downslope projection of the highest layers, there is at least 160 m of missing section – 91% of the estimated 175 m thick deposit that was once there. This estimate further only considers the vertical dimension. Volumetrically, the percent of missing fan is much greater.

Direct evidence for backstepping is represented in both Fig. 28a and b as a NW dipping boundary (shazam line) separating the Claw from the Claw-Nose transition. Further evidence for a back-stepping fan apex is exemplified by the large region labeled ‘Non-Deposition’ on the lower left corner of the diagram (Fig. 28b). The implication is that while material in the Tongue (and material further downslope) was being deposited, the upslope portions of the Basal unit and pediment acted as a transfer zone without substantial deposition or erosion. As the fan apex systematically backstepped – avulsing laterally to fill the

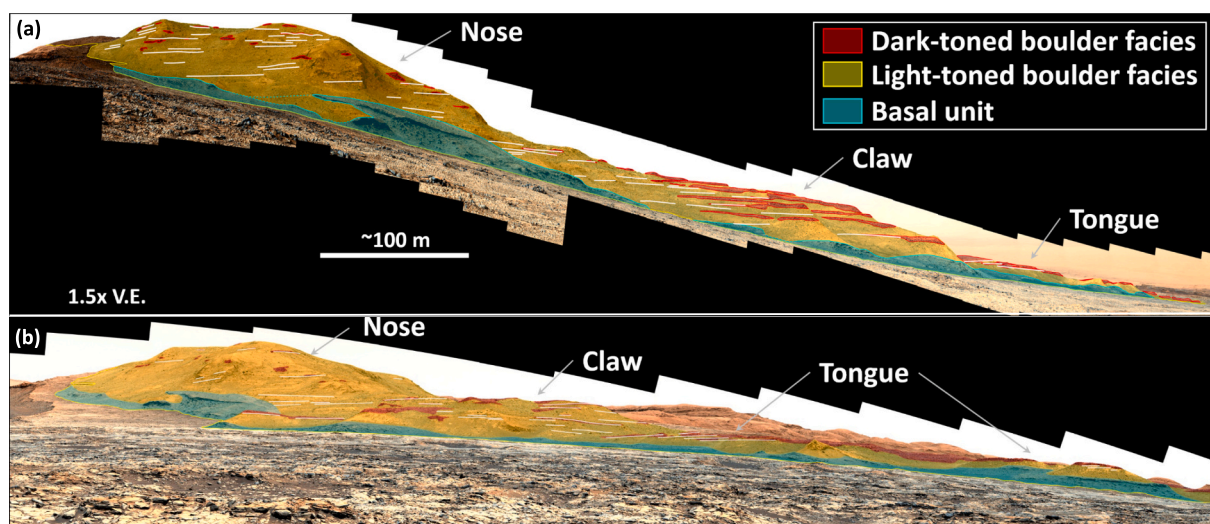


Fig. 25. Facies maps for the Gediz Vallis Ridge showing bedding traces (white), the Basal unit (blue), the LTBF (yellow), and the DTBF (red). a) Facies map overlain on Mastcam data taken from the Blackcraig ascent area. b) Facies map overlain on Mastcam data taken from the Tower butte ascent area. Note the facies map shown of the Transition to uGVR package in Fig. 24c completes coverage of the NE margin of the Gediz Vallis Ridge. See Fig. 3 for ascent area references (i.e., dedicated imaging locations). Mastcam image credit: NASA/JPL-Caltech/MSSS. See Table S1 in Supplemental Information S3 for image sequence identifiers. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

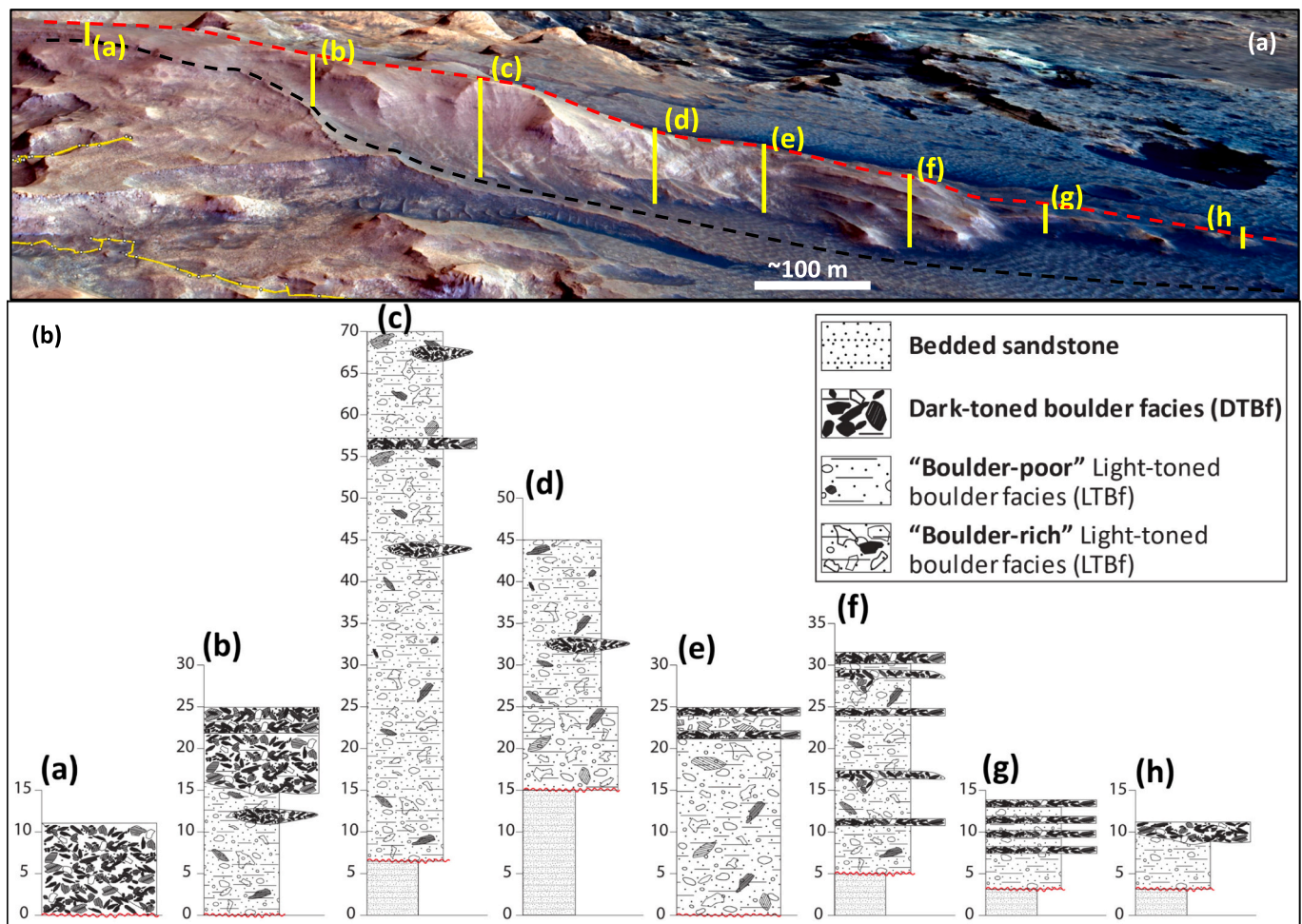


Fig. 26. Compiled stratigraphic columns for the Gediz Vallis Ridge. a) Columns a-h locations overlain on 3D HiRISE data. Topographic profile locations (black and red dashed lines) used in Figs. 27 and 28 are also shown. b) Columns a-h hung on the top of the Stimson. HiRISE image credit: NASA/JPL-Caltech/University of Arizona. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

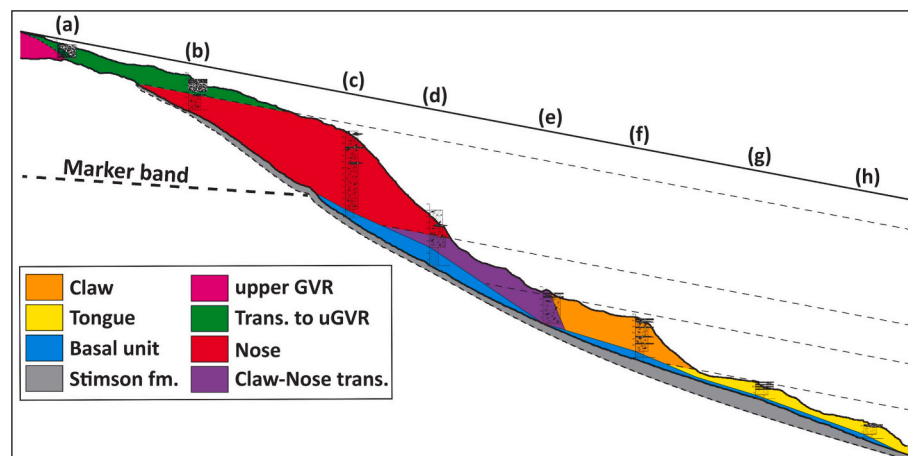


Fig. 27. Interpreted cross-sectional stratigraphic model for the Gediz Vallis Ridge. The top/bottom solid black lines bounding the Gediz Vallis Ridge correspond to the red/black dashed profile lines from Fig. 56a respectively. The base of the Stimson (lowest black dashed line) was extracted from the interpolated basal SPg contact of Bryk et al., 2025). Units are colour-coded and approximate bedding dips are shown as dashed lines dipping at 7.1% grade ($\sim 4^\circ$). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

accommodation space – the next 4 units were deposited. The down-dip facies transitions observed at the boundary between the Claw and the Claw-Nose transition (Fig. 18b; 27; 28) indicate a retrogradational

stacking pattern, which supports the backstepping fan apex interpretation. Down-dip facies changes like this were not preserved elsewhere in the fan, likely occurring more distally and have since been eroded away.

