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Using quantitative topographic analysis to understand the role of water on transport and
deposition processes on crater walls

By

Marisa Christina Palucis

A dissertation submitted in partial satisfaction of the
requirements for the degree of
Doctor of Philosophy

in

Earth and Planetary Science

in the

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of the

University of California, Berkeley

Committee in charge:

Professor William E. Dietrich, Chair
Professor Michael Manga
Professor Ronald G. Amundson

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Marisa Christina Palucis

Abstract

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Professor William E. Dietrich, Chair

The amount of water runoff need to evolve landscapes is rarely assessed. Empirical studies correlate erosion rate to runoff or mean annual precipitation, but rarely is the full history of a landscape known such that it is possible to assess how much water was required to produce it. While this may not seem to be of primary importance on Earth where water is commonly plentiful, on Mars the amount of water to drive landscape evolution is a key question. Here we tackle this question through a series of five chapters, one devoted to field work at Meteor Crater, another to laboratory experiments about controlling processes, and then two chapters on analysis of landforms and implications of water runoff on Mars (associated with the Mars Science Laboratory mission to Gale Crater), and then we complete this effort with a consideration of how we can reliably assign relative timing between events resulting in small depositional features. What follows below is a summary of what is found in each chapter.

Meteor Crater, a 4.5 km² impact crater that formed ~50,000 years ago in northern Arizona, has prominent gully features on its steep walls that appear similar to some gullies found on Mars. At the crater bottom, there are over 30 meters of lake sediments from a lake that disappeared ~10,000 to 11,000 years ago, indicating the transition from the Pleistocene to the current, drier climate. A combination of fieldwork, cosmogenic dating, and topographic analysis of LiDAR data show that debris flows, not seepage erosion and fluvial processes as previously suggested in the literature, drove gully incision during their formation period of ~40,000 years before the onset of the Holocene. Runoff from bare bedrock source areas high on the crater wall cut into lower debris mantled slopes, where the runoff bulked up and transformed into debris flows that carried boulders down to ~5 to 8 degree slopes, leaving distinct boulder lined levees and lobate tongues of terminal debris deposits that crisscrossed on the lower slopes. We hypothesize that the fine material, likely generated in the impact, and deposited with the coarse debris on the lower portion of the crater wall, is key to this bulking up process as flows cut across the deposits. Fluvial processes following the debris flow gullies extended alluvial deposits to the crater floor and contributed to lake infilling. Cosmogenic dating confirms that most of the modification of the crater walls occurred before the early Holocene. To account for the 75 distinct deposits currently lying on the crater floor, debris flow frequency would be about 1 event every 17 years, assuming debris flow activity terminated ~10,000 years ago. Assuming a water-to-rock ratio of 0.2 at the time of transport, it would have taken ~100,000 m³ of water to transport the ~500,000 m³ of debris flow deposits on the crater floor. Given the 4.5 km² size of the crater, this extensive erosion would require less than 0.02 m of total runoff, or the equivalent of just 0.001 mm/year over a 40,000 year period. This insignificant amount of water was likely

packaged into short-lived storm or snow-melt events when debris flows were generated. Much more runoff did occur, as evidenced by the lake and fluvial deposits, as well as the likely cool, wet conditions of the late Pleistocene. This suggests only a small fraction of the total runoff is needed to do considerable geomorphic evolution, producing strongly gully-scarred crater walls. Currently, only minor fluvial modification of the gully networks occurs.

To test the hypothesis that fine-grained material is a necessary component behind sustained mobilization in granular debris flows, we performed a series of experiments in a 4 m diameter vertically rotating drum. Specifically, we assessed the role of fines and coarse grain size distribution on the rate of pore fluid pressure generation and its dissipation upon cessation of movement of the granular mass. We varied the amount of fines, from no fines to amounts found in debris flow deposits, and varied the coarse grain size distribution, from a single grain size (10 mm) to a range found in a natural flow deposit. The mixtures with fines contents close to those found in actual debris flows had elevated pore pressures, indicating they were almost fully liquefied, though the highest excess pore pressures were generated with the combination of fine material and a wide coarse grain size distribution. A boulder was also placed upon the fines-rich mixtures just after cessation of motion and it was observed that the pore fluid pressure rose instantly, bearing most of the grain's weight, and then the pressure slowly declined. When the same boulder was placed on the water-gravel flows, there was no change in fluid pressure, as the mass was supported entirely by grain-grain contacts. Our observations, combined with observations of others in the literature, inform the conceptual model presented herein where we hypothesize that the dilational separation of particles during debris flow shear leads to coarse particles (in this case particles >12 mm) slowly settling through the highly viscous non-Newtonian fluid. As the fines-rich mixture has a yield strength of ~5 Pa, we calculate that particles ~12 mm and smaller remain suspended. This combined effect requires the fluid to sustain the weight of the particles, leading to pore pressures equal to the entire weight of the solid and fluid mass.

As part of the MSL science team, we focused on understanding the geomorphology and its implications for the hydrologic history of the landing site for the *Curiosity* rover, which was located at the distal end of the Peace Vallis fan in Gale Crater. The Peace Vallis fan covers 80 km² and is fed by a 730 km² catchment. Valley incision into accumulated debris delivered sediment through a relatively low-density valley network to a mainstem channel to the fan. An estimated total fan volume of 0.9 km³ matches the calculated volume of removal due to valley incision (0.8 km³). The fan profile is weakly concave up with a mean slope of 1.5% for the lower portion. Runoff (discharge/watershed area) to produce the fan is estimated to be more than 600 m, perhaps as much as 6000 m, indicating a hydrologic cycle that likely lasted at least 1000's of years. Atmospheric precipitation, possibly snow, not groundwater seepage produced the runoff. Based on topographic data, Peace Vallis fan likely overlapped the rise onto which *Curiosity* landed, Bradbury Rise, and spilled into a topographic low to the east of the rise. This argues that the light-toned fractured terrain within this topographic low corresponds to the distal deposits of Peace Vallis fan and in such a setting lacustrine deposits were predicted (and later confirmed by observations from the rover).

Previous studies had suggested that there might have been large lakes in Gale. This history is important to understanding Mars' climate history and the volumes of water that may have passed through or covered the sediments encountered by *Curiosity*. Here we use improved

imagery, topography, and ground-based observations from the *Curiosity* rover to map deltas, fans and gullies, which leads to a proposed history of lakes in Gale. We report evidence for at least three distinct large lakes within Gale, all which occurred after the crater's central mound, Aeolis Mons, reached close to its current topographic form. From corresponding deltaic deposits off the southern rim of Gale crater and the southern flank of Aeolis Mons, we identified the highest lake level at -3280 meters, which would have had a mean depth of 700 meters. The larger context provided by craters near Gale suggests that this lake was derived from regionally sourced water from the south. The next lake level, which was established after a period of drying with subsequent lake level decline before rewetting and lake level rise, is defined by four deltaic features as well as the termination of gullies around the northern rim of Gale. This second lake corresponds to an elevation of -3980 m, has a mean depth of 300 meters, and was likely sourced from water more local to Gale. Lake levels then rose to the third lake level, at -3780 m, with a mean depth of 400 meters. Two deltaic deposits derived from sediment from the rim of Gale define this lake. This last lake declined sufficiently quickly to prevent incision into any of the deltaic forms around the crater, though hydrologic activity likely continued as evidenced by a time of fan building, including the Peace Vallis fan, around the crater. Quantification of the fan and delta volumes and their gully sources on the crater wall and rim, and on Aeolis Mons, suggest that these discrete deposits contain essentially all of the sediment mobilized from the upslope source areas. Importantly for the MSL mission, this suggests that most of the sediments the rover has and will encounter experienced at least a few cycles of drying and wetting.

The depositional features (i.e. alluvial fans and deltas) within Gale Crater are small but they likely record a time history of lakes and other depositional events within the crater. An obvious question to ask is: can crater counting, which has been used to date much of Mars' surface, be used on such small surfaces to provide accurate age dating? Here we introduce a simple simulation model to quantify the effects of sample area size and crater obliteration effects on age estimates derived from crater size-frequency distributions. Our results show, quantitatively, that crater counting to estimate age of small surfaces (<1000 km²) has unavoidable large uncertainties. This arises because the small craters, which are most numerous, are eliminated due to erosion and infilling and there is a low probability of counting large craters on small sample areas. These effects on small and large craters leads to a narrow range of crater sizes in which the correct age may be reflected in the crater density functions that are used to date surfaces. This range decreases with decreasing surface area and increasing erosion and infilling. Steps in the isochron data, in which crater density data are shifted downward for some crater sizes to a different density size relationship, are often cited as indicators of resurfacing events. We find, however, that such steps occurred randomly in $\sim 5\%$ of the crater size-frequency distributions we generated. This can lead to large errors when determining which tangential isochron is used to assign a surface age. Our modeling, which accounts for obliteration of craters and a reduced chance of encountering large craters on small and young surfaces, suggests that in general the least reliable ages occur between 1 and 3 Ga years. Younger surfaces preserve the smaller craters and older surfaces ones collect larger ones, both improving the probability of obtaining the correct age. For areas less than 1000 km² and true ages of 1 to 3 Ga, there is only a 20 to 40% probability of the calculated age lying within 0.25 Ga of the true value, where as for surfaces smaller than 500 km² (most mapped Mars fans are smaller) the corresponding probabilities drop to about 10 to 20%. To highlight the constraints on dating small deposits, we applied this model to cratering data from the Peace Vallis fan (~ 80 km²) and one of the larger deltaic deposits within Gale crater (117 km²), which deposited at different elevations and likely

record different hydrologic events. Even though portions of the observed crater count data in each case follow the proposed crater density isochrons, the uncertainties for such small areas are so large as to make the age assignment unreliable.

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Chapter 1: Introduction

When we look at the surface of our planet, we cannot see tectonic plates sliding past one another or the rebound of continents from the last ice age or acids chewing their way through rock, but we can see the landscapes that result from these processes. The research presented herein has focused on using landforms (e.g. channels, fans, deltas, and gullies) within regions that have experienced large changes in climate (i.e. the American Southwest and Mars) to 1) reconstruct the necessary hydrologic conditions required to create such features and 2) understand how those conditions evolved through time to produce the landscapes we see today.

The following work has three main contributions that are contained within five research chapters: (Chapter 2) field work and mapping at Meteor Crater (Arizona, USA) to understand the dominant processes and water volumes behind gully formation, (Chapter 3) large-scale experimentation of debris flow processes in a rotating drum aimed at assessing the role of fines in the generation and dissipation of excess pore pressures, (Chapters 4 and 5) remote sensing/GIS work within Gale Crater (Mars) to better understand the hydrogeomorphic history of the Mars Science Laboratory (MSL) landing site, and (Chapter 6) assessment of uncertainty in using crater counts for estimating surface ages of geomorphic features on Mars. For each chapter a separate introduction including a literature review is presented.

A central question that arises throughout these chapters is “how much water did it take to form the landscapes we see?” This is not an easy question to address and requires a deep mechanistic understanding of sediment entrainment, transport and deposition processes, often on steep slopes and under varying climate conditions. This debate has been central to Mars exploration, from the first mapping of channels (“canali”) on Mars by Giovanni Schiaparelli in the 1800s [Manara et al., 2014], to the observation of direct evidence of a streambed on Mars by the MSL *Curiosity* rover [Williams et al., 2013]. Currently, much of the water on Mars exists as ice, located either at the polar caps or in the shallow subsurface at more temperate latitudes [Mitrofanov et al., 2003; Bibring et al., 2004; Bandfield et al., 2007]. Transient liquid surface water may have occurred in the more recent geologic past, but only under certain conditions [Clow, 1987; Hecht, 2002; Kite et al., 2013]. But, in Mars’ past, much larger quantities of liquid water may have existed, possibly including a large northern ocean [e.g. Clifford and Parker, 2001; Perron et al., 2007; Di Achille and Hynek, 2010]. There is growing evidence to support that Mars was once a much wetter place, including mineralogical and meteorite data, but perhaps the most abundant dataset lies within the geomorphology. Both orbital data and surface investigations from landers and rovers have shown large outflow channels [e.g. Carr and Clow, 1981; Wilson et al., 2004; Leask et al., 2006; Warner et al., 2013], valley networks [e.g. Craddock and Howard, 2002; Hynek et al., 2010], meandering channels [e.g. Moore et al., 2003; Malin and Edgett, 2003], deltas [e.g. Moore et al., 2003; Mangold and Ansan, 2006; Dibiase et al., 2013], alluvial fans [e.g. Moore and Howard, 2005; Kraal et al., 2008; Morgan et al., 2014] and gullies [e.g. Malin and Edgett, 2000]. In light of this, for the first decade of the 21st century, NASA’s science theme for Mars was to “follow the water”, and for the second decade the theme is to “explore habitability and seek signs of life”. Recognizing that water created these features was just the first step. The challenge now is to estimate the water volumes and discharges that formed these landforms.

Some of the most prominent water-carved features on Mars are its large outflow channels (e.g. Kasei, Ares, Tiu, Simud, and Mangala Valles), which are hypothesized to have formed

either during catastrophic floods [e.g. Baker and Milton, 1974; Leask et al., 2006] or dewatering of evaporite deposits [Montgomery and Gillespie, 2005]. Discharge estimates are often very large, ranging from 10^6 to 10^8 m^3/s [Carr, 1979; Robinson and Tanaka, 1990; see review in Kleinhans, 2005], where flow velocities are calculated from Manning's equation or the Darcy-Weisbach equation [Wilson et al., 2004] and then multiplied by the valley width and depth. Presumably these high discharge estimates are due to large uncertainties in the channel geometry and grain sizes (for comparison, discharges of $\sim 10^6$ m^3/s are estimated to have produced the Channeled Scablands of eastern Washington during the catastrophic breakout floods of Pleistocene Lake Missoula, one of the largest known terrestrial floods [O'Conner and Baker, 1992]). Irwin et al. [2005] presented an empirical estimate for formative discharges in alluvial channels within valley networks that scales with channel width, but often valley widths are used [e.g. Fassett and Head, 2005], and valleys tend to be much broader and deeper than the channels that cut them. It is also difficult to discern from orbital imagery whether the channel banks are actually alluvial or bedrock [Howard et al., 2005], the latter of which does not apply to the Irwin et al. [2005] empiricism. High-resolution imaging has also revealed numerous deltaic and alluvial fan features on Mars [e.g. Moore and Howard, 2005; Irwin et al., 2005; Mangold and Anson, 2006; DiBiase et al., 2013; Morgan et al., 2013], many within crater basins. Delta fronts are used to identify potential paleo-lakes, for which average lake depths and water volumes can be calculated [e.g. Mangold and Anson, 2006; Dietrich et al., 2013; Le Deit et al., 2013]. Formation time scales are estimated by dividing the sediment volume of the delta or fan by an estimated sediment transport rate, but a detailed geomorphic analysis to determine the processes by which sediment was delivered (i.e. fluvial versus mass flows) is often ignored [e.g. Kleinhaus et al., 2010]. Moore et al. [2003] used channel width and meander wavelength of inverted channels on the Eberswalde delta to calculate formative discharges of ~ 700 m^3/s . Morgan et al. [2013] also used inverted channels on fans in Saheki crater to estimate formative discharges, though they used the Darcy-Weisbach equation and assumed a concentrated flow (i.e. high sediment to water ratios) based on their terrestrial analysis of fan deposits in the Atacama.

While there has been a lot of work to estimate flow velocities and peak discharges needed to form these features on Mars, there have been relatively few attempts to identify the total amount of water associated with building an individual landform. Gulick [2001] estimated the volume of sediment removed from valleys off of the side of Martian volcanoes and used two water-to-sediment ratios (3:1 and 1000:1) to estimate water volumes. However, the channel slopes, degree of weathering and cohesiveness of the surface material were relatively unconstrained due to low-resolution topographic data and imagery. Jerolmack et al. [2004] estimated the quantity of water necessary for alluvial fan construction near Holden Crater, but they used their estimates for water discharge (which are based on an unconstrained grain size) and fan formation times (which are based on a terrestrial intermittency factor [Parker et al., 1998]), and they acknowledged that the size of the fan was likely much larger in extent and only an erosional remnant remains. An approach that uses high-resolution topography to perform detailed mapping, process analysis, and sediment budgeting to determine water volumes is addressed in detail in Chapters 2, 4 and 5.

With the availability of high-resolution LiDAR data on Earth and the increasing coverage of HiRISE and CTX data on Mars, we now have the opportunity to use quantitative methods to assess the dominant flow process and hence the water fluxes and volumes to form features like gullies, fans, and deltas. Currently, tools used on Mars to assess water amounts are derived from

empirical studies on Earth, but generally field studies on Earth have not asked the question: “how much water does it take to make the landscape we see”. In Chapter 2, this question is addressed at Meteor Crater, a 1.2 km diameter impact crater in northern Arizona with gully features similar to some of those seen on Mars by Malin and Edgett [2000]. Unlike Mars, where only imagery and in some locations high-resolution topography is available, Meteor Crater affords the opportunity to do field-based measurements and cosmogenic dating in addition to quantitative analysis of its topography. This was extremely useful as we were able to correlate features we had mapped in the field, like debris flow levees and lobate snouts, with the topographic data. We were also able to obtain grain size data, something that is necessary for paleo-discharge analysis, but due to image resolution limitations on Mars, is often impossible to obtain without a landed mission. Perhaps most importantly, Meteor Crater provided us with a general conceptual model for how gully processes work within one crater. Here, runoff, not seepage erosion, is generated on the upper bedrock-dominated crater walls and as this runoff crosses onto unconsolidated sediment mantling the lower crater walls, entrainment occurs. The runoff rapidly bulks up and transforms into a debris flows, which travels downslope until it reaches lower unconfined slopes where deposition ensues. When runoff lacks sufficient duration or intensity to cause entrainment to mass flows, it may be able to transport finer sediment by fluvial processes. While this model likely does not explain how all gullies systems form and are subsequently modified, it highlights the dependence of process (i.e. debris flow versus fluvial) on slope and climate, which is an important consideration when considering gully processes on Mars.

Within the coarse-grained deposits on the lower crater walls of Meteor Crater we observed an abundance of fine-grained material, likely generated from the impact event, which we hypothesized was a necessary condition for debris flow generation and mobilization. Using a 4 m rotating drum located at the Richmond Field Station, we conducted a series of experiments on the role of fines and coarse grain-size distribution on pore fluid pressures generated within steady-state mass flows. We also monitored the rate of dissipation of these pore fluid pressures upon cessation of the drum and measured how they responded to an external force (i.e. a boulder that was placed atop the debris flow mass). Chapter 3 mostly focuses on the monitoring data after the drum had stopped, but Kaitna et al. [in prep] will be publishing the steady-state experimental results. The experimental results within the drum, combined with observations from the literature, inform the conceptual model we present on how pore fluid pressures in excess of pressures due to the weight of the fluid phase (i.e. water and fines) are generated and maintained. We hypothesize that the separation of particles during shearing of the debris flow leads to coarse particles slowly settling through the highly viscous fluid. The cumulative effect of grains settling close to terminal velocities requires the fluid to sustain the weight of these settling particles, leading to pore pressures, under the right conditions, equal to the entire weight of the solid and fluid mass. Post-doctoral work at Caltech will further explore the role of fines and coarse-grained material in generating debris flows from runoff, and the pore fluid pressure response, in a large steep flume.

Chapters 4 through 6 focus entirely on Mars, and were motivated in large part by the Mars Science Laboratory mission to Gale Crater, which we participated on as part of the science team. Upon landing in Gale Crater, at what is now called Bradbury Landing, we performed careful topographic analysis of the terrain local to the landing site. To the north was an 80 km² alluvial fan, the Peace Vallis fan, whose distal end appeared to flow into a topographic low ~1 km to the east of the landing site. The primary mission was to explore the 5 km high mound in

the center of Gale Crater, named Aeolis Mons, that has a distinct mineralogical transition from phyllosilicates to sulfates, hypothesized to be the result of planet-wide changes in climate and hydrology [Milliken et al., 2010]. But, our topographic analysis, combined with geomorphic mapping of the landing ellipse by the team, resulted in the rover first traversing east, to what has been named Yellowknife Bay. Here, as the geomorphology predicted, fine-grained sedimentary rocks, inferred to represent an ancient lake, were found to preserve evidence of a habitable environment during the post-Noachian period of Mars [Grotzinger et al., 2014]. The results of this analysis, which focuses on the origin and evolution of the Peace Vallis fan and its implications on climate and water volumes, were published in Palucis et al. [2014] and are contained within Chapter 4. We extended our work outside of the landing ellipse to the entirety of Gale Crater, which had been hypothesized to have hosted several large lakes in the past [Cabrol et al., 1999; Irwin et al., 2005; see review in Anderson and Bell, 2010; Le Deit et al., 2013]. This analysis is important to understanding Mars' climate history, and importantly for the MSL mission, the volumes of water that may have passed through or covered the sediments encountered by the *Curiosity* rover. Investigating fan and delta deposits across the rim and walls of Gale and on Aeolis Mons, we determined a sequence of rising and falling lake levels with at least three lake stands. The hydrogeomorphic history detailed within Chapter 5 differs from those proposed by others [e.g. Anderson and Bell, 2010; Le Deit et al., 2013] and suggests a shift from regionally- to locally-derived waters sourcing Gale Crater in the post-Noachian. As the timing of these hydrologic events is important for understanding Gale's regional climate history, we probed the question of whether or not the small fan and delta features could be dated with crater counting techniques to get a relative time history between the delta- and fan-building events. In Chapter 6 a simple simulation model was developed to quantify the effects of sample area size and crater obliteration effects on age estimates derived from crater size-frequency distributions. Generally, the probability of obtaining the correct date drops significantly for areas smaller than 10,000 km² and since most fluvial geomorphic features are smaller than this, their age dates from crater counting are unreliable. Two depositional features in Gale, one of which was the Peace Vallis fan and the second a large delta on the southern floor of Gale, which have been the subject of several papers, cannot be assigned a reliable date via the crater counting method.

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Chapter 2: The role of debris flows on the post-impact evolution of Meteor Crater (Arizona, USA)

2.1 Abstract

Both fluvial sediment transport and periodic debris flows account for sediment routing and landscape evolution on steep slopes. In many environments, both of these processes will occur over time depending on the duration or intensity of the storm or snowmelt event, the timing between these events, the availability of entrainable sediment, and surface material properties. Understanding these processes is necessary if we want to use fluvial and debris flow deposits in the field to predict the runoff, and more importantly, the amount and intensity of precipitation or snowmelt needed to create these deposits. This type of analysis can inform us about the paleo-climate and paleo-hydrology for a region, especially when other paleo-environmental records are sparse, and provides key quantitative information on the hydrology and climate of other planets, like Mars.

Meteor Crater, an impact crater that formed ~50,000 years ago in northern Arizona, has prominent gully features on its steep walls that appear similar to some Martian gullies. At the crater bottom, there are over 30 meters of lake sediments from a lake that disappeared ~10,000 to 11,000 years ago, indicating the transition from the Pleistocene to the current, drier climate. A combination of fieldwork, cosmogenic dating, and topographic analysis of LiDAR data show that debris flows, not seepage erosion and fluvial processes as previously suggested in the literature, drove gully incision during their formation period of ~40,000 years before the onset of the Holocene. Runoff from bare bedrock on the upper crater walls cut into debris mantling the lower walls, where the runoff bulked up and transformed into debris flows that carried boulders down to ~5 degree slopes. We hypothesize that the fine material, likely generated in the impact, and deposited with the coarse debris on the lower portion of the crater wall, is key to this bulking up process. Cosmogenic dating confirms that most of the modification of the crater walls occurred before the early Holocene. To account for the 75 distinct deposits currently lying on the crater floor, debris flow frequency would be about 1 event every 17 years, assuming debris flow activity terminated ~10,000 years ago. Assuming a water-to-rock ratio of 0.2 at the time of transport, it would have taken ~100,000 m³ of water to transport the ~500,000 m³ of debris flow deposits on the crater floor. This extensive erosion would require less than 0.02 m of total runoff, or the equivalent of just 0.001 mm/year over a 40,000 year period. Much more runoff did occur, as evidenced by the lake and fluvial deposits. This suggests only a small fraction of the total runoff is needed to do considerable geomorphic evolution, producing strongly gully-scarred crater walls. Currently, only minor fluvial modification of the gully networks occurs.

2.2 Introduction

Meteor Crater, a well-preserved impact crater located on the Colorado Plateau in north central Arizona (35°03'N, 111°02'E, Figure 2.1a,b), formed approximately 50,000 years ago [Sutton, 1985; Nishiizumi et al., 1991; Phillips et al., 1991] during the Pleistocene, when climatic conditions were generally cooler and wetter (compared to present-day) across the southwestern United States [e.g. Ballenger et al., 2011; Cole et al., 2013]. Since Meteor Crater's formation, a network of gullies has developed along its inner walls (Figure 2.2), which were first discussed in

detail by Kumar et al. [2010] and later by Palucis et al. [2011]. While gullies have gained considerable attention on Mars since first being identified by Malin and Edgett [2000], especially in regards to the role of water versus dry granular flows [e.g. Treiman, 2003; Marquez et al., 2005; Head et al., 2008; Lanza et al., 2010; Conway et al., 2011], they have not been as well studied here on Earth, with the notable exception of gullies located in cold-climate settings with temperatures closer to modern Mars, such as Iceland [Hartmann et al., 2003], Greenland [Costard et al., 2002], Lapland [Dixon et al., 2010] and arctic Canada [Lee et al., 2001]. One aspect that has not been acknowledged in these previous terrestrial gully studies, to the best of our knowledge, is the effect of climate change on the evolution of the gully networks. Meteor Crater however, as first pointed out by Kumar et al. [2010], provides us the opportunity to explore how a landscape with a known initial condition (i.e. a fresh impact crater) has evolved over a geologically short period of time, but under significant climate change. The frequency and intensity of the hydrologic events that can significantly alter a landscape are often difficult to ascertain based on limited records, therefore acquiring new sets of paleo-hydrologic records to better constrain past climatic conditions, and their resulting effect on geomorphic processes, can be useful for places like Meteor Crater, where pollen and speleothem records are scarce [Kring, 2007].

Observations at Meteor Crater made by Kumar et al. [2010] lead them to propose a time sequence of events for the formation and modification of the gully network at Meteor Crater. Immediately post-impact, crater wall collapse preferentially occurred along the concentric fracture network mapped by Kumar and Kring [2008], forming the larger gullies exposed on the crater's corners. A lake then formed at the bottom of the crater, based on 30 meters of lake sediments that cover allogenic breccias with no intervening alluvium [Shoemaker, 1960; Shoemaker and Kieffer, 1974]. Since the crater is not breached, Kumar et al. [2010] assume that the source of water for the lake was artesian flooding, groundwater springs, rainfall or some combination of the three. Surface runoff from the crater rim initially washed "crater wall soft sediments" into the lake and then later incised bedrock. As lake levels fluctuated, fluvial processes driven by groundwater seepage associated with radial fractures and tear faults continued to erode the crater walls and deposit sediment, mapped as Pleistocene alluvium by Shoemaker and Kieffer [1974] (Figures 2.3 and 2.4), which inter-fingers with the lake sediments. Later, perhaps during a second phase of enhanced precipitation, this Pleistocene alluvium was dissected, producing gullies lower on the crater wall. The material deposited from this post-Pleistocene gully-forming event is now being dissected and overlapped with modern playa deposits. Current day processes are limited to modification of the gullies by a few debris flow events, as evidenced by "rock channels with rock levees" on the lower slopes of the crater walls [Kumar et al., 2010]. No grain size data, channel slopes, channel geometry or other quantitative metrics that might support that fluvial processes are the dominant erosive and transporting agent are reported by Kumar et al. [2010].

While Kumar et al. [2010] provide valuable observations that contribute towards our understanding of the gully networks lining the inner walls of Meteor Crater, there are some remaining issues to resolve before concluding with certainty how these systems evolved. Our own independent field observations suggest that rather than fluvial processes, defined here as erosion and deposition of alluvium by water flows with low to modest sediment concentrations, being the dominant mechanism for gully incision and sediment transport, the gullies instead are

the result of repeated debris flows, as evidenced by matrix-supported levees lining the gully channels, that have no clear connection to groundwater seepage. The discovery of these levees lead to the acquisition of high-resolution topographic LIDAR data of the crater (~0.25 m per elevation pixel), detailed field mapping and surveying, and cosmogenic ^{10}Be dating of several deposits exiting one of the gully systems. Here we present data that strongly suggests that rather than fluvial erosion and transport, debris flow processes dominated gully incision up until the Holocene. Furthermore, we address the following questions: 1) Why did debris flows happen here and why did they cease, 2) How have debris flow and fluvial processes influenced each other, and 3) How much water is required to accomplish the observed erosion? The third question is motivated in part by the common effort on Mars to use geomorphic features to try and constrain climate history [e.g. Irwin et al., 2005; Parsons and Nimmo, 2010; Morgan et al., 2013; Palucis et al., 2014]. We conclude with a general conceptual model of gully formation processes that is, perhaps, more broadly applicable to steep escarpments that are partially mantled by rock debris and bordered by a relatively flat boundary.

2.3 Study Site

2.3.1 Geology

Meteor Crater is a bowl-shaped depression, about 180 m deep and 1.2 km in diameter (4.5 km² in area), and is encompassed by a rim of ejecta that rises 30 to 60 m above the surrounding plain [Shoemaker, 1960]. Independent dating using $^{10}\text{Be}/^{26}\text{Al}$ measurements [Nishiizumi et al., 1991], cosmogenic ^{36}Cl measurements [Phillips et al., 1991] and thermoluminescence dating of shock-metamorphosed dolomite and quartz [Sutton, 1985], places the age of the impact at about 50,000 years ago, corresponding to the Wisconsin interstadial, which was a relatively warm period [Jacobs, 1984]. Shoemaker [1960, 1987], Shoemaker and Kieffer [1974], Kring [2007], Kumar and Kring [2008], and Kumar et al. [2010] provide a detailed description of the bedrock, ejecta, structure, and surface deposits at Meteor Crater, which are only briefly summarized here. The rocks exposed in the crater range from the Coconino Sandstone of the Permian age to the Moenkopi formation of the Triassic age, which are units in the upper portion of the Grand Canyon Sequence [Grotzinger et al., 2007]. A detailed cross-section of Meteor Crater, taken from Kring [2007], is shown in Figure 2.3. The lowest exposed unit in the crater is the Coconino sandstone, composed of well-sorted quartz aeolian sands [McKee, 1945]. This is overlain by 3 m of Toroweap Formation, composed of sandstone and dolomite. The Kaibab Formation, an approximately 80 m thick unit, is composed of dolomite, dolomitic limestone, and thin calcareous sandstone horizons, and overlies the Toroweap. The Kaibab is exposed along the steep upper wall of the crater. Two beds of the Moenkopi Formation rest disconformably on the Kaibab. The bottom bed is the Wupatki Member, which is 2 to 6 m thick and composed of very-fine sandstone. Atop the Wupatki is the Moqui Member, which is 2 to 10 m thick and is composed of siltstone. The crater rim is underlain by a sequence of Quaternary debris and alluvium, resting on the previously described strata. Beneath the crater floor lies about 1.6 m of Holocene sediments resting on ~30 m of Pleistocene lake sediment that overlie 10.5 m of thoroughly mixed (by source) bedrock debris, which is most likely fallout debris. Beneath this mixed layer is brecciated bedrock, locally over 150 m thick. Figure 2.4 shows a geologic map of Meteor Crater as mapped by Shoemaker [1960]. The steepest and highest part of the crater wall consists of exposed bedrock (Kaibab

formation) with local patches of breccia that is bordered by material mapped as talus by Shoemaker and Kieffer [1974], downslope of which lies Pleistocene and Holocene alluvium, lake deposits, and Holocene playa beds.

2.3.2 *Ecosystem*

Arid conditions able to support grasslands currently dominate at Meteor Crater with only limited occurrences of rain and snow. Kumar et al. [2010] report mean annual rainfall of 200 mm and mean annual snowfall of 290 mm from the Winslow station, located ~30 km east of Meteor Crater (since the crater itself is not gauged), for the past 100 years. To the east of the crater, at lower elevations, there is a sagebrush ecosystem and to the west, at higher elevations, the grassland turns to woodland, dominated by juniper, pinyon and pine [Anderson et al., 2000; Kring, 2007]. Packrat middens have been observed in the crater wall and small burrowing animals are still present at the crater, as indicated by the burrow system on the crater floor [Kring, 2007].

2.3.3 *Paleo-climate*

Cole et al. [2013] collected sixty packrat middens across the southwestern US that ranged from >48,000 years BP to present, where well-preserved pine needles within the middens document the geographical response of three types of pinyon pines to climate over the last 50,000 years. Application of modern climate requirements for these species to survive suggests that from northwestern to central Arizona, during the full glacial Wisconsinan (23.4 to 14.7 ka), winter precipitation was at least 150% greater than current day (measured in mm). During the Bolling-Allerod interstadial (14.7 to 13 ka), rapid expansion of these species into central Arizona suggests warmer temperatures than most of the latest Wisconsinan and summer precipitation that was 120% greater than current day (measured in mm). Rapid warming at the start of the Holocene (11.7 to 9.0 ka) lead to the decline and northward retreat of these populations to their current day distributions [Cole et al., 2013]. Kring [2007] also found pollen deposited in the lake sediments at Meteor Crater, perhaps suggesting that woodlands might have been established near or at the impact site, though the diversity and concentration of pollen was low and could have travelled from long-distances. Cole et al. [2013] found two needles of *P. edulis* (a species of Pinyon Pine) at Wupatki National Park, located approximately 40 miles from Meteor Crater, which had radiocarbon ages of 13.1 to 13.2 ka and 13.3 to 13.4 ka; they interpret these ages to be when *P. edulis* expanded north of its full-glacial limits during the Bolling-Allerod intervals.

More specific paleo-climate inferences have been made for the Black Mesa basin in northeastern Arizona (see region in Figure 2.1) [Zhu et al., 1998; Zhu et al., 2010]. Direct dating, numerical modeling and noble gases analysis suggest that 14,000 to 17,000 years ago the recharge rates increased by three times relative to today. Zhu et al. [1998] propose the pulse of high recharge was due to a northward migration of the southern branch of the split jet stream and that water level fluctuations rose by as much as 60 m relative to current levels. Wagner et al. [2010] reported the results from an oxygen isotope record from a speleothem collected from the Cave of the Bells in southern Arizona (Figure 2.1). Using these data for a proxy for aridity in the southwest during the last glacial period, they show that periods of warmth in the North Atlantic Ocean, like interstadials and the Bolling-Allerod warming, correspond to drier conditions in the

southwest. They propose warming of the ocean diverts the westerly storm track northward and reduces moisture delivery to the southwest. Ballenger et al. [2011] performed an extensive review of paleoenvironmental change in the southwest from a variety of regional records, including vegetation, faunal, stable isotopes and geomorphological change, that are coincident with the Bolling-Allerod/Younger Dryas/Early Holocene transition. In general, they found no consensus on the magnitude or direction of climate change in the southwest, but they acknowledge that high-resolution records during this time are rare. Comparing the speleotherm ^{18}O values from the Cave of the Bells [Wagner et al., 2010] and the Fort Stanton Cave in southeastern New Mexico [Asmerom et al., 2010] to the Greenland ice cores shows the onset of increased winter precipitation and/or cooler temperatures around 12.9 ka, though packrat records indicate increasingly xeric conditions around this same time. The onset of Holocene warming as measured by the loss of pinon correlates with pollen data by Van Devender and Spaulding [1979], but more recent pollen data show a mixture of desert grasslands and pinon-woodlands at mid-elevations between 12.9 and 12.6 ka. Lake level chronologies from across the Midwest collectively show that lake stands during the Younger Dryas were lower than the last glacial maximum, the period of deglaciation right before the Bolling-Allerod interstadial, and the early Holocene [Waters, 1989; Ballenger et al., 2011]. Collectively it is difficult to create a consistent model of oscillations in precipitation and temperature during the past 50,000 years since Meteor Crater's formation, but a general pattern of wetter and cooler conditions seems to have existed throughout much of the late Pleistocene, perhaps with snow being a greater proportion of the total precipitation.