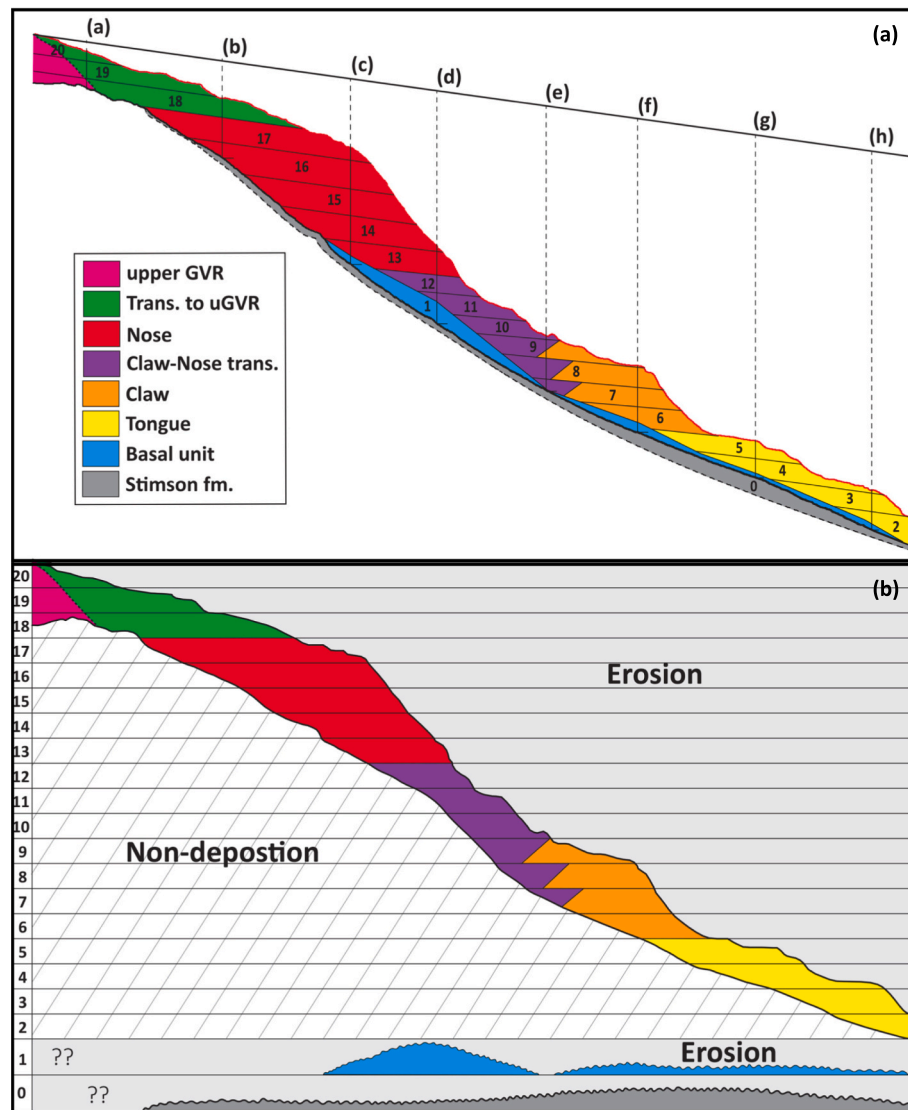


Fig. 28. Chronostratigraphic interpretation for the Gediz Vallis Ridge. a) Vertically expanded (for visualization purposes) cross-sectional profiles (red and black dashed lines) from Fig. 26a including arbitrary timesteps used for chronostratigraphic expansion (b). Note, the unit boundary between the Claw and the Claw-Nose transition is shown here as a zig-zagging shazam line indicating the retrogradational stacking pattern observed by Curiosity. b) Interpreted chronostratigraphic chart (i.e., Wheeler Diagram) for the Gediz Vallis Ridge. Blank areas (indicated by diagonal lines) represent periods of non-deposition or possibly erosion. Areas shaded in gray indicate erosion. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Had these changes been preserved elsewhere in the Gediz Vallis Ridge, the majority would also likely have exhibited retrogradational stacking. One location where retrogradational stacking was not observed was at the boundary between the Claw-Nose transition and the Nose. This boundary is clearly parallel to bedding and does not indicate backstepping. This is likely in part due to the lack of preservation over sufficient section thickness to identify stacking. However, it could also mean the Nose (the largest package in the ridge) represented a more aggregational phase in the overall deposition of the GVR, perhaps through an increase in sediment flux. Ultimately the Nose would have exhibited backstepping architecture as the fan apex continued to retreat upslope towards the mouth of Gediz Vallis depositing the strata above the Nose.

It is possible – perhaps even likely – that some of the ‘Non-Deposition’ region (Fig. 28b) experienced erosion. This could explain why the Basal unit is missing in the highest areas of the pediment and why there appears to be significant erosion along the Basal unit where it is present. However, for the significant volume of sediment to have traversed overtop of the Basal unit, whilst leaving any of it preserved at all, the

unit must have either been lithified, or the transport system must not have incised much despite the steep slope of the pediment. Although the slopes everywhere on the pediment are greater than necessary for debris flow entrainment and possibly incision, erosional wear by debris flows will also depend on development of a coarse relatively mud-free front and will be less on gentler slopes (e.g., Stock and Dietrich, 2006). The preserved Basal unit has an erosional contact with the overlying units and thus likely was lithified before the arrival of these units. It is possible the Basal unit was also extensively eroded prior to the onset of boulder deposition, although some of this erosion (especially upslope) may have occurred as boulders were being deposited downslope.

Despite the difficulty in identifying architectural elements, thickness and lateral extent data were collected for some of the GVR stratigraphy (see Fig. S35 in Supplemental Information S3). Bed thicknesses in general appeared to increase up-section from <1 m in the proximal Tongue, to ~3 m thick in some areas of the Nose (see Fig. S35 in Supplemental Information S3; Table S6 in Supplemental Information S1). In general, beds of the LTbf are thicker than the DTbf throughout the GVR. In the Tongue and the Claw, LTbf beds were the thinnest (typically less than 1

m thick). Higher in the section, LTBF deposits appeared thicker, with some beds appearing up to 3 m thick in the Nose. Similarly, some DTBF lenses in the Transition to uGVR package were at least 2 m thick (see Fig. S35 in Supplemental Information S3; Table S6 in Supplemental Information S1). Some researchers (e.g., Pederson et al., 2015; de Haas et al., 2016, 2018a, 2018b) have suggested the thickness of individual events may be related to the relative level of compensational stacking as the fan aggraded. Thicker bedding in debris flow fans has been interpreted to indicate higher compensational stacking (Pederson et al., 2015). This supports the interpreted change in transport regime at and above the Claw-Nose transition. Thinner interbeds in the Claw and Tongue suggest lower compensation, which is further supported by the laterally extensive DTBF sheets found in those two packages.

Finally, the upper Gediz Vallis Ridge is shown in Figs. 27 and 28 as the magenta region on the left-hand side of each figure. The nature of this boundary and its timing relationship with the ridge downslope is unclear due to poor on-the-ground visibility and ambiguous HiRISE coverage. It is possible that it indicates an erosional truncation of the gently dipping uGVR sediments and onlapping of the boundary by the GVR (see also Fig. 23 and S19b in Supplemental Information S3). If this relationship is real, it would suggest that part or all of the GVR postdates the upper GVR. This seems reasonable in that the upper GV slope is just ~6%, a slope at which debris flows commonly deposit. The preserved layers in uGVR may be associated with the delivery of earlier phases of the fan construction and then were onlapped as the fan apex ascended.

5. Other coarse-grained deposits potentially related to Gediz Vallis

Williams et al. (2018) identified fluvial pebble conglomerates in the Kylie and Kimberley area and proposed them to be associated with the Siccar group, i.e., draping strata that post-dates deposition and exhumation of the lower strata of Mt. Sharp (see Figs. 1 and 2 in Williams et al., 2018 for locations). This now suggests some possible relation to a Gediz Vallis fan. Wiens et al., 2020 documented the composition of three patches of unconsolidated mixtures of pebbles, gravel and boulders located on the eroded floor surrounding the Murray Buttes (Figs. S30 and S31 in Supplemental Information S3). The lithology consisted of sandstone clasts and conglomerates. Chemically the sandstone clasts differ from Stimson bedrock but have similarities to some float rocks in Bradbury rise, whereas the matrix and pebbles in the conglomerates are similar to the Stimson composition. Wiens et al. (2020) proposed 5 possible depositional models to explain their occurrence – three of which involved a combination of fluvial and debris flow transport of material from Mt. Sharp downslope and possible connection to the former Gediz Vallis fan. HiRISE imagery of these patches were used to guide further mapping in this study of likely similar deposits along the northern footslopes of Mt. Sharp (Fig. S30 in Supplemental Information S3). Many of the patches were uncertain, but suggest a wide area of deposition. Where the rover crossed Vera Rubin Ridge, a ~ 40 m long heterolithic boulder deposit (Bressay) was exposed (Fig. S31 in Supplemental Information S3; see Fig. S30a in Supplemental Information

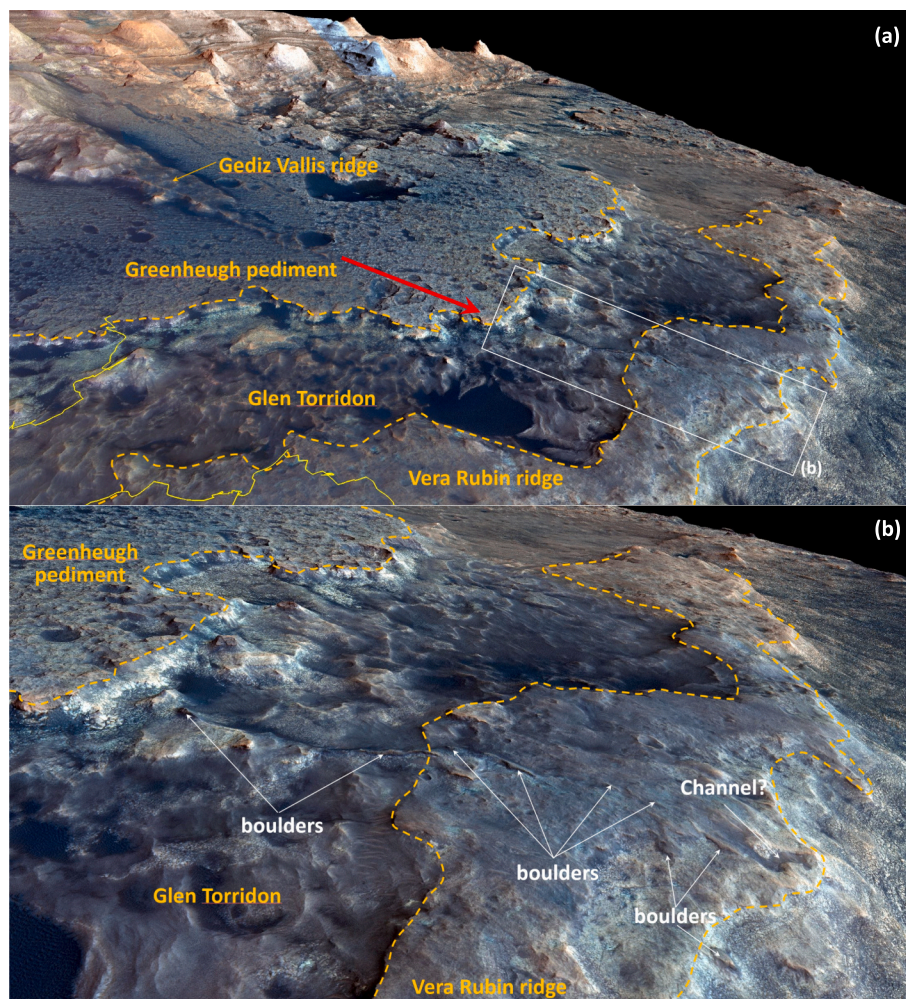


Fig. 29. 3D HiRISE view of the Greenheugh pediment, central Glen Torridon, and Vera Rubin ridge downslope from the Gediz Vallis Ridge. a) Regional view facing SW. b) Close up view showing locations of boulder trains in Glen Torridon and on VRR. HiRISE image credit: NASA/JPL-Caltech/University of Arizona.

S3). Given its location, this deposit likely comprises material shed from Mt. Sharp with possible connection to a Gediz Vallis Ridge fan (Bryk et al., 2025).

Directly (and exclusively) downslope of the modern Gediz Vallis Ridge, numerous linear boulder-rich deposits were observed draped across a region known as Glen Torridon and rising onto Vera Rubin Ridge (Fig. 29a, and b). In Fig. 30, and S32 in Supplemental Information S3, linear boulder ridges comprised of large dark-toned blocks are visible. Some of these features (see Fig. S32b in Supplemental Information S3) appear to be resting on slightly pedestaled country rock, indicating their resistance to erosion.

At 1.7 km to the east of GVR at the same elevation as the transition to upper GVR, there is an exposure of a 115 m long, 6 m thick unconsolidated deposit of coarse boulder-rich sediment that is deposited on a sharp bedrock contact (Fig. 31). The poor sorting and coarse size composition suggest a possible debris flow origin. The exposure must be a truncation of a once more extensive deposit to the north. This similarity in elevation and stratigraphy with GVR suggests that this deposit was formed synchronously, indicating widespread surface runoff from Mt. Sharp during GVR fan development. Finally, dark boulders, are scattered along the footslopes (Fig. S33 in Supplemental Information S3), suggesting significant lateral retreat of the more extensive fan deposits that may have united the GVR with the deposits shown in Fig. 31.

6. Discussion

6.1. Former extent of the Gediz Vallis fan

The exposed stratigraphy of GVR, which indicates no record of lateral depositional boundary to the deposits, the gentle measured depositional slopes, and the evidence of back-stepping fan aggradation up the steep pediment surface all indicate a fan system that grew progressively upslope and developed into a deposit much larger than the present outcrop. Generally, fans once leaving valley confinement build by spreading downslope and radially. Fan slopes may become less steep downslope if debris flow deposition is replaced by fluvial transport further downslope (e.g., Blair, 1999). The slope of fluvially-dominated alluvial fans tends to decrease downslope due to some combination of grain size change and reduced sediment supply (Stock et al., 2008). We have no constraint on what form the fan profile took. The one measure we have is the mean slope of the bedding along its axis, hence the simplest projection is to extend that slope until it intersects the Gale Crater floor.

Fig. 32 shows several Gediz Vallis fan reconstructions with varying surface slopes to show how sensitive the size of the fan is to slope projected. The reconstructions were produced by choosing a terminal fan apex at the hypothesized mouth of Gediz Vallis at the time of deposition (yellow star in Fig. 8), and then linearly projecting downslope at various slope angles. Although alluvial fan geometries in unconfined settings

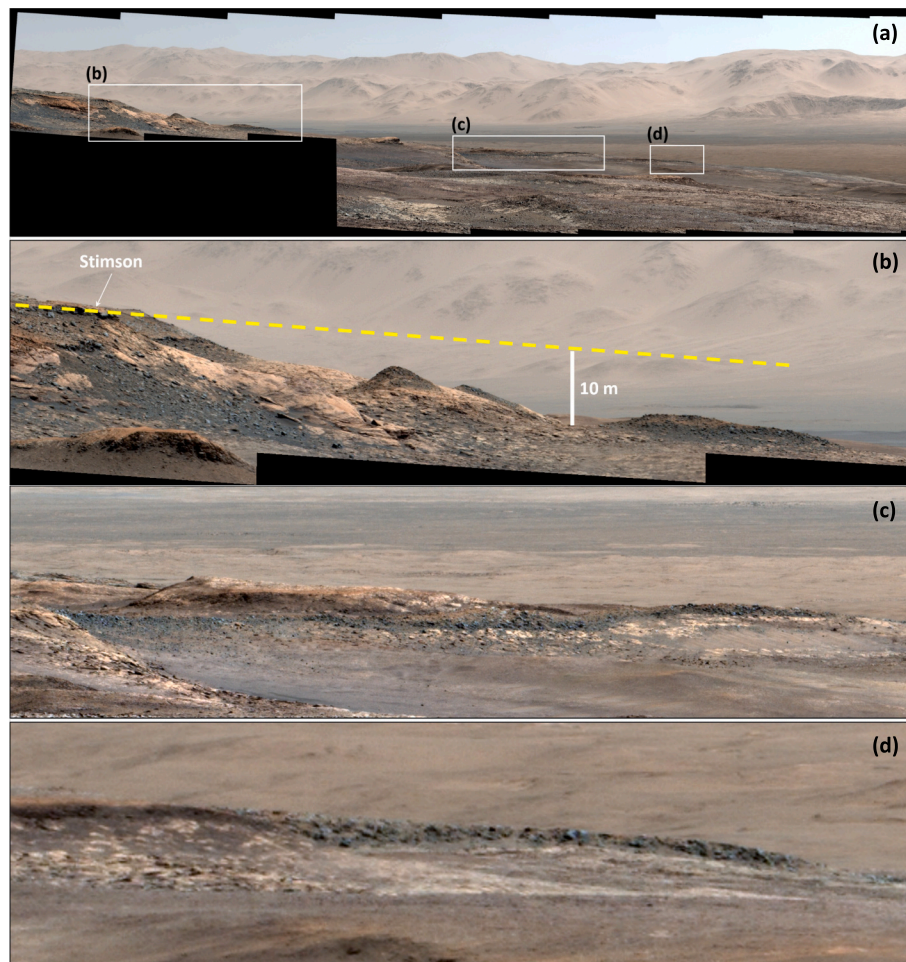


Fig. 30. Mastcam view shot from Vera Rubin ridge showing multiple boulder dark boulder locations in Glen Torridon and on Vera Rubin ridge. a) Wide view with reference locations for b-d. b) Mastcam mosaic showing boulder deposits roughly 10 m below the projection of the Stimson formation approaching and ascending the southern slope of Vera Rubin ridge. c-d) More distal boulder trains approaching and ascending the southern slope of Vera Rubin ridge. Mastcam image credit: NASA/JPL-Caltech/MSSS. See Table S1 in Supplemental Information S1 for image sequence identifiers.