2.3.4 Groundwater at Meteor Crater

In 1978, Roddy [1978] reported that the water level in the Meteor Crater well, located 1050 m north of the point of impact, was 186 m below the ground surface (or 1500 m MSL), putting the water table about 60 m below the current crater floor. Pilon et al. [1991] used ground-penetrating radar to locate the water table, and found it to be ~65 m below the crater bottom (or 1440 m MSL), in good agreement with the well reading. Based on the immediate formation of lake deposits post impact and the presence of a small hill of sediment on the crater floor composed of Pleistocene lake sediments (i.e. Silica Hill), Shoemaker and Kieffer [1974] proposed that at the time of impact, the water table was 30-40 m higher (1536-1546 m MSL) than the current water level. Roddy [1978], however, pointed out that shock compression of the Coconino sandstone could have induced high pore pressures which forced water to flow upward into the crater, allowing for a lake to form without the crater floor intersecting the local groundwater table. The intensive fracturing associated with the meteorite impact may also have enhanced the permeability of the crater wall rocks [Kumar et al., 2010]. If such fracturing were to lead to water directed to the walls, then perhaps very local seepage could occur there. Given that the bedrock was already relatively permeable before impact [Grant and Schultz, 1993; Kumar et al., 2010] it is not clear if a perched water table within the fractured rock and above the regional water table would occur, or if it would drain through the crater walls. In either case, mollusk populations identified by Reger and Batchelder [1971] from coring samples indicate that a majority of the mollusk species are typically found in perennial waters, as opposed to fluctuating waters (i.e. bogs, swamps, drying muds), suggesting that the lake at Meteor Crater was not short-lived and required a sustained source of water.

2.4 Methods

2.4.1 Data Collection and Processing

On March 12, 2010, the National Center for Airborne Laser Mapping (NCALM) performed an airborne survey of Meteor Crater as part of a SEED grant for this work. The survey area was a rectangular polygon, roughly 30 km², which covered Meteor Crater and its ejecta blanket. In order to obtain high-resolution data of the gullies and their deposits, specifically of the meter-scale levee deposits, two point densities were used; 8 points/m² were collected for the crater walls and rim and 4 points/m² were collected for the surrounding area. To achieve this dual density point scheme and maximize our ability to map the crater walls, flight lines for two polygons were flown (Figure 2.5). Flight lines for the larger polygon were oriented east to west and those for the inner polygon (which covers the crater itself) were oriented north to south. The orthogonal lines over the crater walls and rim doubled the nominal point density and reduced the effects of shadowing by the crater walls. The survey was performed 600 m above the local terrain, at a mean flying speed of 65 m/s with a swath overlap of 50%. Because our main objective was to map the relatively small features on the crater wall, the laser Pulse Repetition Frequency (PRF) was set to 70 kHz to balance the point density and accuracy requirements. The scan angle was limited to ± 21 degrees and the scan frequency (i.e. the minor oscillation rate) was set to 40 Hz to ensure uniform across-track and along-track spacing. The beam divergence was set to a narrow divergence of 0.25 mrad. Three GPS reference stations were used during the survey, one was located at the Winslow-Lindbergh Regional Airport (WINS) and the other two were located within 10 km of the survey polygon. All three stations were logged at 1 Hz using ASHTECH Z-Extreme receivers with choke ring antennas mounted on 1.3 m fixed-height tripods.

All data processing, including generation of the final data product, was performed by NCALM and is described in brief below. Reference coordinates for all three NCALM stations were derived from observation sessions taken over the project duration (~4 hours) and then were submitted to the NOAA National Geodetic Survey's (NGS) on-line processor (OPUS) which processes static differential baselines tied to the international Continuously Operating Reference Station (CORS) network. All coordinates are relative to the NAD83 (CORS96) reference frame. Airplane trajectories for the flight survey was processed using Kinematic and Rapid Static (KARS) software, which is a GPS processing software that uses dual-frequency phase history files of the reference and airborne receivers to determine a fixed integer ionosphere-free differential solution. All three GPS reference stations were used to create individual differential solutions and then those solutions were differenced and compared for consistency. The standard deviation of the component differences (easting, northing, and height) between these individual solutions was typically between 5 to 25 mm horizontally and 15 to 55 mm vertically.

After GPS processing, the trajectory data and the raw inertial measurement unit data collected during the flight were processed using APPLANIX software, which uses a Kalman filter to produce a final, smoothed navigation solution that includes both aircraft position and orientation at 200 Hz. This final solution is referred to as the SBET (Smoothed Best Estimate Trajectory). The SBET and the raw laser range data were combined using Optech's DashMap processing program to generate the laser point dataset in LAS (LASer) format. Bore sight

calibration was done by surveying crossing flight-lines over near-by residential neighborhoods, as they provide ideal surfaces (exposed, solid, and sloped in different aspects) for automated calibration using TerraSolid's TerraMatch software. This software used least-square methods to find the best-fit values for roll, pitch, yaw and scanner mirror scale by analyzing the height differences between the computed laser surfaces off of rooftops and ground surfaces from individual crossing and/or overlapping flight lines. TerraSolid's TerraScan software was used to split the entire project area into 36 1-km² areas. No bare earth extraction was required for Meteor Crater since it is sparsely vegetated and the bare-earth classification algorithm would smooth out the small-scale topographic features of interest inside of the crater. The elevation and intensity data were interpolated at 25 cm cell sizes using Golden Software's Surfer 8 Kriging algorithm. The final project was projected in UTM zone 12N with units in meters, and heights are NAVD88 orthometric heights computed from GRS80 ellipsoid heights using the NGS GEOID09 model. The elevation accuracy is 5 to 30 cm (1 sigma). All data can be accessed through NSF's OpenTopography portal.

2.4.2 Volume Calculations

Much of the topographic analysis presented in Section 2.5 (i.e. calculation of the removed volume of material by gully incision and levee and lobate deposit volumes) relies on methods similar to those developed by Warner et al. [2011] for basin volumes. The volumes eroded from the gullies in the source regions were calculated by projecting a smooth surface between the bounding sidewalls on either flank of the gully wall, essentially creating a "lid" over the gully. This projected surface was then gridded and subtracted from the original gridded topography. To estimate levee and terminal deposits volumes we first eliminated the topographic data under the deposit (mapped in planform) and projected the local topography (typically the crater wall or floor) laterally under the deposit to create a pre-deposit surface. We then took a difference between the gridded deposit surface and this projected pre-deposit surface to estimate the volume of the deposit. To estimate the amount of alluvium deposited on the crater floor by debris flows, we used the cross-section of Shoemaker and Kieffer [1974] and fit second order polynomials to the boundaries defining the breccia, talus/alluvium/lake sediments and the modern day crater surface. Shell integration was then used to integrate these functions along the axis perpendicular to the axis of revolution (in this case we are revolving around the y-axis and integrating along the x-axis). As the 'volume' above the breccia had talus, alluvium and lake sediment, we estimated the percentage of each from the cross-section and then applied that to the total volume. This method inherently assumes the crater is a perfect circle and that deposition of each unit was uniform around the crater; as neither are the case, these are rough estimates of the volumes of each depositional unit. Our methodology mainly used ArcGIS tools, though much of the post-processing was conducted in MATLAB using TopoToolbox [Schwanghart and Kuhn, 2010].

2.4.3 Slope-Area Relationships

We calculated slope-area relationships for each individual gully, as Stock and Dietrich [2003, 2006] found that "valleys traversed by debris flows have a plot of slope against drainage area that is curved in log-log space when measured from sufficiently high-resolution contour maps or DEMs." This parameter space is also beginning to be explored for gullies on Mars to try and differentiate the difference between gullies formed by fluvial processes, debris flows, and

dry granular flows [Lanza et al., 2010; Conway et al., 2011]. In order to extract the necessary data, the Meteor Crater gridded DEM was imported into Matlab and the slope gradient at each pixel (0.25 m per pixel) was calculated using TopoToolbox, which utilizes a D8 algorithm (unlike ArcGIS, TopoToolbox calculates the gradient in eight possible directions for each cell). As TopoToolbox was designed for grid sizes less than 2000 columns x 2000 rows, the DEM data for each mapped gully was imported into Matlab separately. The slope maps were imported back into ArcGIS in order to calculate the corresponding contributing area for points along the gully floor. To calculate contributing areas with ArcGIS, the DEM is first “filled” to remove any sinks, which are defined as pixels with elevation values lower than all those around it, in the elevation raster. As Meteor Crater itself is a large sink, we chose a fill limit of up to 1 m deep (i.e. only sinks 1 m deep or less are filled). After all sinks are removed elevation raster, a raster of water flow direction from each pixel was generated (i.e. max fall), and from that, a raster with an accumulation value for each pixel. This value represents the total number of pixels from which water would flow into a given pixel (i.e. the contributing drainage area). Slope and the corresponding drainage area were extracted every meter along the gully floor axis. We only chose points along the axis of the valley floor to prevent including erroneous data points, which can easily occur when doing metadata analysis from DEMs. For example, the sidewalls of a terminal deposit of a debris flow located on the crater floor can be relatively steep and have a very small contributing area, but this not the result of an erosional process, rather it is due to the material properties of the flow that deposited it, and therefore should not be included in the analysis.

2.4.4 Field Surveying

In combination with the LiDAR data (contoured to 25 cm) we performed extensive mapping in the field with a differential GPS system which allowed us to confirm and extend much of what we had mapped remotely (i.e. lateral levee deposits and depositional lobes), as well as identify the lowest position of boulders on the crater floor, determine the chronology of cross-cutting relationships among deposits, and do deposit volume estimates in the field to compare to the volume estimates performed using topography alone. Pebble counts and local field surveys were also performed to determine the grain size distribution and cross-sectional profiles of depositional features within the crater. Following a monsoonal storm in early September of 2011, we mapped the occurrence of fluvial channel development on the floors of several gullies, including local fluvial channel cross-sections and slopes. We also conducted Wolman pebble counts [Wolman, 1954; Bunte and Abt, 2001] on bars that developed in these channels in order to determine the critical discharge for fluvial sediment motion and the source area runoff rate. To conduct these counts we set up a small grid on the bar and counted the size of the particle at each grid node, the particle being chosen using the tip of a pencil, and making sure to count at least 100 grains [Wolman, 1954]. The sample size (n=100) allows for the determination of the median grain size (D_{50}) within tolerable limits, though may be too low to compare particle size distributions from different sites [Bundt and Abt, 2001].

2.4.5 Cosmogenic ^{10}Be Exposure Dating

In an attempt to date the timing of debris flow levee emplacement at Meteor crater, we collected four samples from levee deposits on the northeastern side of the crater, labeled MC-11-

01 through MC-11-04, as shown in Figures 2.6 and 2.7. The samples were derived from sandy dolomite blocks, which are expected to be the most resistant to weathering and hence have the most continuous exposure ages. All samples, except one (MC-11-04), were collected from relatively flat horizontal surfaces on blocks derived from the Kaibab formation; they consist chiefly of dolomite with various amounts of detrital and secondary quartz. Maximum sample depth from the exposed surfaces was 1.5 cm. MC-11-04 was collected on a vertical face, on the same boulder as MC-11-03. The sampled deposits were located under a weathering-resistant knob rising above the oldest Pleistocene talus near the base of the beta member of the Kaibab, which was sampled by Nishiizumi et al. (1991) (M-01, Figure 2.7). We know that once the impact occurred, the rocks we sampled were likely exposed to cosmogenic radiation, but presumably under different shielding conditions. Based on an exposure age of the outcrop (M-01) from Nishiizumi et al. [1991] of ~29,000 years, we know that if our samples were exposed exactly at the surface of the wall or talus, the cosmogenic production would be significant. If the samples were buried approximately a meter from the surface of the wall, the production rate is only ~4%, and if buried more than 2 m it is negligible. When choosing our samples, we looked for rocks that appeared to be visually minimally eroded, allowing for a greater chance that they were not exposed at the surface for a long time before being entrained and deposited by a debris flow. At site MC-11-03/04 we tested for shielding effects by taking samples at both the top of the rock, which is the least shielded and would be expected to have the highest cosmogenic production, and at the side of the rock, which is shielded more of the time than the top, and hence should have a reduced production rate. By observing both a cosmogenic nuclide production that is lower than that of the knob (M-01) and this shielding effect, we have greater confidence that the rock we sampled was exposed near the time it was entrained and deposited within the levee and has not experienced movement since.

The samples we collected were first processed to obtain a pure quartz phase and then the Be and Al were chemically separated and purified (Kohl and Nishiizumi, unpublished data). ^{10}Be and ^{26}Al measurements were obtained using accelerator mass spectrometry (AMS) at the Purdue Rare Isotope Measurement Laboratory (PRIME). The results are shown in Table 2.1, which also lists the altitudes of the samples and the Al concentrations in the purified quartz. The ^{10}Be and ^{26}Al concentrations were calculated from the amount of Be carrier added (~1.5 mg), the Al content in the quartz, and the $^{10}\text{Be}/\text{Be}$ and $^{26}\text{Al}/\text{Al}$ ratios as measured by AMS. Exposure ages were calculated for these samples using production rates of ^{10}Be and ^{26}Al at sea level ($>50^\circ$ latitude) of 6.0 and 36.8 atom/year g SiO_2 , respectively (Nishiizumi et al., 1989) and correcting for sample altitude, latitude, and exposure geometry. These correction factors are based on cosmic-ray measurements, and their associated errors are expected to be less than 10% (Lal and Peters, 1967). In the present work, we use the geographic latitude to calculate exposure ages since the present geomagnetic latitude is not the same as it was during the last 50,000 years. We estimate an overall uncertainty of about $\pm 10\%$ in the production rates of these nuclides, calculated from granitic surfaces with the Sierra Nevada range; the principal uncertainty arises due to the lack of a precise date for the Tioga glacial retreat event [Nishiizumi et al., 1989]. Exposure ages have been calculated assuming no erosion, but we acknowledge that most surfaces experience some erosion with continuous exposure (though no soil development was observed on the rocks we sampled). By making a no erosion assumption, our ages are defined to be minimum exposure ages. These ages also assume no snow cover on the rocks sampled.

Correction for snow cover is negligible at present time, but could possibly have been important in the past. The error of the ages due to sample thickness is less than 2%.

2.5 Results

Here we present our field results in combination with our quantitative analysis of the LIDAR data. We first did an intensive investigation of the role of debris flow processes at one individual gully system, and using the field knowledge gained at this site, referred to as “Kuni’s Thumb”, we extrapolate to the rest of the crater through the use of our LIDAR data. We then document the effects of a single rainstorm event that occurred at the crater in September of 2011 to investigate modern day fluvial processes.

2.5.1 Kuni’s Thumb

In May of 2011 we spent several days investigating Kuni’s Thumb, located near the northeastern corner of the crater (at the stars in Figure 2.4). The site was chosen both because of its well-preserved deposits and its location under the prominent bedrock outcrop (that is shaped like a thumb) that was dated by Nishiizumi et al. [1991]. The gully system has a well preserved pair of levees leading to a distinct lobate deposit that defined a debris flow event which had crossed a pre-existing pair of levees that appeared visually older created by a debris flow that had run further out (see Figures 2.6b and 2.7). Boulders over 512 mm (i.e. with ϕ values ≤ -9) occur in the levees in both the younger and older flows (Figure 2.8 shows the grain size distribution for the younger flow). In both flows, the levee deposits are ~6 m apart and about 0.5 to 0.7 m high (Figure 2.9 is a cross-section from the younger flow, taken from A to A’ as shown in Figure 2.6b), and are detectable in the LIDAR data. The local slope for both flows was approximately 5 degrees (or 10%). Some of the boulders within the levees showed evidence of having a calcium carbonate coating on them before being entrained in the flow allowing us to trace the source of these coated boulders to a local knickpoint in the gully. Exposures in the walls of the levees (Figure 2.10) show them to be predominantly matrix-supported, though careful digging into the deposits revealed isolated regions of stratified sand and gravel. The matrix was predominantly sandy, though finer-grained black material, possibly ash, was observed. Before the shafts at the bottom of the crater were infilled, Shoemaker observed three volcanic ash layers ~5 m below the surface which he correlated to eruptions in the San Francisco Volcanic field near Flagstaff during the late Pleistocene [Kring, 2007]. At cross-section A-A’ (Figure 2.11) we observed that between the levee deposits, on the gully floor, was an approximately 4 m wide and 15 cm deep channel with a gravel bar ($D_{50} \sim 50$ mm, Figure 2.12). The “v-shaped” contours of this channel are detectable in the 25 cm data, and are mapped in Figure 2.6b (we note that in Figure 2.6, 0.5 m contours are shown since 0.25 m contours are difficult to differentiate at the scale of the image, but all mapping of the channels and levees were performed using 0.25 m contours). Incised into this ‘high flow’ channel was a smaller ‘low flow’ 0.4 m wide channel (Figure 2.11); a pebble count on a bar in the smaller channel gave a median grain size of 22 mm (Figure 2.12). The smaller channel appeared to be from a recent event (prior to May 2011) based on disturbances to the vegetation (i.e. flattening of the vegetation in the downslope direction).

Further up the gully, near the headwall, we observed stratified debris supported by a lower layer that was rich in fine-grained white material (Figure 2.13); a very prominent exposure

of these “white fines” was observed near the gully headwall (Figure 2.14). At the headwall of the gully there was no evidence of staining, coarse lag deposits, or evidence for sustained fluvial wash in the channel, nor did we see any cave or fractures from which water would have discharged. There was also a lack of evidence of rilling on the relatively smooth talus slopes on either side of the gully.

For the samples we collected on both the older and younger debris flow levees, whose locations are shown in Figures 2.4, 2.6b and 2.7, the exposure age obtained on the older flow, (sample MC-11-01) was found to be 13910 ± 670 years. The exposure ages taken from the younger flow (MC-11-02 and MC-11-03) were $18,600 \pm 1230$ and $10,050 \pm 380$ years respectively. We have plotted these ages relative to the ages for the late-Pleistocene to the late Holocene as well as other paleo-environmental indicators in the vicinity of Meteor Crater in Figure 2.15. The youngest sample dated, MC-11-03 (Figure 2.6b), was found in the younger flow levee and dates to the early Holocene (EH), which was a time when pinyon stands were starting to retreat north (perhaps an indication of warmer and drier conditions) [Cole et al., 2013], but Lake Cochise in southeastern Arizona was at a high-stand [Waters, 1989]. The second sample within the younger flow, MC-11-02, had the oldest exposure age (~ 19 ka), but as explained in Section 2.4.5, we do not know its full exposure history prior to emplacement. The stratigraphically older flow had an exposure age of 13.9 ka, which coincides with the Bolling-Allerod interstadial, and according to the paleo-environmental data, was a period of relatively wet conditions. While the dataset is small, and there are uncertainties involving the exposure histories of the samples and their degree of weathering, taken together the data suggests that debris flow activity at this site was active during the late Pleistocene, and ceased during or after the early Holocene.

2.5.2 Debris Flow and Fluvial Processes at Meteor Crater

Using our observations from Kuni’s Thumb, as well as several neighboring gullies, we were able to confidently correlate features on the ground (i.e. levees, small fluvial channels, and lobate debris flow deposits (Figure 2.6b)) with their topography on the LIDAR map. We characterized over 75 individual gully systems within Meteor Crater using a combination of field mapping and LIDAR analysis, which is summarized here. We also document systematic changes that occur with decreases in slope and the general trend of debris flows switching channels and their eventual transition to the fluvial domain.

The top of the gully heads are located at approximately the same elevation around the crater wall, $1656 \text{ m} \pm 14 \text{ m}$, which corresponds to the transition from the cliff-forming Kaibab limestone and the Coconino sandstones ($\sim 1660 \text{ m} \pm 5 \text{ m}$ contour). The bottom of the alcoves, defined to be where the alcove begins to narrow and form a single channel incised into slope mantling debris, typically occurs at an elevation of $1603 \text{ m} \pm 11 \text{ m}$. The individual alcoves have contributing areas ranging from 900 to $14,000 \text{ m}^2$ with an average of 2460 m^2 and are typically $103 \text{ m} \pm 37 \text{ m}$ long and $1.5 \text{ m} \pm 1.1 \text{ m}$ deep. An example of what constitutes the contributing area, length and depth for Kuni’s Thumb is shown in Figure 2.16.

The gully channels within the gully source area are “u-shaped” (Figure 2.6a) and have slopes of 26 ± 4 degrees (based on their long profiles) and tend to be slightly concave up (an

example is shown in Figure 2.17). The non-gullied talus, by comparison, has an average slope of 28 ± 3 degrees (based on 20 long-profiles), with profiles that are more linear in form (Figure 2.17). The gully channels typically maintain a fairly constant width and are on average 7 ± 2.4 m wide and 1.4 ± 0.6 m deep (based on 5 cross-sections per gully channel). They are commonly found to either terminate abruptly without fans or end in a lobate deposit of debris. Features suggesting a transition into standing water (i.e. deltaic deposits) have not been observed. Incised into the gully channels are smaller fluvial channels, as was observed at Kuni's Thumb. These channels tend to be straight not sinuous, and are sometimes observed to flow around or through mapped lobate debris deposits. Their average width is $3.6 \text{ m} \pm 1.2 \text{ m}$ and their average depth is $0.3 \text{ m} \pm 0.1 \text{ m}$.

Like the levee deposits found at Kuni's Thumb, most levees around Meteor Crater tend to be paired, but unpaired deposits were also mapped. One pair of levees is often found within a larger pair of levees, sometimes there are numerous boulder snouts within a single levee pair, and the levees tend to crosscut one another, all providing evidence that multiple events came down each canyon (Figure 2.18). Cross-sections of 100 levees show that they tend to be less than $0.6 \text{ m} \pm 0.2 \text{ m}$ high, $1.9 \text{ m} \pm 0.7 \text{ m}$ wide, and are typically rounded (Figure 2.19). Lobate debris flow boulder snouts located between paired levees range in depth from 0.4 m to 1.8 m, with a mean of $0.8 \text{ m} \pm 0.4 \text{ m}$. In some cases, mappable channels do not extend past these lobate deposits, but the levee tracks extend well beyond, sometimes onto the crater floor. These visibly older tracks are detectable at the resolution of the elevation model, but field examination shows these deposits to often consist of a single line of individual boulders (Figure 2.20), where boulders are defined as having a b-axis greater than 265 mm.

In addition to boulder levees, we also mapped the location of individual boulders/large cobbles (average grain size of $250 \text{ mm} \pm 50 \text{ mm}$) found along the floor of the crater. The large grains were found to go down to elevations as low as 1563 m (the crater floor is located at ~ 1561 m), which is approximately 7 m below the highest mapped lake sediments (i.e. Silica Hill) on the crater floor [Kring, 2007]. In Figure 2.4 we show the maximum extent of the paleo-lake (blue contour), the location of these lowest mapped boulders (orange dots) and the geology map of Shoemaker [1960]. The highest concentration of large grains is found on the eastern and southern side of the crater near the transition from Holocene alluvium to playa beds.

In Figure 2.21a,b we show a slope map generated from the LIDAR data overlain with our geomorphic mapping. Slopes steeper than about 32 degrees are either in bedrock or lining the channels cutting through the talus, where the talus itself lies almost entirely on slopes between 20 to 32 degrees. The debris flow levees crisscross slopes down to about 5 to 8 degrees, with localized boulder snouts sometimes found stopped in the tracks of previous leveed debris flows (as was the case for Kuni's Thumb). The boulder snouts typically deposit on slopes above 8 degrees and most are concentrated between 8 and 15 degrees. We note the apparent similarity in elevation of many of the debris flow boulder snouts at the lower end of the debris flow domain, which is ~ 5 to 10 m in elevation above the highest mapped lake level (blue line, Figure 2.21a,b). Extending past the lobate snout and levee deposits onto slopes less than 8 degrees are fluvial channels, some of which extend all the way to the center of the crater (slopes < 2 degrees). Our lowest mapped boulders are found on slopes of 2 to 5 degrees.

From our mapping we estimate the total volume of material mobilized from each individual gully by debris flows and compare it our estimate of material that has been deposited on the crater floor. From our geomorphic slope maps (Figure 2.21a,b) and a plot of slope-area for the gullies, where the curved area-slope form loses its signature (Figure 2.22), we assume that material deposited below 8 degrees was predominately emplaced via fluvial processes and therefore do not include it in our estimate. We calculate that $3.9 \times 10^5 \text{ m}^3$ of material was removed from the gullies compared to $5.0 \times 10^5 \text{ m}^3$ stored on the crater floor that is not considered talus or fluvially deposited Pleistocene alluvium. The fact that there is about 1.5 times more material on the floor than can be accounted for by gully incision could be from infilling from continued bedrock erosion since debris flow activity ceased, but is more likely due to uncertainties in estimating the volume of debris flow deposited alluvium on the crater floor (See Section 2.4.2). However, this approximate mass balance (we are assuming density changes are negligible from source to sink area) suggests that gully formation was likely the result of debris flow incision, not fluvial incision. To get an estimate of debris flow frequency at Meteor Crater we took the average volume of material comprising a single event, based on the exposed deposits, and divided that into the amount of debris flow deposited alluvium on the crater floor. We assume based on the dating from Kuni's Thumb that debris flow activity was active from the time of impact to the early-Noachian, which is $\sim 40,000$ years. The volume of material comprising an individual flow event was calculating by identifying clear relationships between levee and lobate snout deposits. For example, at Kuni's Thumb the prominent lobate deposit corresponds to well-defined levees that were deposited during the same event (volume $\sim 283 \text{ m}^3$). We were able to identify 20 clear flow events, and from these found flow volumes to range from 10 to 1000 m^3 with an average of $200 \pm 103 \text{ m}^3$. We find that debris flow frequency at Meteor Crater was about 0.06 events / year, or 1 event every 17 years. Currently, over the past 100 years at the crater there have been no observed debris flow events, except for a small mudslide near the man-made trail on the northwestern rim in 1906 [Kring, 1997].

2.5.3 Current-day Fluvial Processes at Meteor Crater

In mid-September 2011, a moderate-sized storm occurred in the region near Meteor Crater. The exact amount of precipitation that fell at the crater is not known, as it is un-gauged, but precipitation data was reported for the nearby Winslow Airport, which is ~ 30 km from Meteor Crater. In Winslow, a maximum of 22 mm fell over the course of one day and 85 mm in total fell over 10 days (Figure 2.23). Figure 2.24 shows data from the Winslow airport showing the daily average rainfall over the past ~ 100 years as well as the daily record maximums, which have been as high as 100 mm. Post-storm (~ 1 week) we observed small channels that had incised into gully floors, vegetation that had been flattened in the downslope direction, and small gravel patches that appeared to be recently mobile (Figure 2.25). Rilling and sheetwash was evident all across the inner walls of Meteor Crater, as well as on the outer crater rim. We surveyed one particular gully that had formed a fairly continuous fluvial channel from its headwall to the crater floor. The upslope drainage area at the head of the gully (where the gully encounters exposed bedrock) is 3867 m^2 and the main gully channel is on average $8 \text{ m} \pm 2 \text{ m}$ wide and is approximately 200 m long. At each of the points marked in Figure 2.26, channel dimensions, the grain size distribution of gravel bar deposits, and local slope (both measured in the field and using the LiDAR data) were measured. The upslope drainage area from each of these sites was calculated using LiDAR data. These data are shown in Table 2.2. There was no

clear pattern in changes to channel width or depth going downslope, but it was clear that the fluvial channel was greatly affected by the vegetation and micro-scale topography on the gully floor as it often flowed around vegetated mounds and larger rocks and occasionally would become braided for short distances. At each sample location, there was no evidence for any rilling within the finer-grained material comprising the levee walls, suggesting that the source of water for forming the channels, like the gullies themselves, was runoff from the exposed bedrock slopes above the gully (i.e. Horton overland flow).

2.6 Discussion

2.6.1 Controls on debris flow and fluvial processes at Meteor Crater

It is clear that at Meteor Crater both debris flow and fluvial processes have played a role in the degradation of the crater walls. In both cases, the process appears to be driven by overland flow where water is concentrated on the exposed bedrock walls and delivered to the talus slopes below. Kumar et al. [2010] suggest that exposed fractures and faults, and for one gully system a cave, delivered groundwater to the crater wall. In Figure 2.27, which is the geologic map of Shoemaker [1960] with our gully entrainment areas and his mapping of the fault system exposed around the crater rim highlighted, there is not a one-to-one correspondence between the two (many gullies do not have a fault or fracture above them, as was the case for Kuni's Thumb). We also explored the single cave in the field and there was no staining or lag as would be expected with sustained discharge. Groundwater discharge is also problematic in that the fractures/faults/cave pointed out by Kumar et al. [2010] are located on the bedrock headwalls, which are approximately 95 m above the current day floor of the crater. Even at the paleo-lake's highest stand of ~1570 m (based on the elevation of Silica Hill), this would correspond to ~86 m above the lake. Either the groundwater table was much higher (in which case one might expect to see evidence of seepage throughout the talus as fractures occur around the entire base of the crater, Figure 2.3) or the groundwater table was perched (though based on the fractures mapped by Kumar and Kring [2008], Kumar et al. [2010] suggest enhanced permeability of the basement rock around the crater). Water infiltrating from the surface (so not groundwater) could have also flowed through fractures, and when reaching an impermeable layer in the Kaibab limestone (the dolomite sections are not as permeable as the sandstone strata [Kumar et al., 2010]), flowed along this boundary until being discharged into the crater. While some minor contributions due to seepage may have occurred from surface infiltration, evidence points primarily to surface runoff as the source of water, not groundwater seepage.

Once water is delivered to the talus slopes, infiltration must be minimized such that sediment mobilization can occur. As shown in Figure 2.13, near the gully headwalls, a lower layer of finer-grained white material supports many of the larger clasts. Field studies on debris flows often report what has been called a "fire hose effect" [e.g. Johnson and Rodine, 1984; Berti et al., 1999; Berti and Simoni, 2005; Coe et al., 2008; Gregorette and Fontana, 2008], where "high-energy dilute flows in very steep bedrock channels debouch onto talus slopes and entrain enough loose sediment to achieve debris-flow concentrations [Meyer and Wells, 1997]", and fine-grained sediment is often cited as necessary for producing these flow conditions [e.g. Rodine and Johnson 1976; Major and Pierson 1992; Meyer and Wells, 1997; Griffiths et al., 2004]. The fines (i.e. clay and silt sized material) mix with water to form the highly viscous pore

fluid that mediates inter-granular collisions and prevents separation of the fluid and solid components of the flow [Iverson, 1997; Kaitna et al., *in prep*]; at Meteor Crater the fines near the gully headwall, perhaps generated during impact, are likely necessary for causing entrainment into a mass flow. Once generated, debris flows tend to surge and stall as they flow downslope, often continuing to entrain and gain in size until reaching lower unconfined slopes where deposition begins. In flows with poorly sorted sediment, the coarse fraction will migrate to the sides and front of the flow, leading to levee formation where low fluid pressure and high inertial stress prevails, and a snout which may be pushed downslope by the flow [e.g. Iverson, 1997; Pierson, 1981; Hsu, 2010; Johnson et al., 2012].

Debris flows are self-confining, meaning they create their own lateral margins (i.e. levees), which has important controls on run-out distance. Depending on the flow rheology and sediment supply, debris flows are able to flow over relatively gentle slopes as flow confinement prevents the flow from spreading, thinning and depositing [Whipple and Dunne, 1992]. At Meteor Crater there is clear evidence that multiple events came down the same canyon, evidenced by numerous boulder snouts within the same levee pairs, often flowing out onto the crater floor. These flows were at times observed to overrun, or cross, previous flow paths and create new flow paths for future flows. In addition, flows from neighboring canyons are observed to cross one another. These flow events therefore create shallow, often sinuous, crossing channels that can be distinguished from the deeper main gully channels located within the gully head (Figure 2.18). The boulders we mapped on the crater floor may have been transported there by debris flows, and aeolian erosion has since removed the finer-grained matrix. In general though, for slopes less than 8 to 10 degrees, we observe the handoff to fluvial processes. While fluvial channels are observed to occur on the floor of gullies steeper than 10 degrees, the highest concentration of channels occurs below 8 degrees, where the lobate snout deposits and levees start to disappear and finer grained sediment is observed (we note that at Kuni's Thumb the larger fluvial channel was observed in the LIDAR dataset, but not the smaller channel (~0.4 m wide), so our mapping is not capturing the finest-scale fluvial channels).

Kumar et al. [2010], following Shoemaker and Kieffer [1974] and Grant [1999], suggest that post-impact, fluvial processes from groundwater seepage eroded the crater wall and deposited the material as Pleistocene talus, and that further erosion of the talus was the result of a later period of enhanced erosion (perhaps via rainfall?). They point to their mapping of a branched network of channels (which is actually a map of both debris flow and fluvial channels where the branching pattern mid-talus arises from crossing debris flow paths) that originate on the talus slopes as evidence for this second erosion event. We propose that post-impact, after breccia and talus emplacement, runoff from bare bedrock entrained sediment and incised the gullies predominately via debris flow processes (which is supported by our rough mass balance) and that the branched network of channels observed by Kumar et al. [2010] on the talus slope is due to debris flows created channels that cross and switch canyons, not a second period of gully incision. When water inputs are not sufficient to generate a debris flow, as was observed during the September 2011 event, fluvial processes, also driven by overland flow processes, lead to minor modification of the gully floor and the transport of finer sediment. Dating suggests that most of the modification of the walls of Meteor Crater occurred before the early Holocene; presently, only fluvial modification of the crater and gully floors occurs.

2.6.2 How much water?

Volume estimates suggest that debris flows activity in the past may have been as often as several events every hundred years, which was likely driven by differing climate conditions. Here we take our estimate of the volume of sediment within the terminal snout and levees at Kuni's Thumb and then assume a rock to water ratio based on experimental work [e.g. Major, 1996; Iverson, 1997; Kaitna et al., 2014; Kaitna et al., *in prep*;] and field estimates [e.g Pierson, 1980; Iverson, 1997; McArdell et al., 2007] to determine the minimum amount of water needed to transport the sediment (we are assuming no losses to infiltration and evaporation). At Kuni's Thumb, the volume of sediment comprising the snout and lateral levees was approximately 283 m³; assuming a water-to-rock ratio of 0.2 at the time of transport [Iverson, 1997], a minimum of 56 m³ of water was needed. The bedrock source area above the levees is approximately 1700 m², hence a minimum of 33 mm of runoff from the upslope bedrock was needed to generate the flow. If we use this water-to-rock ratio of 0.2 and apply it to the ~500,000 m³ of debris flow deposits on the crater floor, then it would have taken a total of ~100,000 m³ of water to transport. This extensive amount of erosion would require less than 0.02 m of total runoff, or the equivalent of just 0.001 mm/year (for a 4.5 km² crater) over a 40,000 year period. While a minimum, it suggests that large volumes of water are not required to form these flows, though the intensity of the flow is likely to be important. Cannon et al. [2001] found that debris flows in a recently burned basin in New Mexico were triggered in response to a maximum 30 min rainfall intensity of 31 mm/hr, while McCoy et al. [2010] found that a short-duration (<3 h) rainstorm with low- to moderate-intensity rainfall (<10 mm/h) triggered debris flows in Chalk Cliffs (Colorado).