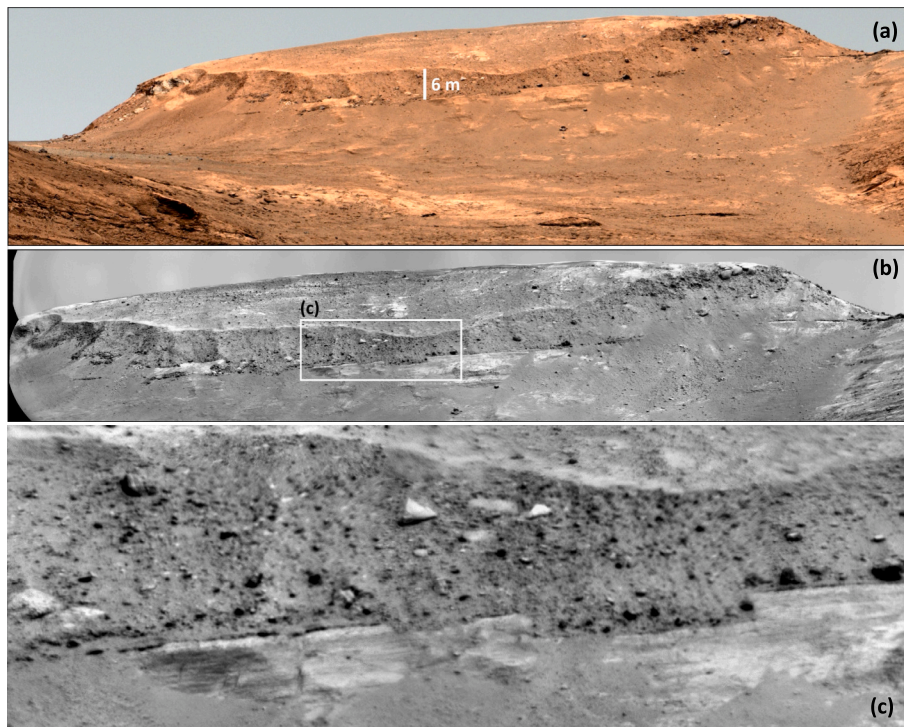


Fig. 31. Mastcam images of unconsolidated boulder deposits east of the Gediz Vallis ridge (see Fig. 3 for reference location). a) Mastcam and b) CCAM LD RMI mosaics of unconsolidated boulder-rich deposit adjacent to the Gediz Vallis Ridge. c) Closeup CCAM LD RMI mosaic showing poorly sorted light and dark-toned boulders. Mastcam image credit: NASA/JPL-Caltech/MSSS. ChemCam image credit: CNES/LANL/IRAP/IAS. See Table S1 in Supplemental Information S1 for image sequence identifiers.

tend to expand laterally over a full 180 degrees, this geometry is typically associated with a fan slope that rapidly declines onto a valley floor. Sediment leaving Gediz Vallis would have descended a much steeper pediment slope than the feeder channel. That may have restricted the lateral spread. Furthermore, some experimental studies on fans built by successive debris flows indicate reduced lateral spreading due to deposition on steep lateral slopes (e.g., de Haas et al., 2018b). Anderson et al. (2023) show the outline of many fans north of the Hellas basin, several with apex angle openings well below 180 degrees. To account for range of apex angles, in addition to changing slope, three different fan apex opening angles and a range of slopes were used (Fig. 32b) (see Table S4 in Supplemental Information S1). The 7% slope (matching mean dip of the GV ridge bedding) and an apex opening of 120 degrees may reproduce the ultimate extent of the fan as it encompasses all the coarse boulder deposits identified by Williams et al. (2018) and Wiens et al. (2020) (Fig. 33a). In Fig. 33b the projected longitudinal profile indicates the fan could have been over 260 m thick as it crossed onto the crater floor, all of which was later likely removed through wind erosion.

The projected fan area of Gediz Vallis can be compared to fan areas as a function of catchment contributing area for fans on Mars and Earth (Fig. 34a). The red bar represents the range of fan areas for the various fan openings (60 to 170 degrees). For the 120 degree opening, the 7% slope fan has an area of 140 km² which corresponds catchment area of 127 km². The calculated fan area matches Mars and Earth values for the same catchment area, and lies along the general trend of the data, but fan area just one-tenth the size would still fall within the range of observations. The smaller red triangle is the current area of GV ridge, which is at least two orders of magnitude, and probably three, below Mars fans. The blue triangles in Fig. 34a show the size of fans at mouths of three canyons carved into Mt. Sharp to the west of Gediz Vallis (see also Fig. S8 in Supplemental Information S1). All three retain a fan form and two have been interpreted as fan deltas. They lie along the bottom of the general data field.

As shown in Fig. 33a, the projected fan area crosses the locations

where unconsolidated coarse sediment deposits were mapped and reaches the proposed fluvial conglomerates on the crater floor. These deposits may be unrelated to the Gediz Vallis fan, but their projected co-occurrence is suggestive.

Another test of the predicted fan size is to compare the volume of the fan with that of the eroded volume that formed the source catchment area. We use a thin-plate method after Bryk et al. (2025) to reconstruct the topography of the catchment area of Mt. Sharp prior to its incision. Similarly, the volume for the 7% sloping fan were calculated by differencing the reconstructed fan surface from modern topography. A hypothetical reconstructed fan volume was also calculated using the former extent of the Greenheugh pediment as the base topography (Bryk et al., 2025) to evaluate if including the likely full extent of the pediment matters to the volume estimate.

The reconstructed fan volumes for the 7% sloping fan were 16.50 km³ for modern topography and 15.11 km³ when using the reconstructed Greenheugh pediment (Bryk et al., 2025). The modern catchment projected missing volume is 8.64 km³. These results suggest the modern catchment is roughly half the volume of the hypothesized fan reconstruction. Anderson et al. (2023) compared fan volumes to eroded catchment volumes on fans in the Hellas basin and found both fan volumes that were much smaller and fan volumes that were much larger than the missing catchment volume (Fig. 34b). While the smaller fan volumes could be attributed to post deposition erosion of the fan, the larger volume they found to be problematic. Our observations are less constrained since we are predicting the original size of the fan. It may be that the projection is too large by a factor of about 2, or that the catchment 3D shape does not record all the missing mass. Given the large nearly planar contributing slope to the catchment, we suggest this later explanation is likely a factor.

Recasting the hypothesized former fan volume as a thickness of material eroded from the catchment indicates that roughly 125 m of material would need to have been eroded across the modern Gediz Vallis catchment to construct the former fan. Given that, since deposition of

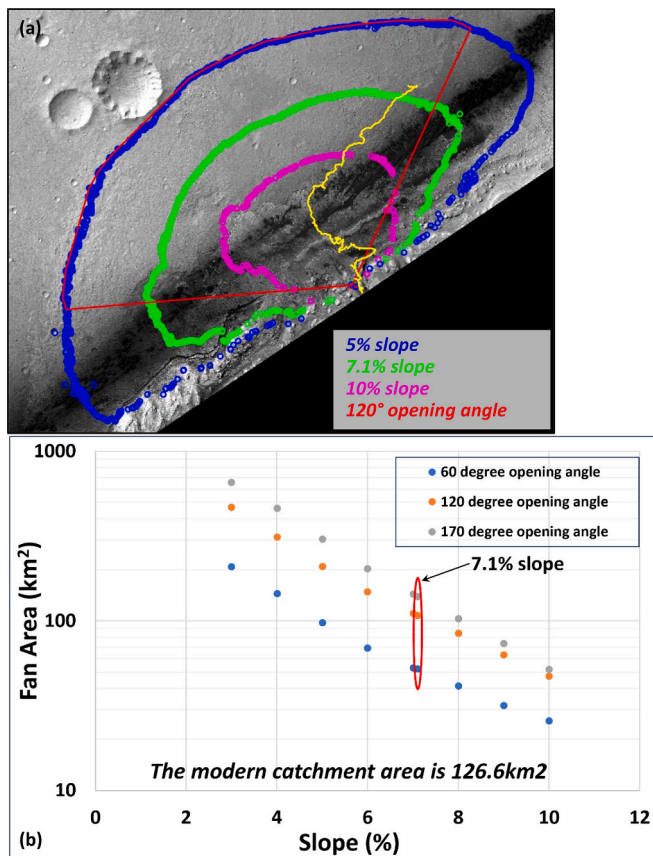


Fig. 32. Areal reconstructions of the Gediz Vallis fan. a) Regional fan extents for fans exhibiting linear surface slopes of 5% (blue), 7.1% (green), and 10% (purple), overlain on modern surface topography. Context Camera (CTX) imagery on board the Mars Reconnaissance Orbiter (MRO) was used for the regional orthoimagery and modern topography (Malin et al., 2007). b) Semi log plot of reconstructed fan area vs. fan surface slope. Three fan apex opening angles were included (60°, 120°, and 170°) to account for possible partial filling of the regional accommodation space. An example 120° opening angle is also shown (red) in (a). The red oval indicates the reconstructed fan areas that correspond to the average dip of the beds measured in the modern Gediz Vallis Ridge. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

the GVR, upwards of 80 m of vertical erosion just by wind scour has occurred, this 125 m does not seem exceptional for erosion by active fluvial and debris flow processes (albeit over a shorter time). Furthermore, the GV fan volume to catchment area falls within the scatter of the relationship for 26 basins north of the Hellas basin reported by Anderson et al. (2023) (Fig. 34c), suggesting similar rates of erosion. Nonetheless, the original areal size and volume of Gediz Vallis are poorly constrained.

What does differ significantly between the predicted GV fan size and volume and that reported across Mars is the current very small residual volume. As discussed above, perhaps this reflects the unique occurrence of a fan that was built of highly erodible aeolian central mound deposits as compared to crater rim wall material. The three fans to the west of GV ridge that issue from canyons cut into Mt. Sharp also are anomalously small compared to their source areas. The Peace Vallis fan volume, formed on the crater wall, can be fully accounted for by the sum of the gully network that fed it (Palucis et al., 2014), yet its fan area (80 km²) is relatively small for its catchment area (730 km²). This is because the fan was built just from a gully network cut into the catchment floor: the entire catchment did not contribute sediment.

6.2. Influence of lakes

We did not identify any stratigraphic evidence for the proposal by Palucis et al. (2016) that lakes influenced the deposition of the Gediz Vallis Ridge sediment. In a few places, in the Light-toned Boulder facies, there may have been fine layered sediments but image resolution was inadequate to draw a conclusion. GV ridge is a small remnant of a much larger deposit, and the characteristic foreset bedding of deltas may have not been preserved. Based on delta deposits found on the crater wall perimeter and on Mt. Sharp, Palucis et al. (2016) identified three lake levels (−3280 m, −3780 m, and −3980 m), associated with mean lake depths of 700, 400 and 300 m, respectively (see also Fig. 5 inset). The highest lake level was associated with a regional large-scale flow system that led to canyon cutting and fan and delta development in Gale and in surrounding craters. This flow entered Gale via groundwater which caused seepage erosion that developed into the canyon Farah Vallis. This lake would have crossed Gediz Vallis well upslope of the Gediz Vallis Ridge deposits and roughly coincident with the downslope edge of the Yardang unit (at −3280 m). The next lake level was the shallowest (at −3980 m), which is coincident with the sheet flood deposits of the Claw. Palucis et al. (2016) proposed the third lake level reached −3780 m, which places it close to the large break in slope between the upper Gediz Vallis Ridge and the steepened descent down the pediment surface. Palucis et al. (2016) proposed a back-stepping climbing depositional sequence in the GVR deposits, just as we found in the stratigraphic record. Based on crater wall gullies that terminate at the two lower lake levels, Palucis et al. (2016) proposed local precipitation- runoff processes generated the erosion and deposition and gave rise to the lakes that each persisted for at least 1000 years. The near exact framing of the GVR deposits by the two lake levels is suggestive, as production and transport off upper Mt. Sharp is more likely a precipitation- induced runoff-driven process (rather than groundwater), and several of the deltas that defined the lake levels were formed on the footslope of Mt. Sharp. This then suggests that GV ridge fan construction could have been synchronous with if not controlled by the hydrologic activity that drove putative lake level rises.

6.3. Sediment source and sediment production

The dominant light-toned sediment in Gediz Vallis Ridge has an abundant potential source in the catchment area which appears to be light-toned throughout. At that transition to the Yardang unit, a large landslide scar is evident on the eastern slope and large blocks are scattered downslope (Fig. 35a). Massive blocks (some >10 m across) line the floor of Gediz Vallis up to the entrance of the valley cut through the Yardang unit (Fig. 35b, c) near the landslide scar. Above the Yardang unit in the upper catchment no such scars exist, and the typical ridge and canyon topography cut by repeated debris flows is absent. Such canyon networks are found in the sources area of Martian crater wall fans (e.g., Morgan et al., 2022). Further to the east, outside of the catchment, large lobate features downslope of an amphitheater setting have been mapped as landslides or glacial or periglacial features. (e.g., Anderson and Bell III, 2010), but no such features are visible within the catchment. To produce the debris flows that swept down the canyon the bedrock needs to be broken, made mobile (separation from the source bedrock), and entrained with sufficient water-rich fines to mobilize the flows on these modest slopes. Sediment production may have been facilitated by periglacial processes, though evidence remains to be identified. The absence of canyons and bouldery colluvium suggests there has been significant wind scour post fan formation, erasing and perhaps burying such features.

There is no clear upslope source for the dark-tone gravel and boulders prominent in the Gediz Vallis Ridge deposits. The clasts appear similar to Stimson formation (e.g. Banham et al., 2022; Bedford et al., 2022). Perhaps some of the clasts in the Claw and Tongue were derived from entrainment by cascading debris flows that swept down the

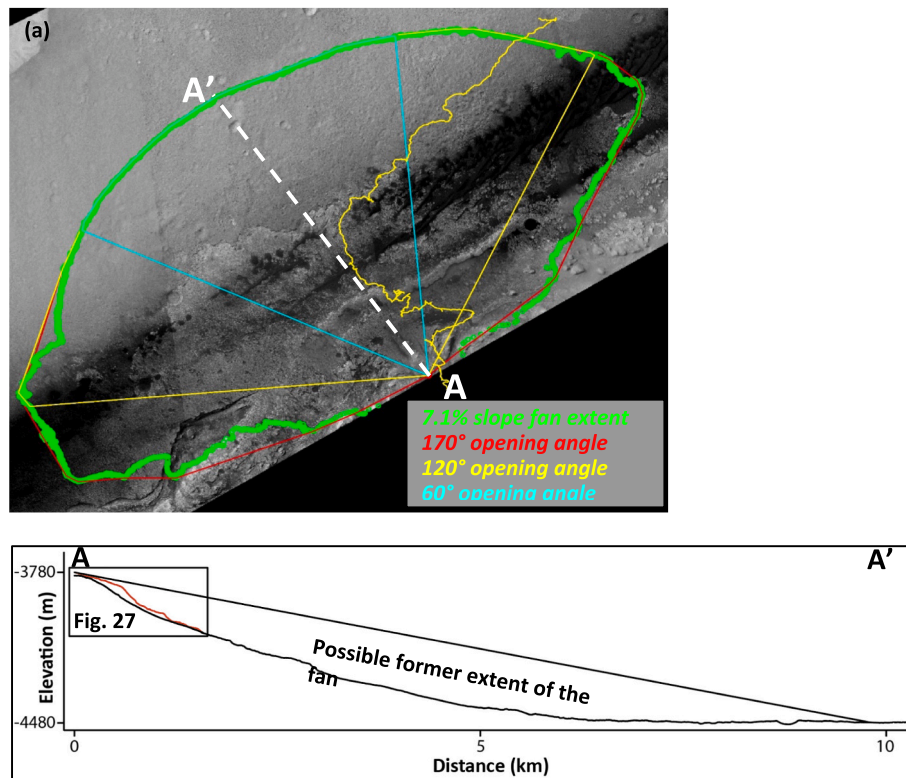


Fig. 33. Likely aerial extent of the former Gediz Vallis fan. a) Estimated fan maximum aerial extent made using a 60°, 120°, and 170° opening angles, and a constant surface slope of 7.1% – representative of the mean slope of the beds measured in the GVR (see Fig. 14). b) Cross-sectional profile A-A' showing the likely former extent of the fan in relation to the remaining deposits (red). Context Camera (CTX) imagery on board the Mars Reconnaissance Orbiter (MRO) was used for the regional orthoimagery and modern topography (Malin et al., 2007). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

exposed steep pediment surface enroute to the floor of Gale. This would not explain the significant increase in dark-toned rocks in the transition unit to the upper Gediz Vallis Ridge, nor the exclusively dark-toned layers deposited on the upper Gediz Vallis Ridge that leads to Gediz Vallis. Inspection of the Gediz Vallis walls and floor showed only no places of dark resistant outcrops that appear to generate blocks. Upslope beyond the large Gediz Vallis, much of the trough through the Yardang is floored by the Yardang unit and no outcrops were found in the upper catchment. One explanation is that the dark-toned rocks were derived from either an upslope extension of the Stimson, or an earlier similar lithified aeolian deposit that was subsequently completely removed by the subsequent climate condition that led to runoff and both debris flows and sheet flood transport events. This would require entrainment of sheets of erosion-resistant bedrock. If prior wind scour had cut bluffs into the rock and exposed underlying weak light-toned bedrock, concentrated flows could undermine and entrain.

As noted earlier, the estimated fan size of 140 km² based on bedding projection and a wide aspect angle requires some 15 km³ of sediment to be derived from the catchment area. The current topography does not suggest such a large sediment production. If just 1% of the total fan was dark-toned rocks, it would still require 1.5 km³ of source bedrock, and if spread uniformly across the catchment would predict a mantling 11 m thick. Such a uniformly distributed layer is unlikely. A smaller size of the fan, perhaps due to a lake-controlled rise would reduce the problem of identifying dark-toned source bedrock, but not eliminate it.