While we do not have data on the storm intensities that triggered debris flow events at Meteor Crater, we can estimate the formative discharge for the September 2011 storm event, which provides us with information on a runoff rate that we know did not produce a mass wasting event but which lead to fluvial incision. For each of the sites within Gully 52 (Figure 2.26), the following analysis was performed to estimate formative discharges, which was converted to runoff rate (in mm/hr) by dividing by the upslope drainage area. For channels with width-to-depth ratios less than 20, the hydraulic radius, R , which is the ratio of the channel area to its wetted perimeter, is used instead of h . Our approach uses a conventional sediment transport relation, the Law of the Wall, where for a turbulent velocity profile for a hydraulically rough flow (i.e. Reynold's number, Re_{channel} , is > 1000 (Chow, 1954)), the average velocity, u , is

$$\bar{u} = \frac{u_*}{k} \ln \left(\frac{11h}{3.5D_{84}} \right), \quad (2.1)$$

where k is the dimensionless von Karman's constant (typically 0.41 (e.g. Furbish, 1997; Garcia, 2007)), h is the flow depth, and the shear velocity, u_* is defined as

$$u_* = \left(\tau_{*cr} \frac{(\rho_s - \rho)}{\rho} g D_{50} \right)^{0.5} \quad (2.2)$$

The Lamb et al. (2008) slope-dependent formulation for the critical shear stress, τ_{*cr}^* , was used

$$\tau_{*cr} = 0.15S^{0.25} \quad (2.3)$$

where S is equal to $\tan b$, where b is the channel bed slope angle from horizontal. The instantaneous discharge, Q , is defined as:

$$Q = uhw \quad (2.4)$$

where u is calculated from Equation 2.1 and w is the channel width.

Discharge, Q , was estimated using Equation 2.4, and the runoff input (in mm/hr) was calculated by dividing discharge by the upslope contributing area. Calculated rates of runoff are shown in Table 2.3 and range from 1.2 to 11.6 mm/hr. The actual precipitation rates may have been higher due to losses to evaporation and infiltration, which are not accounted for in this analysis.

While paleo-environmental data from the late Pleistocene and early Holocene in the region cannot constrain precipitation rates, evidence points to a cooler, wetter period, with precipitation amounts, perhaps, being 120% to 150% greater than current day [Cole et al., 2013]. We know that at the crater a lake was present long enough to deposit 30 m of sediments, and based on mollusk data, was likely a perennial lake. Based on our debris flow frequency estimates and run-off rates from the September 2011 storm, we can only say that storms larger (i.e. more intense or longer duration) than this event were much more frequent. Of course, the slow winnowing of fines from the crater walls onto the crater floor (mapped as Holocene alluvium, Figure 2.4), or widening of the gullies over time as discussed below, may also have lead to more unfavorable conditions for debris flow development.

2.6.3 A conceptual model for gully formation

Our topographic analysis of Meteor Crater suggests a general morphodynamic model for gully incision and deposition processes. This conceptual model has been implied by other studies on gullies, notably by Hartman et al. [2003], but has not yet been made explicit. Generally, three primary process zones can be identified: a runoff generation zone (typically off of bedrock), a sediment entrainment zone (bedrock slope mantled with colluvium, impact breccia (in the case of a crater) and/or eolian deposits), and a depositional zone (Figure 2.28a,b). Empirical studies elsewhere, as well as at Meteor Crater, suggest that each of these zones can be distinguished by characteristic slopes (Figure 2.29), in which the water source region tends to be steeper than 32 degrees, the entrainment zone extends from ~32 to about ~20 degrees and the deposition zone progressively declines from debris flow dominated (down to about ~8 to 5 degrees) and, if present, to fluvial and flat lake sediments. Whether the depositional slopes are self-formed or are pre-determined and force process change, slope appears to be linked to process (i.e. fluvial versus debris flow).

When runoff (either from snowmelt or rainfall) is generated over bedrock it crosses onto a lower, sediment mantled zone and entrainment occurs. During sufficiently high-intensity storms (or rapid snowmelt events), we hypothesize that the runoff generates boundary shear stresses that entrain material, such as the gravel and boulders lining the bed. Bed entrainment and

lateral boundary shear stress destabilizes the gully walls causing a mixture of colluvium (i.e. a mixture of boulders, gravel, sand and mud) to rapidly charge the flow, leading to a debris flow surge. This sediment-rich mass may continue to entrain and increase in size until it reaches lower unconfined slopes where deposition ensues.

Runoff into the channel lacking sufficient duration or intensities to cause entrainment to mass flows may be able to transport sediment by fluvial processes. Gully walls and/or the levees left by the debris flows will confine these flows. Flow of sufficient duration may spill past the debris flow deposits and build a lower gradient alluvial fan downslope. Hence, both debris flow and fluvial processes may occur in gullies depending on the magnitude of runoff. Gully development may also be non-linear, in that initial incision will focus subsequent runoff and mass failure from the upslope sediment source area. The growing walls driven by gully incision would lead to increased sediment flux to the channel, which could serve to enhance entrainment and incision. Over time, given the slowly rising base level of the crater, as the gullies widen and slopes decline, runoff events may be unable to cause sufficient bed and bank erosion (due to reduced shear stresses), thereby stabilizing the channel width and forcing the mode of transport to be fluvially dominated.

2.6.4 Meteor Crater as a Mars analog?

Being an impact crater with well-developed gully features naturally leads to questions about the applicability of Meteor Crater as an analog to understanding gullying on Mars. In some respects, Meteor Crater is very different than Martian gullied craters in that it is much smaller (the gullies identified by Malin and Edgett [2000] had gully heads, often referred to as ‘alcoves’ in the Martian literature, that were several hundreds of meters wide as opposed to tens of meters at Meteor Crater), is not within a basaltic lithology, erosion is driven by rainfall (possibly rapid snowmelt as well), and the “pasted-on” terrain described for Martian gullies, which is thought to have been deposited from the atmosphere and to contain a large fraction of fine sediment and possibly appreciable ice [e.g. Mustard et al., 2001; Christensen, 2003] is absent.

The morphology of the Meteor Crater gullies to those described by Malin and Edgett [2000] on Mars, however, is remarkably similar in many ways. They observed Martian gullies to have theater-shaped heads, often with colluvium collected at the base of these escarpments, and commonly littered with large boulders (several meters to several decameters in diameter). Within the gully head is typically one main channel and at the downslope apex of the gully head the main channel is often associated with adjacent ‘lineations’, suggested to be secondary channels or alternative courses, now abandoned. The main and secondary channels display characteristics common for channels formed via fluid flow, such as sinuous paths, levees, streamlining, super-elevated banking, and incision. At the downslope end are often fan features containing “distally diverging medial to marginal lineations (suggesting flow lines) and channels, lobate margins, and distal thinning [Malin and Edgett, 2000]”. Malin and Edgett [2000] suggest that based on their morphology, Martian gullies, like those at Meteor Crater, are likely the result of numerous processes (i.e. fluvial, debris flow, and other mass wasting processes). They also remark on the fact that the gully heads are often associated with distinct layers within a cliff, which they suggest points to a combination of seepage and surface runoff processes. Since these initial observations, researchers have looked at a wide-variety of gullies across Mars [Dickson et al.,

2007; Dickson and Head, 2009; Goldspiel and Squyres, 2011; Harrison et al., 2014] and several have attempted to determine water volumes and flow characteristics associated with their formation [e.g. Parsons and Nimmo, 2010; Levy et al., 2010; Mangold et al., 2010]. Fluvial models are commonly assumed [e.g. Parsons and Nimmo] resulting in large water volumes that are difficult to reconcile with more recent Martian climate. The work presented here suggests that the dominant flow process changes both with environmental conditions and slope, which aids in our attempts to model gully processes on Mars. Current-day gullying processes on Mars may be limited to only the modification of existing gullies [e.g. Johnsson et al., 2014], but Mars' past climate was much wetter and perhaps more conducive to the initial development of gullies on crater walls.

2.7 Conclusions

Since its formation ~50,000 years ago, Meteor Crater has undergone significant modification as approximately 75 gully systems have incised into its inner walls during the first ~34,000 years of pre-Holocene wetter conditions. Fieldwork and cosmogenic dating on an individual gully system combined with a detailed topographic analysis of LiDAR data suggests an alternative explanation to the previously proposed seepage erosion and fluvial incision and deposition model for the Meteor Crater gullies. We propose rather that debris flow processes drove gully incision, as evidenced by the presence of poorly-sorted, matrix-supported levee deposits and coarse-grained lobate deposits with steep lateral margins. These flows were generated predominately by runoff from bare bedrock, not groundwater seepage, in a manner often described as a 'firehose effect', where entrainment occurs rapidly. When water inputs are not sufficient to generate a debris flow, as was observed during a storm event in September of 2011, fluvial processes driven by overland flow processes lead to minor modification of the gully floor. Cosmogenic dating suggests that most of the modification of the walls of Meteor Crater occurred before the early Holocene, which coincides with a lake on the crater floor and other paleo-environmental factors from the southwestern United States suggesting wetter and cooler conditions. On average, debris flow events at Meteor Crater consisted of 200 m³ of material. To account for the alluvium currently deposited on the crater floor, debris flow frequency was approximately 1 event every 17 years, assuming debris flow activity terminated ~10,000 years ago. Meteor crater provides us with a general conceptual model for how gully processes work, and shows the dependence of process on slope and climate, both important considerations for erosional processes on steep escarpments elsewhere on Earth and Mars.

2.8 Acknowledgements

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2.9 Tables

Name	Elevation (m)	Mass (g)	$^{10}\text{Be}/\text{Be}$ (E-15)	$^{26}\text{Al}/\text{Al}$ (E-15)	Be-10 (atom/g)	Al-26 (atom/g)	Exposure Age (1000 yr)
M-1	1680	30.261	114.8	490	0.425 E+06	.275 E+07	28.85 ± 1.47
MC-11-01	1564	49.925	720.8	0	0.189 E+06	0	13.91 ± 0.67
MC-11-02	1580	50.306	983.5	0	0.257 E+06	0	18.66 ± 1.23
MC-11-03	1600	50.368	539.1	0	0.140 E+06	0	10.05 ± 0.38
MC-11-04	1600	50.232	397.0	0	0.104 E+06	0	7.42 ± 0.31

Table 2.1. Results from ^{10}Be cosmogenic dating of the levee deposits at Kuni's Thumb Site.

Gully 52 Site #	Elevation (m)	Drainage Area (m^2)	Width (m)	Depth (m)	Slope	D_{50} (m)
1	1614	6475	0.96	0.007	0.4	0.0062
3	1611	7575	0.4	0.005	0.48	0.0031
4	1604	8450	1.2	0.008	0.3	0.0042
5	1601	17225	0.55	0.008	0.35	0.0045
6	1600	17800	0.95	0.01	0.22	0.0045
7	1599	18775	1.6	0.008	0.32	0.0037
8	1594	19400	1.65	0.006	0.22	0.0052
9	1586	24650	1.2	0.005	0.17	0.0051
10	1580	28675	0.75	0.009	0.1	0.0048
11	1576.5	31900	2.1	0.004	0.1	0.002
12	1576	35775	0.5	0.009	0.1	0.002

Table 2.2. Channel data collected at each site with Gully 52 (Figure 2.19) post-September 2011 storm.

Site #	LOW u_{bar} (m/s)	LOW Q (m ³ /s)	LOW runoff (mm/hr)
1	0.2	0.001	19
3	0.2	0.0004	13
4	0.3	0.002	22
5	0.3	0.001	13
6	0.3	0.002	16
7	0.3	0.003	18
8	0.2	0.002	14
9	0.2	0.0009	12
10	0.3	0.002	13
11	0.2	0.001	11
12	0.3	0.001	12

Table 2.3. Results of hydraulic analysis to calculate runoff rates for fluvial channels in Gully 52, post-September 2011 storm.

2.9 Figures

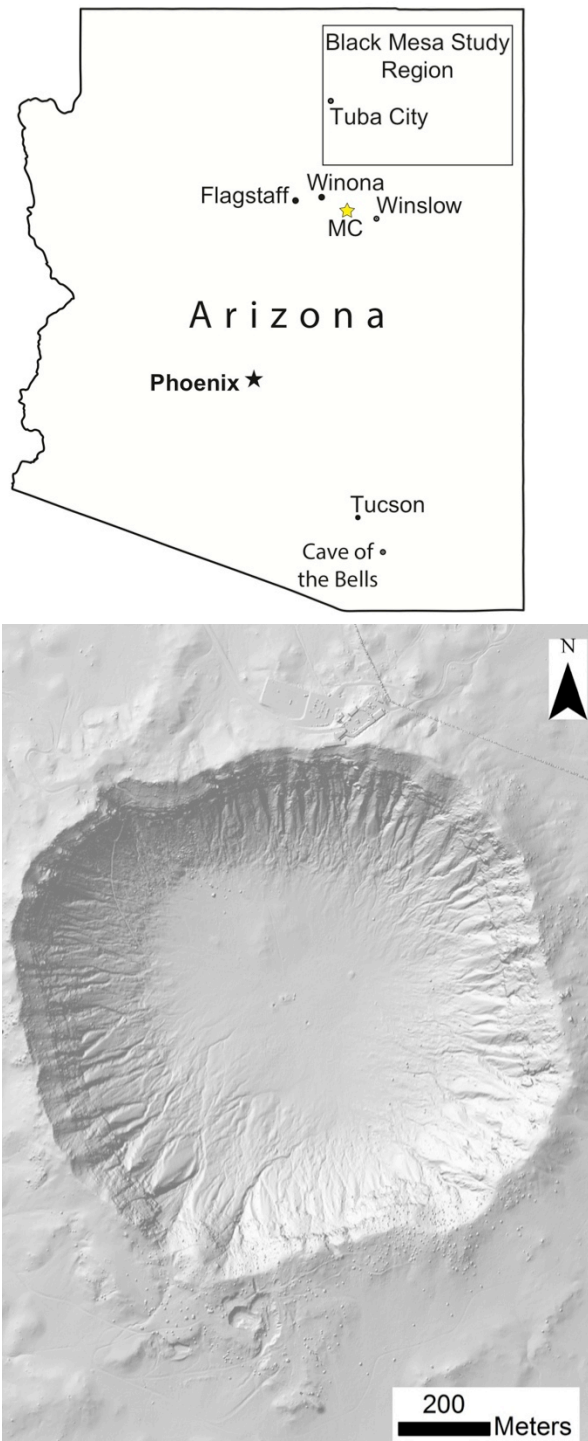


Figure 2.1 (a) Location of Meteor Crater, Arizona (USA) is marked with a yellow star; the location of the Winslow Airport, as well as other relevant locations discussed herein, are marked with black dots. (b) A shaded relief of Meteor Crater created with our 25 cm digital elevation model. The crater is approximately 180 m deep and 120 m in diameter with a rim crest that rises ~30 m to 60 m above its surrounding plains.

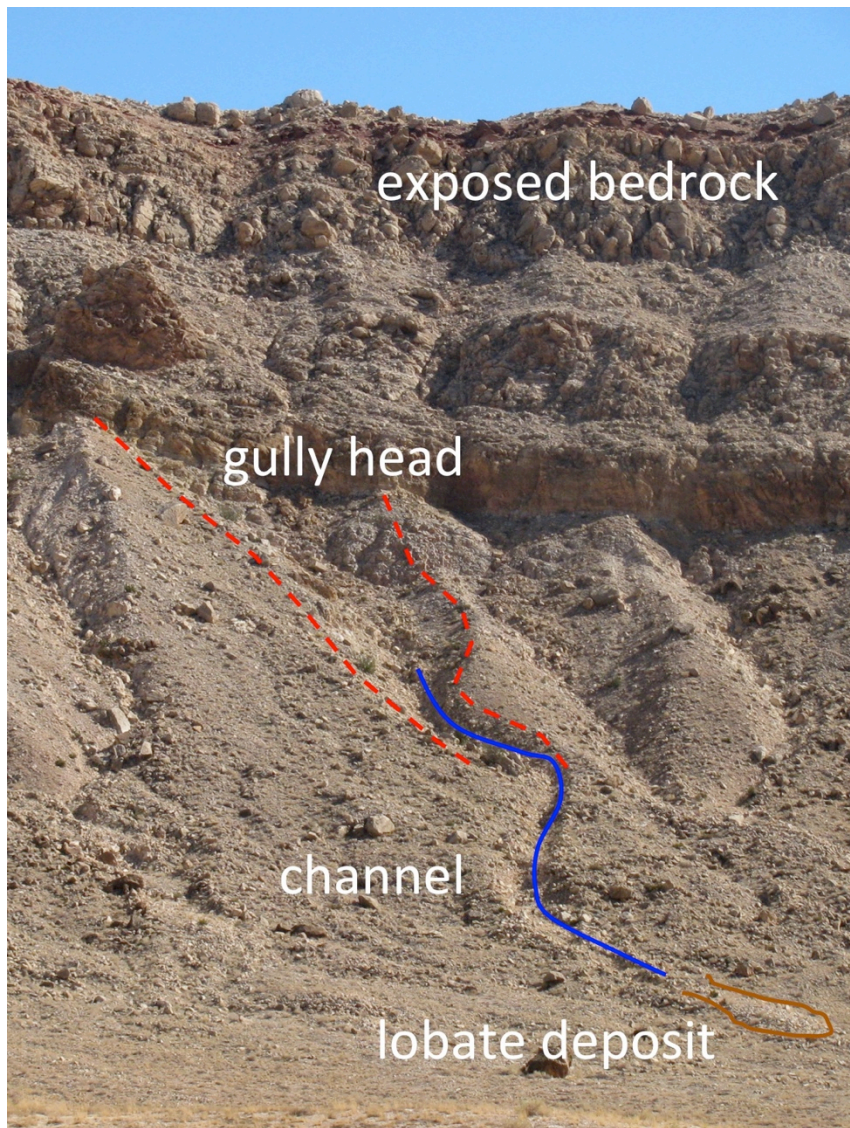
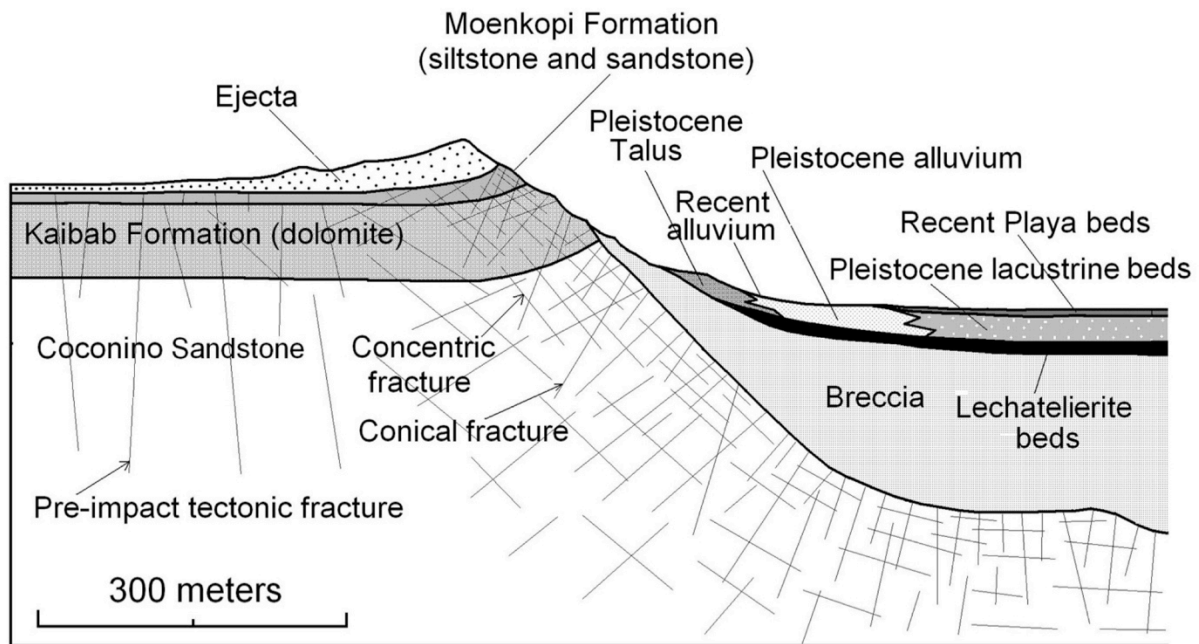


Figure 2.2 An image of the northeastern wall of Meteor Crater, highlighting the morphology of a Meteor Crater gully system, which consists of a well-defined gully head, a channel, and a depositional apron or deposit.



Reprint from Kring [2007]

Figure 2.3 Geologic cross-section of Meteor Crater from Kring [2007].



Figure 2.4 A geologic map originally made by Shoemaker and Kieffer [1974] and reproduced in Kring [2007]. The yellow stars marks the sampling locations of three levee deposits, the purple dots represent the surveying locations of fluvial channels after a storm event in September 2011, the orange dots mark the lowest location of boulders (>265 mm) found on the crater floor, and the blue contour line represents the highest stand of the paleo-lake.

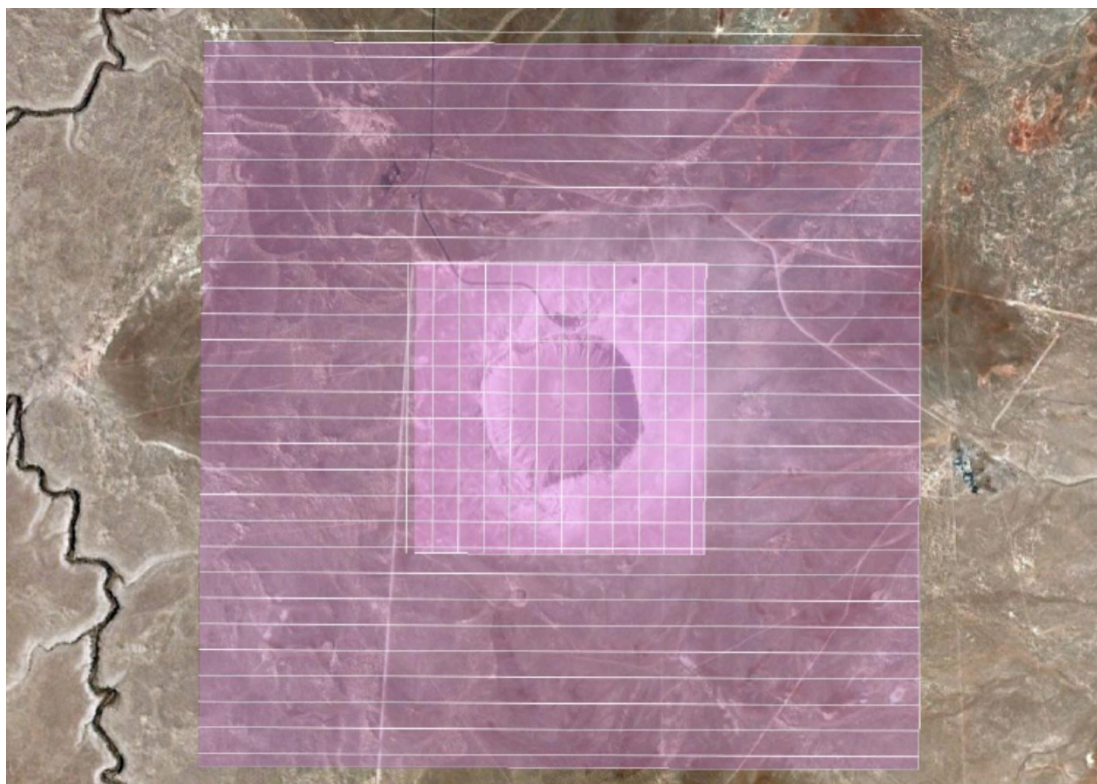


Figure 2.5 Flight lines used by NCALM to collect high-resolution data of the crater interior. Point density in the inner crater was ~ 8 points/m² and that for the rim and surrounding plains was ~ 4 points/m².

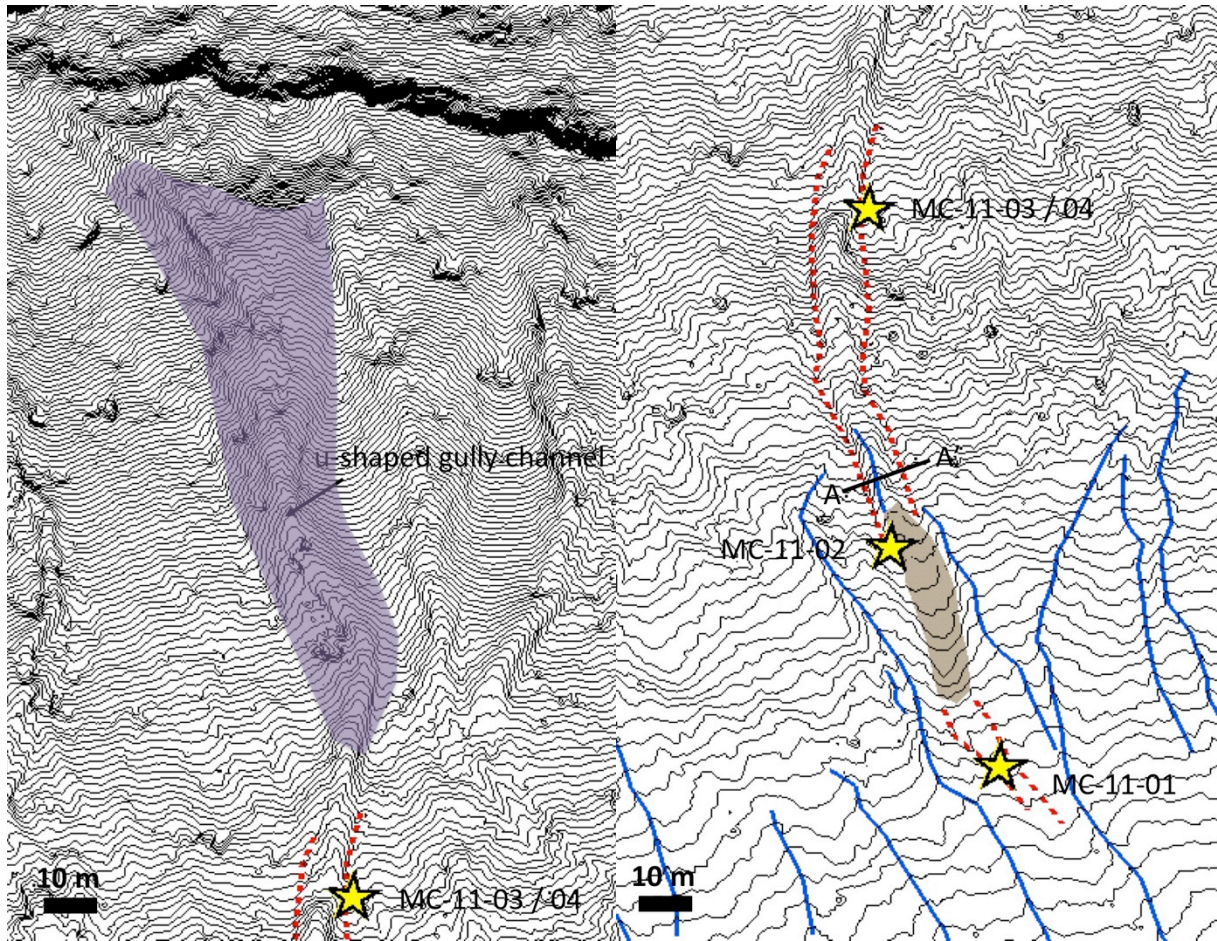


Figure 2.6 (a) A topographic map with 0.5 m contours of the gully head of Kuni's Thumb; the main channel is represented by u-shaped contours and the gully head region is highlight in purple (b) Sampling locations for the ^{10}Be cosmogenic dating of coarse-grained levee deposits are marked with yellow star; we have also mapped the location of the Kuni's Thumb lobate deposit, its levees, and nearby fluvial channels.

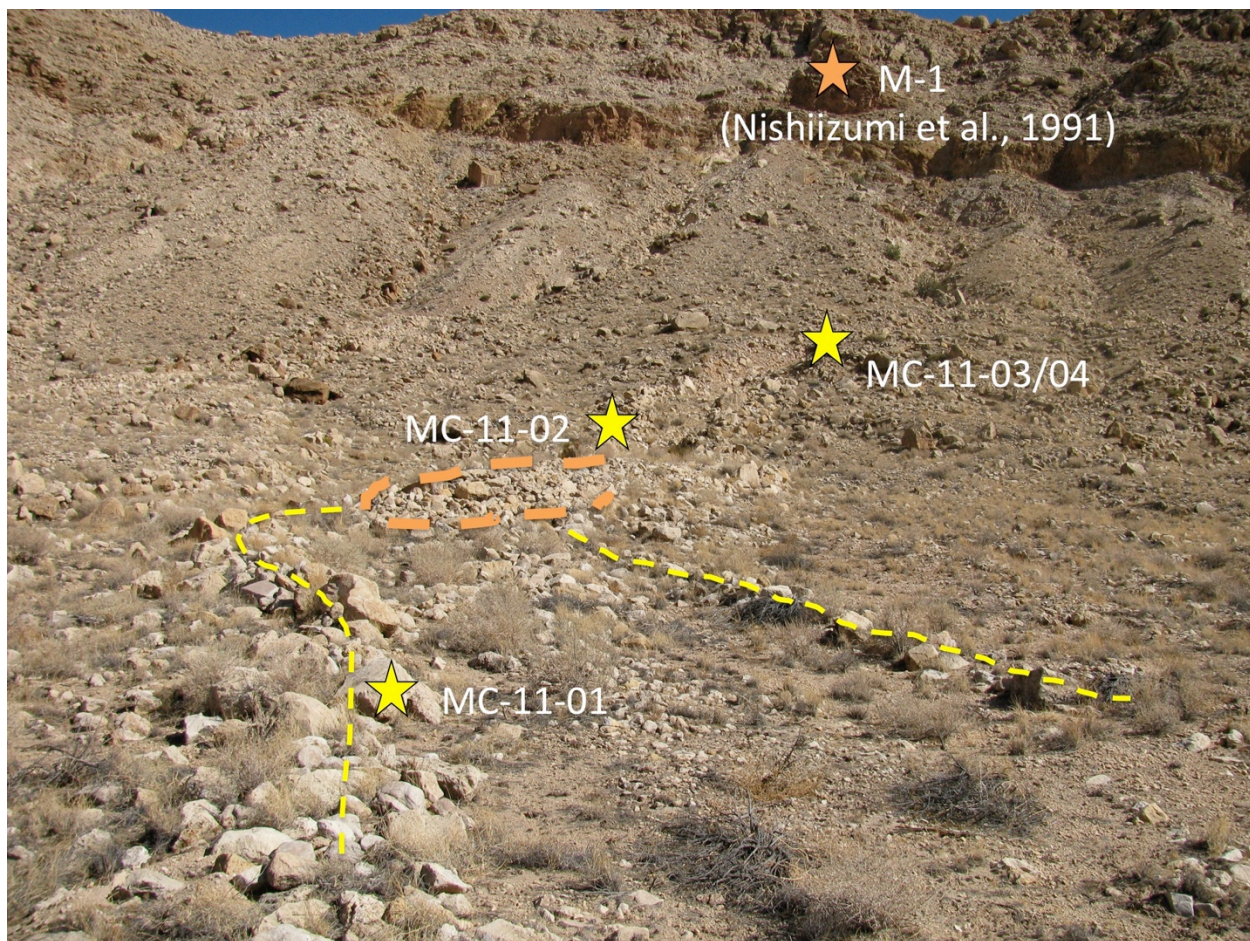


Figure 2.7 Image of the Kuni's Thumb site taken from the crater floor, looking up at the crater wall. The sampling locations shown in Figure 2.6b are marked, as well as the location of the prominent outcrop above Kuni's Thumb that was sampled by Nishiizumi et al. [1991] and referred to as M-1.

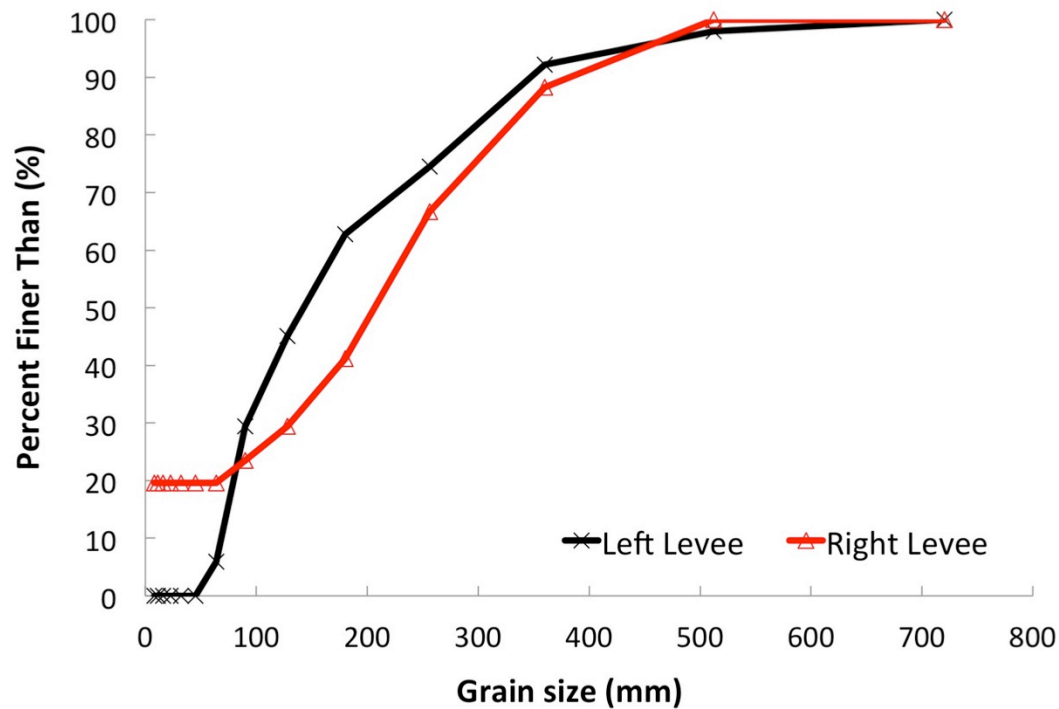


Figure 2.8 Grain size distribution of the levee deposits, located near MC-11-02 at Kuni's Thumb near A-A' in Figure 2.6b. The right levee (looking downslope) distribution is shown in red and the left levee is shown in black.

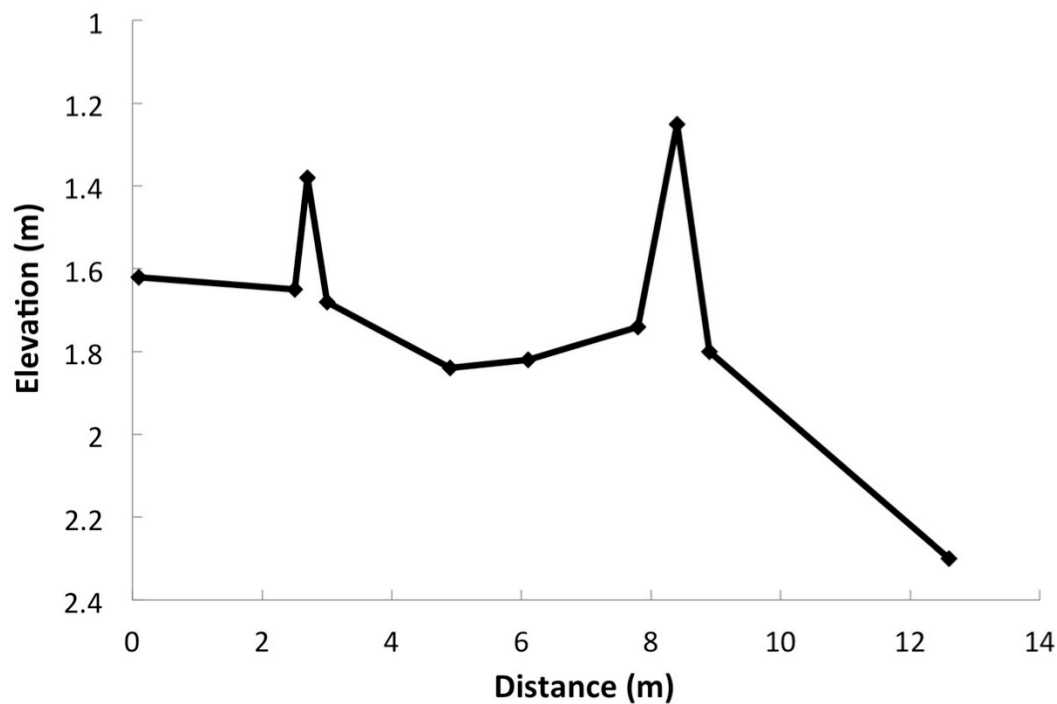


Figure 2.9 A cross-section, performed in the field, of a levee deposit at Kuni's Thumb. The location of the cross-section is shown in Figure 2.6b.