6.4. Implications for climate

The stratigraphy of the Gediz Vallis Ridge deposits indicates a shift from a basal aeolian deposit to liquid water-driven processes that were likely subject to (possibly cyclic) variations in water content and

consequent dominant transport process, from debris flow transport and deposition to fluvial reworking and sheet flood deposits. These variations may reflect internal dynamics, but also the timing and magnitude of surface water runoff. At a larger scale, the distinct packages of the Tongue, Claw, and Nose and the two transitional units indicate step changes in relative frequency and lateral extent of the two dominant processes. Some of this variability may also record internal dynamics of catchment sediment generation processes but given the large difference in water volumes that drive debris flow versus fluvial transport, it seems likely the cyclicity and the distinct packages reflect variable climate conditions controlling runoff production. Though there could be significant time gaps between the packages, we saw no clear evidence of such in the exposed stratigraphy. There is the possibility that some of the step changes in sediment packages is related to the episodic lake activity in Gale and climate on Mars in general. As reviewed by Anderson et al. (2023), studies of Martian fans commonly conclude that they are the result of multiple phases of construction and possibly reflect changes in the relative dominance of fluvial versus debris flow processes. They concluded in their study of fans north of the Hellas basin that fan morphologic variability records a high degree of climate-driven intermittency of fan construction and erosion. Gediz Vallis Ridge observations provide the first field observations on a steep fan that support such satellite-based interpretations.

Commonly, the amount of water required to build a fan is estimated either from discharge data inferences from paleochannels, or from assumptions on the amount of water to mobilize debris flows or transport sediment fluvially (e.g., Morgan et al., 2014; Palucis et al., 2014, 2016, and Anderson et al., 2023). In Gale Crater, Palucis et al. (2014, 2016) have concluded that the Peace Vallis fan on the north crater wall and the Pancake Delta on the south crater wall were deposited by fluvial processes. They derived from terrestrial river data a rock/water ratio of

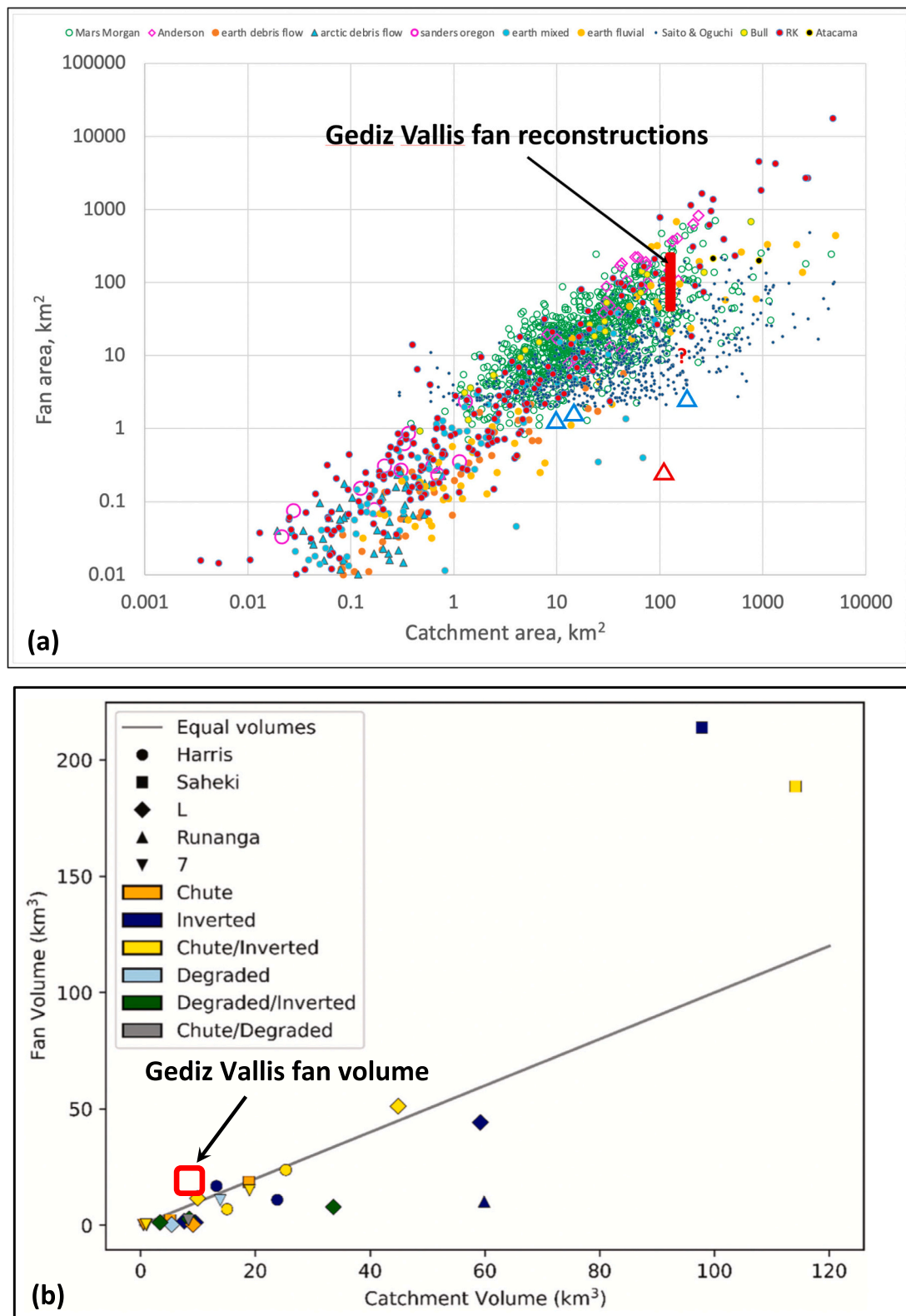


Fig. 34. Comparison of fan/catchment areas and volumes for Earth, and Mars, including Gediz Vallis. a) Plot of fan area vs. Catchment area for Earth and Mars. Data compiled for terrestrial fans are from: Blair debris flow (Blair and McPherson, 2009), debris flow, mixed, blair sheetflood (Blair and McPherson, 2009), fluvial (Dade and Verdeyen, 2007), RK (Arabian fans), Bull (Bull, 1964), Atacama (Haug et al., 2010), Debris flow arctic (Palucis et al., 2022), Fluvial arctic (Palucis et al., 2022), Saito & Oguchi (Saito and Oguchi, 2005), Ashima (Saito and Oguchi, 2005), Kurobe (Tebakari et al., 2013), Shoufeng (Saito and Oguchi, 2005), Kiso (Ohata et al., 2023). Martian fan data compiled from: Mars fans (Morgan et al., 2022; Mondro et al., 2023), Anderson (Anderson et al., 2023), Saheki (Morgan et al., 2014). The range in reconstructed GVR fan extents for the modern catchment area are shown as a red bar. The modern fan extent corresponds to the red triangle. Blue triangles indicate other Mt. Sharp fans west of Gediz Vallis. b) Fan volume vs. Catchment volume (Anderson et al., 2023) and c) fan volume vs. Catchment area reproduced from Anderson et al. (2023) including reconstructed values for Gediz Vallis. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

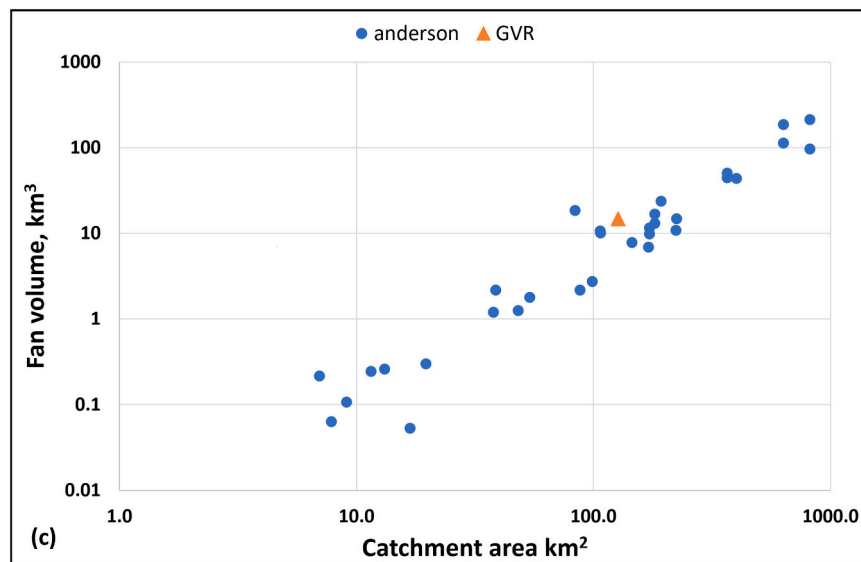


Fig. 34. (continued).

0.00008. If applied to the estimated volume of $\sim 15 \text{ km}^3$ of sediment in the modeled Gediz Vallis fan, a fluvial transport regime would require about 187,500 km^3 , or nearly 1500 km of runoff from the catchment area. Debris flow deposits dominate in the current Gediz Vallis Ridge, however, so a very different water amount is needed to build the fan. Water contents in measured debris flows varies considerably, with a range of less than 10% to greater than 60% (e.g., Barnhart et al., 2021; Zhang et al., 2023; Chen et al., 2023; Pierson, 1981). If we use 30% as an intermediate value, we get an estimated 5 km^3 of runoff, or the equivalent of 35 m of runoff from the catchment. Fluvial processes did contribute to the presently preserved fraction of the fan and may have been significant on more downslope parts of the fan, hence the 35 m is well below the likely total runoff. Unfortunately, given the uncertainty about the fan size we cannot constrain further the water that drove the fan construction, but we can say that liquid water was involved, and that the amount of surface runoff driving fluvial processes varied during fan construction.