Figure 2.10 Image of the lateral deposit found at Kuni's Thumb showing poorly sorted, matrix-supported clasts indicative of mass flow processes.

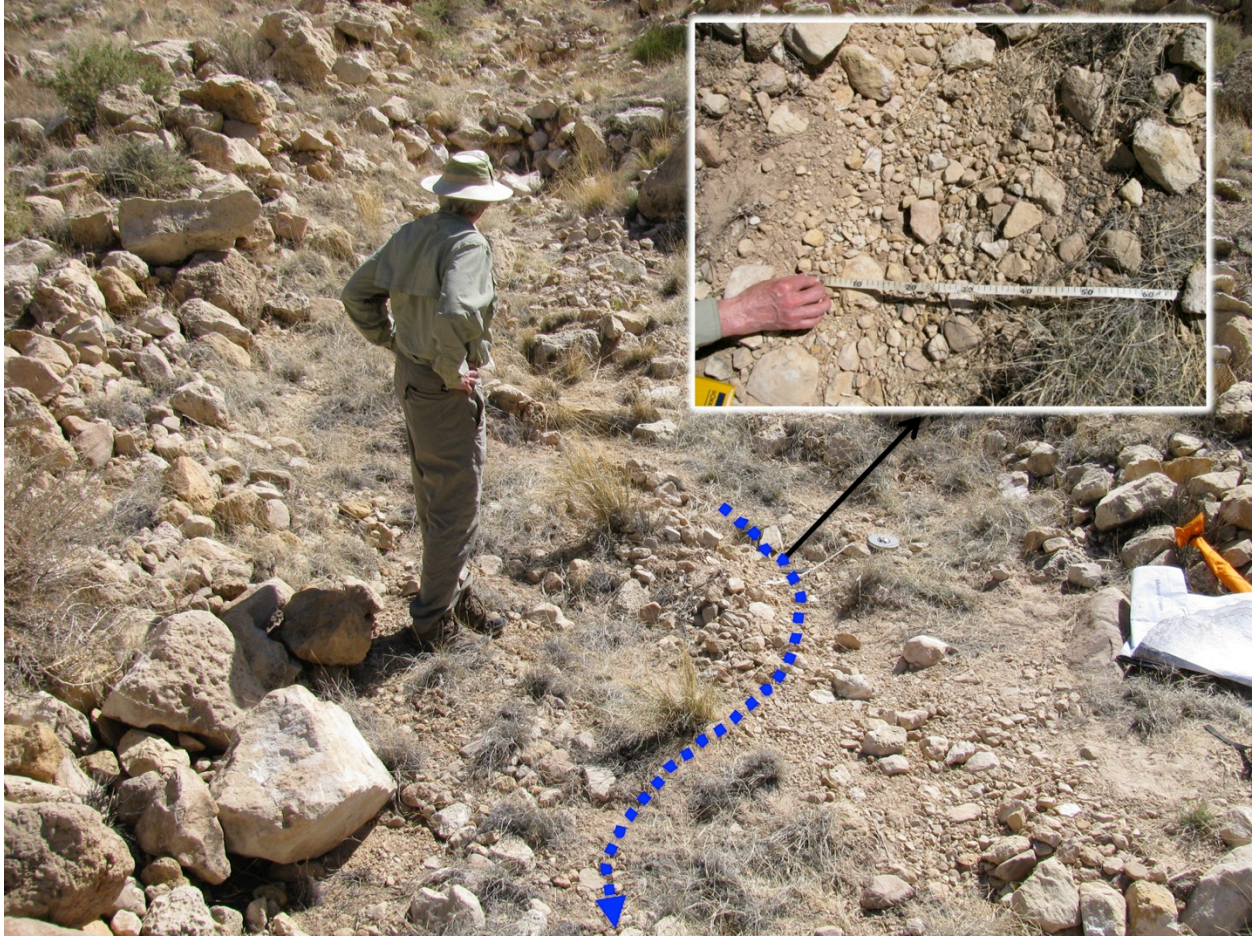


Figure 2.11 Image of the fluvial channel where grain size data was collected (Figure 2.14) is shown in the main panel; the low flow channel is highlighted with a dashed blue line. The low flow channel bed is shown in the inset panel.

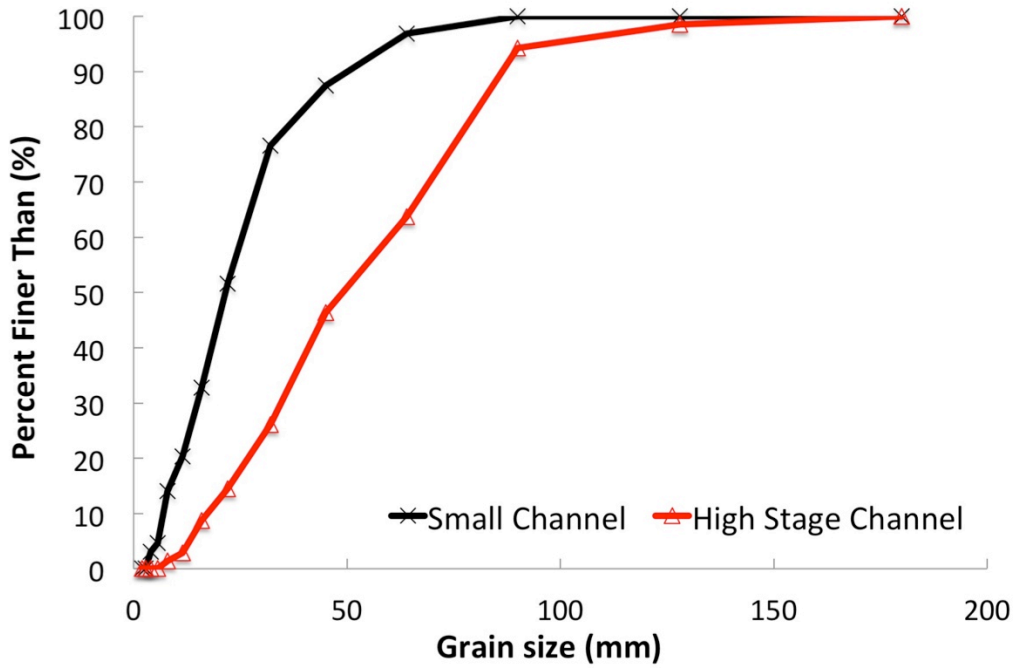


Figure 2.12 Grain size distribution of a small fluvial channel located between the levee deposits on the gully floor. The grain size distribution for the larger, high-flow channel is shown in red and the low-flow channel is shown in black.



Figure 2.13 White fine-grained material located near the gully wall in Kuni's Thumb supporting a coarser gravel layer. These fines may be necessary for both preventing rapid infiltration of water supplied from the bedrock cliffs above, and they act to cause the fluid phase to become more viscous than water thereby preventing the fluid from escaping from the solid phase.



Figure 2.14 Another example of the fine-grained material exposed in the gully walls at Meteor Crater, located approximately at the gully headwall.

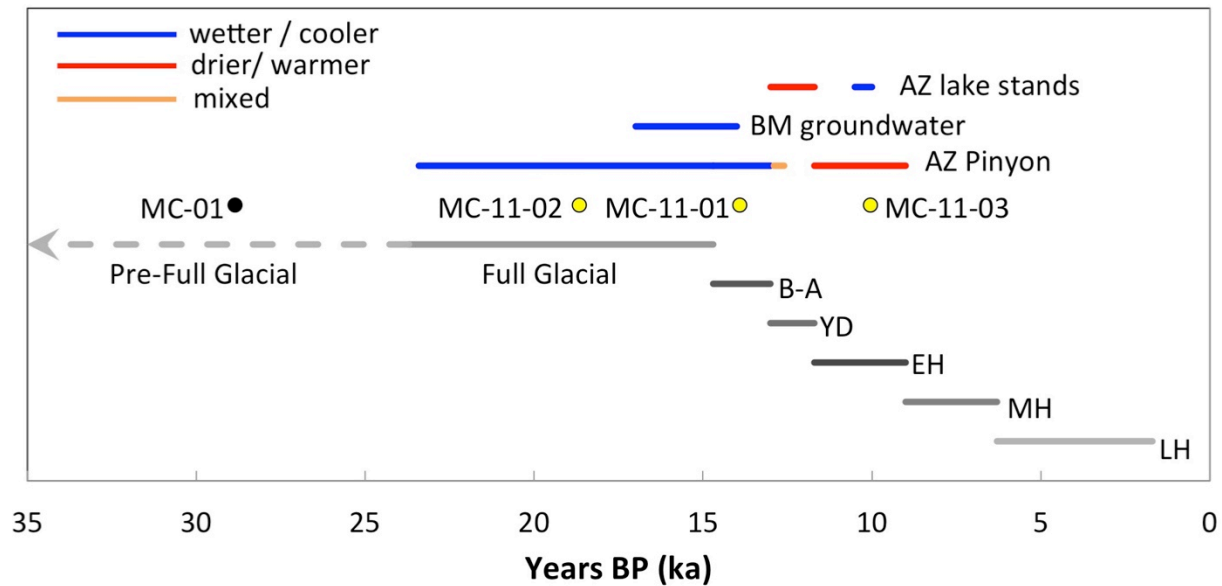


Figure 2.15 A timeline from the late Pleistocene to the late Holocene, showing (1) the years (ka) covered by each of the major periods between the late Pleistocene to the late Holocene, (2) the exposure age dates of the samples taken from Kuni's Thumb, as well as the sample (M-1) collected by Nishiizumi et al. [1991], and (3) periods of time suggested to be wetter/cooler, warmer/dryer, or a mix based on paleo-environmental indicators (i.e. Pinyon tree extent, paleo-lake high stands, and groundwater levels within Mesa Basin (Az); references herein).

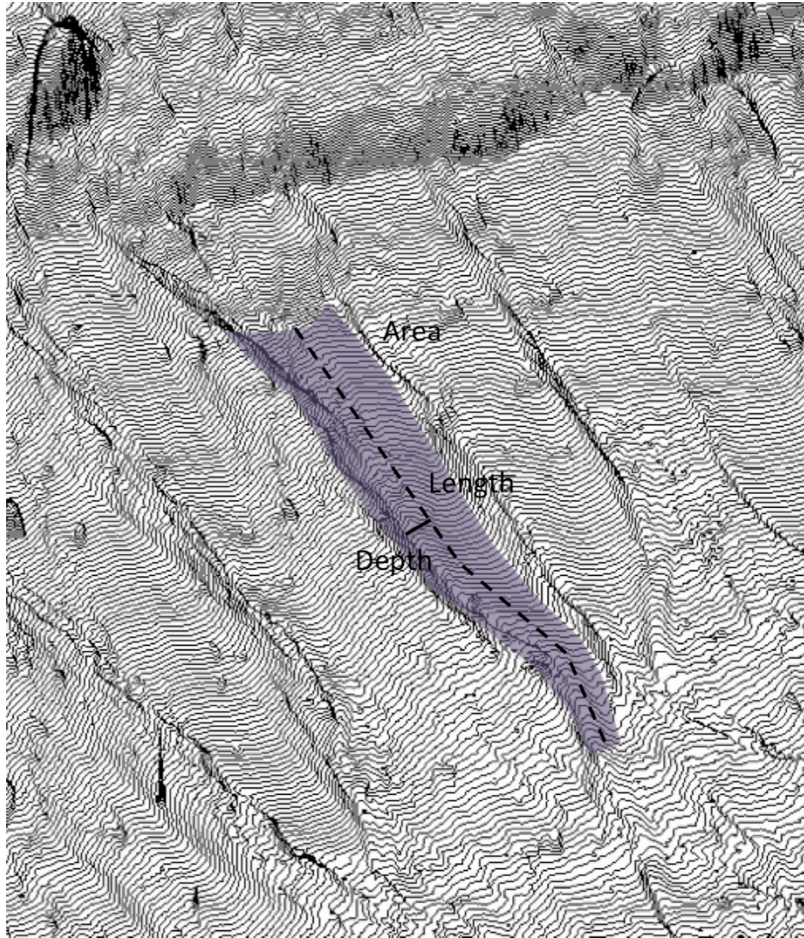


Figure 2.16 Topographic “3-D” map showing our definition of gully head area, length, and depth for Kuni’s Thumb gully.

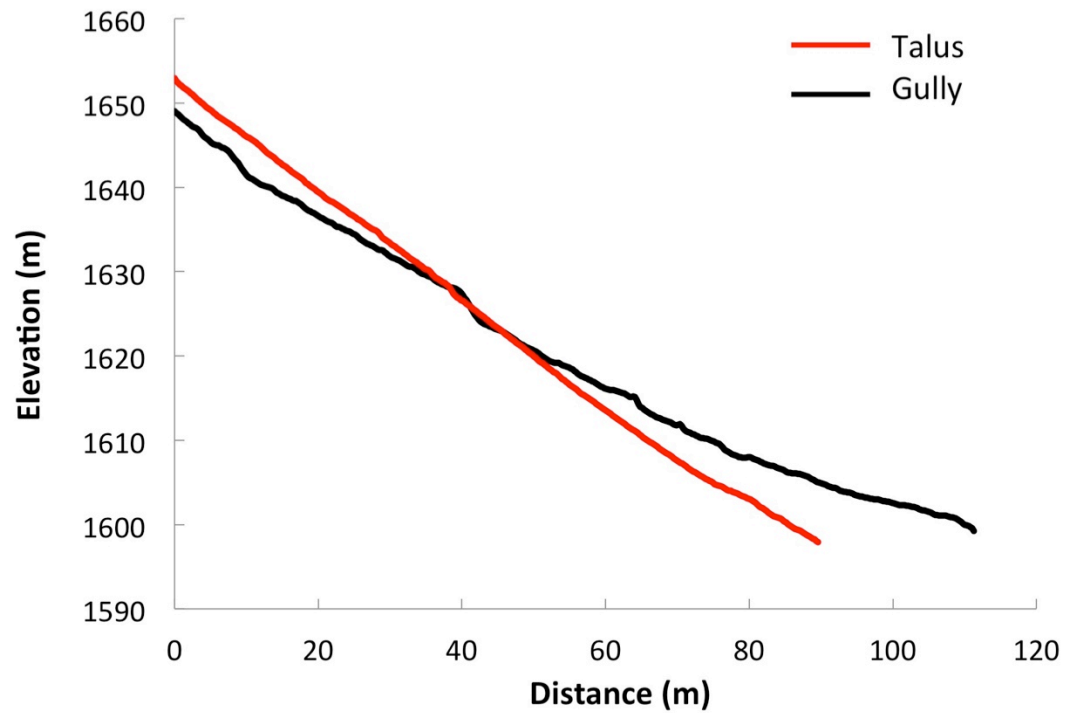


Figure 2.17 Long profiles of typical gully channel (black line) and talus slopes (red line). On average, gully channels tend to be more concave up than their surrounding non-incised talus slopes.

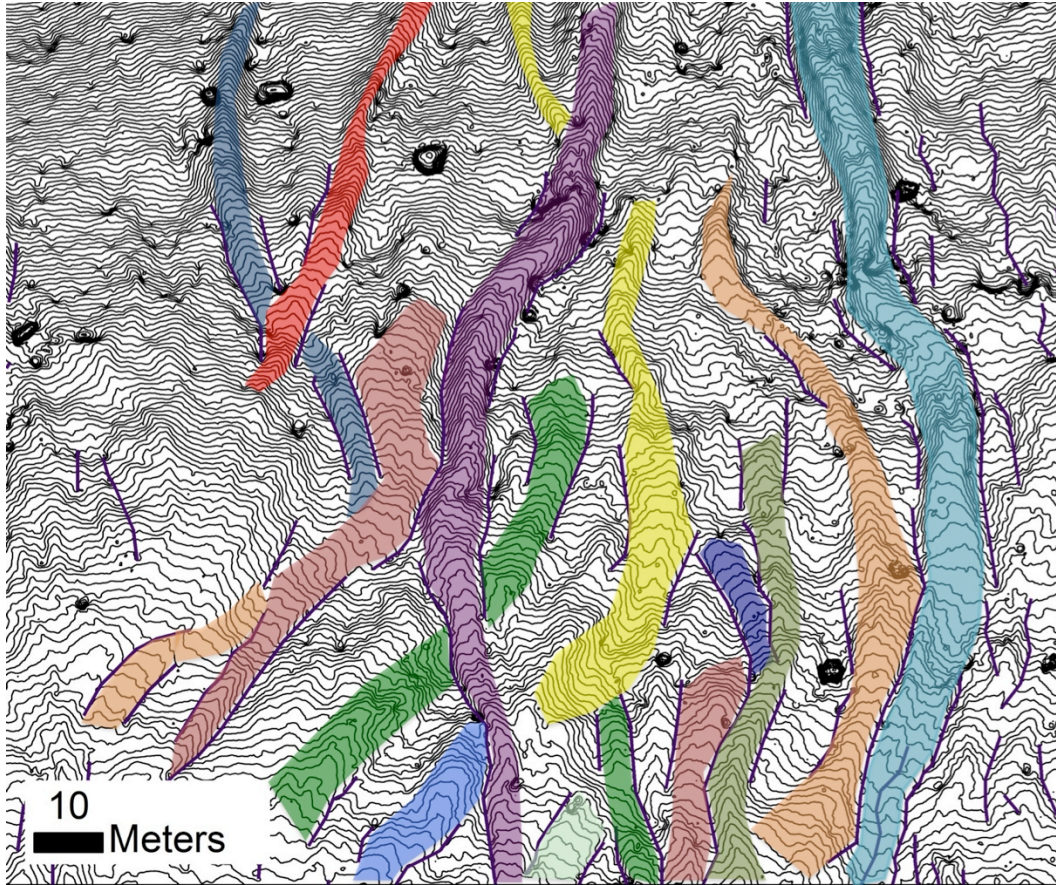


Figure 2.18 Topographic map of leaved channels (shown in solid purple) mapped on the southeastern side of the crater; each channel is color-coded differently to show the sinuous, crossing, “noodle-like” nature of the debris flow channels within Meteor Crater.

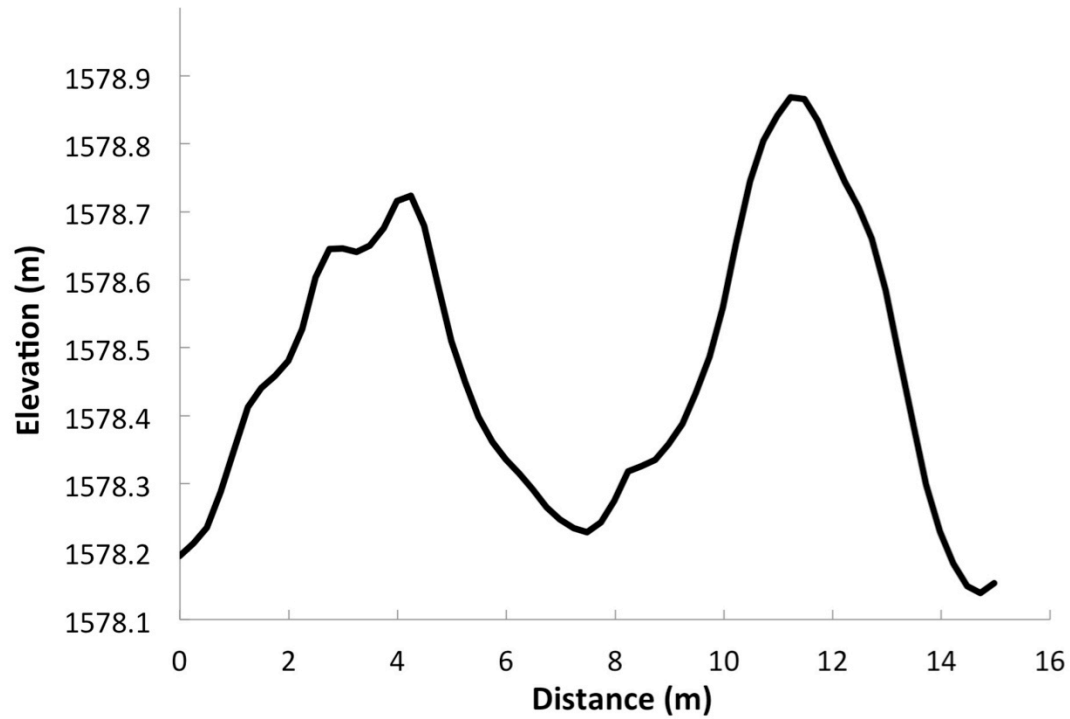


Figure 2.19 Cross-section of a typical paired levee deposited, extracted from the 25 cm DEM. The individual levees tend to be rounded in cross-section, and are often slightly different in height from one another.

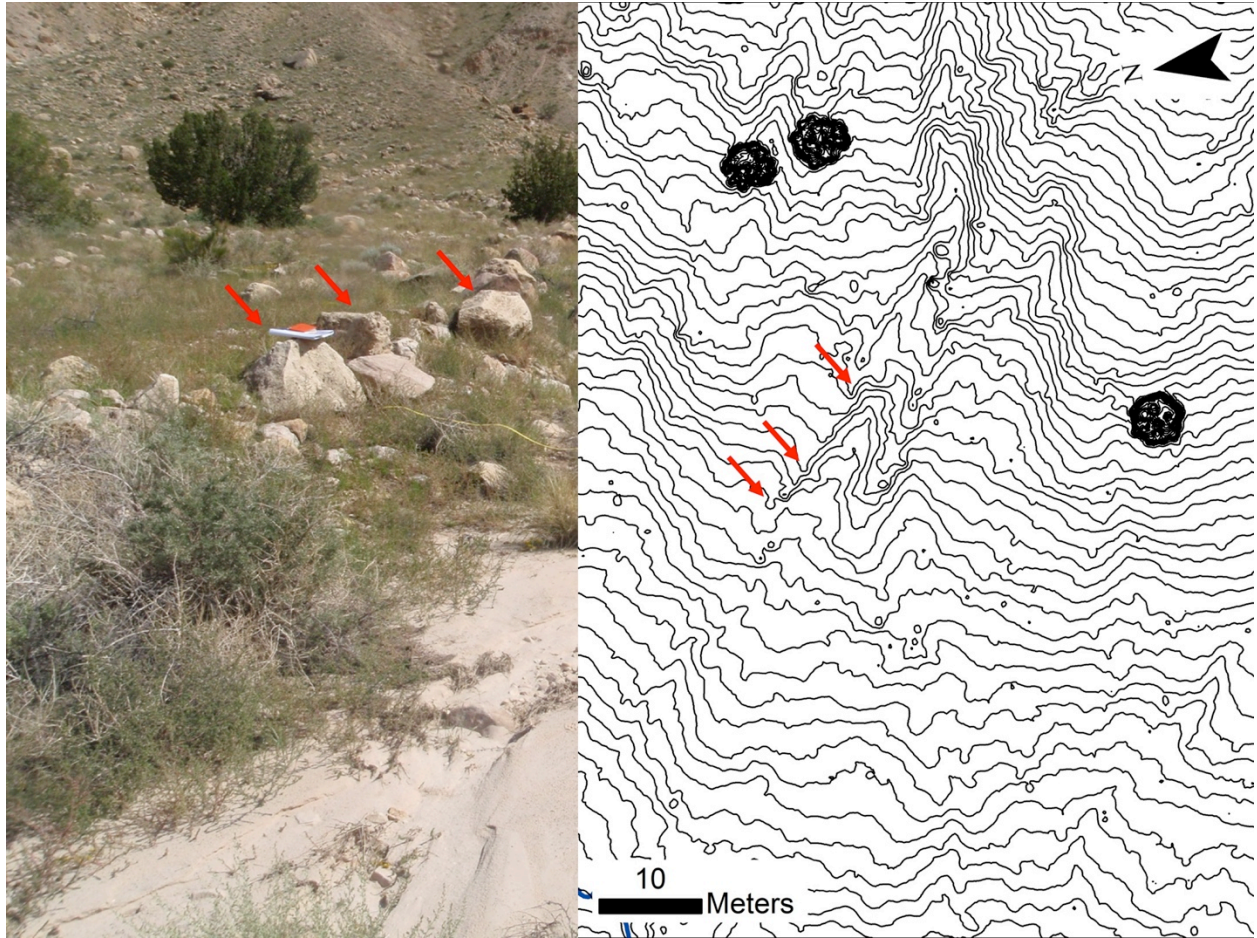
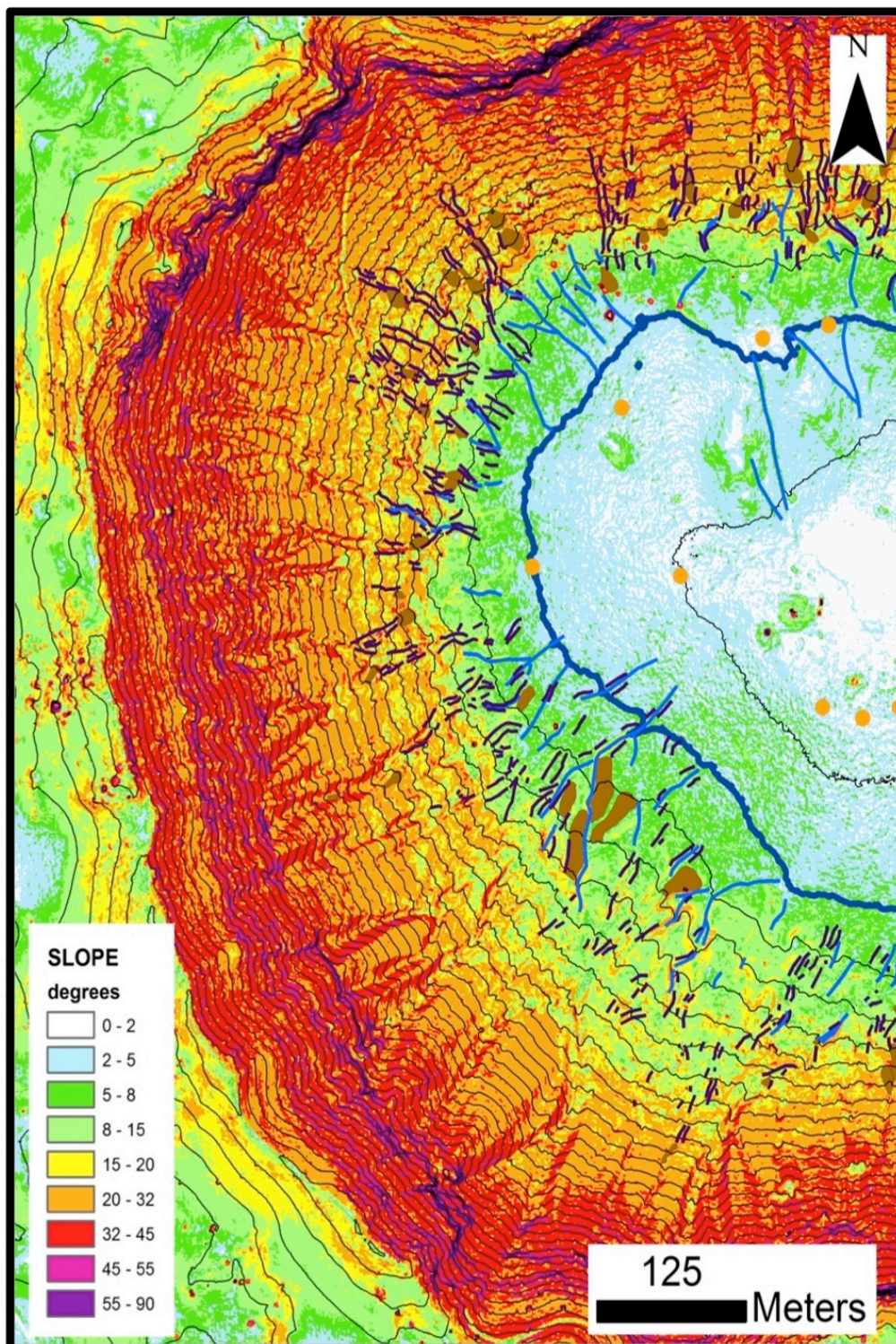


Figure 2.20 (a) Example of a line of individual boulders comprising an older levee at Gully 52 (b) The corresponding topographic map, with 0.25 cm contours, showing that the boulders are visible within the DEM.



a.

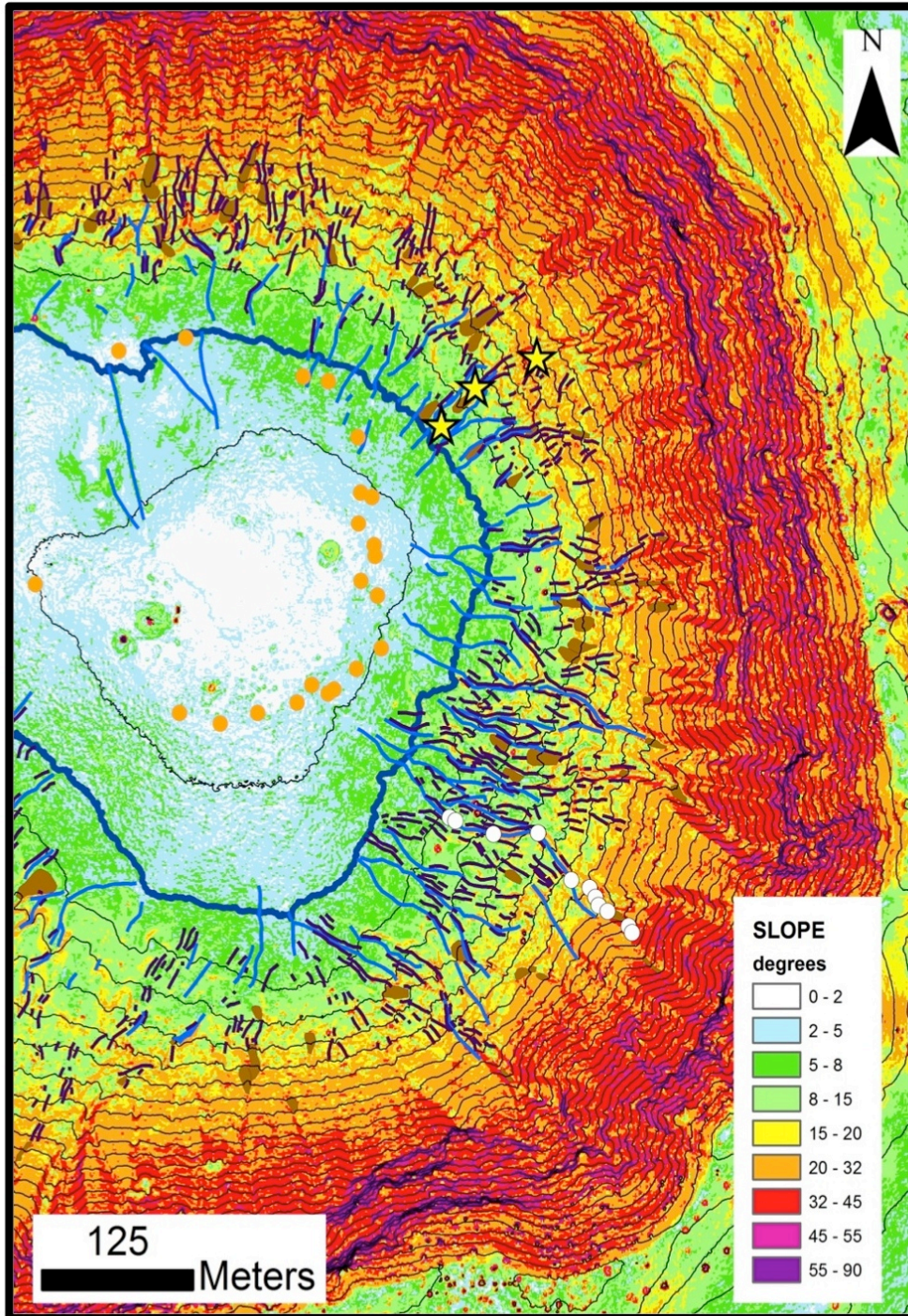


Figure 2.21 Slope map with 5 m contour lines of the (a) western and (b) eastern portions of the crater showing outlines of the gully alcoves/heads (in black), levee deposits (dashed purple lines), lobate deposits (brown), v-shaped fluvial channels (light blue lines), and the lowest mapped position of boulders on the crater floor (orange dots). The highest mapped lake deposits correspond to the 1570 m contour line, highlighted in dark blue. This map highlights the correspondence between slope and process; gully incision occurs between slopes of 20 and 32 degrees, levee and lobate deposit deposition corresponds to slopes between 5 and 15 degrees, and fluvial incision dominates below slopes of 5 degrees.

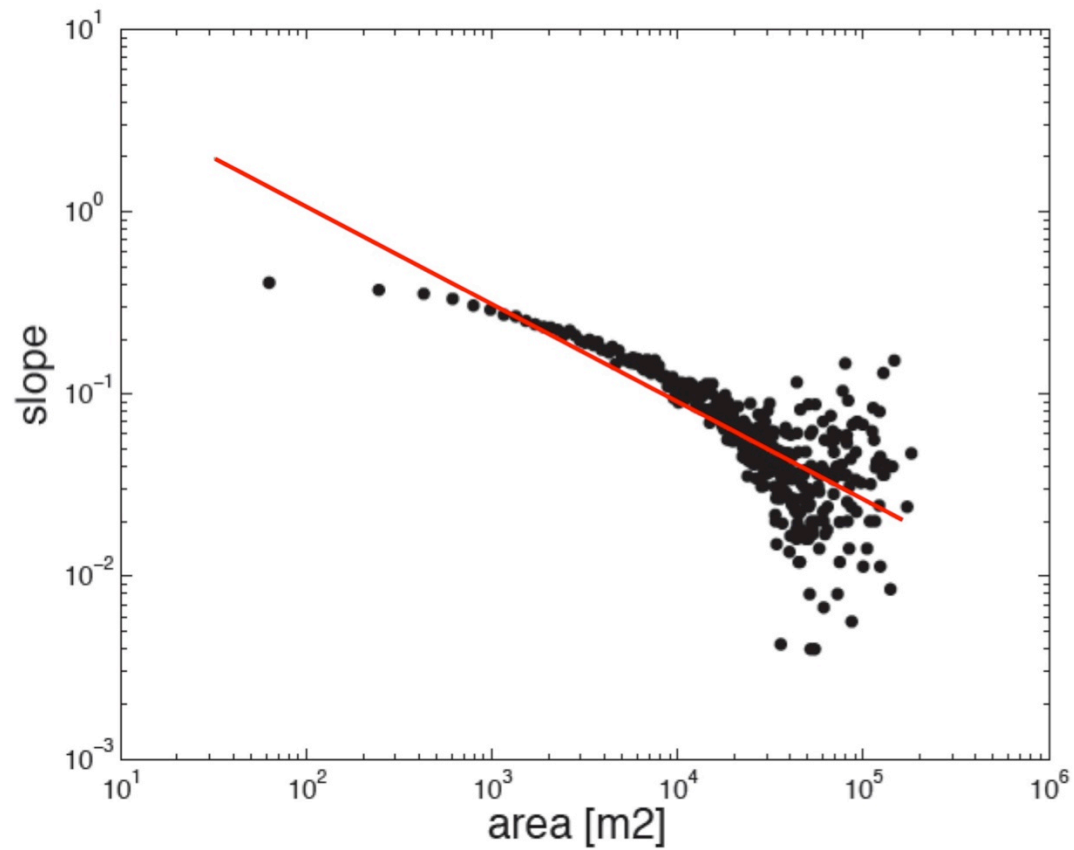


Figure 2.22 (a) Area-slope data for the channels at MC are nonlinear, indicating mass flow processes are important in shaping this landscape (Stock and Dietrich, 2003). Linear power law regression (black line) is shown for comparison.

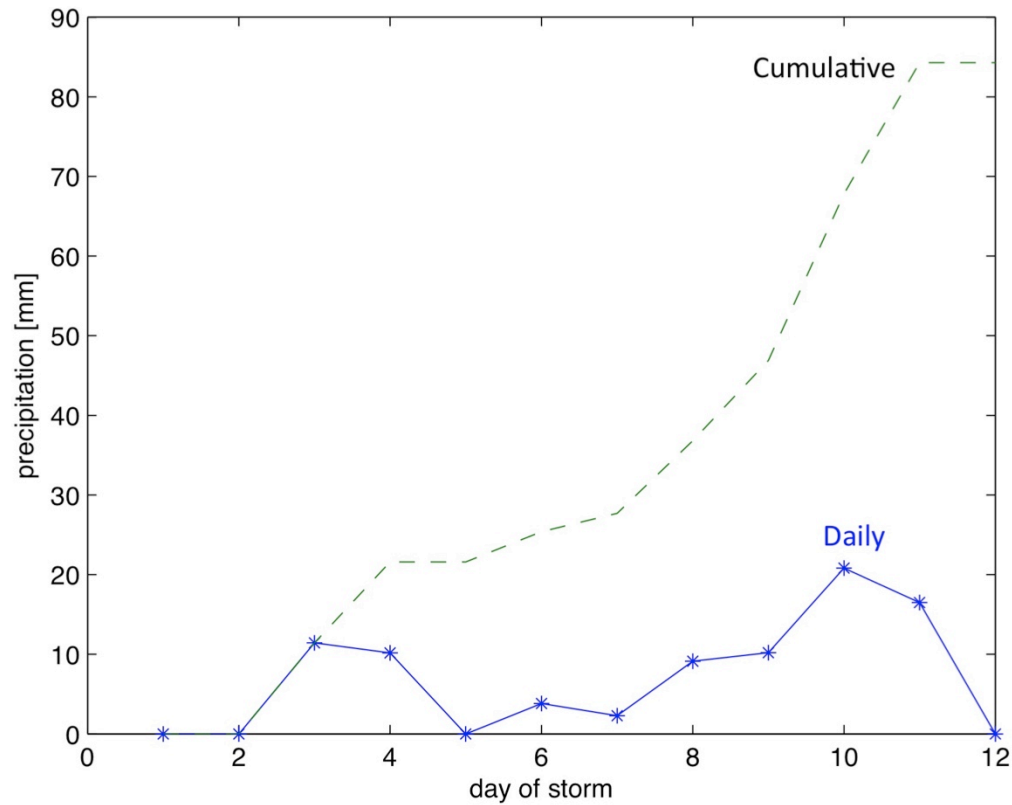


Figure 2.23 Precipitation data from the Winslow Airport station for the September 2011 storm. The cumulative rainfall amount was ~85 mm, while the max daily was ~25 mm. Comparing the storm data to historical precipitation data (Figure 2.4) shows that while this is a relatively large storm for the area, larger storms have occurred in the last 70 years.

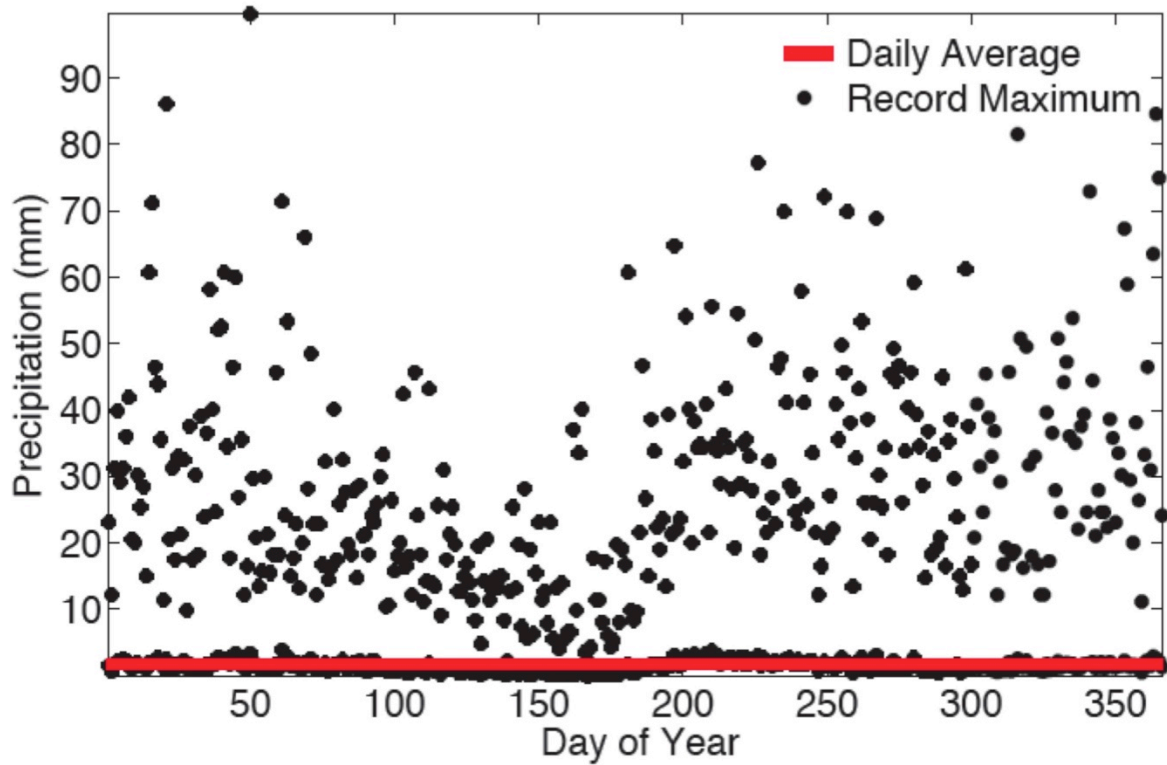


Figure 2.24 Local maximum daily precipitation data from the nearby Winslow Airport (~30 km from Meteor Crater). The daily maximum average is 1.4 mm, and the maximum on record for the past ~70 years of data is 99 mm.



Figure 2.25 Image of the fluvial erosion that occurred within Gully 52 after the September 2011 storm.

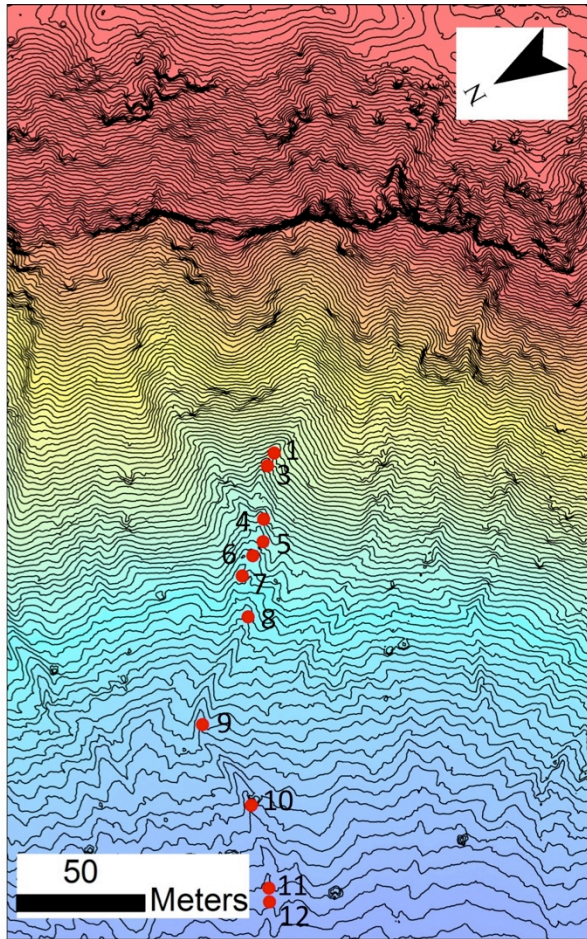


Figure 2.26 Location of the 11 sites visited after the September 2011 storm. At each site, the width and depth of the fluvial channels that had developed as a result of runoff from the storm, as well as the grain size of fluvially transported material and the local slope, was collected and reported in Table 2.2.

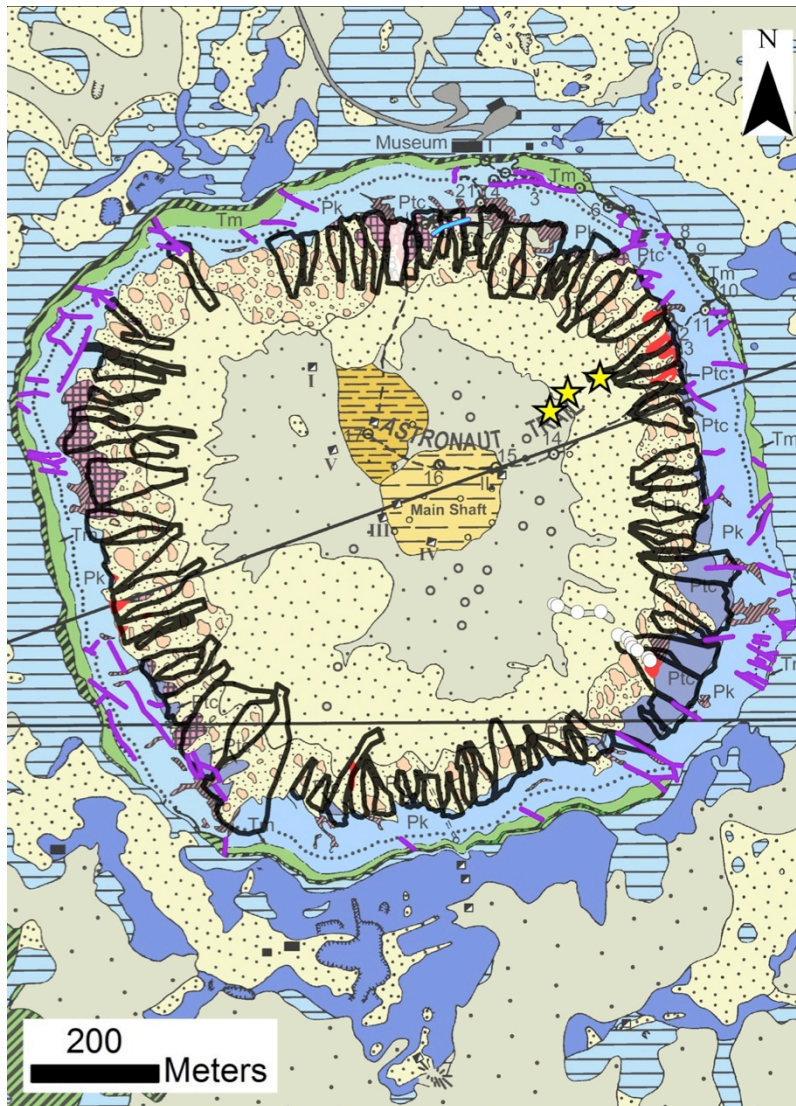


Figure 2.27 Geologic map from Showmaker and Kieffer [1977] with their fault mapping highlighted in purple overlain with outlines of the locations of the gully heads; for each of the gully heads there is not a corresponding fracture suggesting that seepage from fracture flow did not lead to gully incision for the majority of the Meteor Crater gullies.

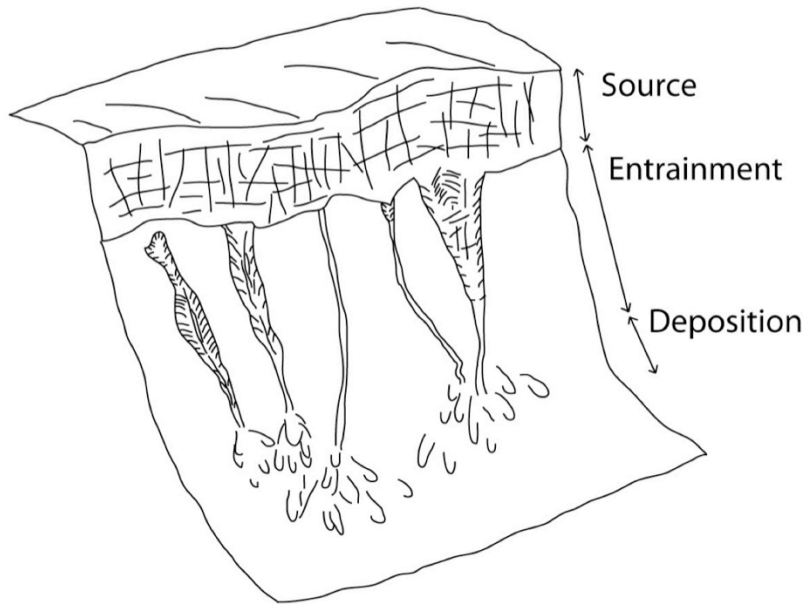
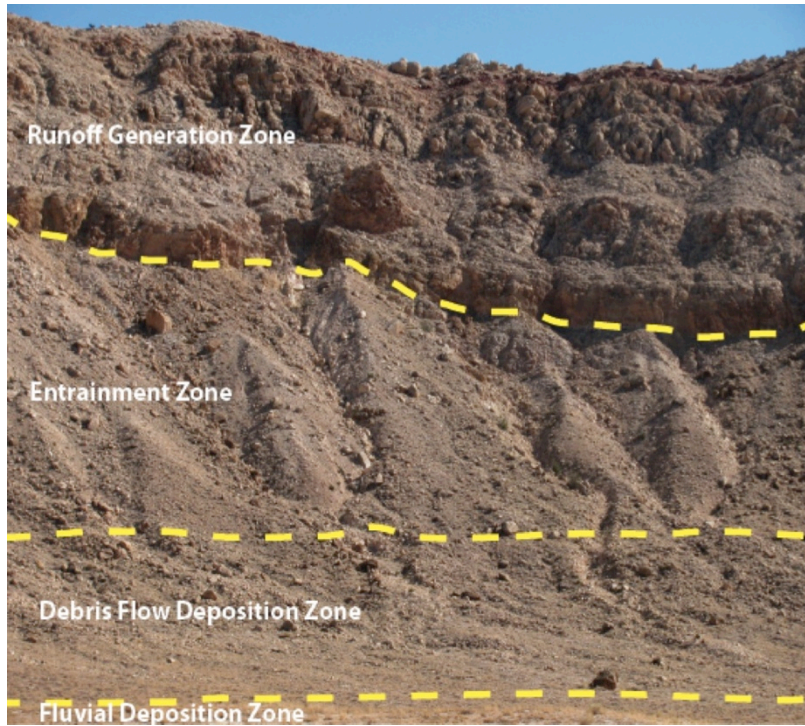


Figure 2.28 (a) An image of the northeastern wall of Meteor Crater showing the 3 primary process zones, which may be applicable to both terrestrial and Martian gullies: a water source area (typically in bedrock), a sediment entrainment zone (bedrock slope mantled with impact breccia (for the case of the crater), colluvium, and/or eolian deposits), and a depositional zone (that progressively flattens as the mode of transport shifts from mass flow to fluvial and lake deposition). (b) A cartoon version of the 3 main process zones discussed above.

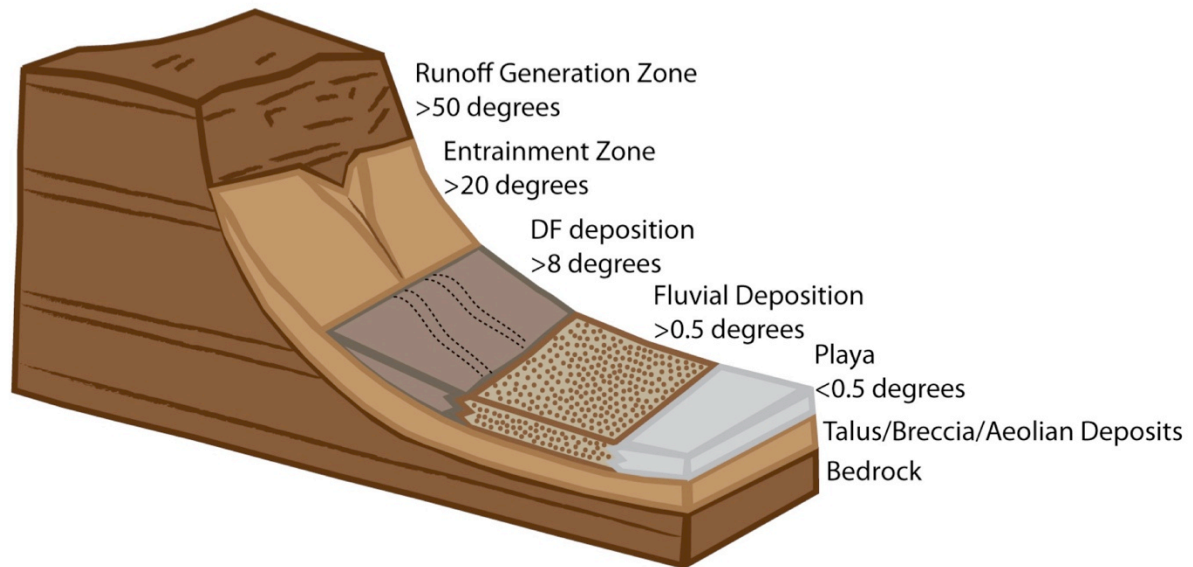


Figure 2.29 When sufficient runoff from the water source area crosses the lower sediment mantled zone, mass flows can develop. This may occur through saturation and mobilization of eroded bedrock material via landsliding processes or the runoff may rapidly entrain channel sediments, undermining adjacent banks that collapse onto the flow and transform it into a surging debris flow. When these flows reach lower unconfined slopes, deposition ensues. In flows with poorly sorted sediment, the coarse fraction will migrate to the sides and front of the flow, leading to levee formation and a snout that may be pushed downslope by the flow.

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Chapter 3: Sustained pore pressures at cessation of debris flow motion: Implications for the role of fines in debris flow mobilization based on large-scale experiments

3.1 Abstract

The dominant mechanisms behind sustained mobilization in granular debris flows are poorly understood, and experiments are needed to determine the conditions under which the fluid can fully support the coarse fraction. However, field-scale studies are difficult to instrument and constrain and laboratory studies suffer from scaling issues. We performed a series of experiments in a 4 m diameter vertically rotating drum to assess the role of fines and coarse grain size distribution on the rate of pore fluid pressure dissipation upon cessation of movement of a granular mass. We varied the amount of fines, from no fines to amounts found in debris flow deposits, and the coarse grain size distribution, from a single grain size to a range found in natural flows. While rotating each mixture in the drum, we monitored fluid pore pressures at the base of the flow, both near the wall of the drum and at the center. Then we stopped the drum and continued to monitor the fluid pressures. Immediately upon stopping, the pore fluid pressure was nearly hydrostatic for the gravel-water flows, and any elevated pore pressure quickly dissipated. The mixtures with fines contents close to those found in actual debris flows had elevated pore pressures, indicating they were almost fully liquefied. While the presence of fines were first order in leading to the generation and maintenance of elevated pore fluid pressure, the highest excess pore pressures were generated with the combination of fine material and a wide coarse grain size distribution. We also placed a boulder upon a fines-rich mixture after cessation of motion above the center pressure sensor, and observed that the pore fluid pressure rose instantly, bearing most of the grain's weight, before the pressure slowly declined. When the same boulder was placed on the water-gravel flows, there was no change in fluid pressure, as the mass was supported entirely by grain-grain contacts. Our observations, combined with observations of others in the literature, inform the conceptual model presented herein where we hypothesize that the dilational separation of particles during debris flow shear leads to coarse particles (in this case particles >12 mm) slowly settling through the highly viscous non-Newtonian fluid. As the fines-rich mixture has a yield strength of ~ 5 Pa, we calculate that particles ~ 12 mm and smaller remain suspended. This combined effect requires the fluid to sustain the weight of the particles, leading to pore pressures equal to the entire weight of the solid and fluid mass.

3.2 Introduction

3.2.1 What is a debris flow and how do we model them?

Debris flows are hazardous events that are common in steep, mountainous regions, but the fact that they typically initiate on slopes greater than 30 degrees, occur within a short period of time, and often are triggered during intense precipitation events [e.g. Costa, 1984; Takahashi, 1991; Iverson, 1997; Coe et al., 2007; McCoy et al., 2010] makes in-situ measurements aimed at understanding the mechanism of debris flow motion and deposition difficult. Generally, debris flows occur when unconsolidated sediment mixes with water and flows downslope in response to gravity. What distinguishes them from floods, where lift and drag forces exerted by the fluid on a grain are necessary for grain motion [Wiberg and Smith, 1989], is that both fluid-grain and grain-grain interactions are important [Takahashi, 1991; Iverson, 1997]. In debris flows, grains interact via frictional or collisional contacts while fluid, often defined as water and suspended fine sediment [e.g. Pierson, 1981; Iverson, 1997; McArdeil et al., 2007], occupies the

corresponding network of pores. The coupled system deforms in response to applied stresses, and this deformation may result in pore volume changes that affect the pore fluid pressure. Gradients in pore fluid pressure can exert forces on grains, in addition to buoyancy and fluid drag, further deforming the granular matrix and facilitating grain motion [Berzi et al., 2010; Goren et al., 2010; Kaitna et al., *in prep*]. The coupling between the granular phase and pore fluid pressure is an important part of debris flow mechanics, as elevated pore fluid pressures can reduce frictional contacts between grains and enhance flow mobility [Iverson et al., 2010; Kaitna et al., 2013; Kaitna et al., *in prep*] and provide grain support [Hampton, 1979; Pierson, 1981], the latter of which is the focus of this paper.

Measurements taken at the base of both real and experimental debris flows and their deposits show common characteristics in regards to pore fluid pressure [e.g. Iverson, 1997; Major and Iverson, 1999; McArdell et al., 2007, McCoy et al., 2010]. Typically, the coarse-grained snout has little to no pore fluid pressure, while the body of the flow often has pore fluid pressures that exceed hydrostatic, sometimes approaching the measured total normal stress [i.e. lithostatic]. The watery tail remains close to hydrostatic. Coarse-grained material that collects at the flow front and along the flow margins often deposits to form levees and lobate deposits [Hsu, 2010; Johnson et al., 2012; Yohannes et al., 2012]. Basal pore fluid pressures from deposits at the USGS debris flow flume were measured to be lithostatic, or nearly lithostatic, and remained so for up to ten hours after deposition, indicating that the pore fluid (which contained 1 to 4% mud) was continuing to support most of the coarse fraction even after cessation of flow [Major, 1996]. Observations of debris flow deposits in the field show abundant amounts of coarse debris, often up to meter-scale boulders, in a fine-grained matrix on slopes as gentle as a few degrees [e.g. Rodine and Johnson, 1976; Hampton, 1979; Major, 1997; Palucis et al., 2011; Chapter 2].

Models proposed to explain the mobility and transport capacity of debris flows and their subsequent deposition can be broadly classified into two groups: single-phase viscoelastic models and two-phase flow models. Johnson [1970] proposed a combination of Coulomb strength and Newtonian viscosity to explain several aspects of debris flows, including their tendency to form U-shaped channels and steep terminal levees. This model, which has been adopted by many others [e.g. Whipple and Dunne, 1992; Coussot, 1994; Whipple, 1997; Chen, 1998; Brufau et al., 2000; Ancy, 2007], assumes that debris flows can be adequately modeled as a Bingham fluid, a viscoplastic material that is rigid until a critical shear stress is applied, after which the mass flows as a viscous, Newtonian fluid. When the applied shear stress is reduced below its critical shear stress, the flow will stop ‘en masse’, an observation not supported by observations [Major, 1997; McCoy et al., 2010]. Much work has been done to determine the critical shear stress (or yield strength) for water-clay and water-clay-sand slurries [e.g. O’Brian and Julian, 1989; Coussot, 1995; Coussot et al., 1998], though realistically, most natural debris flows contain material coarser than sand and have much lower clay and silt contents (~10% or less of the total mass [Major, 1996; see review in Iverson [1997]; Coe et al., 2008; McCoy et al., 2012]) than those used in the rheological studies mentioned above, where experimental clay fractions that lead to measured yield stresses were often 20% or higher. In Johnson’s [1970] model, pore pressure merely serves to reduce the total normal stress, effectively reducing the critical shear stress needed to induce flow, but any coupling between the granular and pore fluid phase is ignored.

Though they model debris flows as a single-phase Bingham fluid, Rodine and Johnson

[1976] consider debris flow strength to be a function of coarse grain packing. When coarse grains are fully dispersed the strength of the debris flow is essentially that of the pore fluid; debris flow strength only increases with grain interlocking. Their theoretical analysis suggests that poorly sorted grains can compose up to 95% (by volume) of the flow, while still not contributing to its gross strength. They also assume that for a given grain size, all grains smaller constitute the fluid, referred to as “pyramiding”, and hence the density of the ‘fluid’ is increased leading to greater buoyancy forces, allowing for flows to simultaneously travel over low slopes while transporting large boulders. Pierson [1981] mixed slurries of different compositions and monitored the decay of pore pressures as the slurries settled under quasi-static conditions. In all cases, the pore pressures were maintained over hydrostatic conditions for a significant period of time. Pierson [1981] found that adding fine particles (i.e. silt and clay) further reduced pore pressure dissipation rate, which he suggested was due to an increase in effective fluid viscosity, which thereby reduces the fluid’s ability to escape through narrow pore spaces. Based on his field and experimental work, Pierson (1981) proposed that the fluid effectively supported the coarse particles within a debris flow by a higher fraction than buoyancy forces would predict, and he suggested that elevated pore fluid pressures may be important for increasing the buoyancy of coarse particles and reducing bulk flow resistance. Hampton [1979] assumes that the total pressure gradient producing buoyant forces on a grain has two components: the density of the water and the loading of the water by clay and/or coarse grains. Effectively, the fluid is the water-clay matrix and any coarse grains whose weight has been transferred to and is supported by the matrix strength. Hence, the vertical pressure gradient is higher than the gradient predicted from density effects alone. Pierson [1981] shows this to be the case for his experimental slurries immediately after mixing, but according to Iverson and LaHusen [1989] unless there is an upward flow of fluid from the bed to the liquefied mass, such a gradient can only be maintained while there is net settling or contraction of the granular phase.

A second class of models takes into account both the granular and fluid phases. Takahashi [1981] derives a dilatant flow model after the work of Bagnold [1954] in which momentum is exchanged via grain-grain collisions and the interstitial fluid plays a minor role. As the grains undergo rapid shear, they dilate and the inter-grain stress is a function of the square of the shear rate. The reverse process occurs when the applied stress decreases, but as Major [1996] pointed out, for realistic inputs, the model predicts deposition at fairly steep slopes (~14-18 degrees), though some deposits are observed to occur on these steep slopes in nature [Chapter 2]. Hutchinson [1986] proposed a two-phase ‘sliding-consolidation’ model in which pore fluid pressures modify the granular flow dynamics. Excess pore pressures are generated from undrained loading, such as that caused from contraction of the debris, leading to mass movement downslope as a result of translational sliding. The base of the sliding ‘sheet’ is saturated or near-saturated, with excess pore pressure mitigating the downslope motion of the mass. Further consolidation brings about the decay of the excess basal pore pressures, resulting in the termination of flow ‘en masse’. Iverson [1997] argues that monotonic contraction, like that described by Hutchinson [1986], cannot be sustained in a steady flow, hence debris flows must be inherently unsteady phenomena. Furthermore, deposition is a special case of unsteady motion, in which the segregated coarse material along the flow perimeter lacks high pore pressures and forms a dam, blocking and eventually impeding the main flow body, which retains pore pressures in excess of hydrostatic [Iverson, 1997; Major and Iverson, 1999]. Kaitna et al. [2012] showed however that excessive pore pressures can be maintained during steady-state flows, and

for muddy, poorly sorted mixtures, these pressures were near liquefaction. In addition to elevated pressures, these mixtures also had a marked rate-dependence of the surface slope to the mean velocity, whereas the surface slope for fines-poor mixtures had a weak rate-dependence.

While models like those of Iverson et al. [2010; 2012] and Pitman and Le [2005] acknowledge the role of pore pressure in modifying debris flow dynamics, their models are depth-averaged. Quantities like pore pressure and depth are functions of only the downstream flow direction and time. Changes in porosity, which can be accompanied by significant pore pressure changes, are due to contraction or dilation of the entire debris mass only, so the model cannot simulate pore fluid pressure gradients with depth due to grain shearing, grain settling, or local porosity changes as a function of grain size segregation with depth, sometimes referred to as kinetic sieving [Iverson, 1997]. Berzi et al. [2010] argue that a two-phase model must distinguish between the depths of the solid and fluid phase, not just the overall flow depth, as it affects the degree of saturation, and hence the local granular rheology, of the flow.

3.2.2 The role of fines in debris flow mobility

If the generation and dissipation of pore fluid pressures in debris flows (and their deposits) is indeed the result of the interaction between the solid and fluid phase, then certain fluid properties, like viscosity, are an important component of debris flow mechanics. The fluid viscosity will affect both the rate of grain settling and the ability of the fluid to advect through the granular matrix. We expect that the concentration of fine material, the type of fines, specifically clay type, and perhaps even the grain size distribution of the fines matters, based on the rheology and flow behavior of different water-clay slurries in previous studies [e.g. O'Brian and Julien, 1988; Phillips and Davies, 1991; Coussot, 1995; Marr et al., 2001; Bardou et al., 2007]. Marr et al. [2001] found that for subaqueous gravity flows, low water content and high clay contents resulted in coherent debris flows, but for coherent flows to form only 0.7 to 5% (by weight) of bentonite clay (a 2:1 swelling clay) was required, as opposed to 7 to 25% kaolinite clay (a 1:1 non-swelling clay), for 25 to 40% water (by weight) and the remainder sand. Bardou et al. [2007] used large-scale rheometers to test mixtures with grains up to 20 mm in diameter, where they altered the clay fraction content from kaolinite to bentonite. They found a general trend of increasing yield stress with increasing solid concentration of the bulk sample, however, below a solids concentration of 78% the samples with bentonite had yield stresses 1.5 to 4 times greater than the kaolinite samples. Above 78%, the samples with the largest amounts of swelling clay (0.014 and 0.021 g swelling clay/g bulk sample) actually had a lower yield stress than the kaolinite or bentonite sample with 0.007 g/g. The authors attribute this to measurement difficulties and an inability to obtain a homogeneous mixture at the highest solids concentration. Coussot [1995] looked at two non-swelling clay types of differing size and found that kaolinite, with a maximum grain diameter of 15 microns, had yield stresses up to an order of magnitude larger for a given solids concentration than those for illite, with a maximum grain diameter of 40 microns. The author does not specifically address why grain size might affect rheology, as he was mainly looking at the effect of solids concentration on the transition from Newtonian to non-Newtonian behavior, but grain size appears to be important.

Although there are abundant field observations and physical and mathematical models describing debris flow behavior (with implications for flow mobility and deposition processes), the dominant mechanisms behind sustained grain support in granular debris flows and their

deposits are poorly understood, and we cannot yet predict the conditions under which the fluid can fully support the coarse fraction. These predictions are important not only for hazard mitigation, which requires knowledge of the volume and size of material in a flow for designing retention barriers and basins [e.g. Hungr et al., 1984], but also for industrial applications, such as slurry transport [e.g. Senapati et al., 2011]. Our goal here is to present a comprehensive conceptual model of grain-fluid interactions in a sheared flow as well as a static deposit and explore the implications for pore pressure generation and dissipation, grain support, and deposit stratigraphy. We base this model on our own experimental results from a large vertically rotating drum, as well as field and experimental observations from others. One element of this model is that the presence of fine material is a necessary condition for the generation and maintenance of elevated pore pressures (i.e. in excess of hydrostatic) in granular flows. We also hypothesize that a wide grain size distribution of the coarse fraction (as opposed to a single grain size) will more effectively ‘trap’ pore fluid, leading to increased pore fluid pressures and an increase in the time taken for the transmission of pressure fluctuations. In recent experimental work by Haas et al. [2014], they found that debris-flow composition had a larger effect on run-out distance than topography, where run-out distance showed an optimum relation with coarse-material and clay fraction. They also suggest there is a clay content “optimum” where small amounts of clay enhances the volume and viscosity of the fluid, thereby liquefying the flow, whereas too much clay increases the yield strength and viscosity to the point of reducing the run-out distance. As such, both the fines content and coarse grain size distribution are varied in our physical experiments to test these hypotheses and further investigate the link between debris flow composition and flow behavior.

3.3 Experimental approach

We performed our experiments in a 4 m vertically rotating drum, and for each granular mixture we monitored the qualitative behavior of the flows (namely grain segregation), and then we made quantitative measurements of the grain velocity and fluid pore pressure with depth, the basal fluid pore pressure, the normal force, and the long profile of the flow [Kaitna et al., *in prep*]. Secondly, we performed a set of static experiments in which we stopped the drum after the dynamic measurements were made and monitored the dissipation of the basal pore fluid pressure with time, as well as the transient fluid pressure response to a boulder emplaced on the static mass. The focus of this chapter is on the static experiments, though some discussion of the dynamic data is necessary for continuity. We also analyzed the rheology of the fluid phase, which has important implications for grain settling.

3.3.1 Experimental Debris flows in a Rotating Drum

Because we were measuring vertical velocity and pore fluid pressure profiles as a function of fines content and grain size in our dynamic experiments, we needed to generate a flow that could be maintained for an extended period of time. In addition, we wanted to use realistic grain size distributions and avoid scaling issues [as discussed in Iverson and Denlinger, 2001]. Hence, we performed our experiments in a four-meter vertically rotating debris flow flume, located at the Richmond Field station in Berkeley, CA, and shown in Figure 3.1. The drum is 80 cm wide and the channel bottom is roughened by rubber treads that are 25 mm tall, 25 mm wide, and spaced 20 cm apart. The instrumentation of the drum includes a laser profiler (for dynamic flow depth), a load plate (for normal force), two basal pore fluid pressure sensors, and a

video camera to monitor flow dynamics. We also instrumented two different probes to measure the grain velocity with depth [Kaitna et al., in prep] and the pore fluid pressure with depth. We describe the pertinent details of the sensors used in the static experiments, as well as the data processing methodology, but more details on the other components of the drum can be found in Hsu et al. [2008], Kaitna and Rickenmann [2007], Hsu [2010], Schneider et al. [2011], and Kaitna et al. [in prep].

The flow geometry is obtained along the flow centerline, as well as 20 cm away from both walls of the drum using a laser profiler (Acuity AR4000). A rotating mirror sweeps the laser along the long profile of the flow at 5 Hz for one minute, yielding 300 individual profiles that are averaged and binned over a 1 degree swath with a vertical resolution of ± 2 mm. In order to make velocity field measurements on the surface of the flow we instrumented the drum with a Canon Vixia S20 high-definition camcorder, which records the flow surface at 30 frames per second. A 15x15 cm load plate (Interface Force Model SWP10-5KB000) measures normal forces on the bed at a frequency of 200,000 Hz, which is then averaged to 1000 Hz. Positioned 86.3 degrees from the load cell are two pressure transducers, like those described in Kaitna and Rickenmann [2007], one at the centerline and one 10 cm from the outer sidewall. The transducers are piezo-resistive pressure transmitters (Keller PR-25Y) that measure pressure fluctuations in an oil filled reservoir that is capped with a flexible membrane. Steel mesh with 2 mm spacing protects the membrane, which is in contact with both the pore fluid in the drum and the oil in the reservoir. The mesh size imposes a constraint on which grains are considered part of the fluid [<2 mm], which is similar to Iverson's [1997] criteria for distinguishing the effective fluid and solid fractions in debris flows.