Given the large volumes of water needed to build and sustain the three lake events proposed by Palucis et al. (2016), the corresponding deltas along the western margin of Mt. Sharp footslopes, and the numerous gullies on the south facing crater wall that terminate at the two later lake levels, it seems possible that Gediz Vallis fan construction occurred during these significant hydrogeomorphic events. The highest (and first) putative lake was likely derived from groundwater influx via the large canyon (Farah Vallis) in the south; consequently the Gediz Vallis fan activity may not have been initiated at the time of its formation. The latter two lakes were associated with the surface runoff that formed mapped gullies, canyons, and small deltas. The climate conditions favorable for lake and fan development may have been associated with snow accumulation and melting (as discussed in Palucis et al., 2016 and references therein).

6.5. Timing of Gediz Vallis fan

Bryk et al. (2025) reports a detailed analysis of the likely sequence of erosion and depositional events leading up to deposition of the Gediz Vallis fan and its subsequent erosion. In essence, at the base of Mt. Sharp, there was an extended period of pediment development via wind scour, then burial and lithification of the pediment by Stimson aeolian sediments, and subsequent wind scour of the Stimson and Mt. Sharp bedrock (and possibly the Basal unit) that ended abruptly with the onset of the Gediz Vallis debris-flow/alluvial fan development. At the cessation of

the wet period of fan construction, there was a return to significant wind erosion that swept nearly all of the fan away. Based on exposure age dates (Martin et al., 2021), wind erosion has been only marginally active over the past billion years.

Palucis et al. (2020) re-examined age estimates in Gale Crater using a probabilistic approach that enables broad estimates to be made for relatively small landforms. They conclude that the age of the Pancake delta (formed during the highest lake level stand) could range from 2 to 3.8 billion years, but with a likely value of 3.2 billion, consistent with the results of crater counting of the 14,200 km^2 floor and wall of Gale (3.3 to 3.5 Ga). Peace Vallis fan on the south facing crater wall (at -4250 to -4500 m), which lies well below the lowest lake level, has experienced much less erosion than Gediz Vallis. Palucis et al. (2020) conclude that this fan could be ~ 1 to 2.5 billion years old, in general agreement with others (e.g., Grant and Wilson, 2019). Palucis et al. (2016) suggest that although significant time may separate the occurrence of the three lakes, that they occurred during a wetter lake-supporting time and thus likely closer to the Pancake delta formation time (Late Hesperian) than to the Peace Vallis construction (Amazonian). These age assignments are consistent with fan ages estimated across Mars (e.g., Grant and Wilson, 2011; Wilson et al., 2021; Morgan et al., 2022).

7. Summary and conclusions

This first rover-based investigation of a steep fan deposit on Mars reveals a vertically varying stratigraphic record of shifting dominance of debris flow and fluvial processes. The deposit drapes an underlying bedrock topographic profile (Greenheugh pediment) that abruptly steepens from a slope of 6% to greater than 30% and then subsequently progressively declines to 10% downslope. At the base of much of the deposit is a dark-toned Basal-unit sandstone which reaches a maximum thickness of 15 m, and appears to have been deposited, lithified, and then eroded prior to or during the deposition of subsequent arriving sediment. It was deposited parallel to the bedrock boundary and is likely of aeolian origin. Above this sandstone, are five distinct stratal packages, four of which had been previously identified based on remote sensing observations (Palucis et al., 2016). The bedding in these five packages lie at about 7%, a slope much gentler than the underlying bedrock topographic profile it mantles, but one consistent with deposition by debris flow and fluvial sheetfloods.

The five units consist of two distinct interbedded facies: a dominantly light-toned coarse (boulder-rich) sediment facies (LTBF), and a dark-

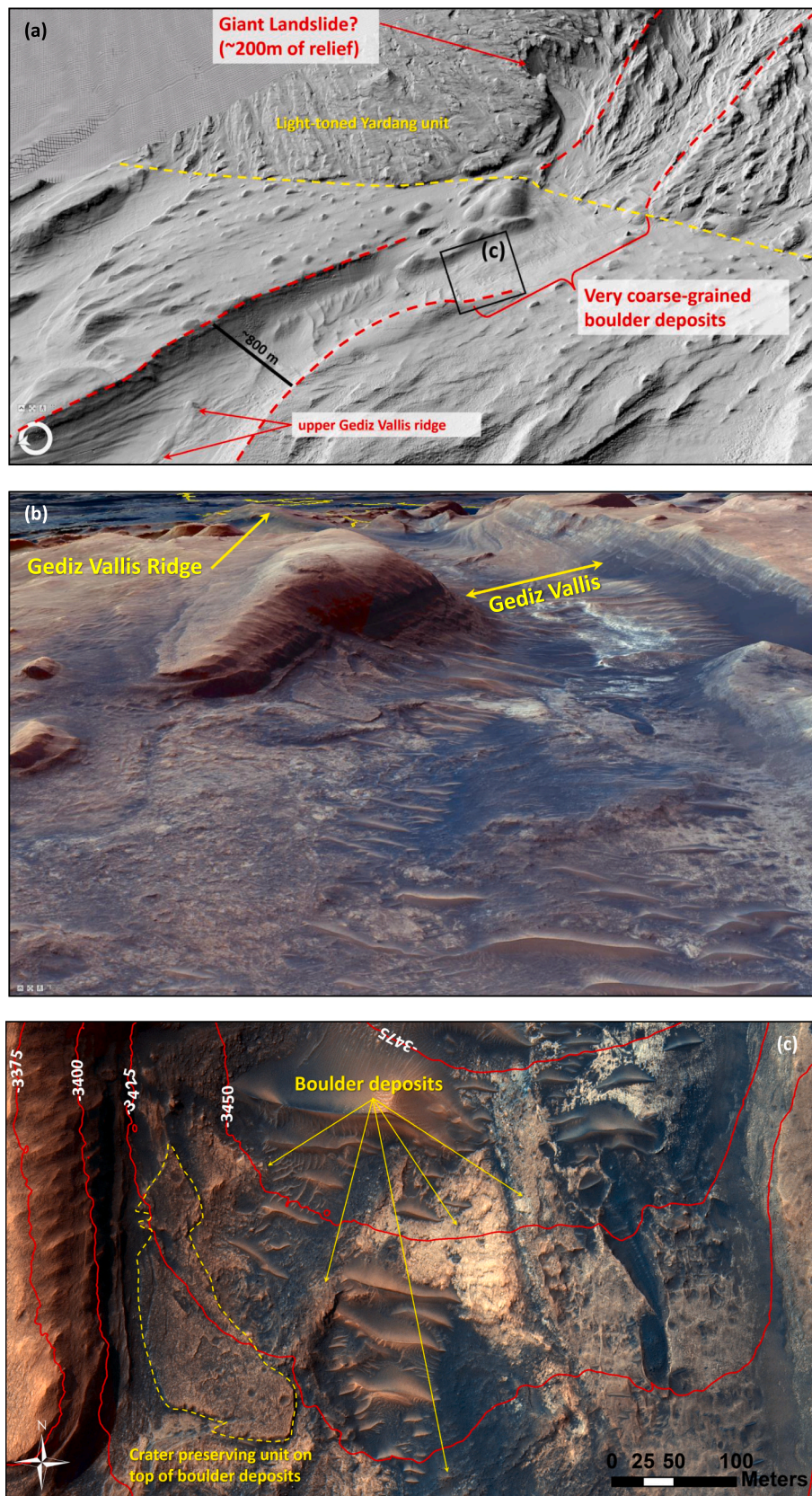


Fig. 35. HiRISE data from Gediz Vallis far upslope from the Gediz Vallis Ridge. a) 3D Shaded relief map of Gediz Vallis upslope from the Gediz Vallis Ridge including parts of the upper Gediz Vallis Ridge and the lower light-toned yardang unit (Anderson and Bell III, 2010). The valley is outlined in red and the yardang unit boundary in yellow. b) 3D HiRISE view looking North down Gediz Vallis towards the Gediz Vallis Ridge. c) HiRISE orthoimagery with 25 m topography overlay showing multiple boulder deposits all overlain by a thin crater preserving deposit (yellow dashed region). HiRISE image credit: NASA/JPL-Caltech/University of Arizona. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

toned coarse (boulder-rich) sediment facies (DTBf). The light-toned facies make up the vast majority of deposit, occurring as 1–4 m thick primarily matrix-supported beds with varying clast concentrations. The light-toned layers consist of poorly sorted coarse gravel and boulders supported by a finer-grained light-toned matrix. They outcrop with steep surfaces in which wind has eroded the light-tone boulders flush to the surface. Scattered throughout are dark-toned gravel, cobbles, and boulders that, in contrast to the light-tone sediment, protrude prominently from the outcrop face. Individual dark-toned boulders are commonly layered and in the larger blocks cross-bedding is occasionally visible. Dark-toned gravel and boulder deposits occur in two forms: Thin (≤ 1 m thick) and laterally extensive (up to 140 m wide) sheet deposits found commonly in the lower (Tongue and Claw) units, and isolated lens deposits more common in and above the Nose unit higher in the stratigraphy. Dark-toned lenses exhibit varying length and thickness, but typically are less than 1 m thick and do not extend laterally more than 30 m. These lenses tend to have concave up lower boundaries, suggesting shallow incision into the underlying light-toned units. The dark boulder concentrations commonly appear better sorted than the light-toned facies, and typically display grain-to-grain contacts.