3.3.2 Experimental Runs in the Drum: Materials and Procedure

For our pore pressure dissipation experiments, we used four different mixtures. The grain size distributions are shown in Figure 3.2. Our 'wide-fines' distribution (solid line) is based on an actual debris flow deposit from the Tungman debris flow in Taiwan [Chen et al., 2001; Hsu, 2010], though our maximum grain size is 81 mm, whereas the Tungman flow had up to meter sized boulders in its deposits. We then removed the fines (defined as grains < 2 mm) but kept the wide coarse grain size distribution (our 'wide-water' distribution, dash-dot line). To see the effect of the coarse grain size distribution, we also replaced the coarse fraction with its D_{50} , 10 mm gravel (PAMI gravel), and ran two runs, with and without fines (dashed and hatched lines respectively). The fine fraction consists of kaolin clay, sieved to a mean grain size of 0.0015 mm (Hydrite 121, Pacific Coast Chemical) and 0.0077 mm (DB Float, Pacific Coast Chemical), and ground silica, sieved to 0.022 mm (Silcosil 125, Pacific Coast Chemical). In order to avoid plugging the pressure sensors, grain sizes between 1.5 and 3 mm were left out of all our distributions.

For all four runs, we kept the concentration by volume of coarse solids ($C_{v,coarse}$) constant at approximately 0.60. Due to porosity differences between the 10 mm well-sorted gravel and the poorly sorted gravel mixture, the total mass of coarse material used per experiment ranged from 533 kg to 727 kg. Table 3.1 shows an overview of the mixture properties. Prior to each run, the drum was loaded with the coarse grain fraction. The fluid fraction, defined as either water or water-fines, was prepared in a large mixing tank and then added to the drum. The drum was rotated for 60 seconds at 1.25 m/s to thoroughly mix the coarse and fluid fractions, after which

we measured long-profiles, vertical velocity and pore pressure profiles, and normal forces on the bed. The basal pore pressures were measured for 60 seconds while the drum was spinning, after which the drum was slowly allowed to come to a complete stop. Upon stopping, the drum was positioned such that the basal pore pressure sensors were located at 6 o'clock, allowing us to continue monitoring pore pressure dissipation with time in the main body of the flow. We recorded basal pore pressures for 30 to 180 minutes before placing a 6.7 kg boulder upon the debris flow mass, which allowed us to see how fluid pressures respond to an external force.

3.3.3 Data Processing

Because the basal pore pressure sensors are attached to the drum bottom, and hence rotate with the drum, they are closed system transducers with an internal calibration (as opposed to differential pressure transducers, which measure the difference in pressure between two points). The raw data for the dynamic experiments was smoothed and filtered using the methods developed by Hsu [2010] and Schneider et al. [2011]. For the dynamic experimental runs the data were smoothed and a sine function was fit using two turning points (-90° and 90° positions) and a minimum (180° position) for each rotation; this generated sine wave was then subtracted from the original signal. Essentially, our processing methods set the probes to atmospheric when the pressure transducer is not passing under the flow mass. This processing method becomes a problem when the sensors are used under the static mass, as we cannot calibrate them to atmospheric. For this reason, we need to set the initial pore fluid pressure upon cessation of the drum equal to the average dynamic pore fluid pressure during rotation, scaled by the flow height, which we discuss below. Typically, upon rotation, the flow forms between -15° and 90° , but its position can vary with water and fines content. When the drum is rotated for 60 s at 1.25 m/s, the pore pressure sensors pass under the flow three times, collecting a total of 84 data points per 1° interval. To calculate the dynamic pore pressure, p_d , per 1° interval, we took the average of all 84 data points. The corresponding dynamic flow depth, h_d , was calculated from the 2-D laser scanner positioned over the center of the flow (and 10 m from the outer wall, corresponding to the side pore pressure sensor). A total of 300 sweeps were recorded per minute, giving ~ 167 data points per 1° interval. We cutoff all flow data past 70° due to a laser shadow caused by our curtain. The curtain was needed to protect the laser from splashing, but as a consequence, we do not have depth data for the very rear of the flow.

To calibrate our static pore fluid pressure data using our smoothed and calibrated dynamic pore fluid pressure data, we first define the dynamic drum pressure, p_d , as

$$p_d = \rho_x g h_d + p_c \quad (3.1)$$

which is the sum of the static ‘fluid’ pressure (first term on right where ρ_x is defined in Equation 3.3) and the centripetal force associated with rotating particles (p_c). The drum is rotated at a relatively low speed so that the effect of centripetal acceleration on the measured values is low [cf. Hsu, 2010 and Kaitna et al., *in prep*]. We can non-dimensionalize Equation 3.1 by the total normal pressure, which is the pressure imposed by the weight of the entire bulk (b) mass (i.e. water, fines and all coarse material), $\rho_b g h_d$.

$$\overline{p_d} = \frac{\rho_x g h_d + \alpha v^2}{\rho_b g h_d} \quad (3.2)$$

If $\overline{p_d}$ in Equation 3.2 is 1.0 or greater then the fluid is supporting the entire mass. If it is less than one, some fraction of the solid phase is being supported by grain-grain or grain-boundary interactions. In Equations 3.1 and 3.2, we define ρ_x as

$$\rho_x = \frac{\rho_f V_f + \rho_{ss} V_{ss}}{V_f + V_{ss}} \quad (3.3)$$

where V_f is the volume of water and all grains less than 2 mm and V_{ss} is the volume of all solids being “supported” by the fluid phase (i.e. grains are either suspended in the fluid phase or are moving at terminal settling speeds). The volume of solids being supported by the fluid is not known *a priori*, but can be back calculated from measured pore fluid pressures. Just upon stopping, any dynamic pressure due to fluid motion ceases, and the non-dimensional dynamic drum pressure becomes the non-dimensional static pressure for an instant, before the static pore pressure begins to decline.

$$\overline{p_d} = \overline{p_s} = \frac{\rho_x g h_s}{\rho_b g h_s} \quad (3.4)$$

We can substitute Equation 3.2, where $v=0$, into Equation 3.4 to get

$$\left[\frac{\rho_x g h_d}{\rho_b g h_d} \right]_{dyn} = \left[\frac{\rho_x g h_s}{\rho_b g h_s} \right]_{stat} \quad (3.5)$$

$$\left[\frac{p_x}{h_d} \right]_{dyn} = \left[\frac{p_x}{h_s} \right]_{stat} \quad (3.6)$$

Where $p_{x,dyn} = \rho_x g h_{dyn}$ (left-side numerator of equation 3.5) and $p_{x,stat} = \rho_x g h_{stat}$ (right-side numerator of equation 3.5).

We then solve for $p_{x,stat}$ at the cessation of motion in Equation 3.7

$$[p_x]_{stat} = \frac{[p_x]_{dyn} h_s}{h_d} \quad (3.7)$$

Our pore pressure signals are recorded at a frequency of 1000 Hz, and for our particular data collection system we are limited to outputting a maximum of 60 seconds of data at that frequency. We were also limited by the fact that for the velocity at which we ran the drum, 1.25 m/s, we could only collect a maximum of 84 data points per 1° interval. In order to get as much dynamic pore pressure data as possible, while at the same time not overloading our data collection system, we stopped collecting pore pressure data as the drum was being slowed to a stop. As soon as the sensors were positioned at, or nearly at, 0°, we re-started our pore pressure

data collection. We are using the 1° degree binned dynamic depth and pore pressure data at the stopping location of the pore pressure sensors, which typically were between 0 and 5°. We calculated the initial static pore pressure values upon stopping using Equation 3.7, and divided this value (from Equation 3.7) by the initial value output by our sensor. In all cases, this ratio ranged from 0.97 to 0.99, indicating that the sensors did not drift much over the course of transitioning from the dynamic to static experiments.

3.3.4 Fluid Rheology

In order to quantify the effect of fluid strength on the coarse particles with regards to particle settling, we measured the rheology of our fines-rich fluid (with a concentration of volume of solids of 0.49) using a cone-plate rheometer (Haake Rheoscope1). The cone has a bevel angle of 4° and a minimum gap thickness of 139 μ m, larger than the largest particle diameter within the fluid (see Figure 3.2 for grain size distribution). We measured the response of the fluid under steady shear only. The unconfined sample was loaded and the applied stress as a function of strain rate was measured. The strain rates varied from 10^{-4} to 10 s⁻¹. The measurement lasted for 153 seconds, at a constant temperature of 24.8 ± 0.01 °C. The results are shown in Figure 3.3. When fit using a Hershel-Bulkley model, the yield strength, τ_y , is ~ 5 Pa and the power law exponent, n , is less than one (i.e. 0.4), suggestive of a shear thinning fluid.

3.4 Results and Discussion

3.4.1 Drum Data

Here we present our results for the four experiments in which we collected static pore pressure data. Figure 3.4a-d shows the first 600 to 1000 s for each of the four experiments, where the fines-rich mixtures are shown in the upper right and left hand corners (3.4a and 3.4b), and the water mixtures are shown in the lower right and left hand corners (3.4c and 3.4d). The wide coarse grain size distribution was used in Figures 3.4a and 3.4c and the uniform grain size distribution (i.e. 10 mm gravel) was used in Figures 3.4b and 3.4d. In each plot, time is on the x-axis and the measured pore pressure, normalized by the total normal pressure, is on the y-axis (Equation 3.2 where $v=0$). The total normal pressure is defined as the pressure imposed by the weight of the entire mass (water, fines and all coarse material), sometimes referred to as the basal normal stress [Iverson, 1997]. The measured dynamic pore pressure is plotted from -60 to 0 seconds, after which the normalized static pore pressure is plotted for both the center and side sensors. For reference, the normalized hydrostatic pressure is also plotted. For the ‘wide-fines’ mixture (Figure 3.4a), the normalized dynamic pore pressure for the center sensor was on average 0.95 and for the side sensor 0.89, meaning that most of the weight of the mass was being supported by the fluid. A normalized pressure of 1.0 means the entire mass (water, fines, and all coarse material) is supported by the fluid phase, and anything less than 1.0 means that some of the weight of the mass is being supported by grain-grain or grain-boundary contacts. If the measured normalized pore pressure is close to hydrostatic (0.44 for the fines-rich mixtures, Figures 3.4a and 3.4b, and 0.50 for the water mixtures, Figures 3.4c and 3.4d), then the solid fraction is being supported entirely by grain-grain or grain-boundary contacts. Our ‘D50-fines’ mixture (Figure 3.4b) had slightly lower normalized pore pressures than the ‘wide-fines’ mixture, 0.90 and 0.88 compared to 0.95 and 0.89, for the center and side sensors, respectively. When the

finer are removed (Figures 3.4c and 3.4d), we see that for both the wide and uniform coarse grain size distributions, the measured normalized pore pressure is close to hydrostatic immediately upon stopping. For the wide coarse grain size distribution (Figure 3.4c) the dynamic pore pressure was above hydrostatic (0.55 versus 0.5 for both sensors), and it took about 1000-s for the pore pressure to drop to hydrostatic. For the D50-water mixture (Figure 3.4d), the pore pressure was barely above hydrostatic for the entire experiment, with both sensors rising to 0.52 after stopping the drum and then declining within the first 200-s to 0.51 and 0.505 for the center and side sensors, respectively.

For the mixtures with fines, there is a noticeable pressure gradient between the center and side sensors, as compared to the water mixtures, where the center and side sensor readings are barely distinguishable from one another (one reason that the center sensor may be elevated relative to the sides could be due to higher boundary friction effects on the side of the drum). We hypothesize that the increased fluid viscosity due to the presence of fines prevented the rapid dissipation of pore pressure, as diffusivity is inversely proportional to viscosity. Changes in permeability, which is directly proportional to diffusivity, can also alter pore-fluid hydraulics, which we see when comparing Figures 3.4a and 3.4c to 3.4b and 3.4d, respectively. The wide grain size distribution with fines leads to higher overall pore pressures and a larger pressure gradient between the center and side pressure sensors than the D50 with fines, which can be accounted for by increased packing and reduced permeability, both of which prevent the viscous fluid from flowing through the gravel matrix pore space. In the case with just water, it also appears that the wide grain size distribution (Figure 3.4c) leads to higher dynamic pore pressures and a longer relaxation time in returning to hydrostatic pore pressures, as compared to the 10 mm gravel (Figure 3.4d), where pore pressures essentially remained hydrostatic. If elevated pore pressures were purely the result of yield strength due to the addition of fines, we might expect that for both the ‘wide-water’ and ‘D50-water’ case, the pore pressures would both remain hydrostatic, but the grain size of the coarse grain-size may also need to be accounted for.

Figure 3.5 shows the three major components contributing to the total measured pore pressure at the center pressure sensor, normalized by the total measured pore pressure, as a function of time for the same four mixtures described in Figure 3.4. The pore pressure contributions due to the weight of water (P_{water}) and the weight of the fines plus water (P_{susp}) are calculated using Equations 3.8a and 3.8b, where $\cos\theta$ takes into account the local surface slope. The density of water is 1000 kg/m³ and the density of the fluid was calculated based on the amount of fines added (see Table 3.1).

$$P_{water} = \rho_{water}gh_s\cos\theta \quad (3.8a)$$

$$P_{susp} = (\rho_{fluid} - \rho_{water})gh_s\cos\theta \quad (3.8b)$$

The difference between the total measured pore pressure and the pore pressure contribution from water and fines is defined as the excess pore pressure, P^* .

$$P^* = P_{measured} - (\rho_{fluid}gh_s\cos\theta) \quad (3.9)$$

For the fines-rich mixtures (Figure 3.5a and 3.5b) ~18 to 20% of the measured total pore

pressure is in excess of pressure contributions from the fluid pressure, where the fluid is the water and the suspended fine material. Essentially, this is the pressure contribution in excess of what one would predict based on density effects alone. This means that the fluid, in addition to supporting its own weight, is also partially supporting the weight of the coarse material. In comparing Figure 3.5a and 3.5b to 3.5c and 3.5d, excess pore pressures are higher for the wide grain size distribution (~ 0.20) as compared to the 10 mm uniformly sized gravel (~ 0.18). When the fine material is removed, any excess pore pressure is negligible for the uniform grain size distribution (Figure 3.5d), and decreases from approximately 9% down to 2% during the first 500 s for the wide grain size distribution (Figure 3.5c).

In Figure 3.6, we show the three experiments in which we placed a 6.7 kg boulder, which was 260 mm in diameter, on top of the static mass, directly over the center pore pressure sensor. Figure 3.6a is the ‘wide-fines’ mixture, Figure 3.6b is the ‘D50-fines’ mixture, and Figure 3.6c is our ‘D50-water’ mixture. In each plot the measured pore fluid pressure, normalized by the total normal pressure, is shown as a function of time. The data are plotted for both the center and side sensors. In the case of the fines-rich mixtures, when the boulder was placed on the static mass, the pore fluid pressure at *both* sensors rose instantly. If we assume that the boulder exerts a force over an area equal to the area of the base of the boulder (which was measured to be 0.058 m^2), then for a 6.7 kg boulder, it would take 1133 Pa to support the boulder entirely. As both the center and side sensors registered the same magnitude rise in pressure at the same time, using the area of the boulder is a conservative estimate, as the boulder is likely exerting a force over a larger area. If we normalize the theoretical pressure exerted by the boulder by the total normal pressure, we would expect to see a 0.15 rise for the ‘wide-fines’ mixture, a 0.16 rise for the ‘D50-fines’ mixture, and a 0.22 rise for the ‘D50-water’ mixture if the entire weight of the boulder is supported by the fluid. For the ‘wide-fines’ mixture (Figure 3.6a) we see a 0.08 rise, for the ‘D50-fines’ mixture (Figure 3.6b) we see a 0.06 rise, and for the ‘D50-water’ mixture (Figure 3.6c) we see no rise. This suggests that for the fines-rich mixtures, the fluid supports $\sim 53\%$ of the boulder’s weight with the wide grain size distribution, while $\sim 38\%$ is supported with the uniform grain size distribution. The weight of the boulder is supported entirely by grain-grain contacts when the fines are removed, as there is no change in fluid pressure for the D50-water case. For the ‘fines-rich’ mixtures (i.e. ‘wide-fines’ and ‘D50-fines’), the rate of dissipation before the boulder was added was approximately $9.3 \cdot 10^{-6} \text{ Pa/s}$ and $1.4 \cdot 10^{-5} \text{ Pa/s}$ for the wide and uniform grain-size distributions respectively, and after the boulder was added they increased to $2.8 \cdot 10^{-5} \text{ Pa/s}$ and $2.5 \cdot 10^{-5} \text{ Pa/s}$, respectively. For the ‘D50-water’ case, the rate of dissipation remained constant at $2.5 \cdot 10^{-5} \text{ Pa/s}$.

As shown in Figure 3.3, the fluid also exhibits a yield strength, which needs to be taken into account, as it affects which grains will move through the fluid. Taking a simplistic approach by assuming the grains are spherical, the buoyancy force, F_B , can be calculated for a given grain size using Equation 3.10

$$F_B = \frac{4}{3} \pi (\rho_s - \rho_{fluid}) g R_o^3 \quad (3.10)$$

where ρ_s is the density of the sediment (2650 kg/m^3), ρ_{fluid} is the density of the fines-rich fluid ($\sim 1800 \text{ kg/m}^3$) and R_o is the radius of the sediment. The sphere will sink when the yield stress parameter, Y_g ,

$$Y_g = \frac{2\tau_y \pi R_o^2}{F_B} \quad (3.11)$$

is less than 0.143 [Beris et al., 1985]. Solving for the critical radius at which grains will move, R_c ,

$$R_c = \frac{2\tau_y}{0.143(4/3)(\rho_s - \rho_{fluid})g} \quad (3.12)$$

we find that grains ~12 mm in diameter or less will be supported by the fluid (which is the median grain size of both mixtures), and those greater than 12 mm will sink.

One explanation for the results seen in the fines-rich boulder experiments, where the boulder would be predicted to sink following Equation 3.12, is that when the boulder is placed on the static mass, there is some re-arrangement of the coarse-grained network underneath it. As the fluid cannot respond instantaneously, and the coarse-grained particles rearrange suddenly, the fluid is squeezed and accelerated in and out of pore spaces within the network, leading to local instantaneous fluid pressure changes [Iverson and LaHusen, 1989]. When the fluid is viscous and pore spaces are smaller (as may be the case with the wide grain size distribution compared to the D_{50} only distribution), it may take longer for the fluid to move in response to a local increase or reduction in pore pressure gradient, and subsequently a local elevated pore pressure may be maintained over a longer time. Under the right conditions, Pierson [1981] showed that the average pore pressure could be elevated above the hydrostatic case for timescales greater than that of a debris flow event. We can also imagine that as the fluid is being displaced, some of the particles may get pushed upward, and these individual particles, depending on their size, may take some time to settle. Before they settle back into contact with other particles, they may be partially or entirely supported by the fluid rather than other solid particles alone. In these cases, a reaction force, effectively an excess fluid pressure, must be generated to support the solid particles that cannot be supported by yield strength alone within the mass.

3.4.3 Conceptual model

Based on the observations presented here, as well as those presented in the literature [i.e. Pierson, 1981; Beris et al., 1985; Iverson and LaHusen, 1989; Major, 1996; Iverson, 1997; Major, 2000; McArdell, 2007; Berzi et al., 2010; Kaitna et al., 2012; Kaitna et al., *in prep*], we propose a conceptual model for the generation and dissipation of excess fluid pore pressures in both sheared flows and static deposits. We illustrate this model in Figure 3.7a and 3.7b, where 3.7a is an example of a sheared flow consisting of a viscous fluid and coarse grains, flowing from left to right, and 3.7b is the same mixture upon deposition. In a sheared flow, grains interact with each other via grain-grain collisions, in which one grain imparts momentum to another grain, or by grain-grain contacts, in which one grain rests on another grain. The latter may occur more at the lateral margins of the flow, where there is greater frictional resistance than at the center of the flow. These coarse grain interactions occur in, and are modified by, the presence of a viscous, interstitial fluid. Often, the viscous fluid is non-Newtonian due to the presence of fine material, as is the case in our experiments (see Figure 3.3), likely strongly controlled by the clay size fraction [e.g. Coussot, 1995]. As coarse grains move through this viscous fluid, the fluid rheology determines which particles will settle and the rate of settling for those grains, as

increasingly viscous fluids will increase the drag force experienced by each individual grain. If the fluid exhibits yield strength, some grain size fractions may not settle through the fluid as the fluidity of the material adjacent to a grain is the result of stresses created by the grain's own weight. Only when there is development of a slip surface does the grain begin to settle [Beris et al., 1985]. When coarser grains do settle back into contact with other particles, they may be partially or entirely supported by the fluid rather than other solid particles alone, such that an excess fluid pressure must be generated (in order to support the solid particles). In a sheared flow, grains may experience a 'push' by the fluid that is squeezed out of constricting pore spaces due to grain shearing and settling, such that settling forces are being transmitted some distance away (i.e. 'force at a distance'). While the flow is dynamic, it can be at steady state, and a net contraction or settling of the particles is not necessary for the generation of dynamic excess pore pressures. Excess fluid pore pressures may therefore be the result of both a buoyancy-like effect (due to the fluid having a yield strength) and the result of the cumulative effect of grains settling against a highly viscous fluid that can sustain and transmit settling forces through out the flow, as the fluid itself slowly advects between the grains due to local pressure gradients induced by the re-arrangement of the coarse grains through shearing and grain collisions. Upon deposition, the grains are no longer colliding and shearing past one another, though many of the coarse particles may be supported by the viscous fluid, as indicated by Figure 3.6a and 3.6b. Depending on their size and the fluid rheology, some of these particles will slowly settle, and remain part of the fluid phase until coming to rest on a boundary or another grain-grain chain that is ultimately resting on a boundary. As in the sheared flow case, the movement of grains as they settle through the viscous fluid can be impeded by the fluid itself, the presence of other grains, the upward flow of fluid resulting from the downward motion of neighboring grains, and forces transmitted at a distance as fluid squeezes through the granular pore space due to imposed local pore pressure gradients. Fluid slowly drains from constricting pore spaces and flows upward, and smaller and smaller grains settle out, further decreasing the permeability of the granular matrix. The combination of a highly viscous fluid phase and the presence of poorly sorted coarse material prevents a well-graded deposit from forming, even though one may expect larger grains to preferentially settle faster than smaller grains.

3.5 Conclusions

Based on our simple experiments, we observed that mixtures with fines contents close to those found in a real debris flow deposit [i.e. Chen et al., 2001] generated elevated pore pressures upon mixing in the drum. Normalized measured pore pressures ranged from 0.85 to 0.95, indicating that the mixtures were close to liquefaction (a normalized pore pressure of 1.0 indicates that the entire coarse fraction is being fully supported by the fluid). These pore pressures remained elevated, often for the entire duration of the experiment, which could last several hours. While the presence of fines was first order in leading to the generation and maintenance of elevated pore fluid pressures, the highest excess pore pressures were generated with the combination of fine material and a wide coarse grain size distribution. We hypothesize that this is most likely due to a combination of increased fluid viscosity, which prevents the fluid from rapidly advecting through the granular pore matrix in addition to slowing grain settling due to increased drag and buoyancy, a decrease in permeability of the granular pore matrix due to greater packing as the grain size distribution becomes less uniform [Rodine and Johnson, 1976], and fluid yield strength, which can support grains up to 12 mm in size. When the fines are

removed completely, even in the case with the wide coarse grain size distribution, the pore fluid pressure was nearly hydrostatic for the entire duration of the run, and any elevated pore pressures quickly dissipated. Normalized measured pore pressures ranged from 0.5 to 0.54, where 0.5 is hydrostatic, meaning the coarse fraction is being supported entirely by grain-grain and grain-boundary contacts. For the fines-rich mixtures, approximately 20% of the measured total pore pressure was in excess of what one would predict based on density effects alone. This means that the fluid, in addition to supporting its own weight, is also supporting the weight of particles suspended in the fluid (i.e. grains ~ 12 mm or smaller) and slowly settling coarser material. With the fines-rich deposits, the weight of an emplaced boulder was partially supported by the fluid, almost 53% in the case with the wide grain size distribution and 38% with the uniform grain size distribution, using a conservative estimate for the area over which the weight of the boulder acts. In the gravel-water deposits, the weight of the boulder was supported entirely by grain-grain contacts. These observations lead us to conclude that dynamic interactions between solids settling in the fluid and the fluid “pushing back”, in addition to the yield strength within the fluid, may result in sustained excess pore fluid pressures in debris flow deposits.

3.6 Acknowledgements

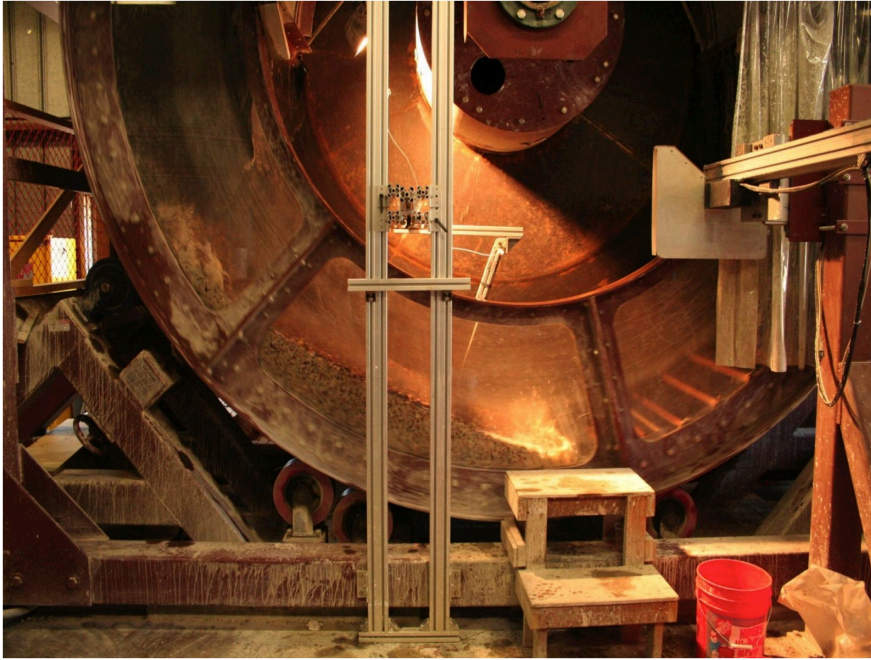
The authors would like to thank Leslie Hsu, Stuart Foster, Roland Kaitna, Bereket Yohannes, Kimberly Hill and Leonard Sklar for insightful discussions and help with the drum experiments. We also thank Michael Manga and Max Rudolph for training and use of the Haake Rheoscope1 for our rheology measurements. This work was supported by the STC program of the National Science Foundation via the National Center for Earth-surface Dynamics under the agreement EAR-0120914.

3.7 Tables

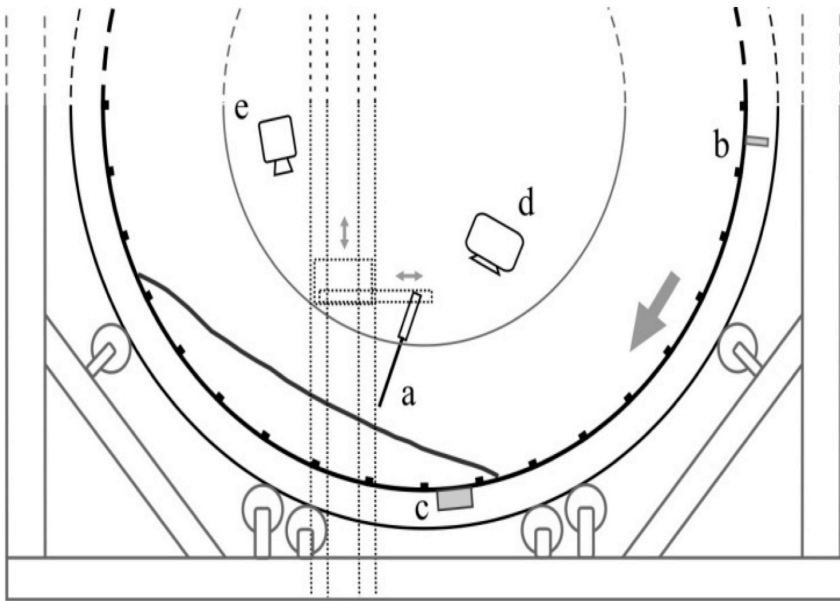
Parameter	Units	Wide- fines	Wide-water	D50- fines	D50-water
$D_{50, \text{coarse}}$	mm	10	10	10	10
m_{coarse}	kg	727	727	533	533
m_{fines}	kg	229	0	94	0
m_{water}	kg	91	177	99	134
$C_{v, \text{coarse}}$	-/-	0.61	0.61	0.60	0.60
$C_{v, \text{fines}}$	-/-	0.49	0	0.26	0
C_v	-/-	0.80	0.61	0.71	0.61
ρ_{bulk}	kg/m^3	2318	2003	2264	1990
ρ_{fluid}	kg/m^3	1804	1000	1760	1000
μ_{fluid}	Pa-s	*	0.001	*	0.001

Table 3.1. Mixture properties for our four experimental runs.

3.8 Figures



a.



b.

Figure 3.1 Photo of the rotating drum (left) and sketch of the instrumented drum (right): (a) velocity probe (vertical velocity profiles), (b) pore fluid pressure sensors (basal pore pressure measurements), (c) load cell (normal forces on the bed), (d) 2D laser scanner (flow long profile), and (e) digital video camera (flow surface velocity and segregation observations) (images taken from Kaitna et al. [*in prep*]). By our convention, 0° is at 6 o'clock, 90° is at 9 o'clock, 180° is at 12 o'clock, and -90° is at 3 o'clock.

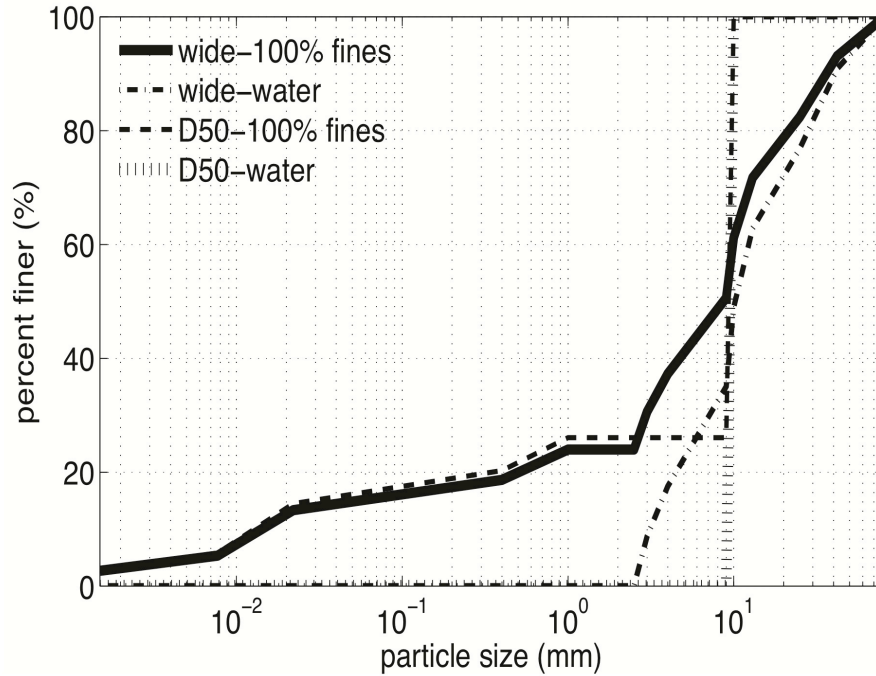


Figure 3.2 Grain size distributions for the four mixtures used in our pore pressure dissipation experiments.

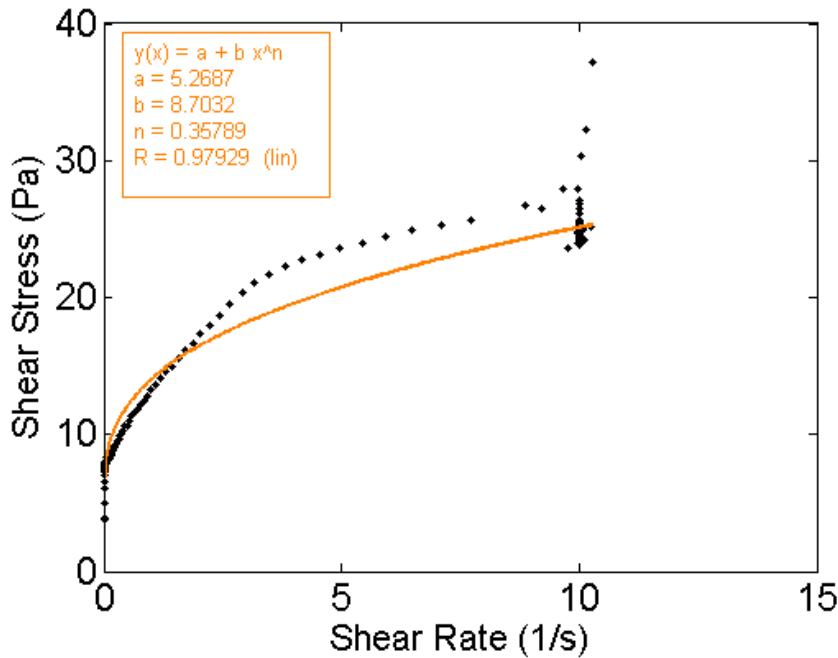


Figure 3.3 Plot of shear stress as a function of strain rate for the fines-rich fluid used in ‘wide-fines’ and ‘D50-fines’ experiments. The concentration of solids per volume was 0.49. The data were fit to a Hershel-Bulkley model where $\tau = \tau_y + k \cdot \text{shear_rate}^n$ where τ_y , the yield strength, is ~ 5 Pa and n is ~ 0.4 , characteristic of a shear thinning fluid.

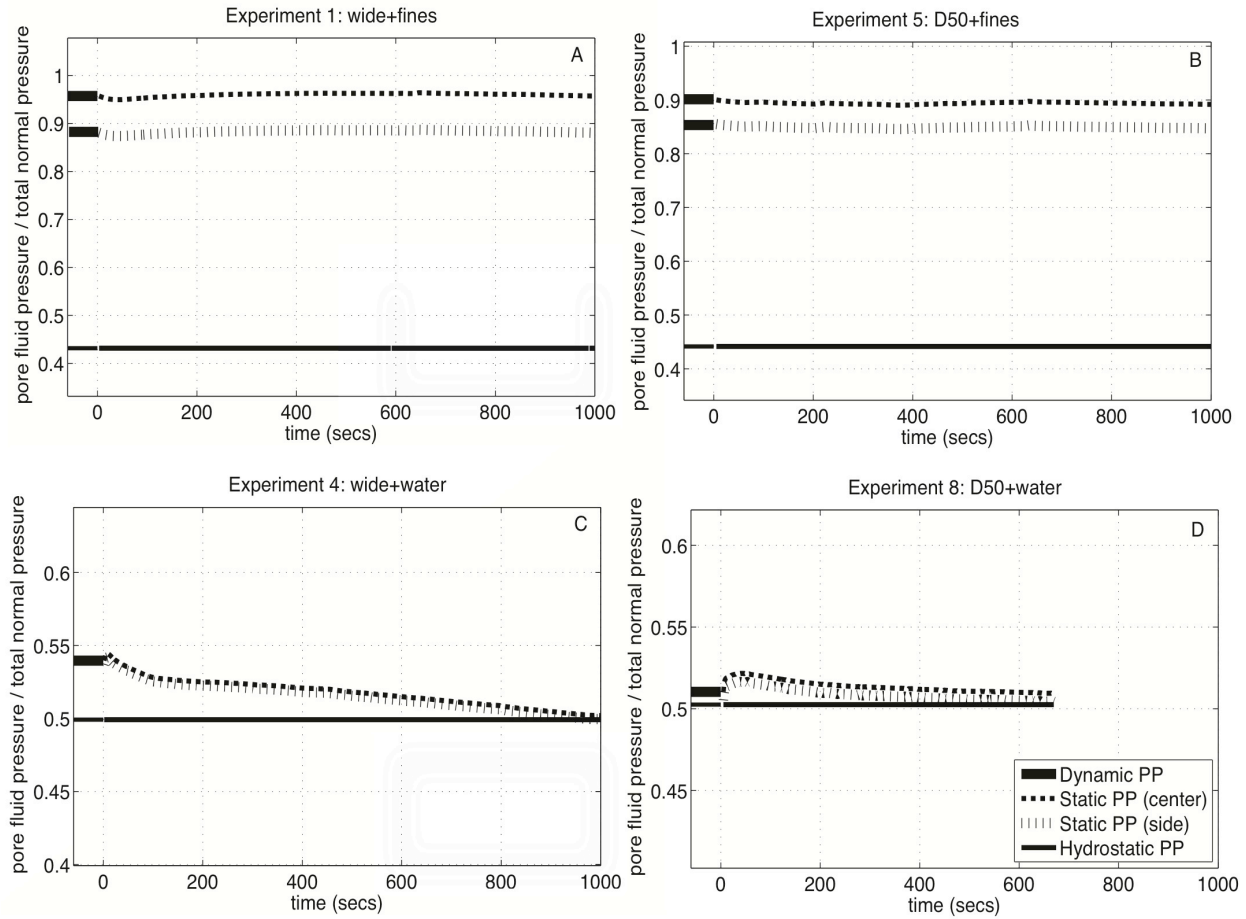


Figure 3.4 Normalized pore fluid pressure as a function of time. The measured pore fluid pressure is normalized by the total normal pressure, which is the product of the bulk density of the entire mass above the sensor (water, fines and coarse grains), its depth, surface slope, and gravity. Both the center and side pressure sensor responses are plotted for when the drum was spinning at 1.25-m/s (from -60 to 0 s) and after it was allowed to come to a stop (a) wide coarse grain size distribution with fines (b) 10 mm gravel with fines (c) wide coarse grain size distribution with water and (d) 10 mm gravel with water.

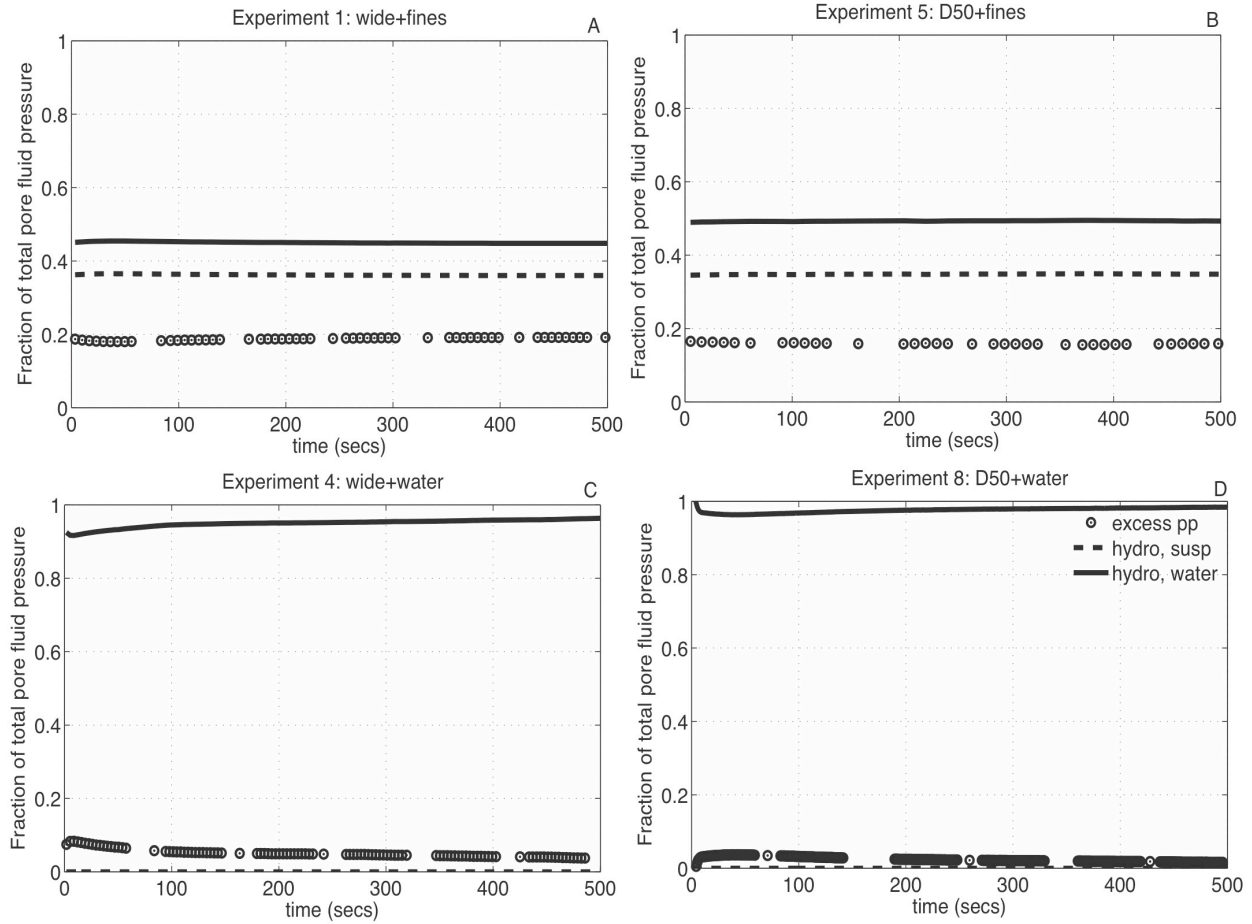


Figure 3.5 Major components contributing to the total measured pore pressure (hydrostatic pressure, pressure due to the presence of fine material, and excess pore pressure) normalized by the total measured pore pressure, as a function of time. (a) wide coarse grain size distribution with fines (b) 10 mm gravel with fines (c) wide coarse grain size distribution with water and (d) 10 mm gravel with water.

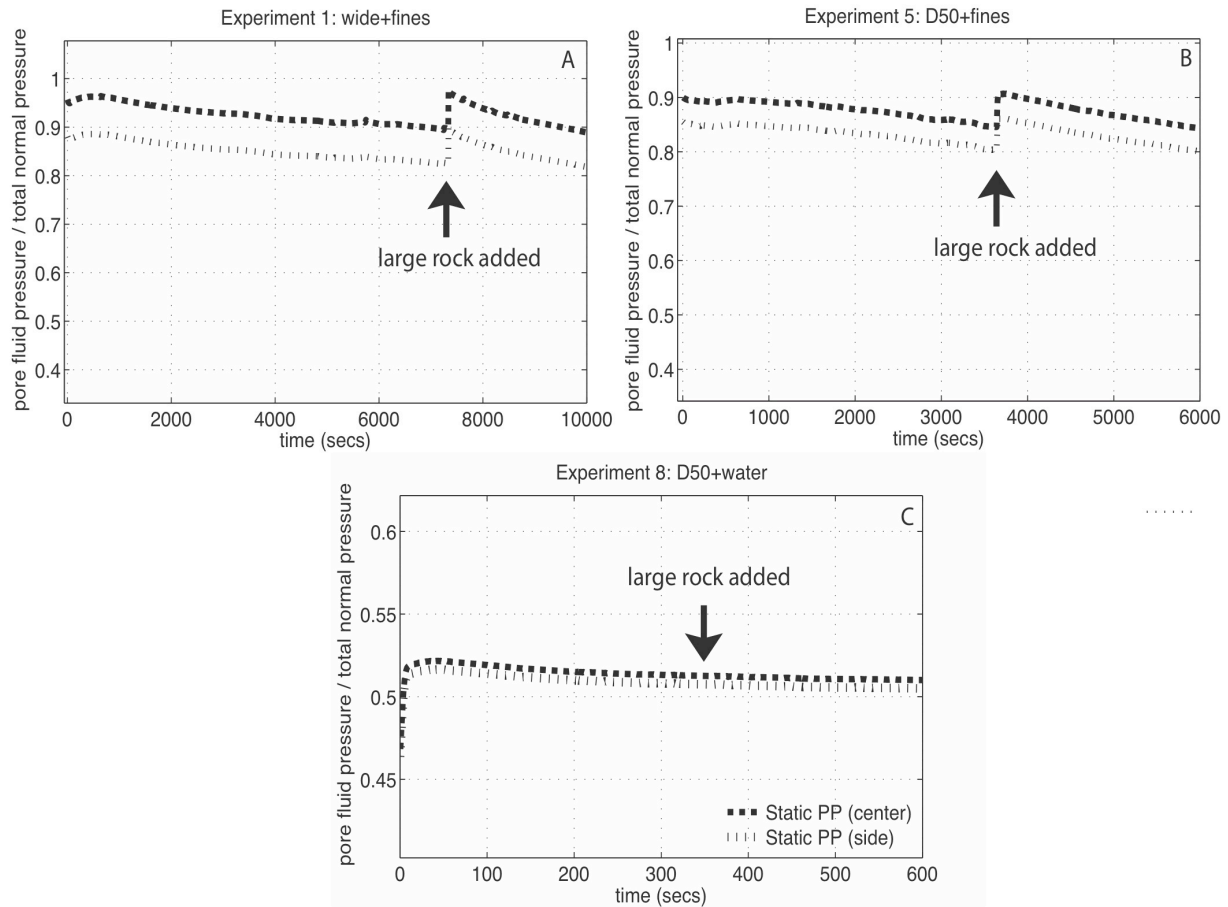


Figure 3.6 Normalized pore fluid pressure as a function of time. The measured pore fluid pressure is normalized by the total normal pressure, which is the product of the bulk density of the entire mass above the sensor (water, fines and coarse grains), its depth, surface slope and gravity. Where indicated by the arrow, a 6.7-kg boulder was placed on the static debris mass directly above the center pressure sensor. Both the center and side pressure sensor responses are plotted. (a) wide coarse grain size distribution with fines (b) 10 mm gravel with fines with water and (c) 10 mm gravel with water. We did not add the boulder to the wide grain size distribution with water.

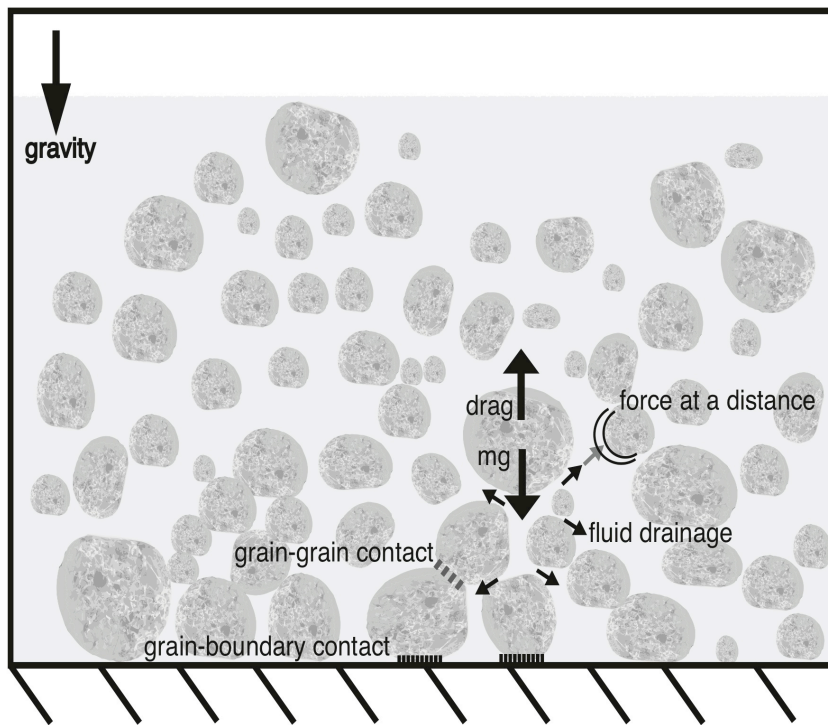
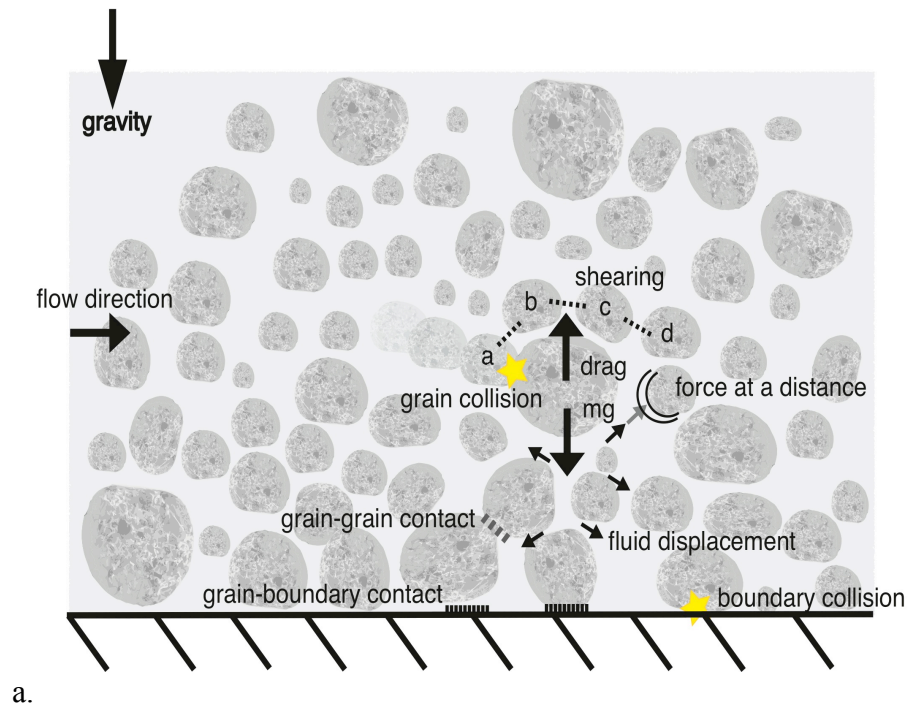


Figure 3.7 Conceptual model of grain-grain and grain-fluid interactions in a (a) sheared flow and a (b) static deposit.

3.9 References

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Chapter 4: The origin and evolution of the Peace Vallis fan system that drains to the Curiosity landing area, Gale Crater, Mars

4.1 Abstract

The landing site for *Curiosity* rover is located at the distal end of the Peace Vallis fan in Gale Crater. Peace Vallis fan covers 80 km² and is fed by a 730 km² catchment, which drains an uplands plains area through a 15 km wide gap in the crater rim. Valley incision into accumulated debris delivered sediment through a relatively low-density valley network to a mainstem channel to the fan. An estimated total fan volume of 0.9 km³ matches the calculated volume of removal due to valley incision (0.8 km³), and indicates a mean thickness of 9 m. The fan profile is weakly concave up with a mean slope of 1.5% for the lower portion. Numerous inverted channels outcrop on the western surface of the fan, but on the eastern portion such channels are rare suggesting a change in process from distributary channel domination on the west to sheetflow on the eastern portion of the fan. Runoff (discharge/watershed area) to produce the fan is estimated to be more than 600 m, perhaps as much as 6000 m, indicating a hydrologic cycle that likely lasted at least 1000's of years. Atmospheric precipitation (possibly snow) not seepage produced the runoff. Based on topographic data, Peace Vallis fan likely overlapped Bradbury Rise and spilled into a topographic low to the east of the rise. This argues that the light-toned fractured terrain within this topographic low corresponds to the distal deposits of Peace Vallis fan and in such a setting lacustrine deposits are expected.

4.2 Introduction

On August 6, 2012, UTC, the Mars Science Laboratory (MSL) rover, *Curiosity*, landed in Gale Crater (5.3°S 137.7°E, Figures 4.1 and 4.2) within a landing ellipse (Figure 4.3) that partially included a hypothesized alluvial fan deposit [e.g. Anderson and Bell, 2010], now named the Peace Vallis fan. The actual landing site (Bradbury Landing) was a few hundred meters east of the distal end of this fan. The MSL team observed that the distal end entered an enclosed topographic low area and predicted that fine-grained sediment, including possibly lake deposits, would be found there. This contributed to the decision to first send *Curiosity* to the east rather than head directly to the base of Aeolis Mons [Grotzinger et al., 2014]. En route, water-transported gravels, outcropping as conglomerates [Williams et al., 2013a] were found and clay-bearing mudstones were discovered in Yellowknife Bay [Grotzinger et al., 2014; Vaniman et al., 2014]. These in situ observations confirm the ability to identify fluvial/alluvial facies from orbital data using geomorphic observations and will provide insight for the 100 other large alluvial fan deposits (> 10 km long downslope) that have been identified on the southern highlands of Mars, most originating in alcoves near crater rims [Moore and Howard, 2005; Kraal et al., 2008; Wilson et al., 2012].

These findings by MSL and previous studies of the other alluvial fans provide strong context for analysis of the Peace Vallis fan. Moore and Howard [2005] showed that the slopes of these crater-rim-derived fans averaged about 2%, were weakly concave up, and combined with fan size and catchment relief, suggested deposition by gravel-bedded rivers. They proposed that the lack of fan-head trenching in these features indicates an abrupt end to climatic conditions favorable to fan generation. Runoff from snowmelt was proposed to be the driver for erosion and fan generation.

Irwin et al. [2005], building upon earlier work by Cabrol et al. [1999], hypothesized that in Gale Crater, specifically, there was a brief period in which highland valley networks were reactivated, drained towards Gale, and breached the crater rim, delivering water that caused lake and delta formation (Figure 4.3). Simultaneously, Aeolis Mons was gullied, the crater wall was cut by gullies and fans formed. This would have occurred, therefore, after the Aeolis Mons sediments were deposited and then partially removed by wind erosion [Malin and Edgett, 2000]. Irwin et al. [2005] suggested the brief hydrologic activity occurred in the late Noachian to early Hesperian.

In the region to the south of Gale Crater, similarly-sized craters appear substantially more degraded, are partially to nearly buried, and have subtle rim topography, indicating that they likely pre-date Gale Crater. Farah Vallis (the large canyon entering Gale from the south, Figure 4.3) may have been part of an extensive regional drainage system (270,000 km²) originating near Hershel Crater (Figure 4.1). Ejecta from Gale Crater may have partly buried this valley network, implying that the network is superimposed by Gale crater and was active before crater formation [Irwin et al., 2005]. It is this network that Irwin et al. [2005] hypothesize was reactivated and subsequently delivered water to Gale Crater. More recently, Newsom et al. (*submitted*) have mapped the valleys superimposed by ejecta and found that the ejecta are generally poorly incised and many of the local slopes of the older valleys were altered by the Gale-forming impact, leading them to conclude that the valleys leading to fan and deltas within Gale crater are fresh and unrelated to past activity.

Anderson and Bell [2010] produced the first detailed geomorphic map of Gale Crater (Figure 4.3), exploiting data from both Mars Reconnaissance Orbiter (MRO) Context Camera (CTX) and High Resolution Imaging Science Experiment (HIRISE). They mapped features they labeled as “valley/canyons” on the crater rim and found that they are commonly branched and many of them along the northern rim lead to “fan-shaped mesas on the crater floor”. Several of their mapped canyons also cut into Aeolis Mons in the center of Gale crater. They propose that the valley/canyons and associated mesas are fluvial features and they interpret the sinuous ridges found on the valley floor to be inverted channels (also referred to as exhumed channels [e.g. Clarke and Stoker, 2011]). In some cases, canyons lead downslope to sinuous ridges and at several locations the sinuous ridges show a sinuosity characteristic of meandering rivers (e.g. a sinuosity of 2.0). Throughout the crater wall and floor, Anderson and Bell [2010] map a unit they label as the “mound-skirting unit” which they describe as a mesa-forming material that is intensely pitted and/or has parallel ridges. The mesas bordering the Peace Vallis fan are shown in Figure 4.3.

Anderson and Bell [2010] divided the 80 km² Peace Vallis fan into an upper, smooth lower thermal inertia unit and a lower, rocky light-toned and fractured, high thermal inertia unit. On the upper fan they noted ridges that they interpreted to be inverted channels or remnants of debris flows. They proposed that the high thermal inertia portion of the fan has been exposed by erosion of the overlying low thermal inertial unit.

Here, we exploit HIRISE and CTX imagery, including photogrammetrically derived digital elevation data, to explore three questions: 1) What is the source of sediment comprising

the Peace Vallis fan, 2) What processes formed the fan, and 3) How did the fan evolve through time? A detailed topographic analysis of Peace Vallis fan and its source area is presented. The eroded volume resulting from valley incision in the source area is then compared with the estimated volume of the fan. These analyses and observations are then collected to explore the three posed questions.

4.3 Methodology

4.3.1 Topographic Data

As part of the entry, descent, and landing (EDL) modeling effort for the MSL mission, comprehensive visual and topographic datasets were assembled for the landing ellipse and Peace Vallis fan in Gale crater [Golombek et al., 2012]. Twelve MRO HIRISE 0.25 cm/pixel stereo pairs (created by the USGS Astrogeology Center in Flagstaff, AZ), three 6m/pixel stereo pairs from the MRO CTX, and a crater-wide 50 m/pixel digital elevation model (DEM) made from Mars Express Orbiter HRSC images were geo-referenced and mosaiced to form the base of a geographic information system (GIS) dataset. These combined datasets form a “resolution pyramid” to allow mapping within Gale crater at scales all the way down to a few meters or less. Datasets were projected into an Equidistant Cylindrical projection with zero as the center longitude and latitude, on a spherical Mars with a radius of 3396190-m. The resulting grid was used to extract 0.25 cm/pixel HIRISE visible, 1m/pixel elevation, and 100m/pixel thermal inertia data [Parker et al., 2013] and these data were provided to the MSL team. Most of the analyses presented here were derived from this topographic dataset (see Table 1). These data were also georeferenced relative to a spheroid, as opposed to a geoid, to get delta radii elevations, which were used mainly by the Entry, Descent, and Landing (EDL) team for EDL simulations [Golombek et al., 2012]. We used this dataset when working with the lower fan and in the vicinity of the landing site/Yellowknife Bay, as it had fewer topographic artifacts than the MSL DEM.

In addition to the DEMs provided to the MSL team by the USGS at Flagstaff, we also created a 1 m resolution topographic dataset using a controlled network of 9 HIRISE stereo images, herein referred to as HCN (HIRISE Controlled Network) DEM. The images were controlled using 280 MOLA shots and 240 tie points spread throughout the HCN DEM boundary (Figure 4.4). The tie points were given z-control (elevation) using a 4-image CTX DEM. This HCN DEM was developed to improve upon the MSL-provided DEM, which had several major artifacts along image boundaries down the center of the Peace Vallis fan, preventing an accurate analysis of the fan’s planform topography. As shown in Figure 4.4, MOLA points along the edges of the HIRISE DEMs are generally sparse, making it difficult to mosaic the various HIRISE DEMs together without topographic artifacts. Several points on the fan were chosen to compare the MOLA elevation values with those extracted from the MSL DEM and our HCN DEM. The MSL model varied by 2-5 m compared to the MOLA data while our HCN DEM was typically within a meter of the MOLA values. In comparing relative elevation differences though, such as when determining the deflation depth of ridges on the fan, both the MSL and our HCN model provided similar results.

For areas outside of the landing ellipse and alluvial fan, CTX and HRSC data were primarily used. A radiometrically calibrated CTX mosaic at 6m/pixel was provided by the MSL team, and was used mainly to identify topographic features. For crater-wide context at Gale, an HRSC-derived topographic map, also provided by the MSL team, at a resolution of 50-m/pixel, was used.

4.3.2 Volume Calculations

Much of the topographic analysis presented here relies on methods similar to those developed by Warner et al. [2011] for basin volumes and Chaytor et al. [2009] for landslide volumes. The volume eroded from the lower mainstem of the valley network in the source area was calculated by projecting a smooth surface between the bounding sidewalls on either flank of the valley, essentially creating a “lid” over the lower incised valley network. This projected surface was then gridded and subtracted from the original gridded topography. Our methodology used ArcGIS tools, though all post-processing was conducted in MATLAB. To estimate the fan volume we first eliminated the topographic data under the fan and projected the topography laterally from the east and west to create a pre- fan surface. We then took a difference between the gridded fan surface and this projected surface to estimate the volume of the fan.

We used three different interpolation methods for determining the pre-fan topography. The first uses a nearest neighbor interpolation using only points within the first 500-m outside of the fan’s edges, the second projects polylines connecting contours on the eastern edge of the fan to the same contours on the western edge, and the third uses the contours on either side of the fan as a guide to the topography beneath the fan. For the second and third interpolations, we assign elevation values to points created from polylines and then use those points, in addition to the points around the fan’s edges, to perform a nearest neighbor interpolation. In the first case the pre-fan topography has an unrealistic bulge where the crater wall meets the floor, the second creates a more planar surface under the fan, and the third creates a concavity (in planform) in the crater wall.

In order to reconstruct the fan topography immediately after its deposition (pre-wind erosion), all of the point elevation data from areas that show extensive erosion were excluded. These excluded data mostly occurred in the distal fan area where extensive wind erosion is evident. Here just the local erosion resistant mesas and outcrops were included to guide construction of the original fan surface. For the fan surface reconstruction we solved Laplace’s equation in three dimensions and evaluated the resulting topography in the areas where the point elevation data had been removed. Comparison of the 5 m contours generated from the modeled surface to the actual 5-m contours over the un-eroded terrain showed good agreement. The gridded reconstructed surface was then subtracted from the original gridded topography to create an erosion map for the lower Peace Vallis fan. In order to reduce computation time, all computations were done at a slightly coarser resolution of 2 m, instead of the native 1 m.

Topographic data to determine erosion caused by valley incision in the source area are limited. There is only HIRISE DEM coverage for the lower main stem of Peace Vallis, and CTX DEM data are too coarse to resolve the smaller channels higher up in the drainage network. We therefore used the area for which we had high-resolution DEM data to determine width-to-depth

relationships for the incised valleys. We then extended this relation to the entire channel network, so that in areas where coverage was limited to visual data alone, measured channel widths could be used to estimate a channel depth. By doing this along the entire channel network (a length of over 107 km), volume of material missing from the headwaters was estimated. Since cessation of channel incision, dust and colluvium have partially infilled these channels. The limited coverage of HIRISE imagery of the source area suggests that there may have been some finer-scale branching network. Hence, our estimate of sediment produced by channel incision may be somewhat low.

4.4 The Peace Vallis Fan

4.4.1 Planform, Materials, and Topographic Boundary Conditions

Peace Vallis fan is a depositional feature on the northern inner wall of Gale Crater that is expressed by convex contours relative to the surrounding crater walls (Figure 4.3 and 4.5, Table 4.2). Anderson and Bell [2010] mapped the outline of the fan using geomorphologic characteristics and thermal inertial data. We independently mapped the fan boundary based on the topographic boundary of the characteristic contours. Our map, which shows minimal differences to that of Anderson and Bell [2010], is used here. The fan is modest in size compared to other alluvial fans on Mars [Moore and Howard, 2005], covering approximately 80 km².

The fan's apex is located about ~10 km downslope from the crater rim and 900 m down in elevation, below which the fan spreads 60 degrees down to a complex valley floor. Here the fan encountered three lower topographic boundary conditions ("Ridge", "Bradbury Rise", and "Low", Figures 4.5 and 4.6). The western side of the fan extends to a flat-topped broad ridge that continues across to Aeolis Mons. This ridge defines the western border of a 64 km² an enclosed topographic low area (up to 45 m deep) that extends 14.5 km to the east (Figure 4.6a). The enclosed low contains two highs: Bradbury Rise (the *Curiosity* landing area) and a second topographic rise to the east that is somewhat lower in elevation than Bradbury Rise. These are separated by a local low (15 to 25 m lower than Bradbury Rise) (Figure 4.5a, 4.6a), which includes the informally named Yellowknife Bay, and a second, larger low area that lies at the eastern end of the enclosed area (Figure 4.6a). The distal end of the fan extends into the enclosed topographic low area potentially onlapping Bradbury Rise, as well as descends into the adjacent local low (in the vicinity of Yellowknife bay). Bradbury Rise is relatively flat, but it does have a mean southeastern slope towards Aeolis Mons. The relatively flat surface south of Yellowknife bay also slopes weakly towards Aeolis Mons. The convexity of the fan contours flatten out (but remain convex) just as the central and eastern portion of the fan enter this enclosed topographic low (Figure 4.6b). All three of these lower topographic boundaries (i.e. Ridge, Bradbury Rise and Low) are bordered by nearly flat-lying dark cratered surfaces (referred to as the "mound skirting unit" by Anderson and Bell [2010], and as "cratered surface unit" by Grotzinger et al. [2014] and described above, as shown in Figure 4.3 and 4.7a.

Anderson and Bell [2010] divided the fan into an upper (northern) "smooth, lower thermal inertia unit" and lower (southern) "rockier, high thermal inertia unit". They extensively illustrate their descriptions with HIRISE and Themis images. The boundary between the upper and lower fan was remapped by the MSL team using these and other properties, including bedding and fractures, and are reported in Grotzinger et al. [2014] as "alluvial fan" and "light-

toned, bedded fractured unit”. Here we simply refer to the two distinctive areas that comprise the Peace Vallis fan as the upper fan (UF) and lower fan (LF) areas, with Anderson and Bell’s [2010] light-toned, bedded fractured unit comprising the distal portion of the alluvial fan. The upper fan outcrops over 49 km² and extends from the fan apex down to two-thirds of the way to the distal margin (Figure 4.5a,b). It is characterized by a relatively smooth, mottled surface (Figure 4.8a) that corresponds to the low thermal inertia unit. Frequent, lighter toned, roughly circular patches of presumably infilled craters create the mottle appearance. Boulders, perhaps mostly produced by impacts, are scattered across the surface. Subtle blocky fractures, locally strongly expressed, can be found across the upper fan. With the exception of a few sharp-edged small craters which are bordered by ejecta of boulders and darker toned material, the craters have rounded edges and are partially to mostly infilled.

The lower fan, which covers 31 km², is characterized by lightly to heavily fractured, laterally extensive bedrock with scarps interpreted as bedded strata (Figure 4.8b). Extensive wind scour has etched the fractures and exposed the strata. In the transition zone between the upper and distal fan, the mottled and fractured surfaces are intermixed (Figure 4.8b) [Sumner et al., 2013]. Further downslope to the southeast the lower fan becomes more deeply fractured, though the sizes of the polygonal fractures do not vary downslope systematically, either with distance or stratigraphy [Hallet et al., 2013; Sumner et al., 2013]. Boulders are rare and appear to be associated with the remnants of impact ejecta. Direct observation by MSL of the lower fan, high thermal inertia unit, in Yellowknife bay indicates the presence of mudstones (Sheepbed member) and sandstones (Gillespie member) consistent with a distal fan location or lacustrine environment [Grotzinger et al., 2014]. In contrast to the upper fan, the crater walls tend to be sharper edged, as controlled by underlying bedrock resistance. Wind-scoured elongated troughs across the exposed bedrock may originate as small craters (Figure 4.8c).

Topographically, the Peace Vallis fan differs from west to east (Figure 4.5b). The western portion of the fan (55 km²) extends mostly from the apex towards the south and has numerous elongated, discontinuous ridges that are well expressed at the distal end. As discussed, these ridges appear to record exhumed paleo-channels. The lower fan material (light-toned, fractured unit of Anderson and Bell [2010]) is absent near the far western margin. The eastern portion of the fan extends towards the southeast and has only a few subtle ridges that may record paleo-channels. Finer-scale topographic domains in the upper fan may record specific phases or process changes in fan development, whereas the lower fan is composed of a single topographic domain (Figure 4.5c). The distinction among domains is defined by areas that share a common planform curvature, as determined by estimating a focus of radiation and fitting lines of constant arc to successive contour lines. The “west” domain is the most southward directed, contains most of the paleo-channels, and has larger scale ridge and swale topography across the lower end (Figure 4.5c). The “steep” domain is a local area of common curvature (distorted by large impact crater) near the fan apex. Below it the “east” domain shows common curvature, has limited paleo-channels and ends in the light-toned, fractured bedded material composing the “south” domain. Bordering the fan, but still part of the fan based on mean contour orientation and surface texture, is a weakly topographically convergent area. This area appears to connect to a trough that runs along the lower eastern margin of the fan. A narrow transition area down the center of the fan does not correspond in form to either the west or east. In the “south” fan domain,

contours share a common curvature for nearly the entire fan width. This domain corresponds to the most of the light-toned, fractured bedded unit with high thermal inertia.

4.4.2 Profile Analysis

The fan slope decreases progressively from 3% at the apex to 1% at the upper-lower fan transition with significant lateral variability across the lower fan (Figure 4.7). Longitudinal profiles through the west, central, and east portions of the fan are similar from the apex down to the transition to the lower fan. Further downslope the profiles differ greatly, depending on the pre-existing topography upon which the fan encroached. The lower portion of the shorter western profile is steeper than the others at equivalent distance downslope, but then abruptly decreases as it crosses the cratered surface lying on the broad ridge at its base. The central and eastern profiles are nearly identical from the apex to the transition from the upper to lower fan, but the elevation and location of the topographic variations at the base of the fan clearly influence the profile shape (Figure 4.7b). The plot of the logarithm of slope versus distance shows a nearly linear relationship from the apex to the upper-lower fan transition, consistent with the profile being weakly exponential in form. Below this transition the current topography steepens into an enclosed topographic low with exposed bedding that is attributed to wind deflation. The topographic profile does not show abrupt changes when original topography is reconstructed (see Section 4.2.2). The lower boundaries beyond the mapped extent of the fan have surfaces slope towards Aeolis Mons (positive slope), and in the case of the profile across Bradbury Rise (the center profile), the slope progressively steepens from a low of about 0.4 % (across the cratered surface) to greater than 1% (Figure 4.7). As discussed below, this suggests that although the distinct fan morphology does not extend across these surfaces, transport from the fan across these areas seems probable.

4.4.3 Paleo-channels on Peace Vallis fan

A system of downslope-oriented ridges in the western portion of the fan are visible in CTX imagery (Figure 4.2). The ridges are weakly sinuous, discontinuous, variable in width, and appear threaded together as compared to forming a branching network. The most continuous and uniform in width (35 ± 5 m) is a ridge about 3950 m long that originates near the fan apex and descends past the largest impact crater on the fan. We interpret these ridges as recording the coarser beds of various channels, preserved from wind deflation relative to finer grained adjacent sediments (Figures 4.5-4.7). Collectively, the ridge topography in the lower western fan suggests either a braided channel system or intersecting stacked channels cut to a similar depth. Either geometry differs markedly from the sinuous, uniform width features that radiate across other Martian fans [e.g. Morgan et al., 2014]. In the eastern fan, there are few, relatively narrow and uniform width ridges, one of which branches as it crosses from the upper to low fan units (Figure 4.5). Elsewhere (Figure 4.3) on Gale wall and floor, and on the south side of Aeolis Mons, well-defined, uniform width, branching, sinuous ridges likely record single-thread channels (as mapped by Anderson and Bell, 2010).