We propose that the light-toned deposits are dominantly debris flow deposits and that the dark-toned lenses primarily record erosion of the underlying debris flow deposits by fluvial incision and concentration of entrained resistant dark-toned clasts. Fluvial transport and abrasion results in the destruction of light-toned clasts, reducing their presence in the fluvial lenses. In some cases dark-tone clasts may have been transported to the fan by fluvial processes.

The upper five units are distinguished by the relative frequency and lateral extent of dark lenses. In the lower two units (Tongue and Claw), most of the dark lenses form sheet flood deposits that probably dispersed across a braid plain. In both of these units the frequency of dark lenses increases near the top of the section. In the overlying two units (Claw–Nose transition, Nose), the dark-boulder facies is less prevalent, but where present, forms lens deposits that likely record patches of coarse sediment, and the infilling of fluvial channels in others. In the highest unit (Transition to the uGVR), dark-toned boulder deposits increase in frequency but remain isolated lenses never recovering the tabular form found in the lowest two coarse units (Tongue and Claw). These clear shifts in the abundance and form of the dark-toned facies perhaps suggests temporal changes in runoff and/or sediment supply within individual units and among the successive units. We found no evidence of extended periods of non-deposition between units. Our observations argue against a fan built via single short pulse of deposition, but rather indicate fan construction that occurred over the course of thousands of individual events. However, with nearly all of the former fan missing and no absolute dates, we cannot accurately estimate the duration of the depositional period.

Grain size analysis in each of the units above the Basal sandstone reveal a median overall grain size of 180 mm, with little variation between the dark- and light-toned facies. The largest clasts in the GVR were roughly 3 m in diameter and included both light- and dark-toned boulders. The largest median grain size (336 mm diameter) was found in the center of the ridge in the unit known as the Claw–Nose transition.

Bedding dips of 7° in the ridge are everywhere lower than the underlying bedrock on which the beds were deposited. The result is an onlapping relationship between the boulder-rich strata and the underlying Basal unit and Greenheugh pediment. This persists through all 5 boulder-rich units above the Basal unit and forms a retrogradational (back-stepping) stacking pattern. This pattern would arise by deposition through a systematically back-stepping fan apex (depocenter). Given the steep basal bedrock topography, debris flows and fluvial transport may have been non-depositional or erosive until they reached lower gradients on the crater floor. The slope of the boulder-rich stratigraphy is approximately constant up section, suggesting a characteristic slope of debris flow deposition (of about 7°), across which fluvial reworking occurred.

The original size of the fan is uncertain. A simple downslope projection of this characteristic stratigraphic slope would extend the fan roughly 10 km to the crater floor. Using an apex opening angle of 120 degrees, the 7° slope projects out a fan to have been roughly 140 km². Such a fan size for the observed slope matches that found on Mars fans generally. It is likely the more distal sediments were fluvially dominated (as commonly occurs in fans) and it is conceivable that finer sediments may have been dominated by light-toned sediments (with consequently less steep slopes). A smaller fan apex angle (as can be seen in some Martian fans), however, would reduce this size to about 50 km². The 140 km² fan would have had a volume of about 15 km³, equivalent to about 125 m (ignoring density changes) of erosion from the source catchment, a value that exceeds estimates of the missing volume generated from the catchment. Furthermore, even if the dark-toned sediment is only 1% of the deposit, it would require 1.5 km³ of bedrock upslope, equal to 11 m thickness across the entire catchment. We have found no significant outcrop of the dark-tone unit in the catchment area, which suggests either it was not a large volume, or that it was spatially restricted. Two of three late stage (late Hesperian to Amazonian) lakes (proposed by [Palucis et al., 2016](#)) that formed in Gale Crater correspond to the Claw and the transition to the upper Gediz Vallis Ridge. A rising lake level could limit the preserved fan size. No foreset beds nor fine grained deposits consistent with a lake influence, however, were found in our imagery. [Palucis et al. \(2016\)](#) proposed a rising lake level that drove a back-steeping fan aggradation. This is consistent with the stratigraphy that indicates a back-stepping fan apex. The exceptional small residual deposit of the fan may have been driven by a dominance of weakly lithified light-toned sediment that likely made up the bulk of the deposit and strong abrasive winds. The erosional form of these residual deposits, in which we could detect no evidence of veins, fractures or tendency to form blocks (except for the Basal unit) suggest that the upper five units are unconsolidated.

While one landslide scar was observed in the Yardang unit, further upslope in the majority of the catchment, the topography does not display typical ridge and canyon morphology that produces debris flows, and the much of the catchment is less than 14 degrees. No evidence was visible for the processes that would produce boulders and transport them out to the fan in the upper catchment.

The reconstructed fan size, measured bed slopes, and catchment area places the modeled fan in the field of mapped fans on Mars and consistent with terrestrial fans built by debris flows and fluvial processes. Unlike most terrestrial and Martian alluvial fans, the initial conditions are unusual, with the mouth of Gediz Vallis at the time of deposition located hundreds (< 650) meters above the crater floor. It also differs from mapped Mars fans in that it formed on the inner mound rather than on the crater wall and is mostly eroded, leaving no fan shape to map.

This first rover-based investigation of a steep alluvial fan deposit adds strong support for interpretations derived from remote sensing that Mars fans have been built episodically through a combination of debris flow and fluvial processes. This points to possible repeated surface water generating climate events that may have been short lived rather than a singular long period. The construction of the fan by eroded blocks of exceptionally weak aeolian deposits led to nearly complete erosional removal of the fan by wind. This massive erosion probably requires a denser atmosphere with stronger winds than current and was likely accentuated by the fan being located on the north facing mound slope of Gale Crater where direct strong winds would develop. The erosional removal of a possibly large original fan suggests extended periods of elevated wind erosion strength, perhaps associated with major obliquity shifts. Hence, the construction of the Gediz Vallis fan and its destruction provides clues about Mars episodic and progressively changing climate, perhaps during the late Hesperian and early Amazonian.

CRedit authorship contribution statement

A.B. Bryk: Writing – review & editing, Writing – original draft, Visualization, Supervision, Methodology, Investigation, Formal analysis, Conceptualization. **W.E. Dietrich:** Writing – review & editing, Writing – original draft, Visualization, Supervision, Project administration, Methodology, Investigation, Formal analysis, Conceptualization. **K.A. Bennett:** Writing – review & editing, Methodology, Investigation, Formal analysis, Conceptualization. **V.K. Fox:** Writing – review & editing, Methodology, Investigation, Formal analysis, Data curation, Conceptualization. **C.M. Fedo:** Writing – review & editing, Methodology, Investigation, Formal analysis, Conceptualization. **A.R. Vasavada:** Writing – review & editing, Supervision, Resources, Project administration, Methodology, Investigation, Funding acquisition, Formal analysis, Conceptualization. **A.A. Fraeman:** Writing – review & editing, Supervision, Project administration, Methodology, Investigation, Formal analysis, Conceptualization. **M.P. Lamb:** Writing – review & editing, Investigation, Formal analysis, Conceptualization. **S. Le Mouélic:** Writing – review & editing, Visualization, Software, Methodology, Investigation. **G. Caravaca:** Visualization, Investigation. **J.M. Davis:** Writing – review & editing, Investigation. **J.A. Grant:** Writing – review & editing, Methodology, Investigation, Formal analysis, Conceptualization. **R.C. Wiens:** Writing – review & editing, Visualization, Methodology, Investigation, Data curation. **R.M.E. Williams:** Writing – review & editing, Validation, Investigation, Conceptualization. **O. Gasnault:** Visualization, Methodology, Investigation, Formal analysis. **K.M. Stack:** Methodology, Investigation, Conceptualization. **R.E. Arvidson:** Supervision, Methodology, Investigation, Formal analysis, Conceptualization. **L.A. Edgar:** Methodology, Investigation, Formal analysis, Conceptualization. **J.P. Grotzinger:** Supervision, Resources, Project administration, Methodology, Investigation, Conceptualization. **S.G. Banham:** Writing – review & editing, Methodology, Investigation. **S. Gupta:** Validation, Supervision, Investigation. **K.W. Lewis:** Validation, Methodology, Investigation, Formal analysis, Conceptualization. **M. Turner:** Methodology, Investigation. **W. Rapin:** Visualization, Methodology, Investigation. **M.N. Hughes:** Methodology, Investigation. **E.S. Kite:** Writing – review & editing, Conceptualization. **S.W. Purdy:** Writing – review & editing, Conceptualization. **R.Y. Sheppard:** Writing – review & editing, Investigation, Conceptualization. **D.M. Rubin:** Validation, Methodology, Investigation. **M.C. Malin:** Visualization, Validation, Supervision, Software, Resources, Project administration, Methodology, Investigation, Funding acquisition, Formal analysis, Data curation, Conceptualization.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.icarus.2026.117051>.

Data availability

Image data are available on NASA's Planetary Data System.

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