Detailed cross-sectional and longitudinal profile analysis was done (using 1 m HIRISE digital elevation data) across these ridges to estimate paleo-channel width and slope, and deflation depth (local ridge height relative to adjacent plain). For 122 cross-sections, we

classified cross-section types following Williams et al. [2013b, Figure 4.4]. The ridges range in shape from flat-topped to rounded to pinnacle, with either single, double, or triple peaks. Most of the ridges are predominately rounded to pinnacle shaped, as opposed to flat-topped, in profile. If the multiple peaks preserve multiple channel threads (consistent with braided channel network where the braid bars have eroded relative to the channel floors), the average channel width and standard deviation is $27 \text{ m} \pm 16 \text{ m}$, but if the multiple peaks represent a single channel the average width and standard deviation is $34 \text{ m} \pm 18 \text{ m}$. The distribution of widths is approximately Gaussian, and width does not vary systematically with slope or distance downstream (Figure 4.9a). Deflation depth, determined from ridge height, averaged 2.4 m on the western fan and only 0.6 m on the eastern fan (Figure 4.9b), which is approaching the vertical resolution of the MSL DEM elevation data. The height of the ridges above the surrounding fan surface does not necessarily reflect the depth of the paleo-channel, but rather the deflation of the less protected fan surface. The inverted channel ridges are presumably resistant to erosion due to a lag of coarser rocks, or cementing of channel deposits (Newsom et al., 2010). The ridge profiles parallel the local fan slope. For the channels on the western fan, the average slope and standard deviation obtained from our long profile analysis is $1.9 \pm 0.6\%$. The channel longitudinal profile for the 3950 m long paleo-channel declines in slope from 2.5% near the apex to 1.5% near where the channel terminates (Figure 4.10). Hence, the downslope decrease in slope that typifies this fan occurs within a single channel.

Exposures on the sides of the ridges marking the paleo-channels reveal some characteristics of the underlying deposits on the upper fan. Stratified layers of light-toned material and darker boulder-rich deposits are visible within many of the ridges on the upper and distal western portion of the fan (Figure 4.11). The alternating layers are on the order of 0.25 to 1 m wide in plan view and there are typically 3 to 5 visible layers in most ridges. The individual boulders are meter to sub-meter scale, and their size distribution within and near the inverted channels appear to be fairly uniform, in contrast to variable size distributions of boulders observed in the vicinity of impact craters elsewhere on the fan. There is an apparent decrease in the number of boulders with distance away from the inverted channels, which is a pattern that has been observed elsewhere on Mars (see Figure 4.3 in Williams et al. [2011]). On the western rim of the largest impact crater on the upper fan surface is a dark band that may be an exposed boulder-rich layer. Boulders are found less frequently in the lower inverted channels on the western portion of the fan and we have not observed boulders associated with the channels on the eastern fan. Boulders may have been transported down the fan as coarse-grained sediment or may have originated as weathering products of more indurated interbeds.

4.5 Peace Vallis Source Area

The northern Gale crater rim is well defined by elevated topography relative to the Gale walls and floor to the south and to the adjacent plains beyond the crater to the north. Peace Vallis fan lies downslope from the largest extent of crater rim that is set back or breached (about 6 km) relative to general position of the rim along the north (Figure 4.12a). This roughly 15 km long set-back causes a topographic break in the rim that provides a drainage pathway from the northern plains; this extends more than 10 km to the north of the Gale rim (Figure 4.12a). Using the current surface topography to infer the direction of transport of water and sediment, we can delineate the equivalent of a watershed area (or catchment) of about 1000 km^2 that may drain to

the Peace Vallis fan (Figure 4.12a). This potential source area consists of steep hills along the offset rim, and then further north, gently sloping plains bordered and interspersed with local hills roughly 100 to 350 m in relief. The hills are not systematically cut by a network of canyons, suggesting no fluvial organization of this topography since impact, and are perhaps the result of an irregular distribution of impact ejecta. In contrast, the broad plains to the west and east of the back-set rim show a mean slope towards the two entrance points to the steep canyon rising up from the fan apex (Figure 4.12a). CTX images of this area (no HIRISE coverage is available) reveals limited bedrock outcrop only at local steep portions of hilltops. Instead, these irregular hills are mostly mantled with debris, which appear to be some combination of eolian deposits and colluvium. The gross slope of the plains surface towards the outlet, the apparent ramps of colluvium off adjacent hills to the plains, the basin-like topography in which the plains are located, and the occurrence of local ridges (both within the source area and adjacent regions) that may record depositional features elevated due to deflation all suggest that plains are underlain by mixtures of colluvium, fluvial, and eolian sediments. These areas may also collect ash fall, as a large area of volcanic plains lies to the north.

Cut into the debris mantle in the source area is a spatially limited, relatively deep valley network that feeds the canyon that drains to the fan apex (Figure 4.12a). We use the term “valley” rather than “channel” because the features are much broader and deeper than the channels that cut them. The valleys are cut 80 to 900 m wide and 5 to 107 m deep into the debris mantle. CTX imagery can resolve features on the order of 10 m, but no valley widths smaller than ~80 m were detected, suggesting that the valley network does not continue to extend throughout the entire source area. There is a 291 km² eastern portion of the catchment (Figure 4.12a) that joins the 730 km² portion of the catchment through a relatively small valley and at the junction is a partially eroded fan shaped deposit. This suggests a limited contribution of water and sediment to the 730 km² catchment where the valley network has developed.

The main valley that feeds the fan apex ascends across the base of the steep hills defining the back-set crater rim. Tributaries enter from the western upland plains, and from the north along the steep hillslopes. The main valley heads up to the east around the back-set hills and forks across the sloping plains in the northern uplands (Figure 4.12a). A limited but well-developed branching network is established across the northern plains. The total valley length is 107 km, equivalent to a drainage density of 0.15 km/ km² for the 730 km² area likely contributing water at the time of dissection. Coincidentally, this drainage density value matches the highest valley network drainage densities reported in a global survey by Hynek et al. [2010] of over 82,219 valleys, with most of the highest drainage density valleys forming between the latest Noachian and earliest Hesperian epochs. The longitudinal profile from the northern branch of the valley network down to the fan apex (Figure 4.12b) shows the upland channels to have a slope of 2% or more and that the channel progressively steepens as it descends and enters the main canyon. Here slopes locally reach 10% before the channel exits onto the crater wall and declines to 5%.

The entire incised valley network that leads to Peace Vallis fan is cut into previously accumulated deposits, likely composed of colluvium, fluvial and eolian debris. Although relatively sharp-edged, the incised valleys are floored with dunes and colluvium, indicating they are partially infilled. HIRISE images of the lower part of the mainstem valley shows boulder-

lined banks in some exposures, suggesting mobilization of coarse debris to the fan. Craters 500 m to 1000 m in diameters are partially to mostly infilled and scour tracks of wind deflation are common. In the limited HIRISE data available some finer-scale infilled channels are visible along the upper crater walls adjacent to the Peace Vallis fan.

4.6 Sediment Budget

The direct connection of the incised valley network to the fan, and the relatively strong morphologic preservation of both the valleys and the fan suggest that the fan sediment is derived from incision of the valley network. Here we test the hypothesis quantitatively by comparing estimates of the eroded and the depositional volumes.

4.6.1 *Volume of erosion due to valley incision*

HIRISE elevation data extends upslope from the fan apex 10 km and the volume needed to fill the valley in this region can be accurately calculated. Further upslope, CTX images must be relied upon to estimate eroded volume. For where there is HIRISE coverage, the valley width to incision depth ratio is 18, based on 20 cross-sections along the valley. This ratio was used to calculate the depth of erosion for valleys identified in CTX images that lack topographic data to measure depths. The valley network was divided into boxes of uniform width in order to calculate the volume of sediment that had been removed to create that section of the valley. As a test of this box-method, the same analysis was done on the lower portion of the Peace Vallis valley network (where HIRISE topographic data are available). Volumes using these high resolution data were calculated following the methods developed by Warner et al. [2011], yielding 0.10 km^3 of material removed from the lower 10 km of the Peace Vallis valley network. The box-method gave a similar result of 0.11 km^3 . The box-method applied to the entire incised valley network gave a volume of 0.8 km^3 . This volume is probably an underestimate because the valleys that are partially infilled are not accounted for, nor are smaller valleys that have been erased either through erosion or complete infilling. The estimate also does not account for erosion that occurred outside of the valley network, but within the drainage basin.

4.6.2 *Volume of the fan*

Following the methods described in Section 4.2.2, the current volume of the fan is estimated to be 0.7 km^3 (0.5 km^3 for the upper fan and 0.2 km^3 for the lower fan). The different assumptions made about the pre-fan topography result in preserved volumes ranging from 0.4 to 0.8 km^3 . For the 0.7 km^3 value, the mean thickness of the fan deposit (for the 80 km^2 fan) is just 9 m. If one reconstructs fan topography (see Section 4.2.2), the amount of wind erosion can be calculated, to produce estimates of the original volume of the fan. About 0.14 km^3 has been lost from the lower fan due to wind erosion (Figure 4.13a). The upper fan has also experienced deflation (hence the paleo-channel ridges). For the average depth of 2.4 m of erosion over the entire western fan (25 km^2), a volume of 0.06 km^3 has been removed. The 0.6 m of deflation over the entire eastern fan (55 km^2) equates to 0.03 km^3 of eroded sediment. Hence, the total original fan volume was 0.93 km^3 .

4.6.3 *Sediment Budget*

Within the error of these estimates, the calculated volume of eroded sediment (0.8 km^3) matches well the volume of the original fan (0.9 km^3). The eroded volume was expected to be an underestimate due to infilling and imperfect preservation of finer valleys. This suggests that the full extent of significant fan deposition has been mapped. This does not rule out transport across Bradbury Rise, but it would argue against the general elevated topography of that rise and of the adjacent broad ridge to the west as originating from deposition of sediment generated from the observed valley incision. Note that since the source material is a granular deposit (not bedrock), the bulk density of the eroded and deposited material was considered the same, such that volumes can be directly compared. The volume agreement also suggests that the valley incision and fan formation were coincident. Gale Crater rim is laced with incised valleys (Figure 4.3). Irwin et al. [2005] proposed that this incision, the associated fan and delta formation and the possible presence of lakes all occurred in a brief period. Alternatively, lakes, valley incision and fan and delta formation could have been episodic and not synchronous across Gale. As discussed in the next section, this suggests a need for stronger evidence of the timing of events.

4.7 Relative Timing

Thompson et al. [2011] proposed that the impact age for Gale, based on crater counts in the surrounding ejecta, occurred in the Late Noachian/Early Hesperian, with uncertainty arising from the occurrence of a few large craters for which it is not clear if they pre or post-date the impact. They counted craters separately on the crater “floor” (which included the crater wall and floor across which the Peace Vallis fan advanced), and concluded that the “floor unit” was also of Early Hesperian age. These counts and stratigraphic relationships suggested to them that the Aeolis Mon deposits were formed soon after impact, about 3.6 to 3.8 billion years ago. They point to the proposal by Irwin et al. [2005] that Farah Vallis breached along the south rim and formed two small deposits around what must have been the pre-existing mound deposit, and conclude that most deposition and channel erosion in Gale crater occurred in a period of 0.2 to 0.3 Ga post Gale impact. Le Deit et al. [2013], working independently and using only crater counts on the ejecta sheet to the south and east, also arrive at an age of ~ 3.61 Ga.

Crater counts in the 730 km^2 source area for Peace Vallis fan, much of which lies in the ejecta blanket of Gale, also yielded an age of about 3.6 Ga [Chapter 5]. This suggests that the source area has not been re-surfaced due to prior erosion and is consistent with the interpretation that the erosion that produced the fan came principally from the observable incised valleys. Early (shortly after impact) erosion and delivery of sediment to the floor of Gale cannot be ruled out. Peace Vallis fan may be too small to obtain reliable crater count age dating directly (e.g. McEwen and Bierhaus [2006]; Michael and Neukum [2010]) and we see no definitive observations to demonstrate that craters within the fan have preceded the arrival of the relatively thin Peace Vallis fan. Le Deit et al. [2013] however report crater counts for a selected area on the northern crater floor (not including the Peace Vallis fan) that they interpret as suggesting a crater floor age of about 3.4 Ga and then subsequent resurfacing at about 1.1 Ga.

Suggestions of lake development may provide further constraints on Peace Vallis formation timing. Several groups have proposed that after Aeolis Mon formation and its erosion to topography similar to present, lakes formed to various depths [e.g. Cabrol, et al., 1999; Irwin

et al., 2005; Dietrich et al., 2012; Le Deit et al., 2013 and see review in Anderson and Bell, 2010]. The deltaic-like forms that have been previously mapped [Irwin et al., 2005 - Figure 4.3)] as well as fan shaped deposits at the mouth of the large canyons cut into Aeolis Mons are deposited at elevations far above the Peace Vallis fan. Anderson and Bell [2010] documented the canyons on Aeolis Mons, the numerous gullies descending across the crater wall, fan shaped deposits, and inverted and incised channels, that indicated spatially extensive water-driven erosion and deposition processes. If Peace Vallis was active when the other large canyons were depositing sediment, deposition at a similar elevation on the crater wall might be expected (especially if that elevation represents a lake level). Instead, Peace Vallis fan extends to floor of the crater, lower than any other fan like form. This low position and the apparent lack of post deposition burial and scour by other events (other than wind erosion) suggest that Peace Vallis fan may be one of the last water-driven depositional events in Gale Crater. Nonetheless, the extensive wind deflation of the lower portion of the fan, the inverted paleochannels on the upper part of the fan, and the generally poorly preserved crater forms on the upper fan all indicate that considerable time has passed since fan formation.

Crater count dating of much larger ($\sim 1000 \text{ km}^2$) fans in craters [e.g. Grant and Wilson, 2011 and 2012] have suggested a possible common development time in the Amazonian. Mangold et al. [2012a] concluded, based on crater counts on individual fans in the Margaritifer Terra region, that a late-stage episode of fan construction was Hesperian to Early Amazonian in age. Morgan et al. [2014] also report fan construction periods of mid to late Hesperian and early Amazonian. Hence, dates elsewhere (on much larger features with more reliable crater counts) point to a timing of Peace Vallis fan (if it were synchronous) as being possibly late Hesperian to early Amazonian.

4.8 Discussion

Based on the observations reported above, the three key questions can now be explored. Here we find many similarities but some important differences with fans investigated elsewhere on Mars.

4.8.1 *Where did the sediment come from?*

The sediment budget analysis strongly suggests that the fan deposits are derived from the visible valley incision into previously deposited debris in the source basin. The origin and process of sediment delivery of this previously deposited debris are less well defined. Most of the sediment in the fan was derived from the wide, deep mainstem valley (Figure 4.12a). This setting would likely favor deposition of coarse debris (including boulders) descending from the adjacent steep valley walls. Debris production could occur by many processes including immediately after the impact (e.g. impact breccia), freeze-thaw activity, or marsquakes. One of the tributaries originating on the steep slopes appears to be marked by parallel-elevated ridges, perhaps levees tracking the path of mass flow deposits to the mainstem. Some, if not much of this sediment may simply be outfall of impact debris, and subsequent remobilization from adjacent slopes. The mantling of the hillslopes with fine debris (such that bedrock is rarely visible) points to mobilization, deposition, and perhaps some in-situ production of fines. Exposures in the walls of the lower part of the mainstem (from HIRISE) show boulders, but the

exposures are dominated by sediment finer than boulders. The incised valleys that cut across the plains may have brought finer sediment to the fan, though the 2% valley slope there would still favor gravel-bedded systems.

The mainstem canyon may have been a path of water and sediment delivery before Peace Vallis fan formation. The apparent old age of the crater wall adjacent to the fan [Thomson et al., 2011], however, does not suggest significant delivery there, and the reconstructed pre-fan surface would seem to rule out significant deposition from a prior, older fan.

Only two other fans developed within craters on Mars have reported distal light-toned deposits like that found in the Peace Vallis fan. Mangold et al. [2012b] noted light-toned, high thermal inertia material distal deposits in Majuro crater. The fan is located in the NE Hellas region and is dated to be Late Hesperian by crater counts. It shows small inverted channels in the northern lower section and phyllosilicates are indicated based on spectral mapping using CRISM data. Morgan et al. [2014] observed a distal light-toned deposit in Runanga crater, which is crossed by numerous paleo-channels. Newsom et al. [2010] documented a light-toned polygonal fractured rock that likely bears Fe/Mg phyllosilicates (also based on CRISM data) on the western floor of Miyamoto crater along with extensive inverted paleo-channels, though no fan or delta deposit was observed. Extensive wind deflation of fans has been documented elsewhere [e.g. Morgan et al., 2014] suggesting fine-grained deposition is prevalent, but the presence of the light-toned fractured pattern may occur only with significant hydrothermal alteration, as suggested by Newsom et al. [2010] and Mangold et al. [2012b]. Investigations resulting from the *Curiosity* rover investigations may provide constraints on the origin of the fracturing and what that implies about source material and post-depositional processes.

4.8.2 What processes formed the fan?

The processes of sediment delivery to the fan, transport across it, and net deposition appear to differ spatially and, perhaps, through time. The mainstem valley that feeds the fan originates in tributaries lying in 2% sloping valleys in broad upland plains. Channels on 2% slopes likely carried mixtures of sand, gravel and boulders and may well have been braided. Given the distance to any steep slopes, it is unlikely that mass transport events (e.g. mudflows) occurred in these upland channels. However, as the upland channels collected and drained into the mainstem from the northeast, the convex profile progressively steepened (Figure 4.11b) to local reaches of 10% and an average slope of 8%. This is a very steep slope for such a large drainage area (730 km²), which means that flows through these valleys were capable of producing shear stresses of exceptional magnitude for relatively modest discharges. Such flows likely caused rapid channel incision into finer material until slowed by bedrock resistance or large grains. Current burial by colluvium and eolian deposits and insufficient HiRISE coverage at the time of this publication prevents evaluation of the bed conditions. However, the preservation of such a steep mainstem suggests limited water runoff, as sustained flow would cause incision that would tend to drive the form to concave up, and advance the steep reach into the upland plains. In the local steep tributaries, debris or mudflows may have been generated, traveling into the mainstem. Such a steep mainstem channel would have transported these flows to the fan. Within the source area channels, sediment could have travelled up to 35 km before

reaching the fan, providing ample travel distance to round sediment and breakdown weak material.

Runoff that drove valley incision in the source area could be due to groundwater flow, rainfall or snowmelt. Groundwater discharge is unlikely as a source mechanism for runoff based on the geometry of the channels and the channel network. The channel tips in the uplands originate in broad plains with distant hills rising typically less than 350 m above the surrounding topography. On the steep tributaries along the mainstem, the channels extend well up the hills. However, none of the channel heads take the characteristic amphitheater shape associated with seepage erosion in loose material, and the valley network is a branching system, not a single unbranched valley, which is typical of groundwater discharge (e.g. Laity and Malin [1985]). All of these observations argue against groundwater flow as a runoff generation mechanism. This leaves runoff from rain or snow. Snow could accumulate in the uplands and funnel water to the fan upon melting. Rain would likely occur over a larger regional scale, perhaps across the entire crater rim, but there is no evidence of preserved extensive channel networks to confirm that rainfall was the dominant source of water for fan formation (although incised valleys do ring the crater wall (Figure 4.3). The Hesperian to Amazonian age of the fan may make rain less likely as both the atmospheric pressure and temperature are thought to be too low to support rainfall [Forget et al., 2013; Wordsworth et al., 2013], though Craddock and Howard [2002] suggest that the morphology and drainage density of highland valley networks point to rainfall occurring possibly into the Amazonian.. Hence, we propose, as others have done for fans across Mars [e.g. Howard and Moore, 2011; Grant and Wilson, 2012; Mangold et al., 2012; Morgan et al., 2014] that runoff was more likely dominated by snowmelt. A terrestrial analog would be the McMurdo Dry Valleys, in which most precipitation falls as snowfall (though small amounts of rainfall have been known to occur in the summer) and channelized ephemeral streams are sourced completely from meltwater [Conovitz et al., 1998; Fountain et al., 1999].

Once runoff and the sediment it carried reached the fan, sediment transport likely occurred by several mechanisms. Clearly fluvial channel-bound processes occurred, as evidenced by the numerous paleo-channels. The elevated channels beds (now ridges) indicate that finer-grained overbank sediments in flood flows, capable of being swept away by wind, must have also been deposited. Boulder abundances associated with the paleo-channels may decrease downslope. Although it is often assumed that grain size decreases down alluvial fans, as Stock et al. [2008] and Stock [2012] report in their review of fan processes that it is often not the case on terrestrial fans. Sheet floods may have also occurred across the fan where relatively dilute flows spilled out of channels and travel across the fan, causing both scour and deposition. Lastly, mudflows could have traveled the entire length of the fan if sufficient fines of sand, silt and clay were present to generate them. We see no morphologic evidence of mudflow processes, but such evidence may be subtle. Morgan et al. [2014] describe mudflows in relatively low gradient fans in the Atacama Desert. These flows travel down channels on the fan, exceed the channel carrying capacity, and then spill out across the fan surface for a considerable distance. This would be difficult to detect using HIRISE data.

The transition from the upper fan, where the channels are presumably armored against significant deflation by pebbles and gravel, to the channel-less lower fan, suggests that although gravel units may occur in the lower fan, the gravel abundance is not sufficient to form a

protective armor. The absence of elevated channel paths in the lower fan may be due to either the channels being sand bedded and thus incapable of armoring against wind deflation or that channelized flow was absent and instead sheetflood transport conditions prevailed. We note that the lower portion of the fan descends into an enclosed topographic basin (Figure 4.5a), suggesting the lower fan may have periodically encountered a shallow lacustrine environment. Direct observations of the lower fan deposits in Yellowknife bay indicate the presence of laterally extensive mudstones indicating lake sediments [Grotzinger et al., 2014].

Nearly all other Martian fans are characterized by a radial pattern of distributary channels exposed as ridges across the fan [Moore and Howard, 2005; Morgan et al., 2014]. The absence of paleo-channels for much of the eastern portion of the upper fan, yet the apparent lack of deep wind deflation, suggests that flows on the eastern part of the fan were either not channelized or the inter-channel sediments were coarser grained and harder to deflate. If the former occurred, one possibility is that the flows that crossed and built the eastern portion of the fan, and deposited much of the finer grained light-toned unit of the lower fan, were smaller in magnitude and simply spilled and spread as episodic patchy sheet floods across the fan. Terrestrial fans often lack well-defined continuous channel systems, and we know from observations at Yellowknife Bay [Grotzinger et al., 2014] that the lower fan could be a mixture of fluvial and shallow lacustrine deposits. Consistent with this sheetflood hypothesis is the presence of pebble conglomerate outcrops across Bradbury Rise [Williams et al., 2013a]. These deposits may record the transport of gravels from the Peace Vallis fan towards Aeolis Mons. No channel paths or elevated channel beds have been detected here, and the absence may reflect sheetflood transport processes.

To assess how much runoff may have been involved in building a depositional feature like Peace Vallis fan, the common practice has been either to assume a water to rock ratio [Mangold et al., 2012c] or to perform a hydraulic analysis based on estimated channel and grain size characteristics to calculate channel filling flows [e.g. Wilson et al., 2004]. Both approaches are very approximate. Typical rock/water ratios vary by orders of magnitude and depend on grain size and climate characteristics. The hydraulic calculations require many assumptions that cannot be tested and only provide instantaneous discharge estimates. Perhaps the most relevant terrestrial study in providing an estimate of rock/water ratios for the Peace Vallis fan system is the work by Pepin et al. [2010] across Chile. They analyzed the suspended sediment annual yield as a function of mean annual runoff. For low vegetation density in very dry regions, the annual rock/water ratio was about 0.0002. For a fan sediment volume of 0.93 km^3 , the ratio suggests that 4650 km^3 of water runoff would be needed. This is equal to about 6 km of water across the active catchment. Syvitski et al. [2003] collected runoff and sediment flux data for 340 terrestrial river basins around the world, and the log transformed rock/water ratio yields a roughly Gaussian curve with a mean value of 0.0001 ± 0.00008 . The estimate of the rock/water ratio on Mars to drive fluvial sediments at other locations has varied greatly. For example, Moore et al. [2003] assume 0.05. Mangold et al. [2012c] proposed 0.0005 for dilute flows and 0.02 to 0.3 for dense flows. Morgan et al. [2014] chose a concentration of 0.2, reflecting the inference that the fans in Saheki crater were built in large part by mudflows. The very conservative 0.2 concentration applied to the Peace Vallis fan would still require 6 m of water to be produced over an area of 730 km^2 in the active catchment. The observations of fluvial gravels reported in Williams et al. [2013a] across Bradbury rise, however, show that dilute transport did occur and the effective

concentration would be well below 0.2. Drilling by the Curiosity rover into the light-toned unit revealed a mudstone (i.e. Sheepbed member) that likely extends at least 4 km² and records a lacustrine environment that may have persisted for hundreds to thousands of years [Grotzinger et al, 2014]. Considering the extent of exposure of what appears to be fluvial/lacustrine deposits in the lower fan, Grotzinger et al. [2014] hypothesize the sediments could have been deposited over millions of years. The abundant evidence of sorted sediments and lacustrine deposits strongly suggests that the more likely rock to water ratio is closer to that observed in Chile (i.e. 0.0002) or even the global average on Earth (0.0001). Conservatively selecting one order of magnitude higher concentration than the Chilean value still requires about 600 m of runoff from the 730 km² source basin (i.e. 465 km³ of water). This hydrologic analysis supports the sedimentologic inference by Grotzinger et al. [2014] that the fan was active over an extended period of time, and it requires the establishment of a hydrologic cycle (rather than a single melt event) to generate this large amount of runoff.

Another constraint on runoff is to estimate the instantaneous discharge in a paleo-channel crossing the fan [e.g. Wilson et al., 2004; Williams et al., 2009; Williams et al., 2013a; Morgan et al., 2014]. We use standard hydraulic equations and threshold channel concepts along with slope data and our measurements of channel width to determine the likely grain sizes that comprised the channel bed and the instantaneous discharges and flow velocities at bankfull conditions for a ‘representative’ cross-section.

The fluid discharge, Q , within a channel can be derived by conservation of mass,

$$Q = uhw \quad (4.1)$$

where u is the average flow velocity, h is the channel depth and w is the channel width. For channels with width-to-depth ratios less than 20, the hydraulic radius, R , which is the ratio of the channel area to its wetted perimeter, is used instead of h . For a turbulent velocity profile (i.e. Reynold’s number, Re_{channel} , is > 1000 [Chow, 1954]), the velocity, u , is found at a given height above the bed, z , using the law of the wall formulation,

$$u = \left(\frac{u_*}{k} \right) \ln \left(\frac{z}{z_o} \right) \quad (4.2)$$

where z_o is the roughness coefficient, k is the dimensionless von Karman’s constant (typically 0.41 [e.g. Furbish, 1997; Garcia, 2007]) and u_* is the shear velocity, defined as

$$u_* = \left(\frac{\tau_b}{\rho} \right)^{0.5} \quad (4.3)$$

In equation 4.3, τ_b is the boundary shear stress, defined as

$$\tau_b = \rho ghS \quad (4.4)$$

where ρ is the fluid density, g is the gravitational constant (3.711 m/s^2 for Mars), and S is water surface slope (assumed to equal measured topographic slope in this analysis). The roughness coefficient for hydraulically rough flow is

$$z_o = \frac{k_s}{30} \quad (4.5)$$

where 30 is the Nikuradse roughness coefficient and k_s is the effective roughness height, typically approximated as $k_s = 3.5D_{84}$, (e.g. Garcia, 2007) giving

$$z_o = \frac{3.5}{30} D_{84}. \quad (4.6)$$

The vertically averaged velocity is:

$$\bar{u} = \frac{u_*}{k} \ln \left(\frac{11h}{3.5D_{84}} \right) \quad (4.7)$$

where the 11 in equation 4.7 is the result of integrating the log profile constant and multiplying it by the Nikuradse roughness (i.e. 30). Assuming that the boundary shear stress needed to move the median grain size on the bed (D_{50}) is the critical boundary shear stress [e.g. Buffington and Montgomery, 1997],

$$\tau_{*cr} = \frac{\rho g h S}{(\rho_s - \rho) g D_{50}} \quad (4.8)$$

(ρ_s is the sediment density) and the shear velocity, u_* is

$$u_* = \left(\tau_{*cr} \frac{(\rho_s - \rho)}{\rho} g D_{50} \right)^{0.5}. \quad (4.9)$$

Vertically average velocities reported here are derived from Equation 4.7 and the relationship given for shear velocity given in Equation 4.9. Instantaneous discharge is then calculated from Equation 4.1.

The average paleo-channel bed width on the western fan is 27 m. Channel slope is set to 0.01 and 0.02 (the steeper slope is more characteristic of where the channel widths were measured). Martian gravity is 3.711 m/s^2 , the sediment density is 3000 kg/m^3 (typical of basalt) and fluid density of 1000 kg/m^3 (low concentration flow). We assume a range of critical shear stresses, τ_{*cr} , from 0.02 (for sandier beds) to 0.06 (for coarser beds) [e.g., Parker et al., 1998; Parker, 2007; Garcia, 2007] and a range of width-to-depth ratios to determine the mobile median grain size under critical conditions by rearranging Equation 4.8 to solve for D_{50} . To constrain the likely grain size in the channels on Peace Vallis fan, we assume width-to-depth ratios between 10 and 100, which are the range of aspect ratios observed for terrestrial alluvial rivers [Knighton,

1998] and for channels on several arid to semi-arid alluvial fans [Stock et al., 2008]. Given the possible braided form in the western portion of the fan, a value closer to 100 is more likely. The roughness length scale D_{84} is set equal $1.5 \cdot D_{50}$ [Morgan et al., 2014]. Also, when the width-depth ratio is less than 20, the hydraulic radius, R , is used in place of channel depth, h . As a separate constraint, flow is also assumed to be critical, i.e. the Froude number is 1, which has been observed on terrestrial fans [see review in Stock, 2012].

The predicted grain size for the western fan paleo-channels of 2% slope lie in the gravel to boulder range (Figure 4.14). For a given flow depth (i.e. W/h) the lower critical Shields numbers predicts a larger grain size. Predicted grain size is proportional to slope, so for a given flow depth, if the slope were 1% the predicted values would all decrease by a factor of 2, but still remain in the gravel to boulder size range. Given the possible braided channel state (with a high W/h) and the lack of abundant boulders visible on HIRISE, the likely median bed surface grain size is probably closer to the low value of 45 mm for the 0.06 critical Shields number. An independent check, which assumes a critical flow depth ($Fr = 1$), predicts a still smaller median grain size of 30 mm at a W/h of 100. Data from an analogue fan in the Atacama Desert (Chile) [Morgan et al., 2014], lies along the $Fr=1$ line, supporting its use in this analysis. For W/h of 100 and a slope of 2% for the paleo-channels on the western fan, the predicted flow velocity ranges from 0.5 to 0.9 m/s for low to high critical Shield numbers respectively. This equates to instantaneous discharges of 3.7 to 6.5 m^3/s . If the channels were much deeper, i.e. W/h of 10, the velocities would climb to 1.6 to 2.8 m/s, respectively, and the discharges would be 117 m^3/s to 207 m^3/s .

For the Peace Vallis catchment area of 730 km^2 , these discharges require water supply rates between 0.02 mm/hr (high W/h) to 1 mm/hr (low W/h), which assumes the entire catchment simultaneously melts and produces a steady runoff rate. During any flow event, there may have been more than one channel on the fan receiving flow, so discharge may have been higher. These input estimates are also conservative, as they assume no losses due to refreezing, evaporation or infiltration. If we assume these losses to be relatively minor, then input rates need to be sustained on a timescale comparable to that needed for water to travel through the valley network (length ~ 50 km). Assuming flow velocities on the fan are comparable to velocities in the incised valley network, flow would need to be sustained between 5 (high W/h) and 27 (low W/h) hours in the headwaters.

Based on geomorphic evidence and the likely deposition of the fan in the Hesperian or Amazonia, we assume discharges could have been generated by snowmelt. Global climate models show that snow accumulation is possible in the equatorial highlands of Mars, assuming a faint young sun and a denser CO_2 atmosphere [e.g., Woodsworth et al., 2012; Kite et al., 2013]. Williams et al. [2008, 2009] use a snowpack model to show that for the past 5 Ma latitudes between $70^\circ N$ and $70^\circ S$ correspond to significant seasonal snowmelt runoff (>1 mm) on the order of several days a year. Kite et al. [2013] suggest that over Mars' history, peak snowmelt rates were likely between 2 and 3 mm/hr, which is adequate to drive the highest estimated discharges and are well above the more probable lower values. The topography of the Peace Vallis source area (Figure 4.11a), where large plains are bordered by hills, appears particularly well suited to collect snow and release snowmelt.

4.8.3 *How did the fan evolve?*

Here we use the observations reported above to propose a possible evolutionary history of Peace Vallis fan. It is hoped that this hypothesis may eventually be explored through modeling. The initial conditions for Peace Vallis formation appear to be an established large uplands catchment variably mantled in debris, a concave up south facing crater wall (lacking a prominent prior fan form) and a complex crater floor consisting of a broad relatively flat ridge, a rise and a low (Figures 4.11a and 4.6b). Runoff, possibly due to snowmelt, would follow the maximum fall path and concentrate along the steep canyon, cutting into the accumulated debris. Permeability of this debris would influence how much surface runoff would be generated and the resulting valley incision. The developing channel system would expand along the topographic convergent areas through the uplands, down the steep canyon and onto the crater wall. As the channel cut across the lower wall it took a diagonal path. When the main valley opened up into the crater, sediment-depositing flows likely first headed down the maximum fall direction along the eventual western margin of the fan. Hence, the channel-rich western fan may have been the first path of deposition.

If flow was limited in duration or peak volume, the fan may have advanced episodically down the relatively gently sloped crater wall leaving patches of coarse material bordered by finer grained sediment and perhaps local playa lake deposits in enclosed lows. Alternatively, sufficient flows may have produced continuous channels across the fan to the crater floor. In this case, the fan form may have advanced upslope from the lower topographic boundary condition. Clearly such channels did eventually develop as evidenced by the numerous preserved paleo-channel beds in the western fan. These channels do not radiate or show signs of a distributary behavior, and instead may record a period of braided channel transport conditions along the western fan. Overbank events would spread finer sediment across adjacent plains that subsequently would be subjected to deflation, leaving gravel bedded channel paths elevated.

Aggradation causes epicenters of fan deposition to shift, and the hypothesized initial deposition along the western margin would cause the fan construction to shift eastward. This proposed time sequence, suggests that the mode of transport changed from significant channelized flow and sediment transport to more dispersed possible sheetflood transport. Small discontinuous gravel bedded channels (a few meters across) could transit this area, but deflation would not preserve the channel traces. The transition from the upper fan's coarse mantle material to the fine-grained, fractured lower fan deposits [Sumner et al., 2013] occurs where the slope declines to 1% (at about 10 km from the apex) (Figure 4.7c). Although still relatively steep, this slope reduction may favor spreading of the flows, transition from suspension to bedload transport of sand, and a predominance of sand and finer sediment deposition in general. These changes in sedimentation style may have caused the change in surface properties from mottled exposures in the upper fan to the light-toned outcrops with scarps and higher thermal inertia characteristic of the lower fan. Patches of gravel associated with discontinuous channels may still have formed on the lower fan but not enough to later prevent deflation. Below the transition, the fan descended into an enclosed basin, and down to a local topographic low where the fan form terminates (Figure 4.11a). Given this topography and the distal position on the fan, lake conditions may have developed.

The center part of the fan likely prograded onto Bradbury Rise and the sediment transported in these flows would possibly continue downslope towards Aeolis Mons (Figure

4.15). If this occurred, the gravels reported by Williams et al. [2013a], may be a record this process. It is possible that some of these flows spilled off Bradbury Rise onto the local cratered surface, creating a local fan-shaped deposit. Detailed mapping is still being done to assess whether the sands and gravels of the Bradbury rise area are best considered ancient (preceding the fan) or derived from it directly.

There are distinct topographic components to the fan with different focal locations associated with distinct sets of arc-shaped contours (Figure 4.5c). Only the lower most convex contours appear to define a continuous topographic feature across the entire fan. Although these different components grade smoothly into each other, they may record variations in both space and time of fan development. The thin beds of sediment visible in HIRISE data in the lower fan, the numerous paleo-channels, and the topographic complexity of the fan all suggest that hundreds to thousands of flow events were needed to build the fan.

Even the most conservative of estimates requires at least 600 m of water over the 730 km² active source area to build Peace Vallis fan (i.e. ~500 km³ of water). Likely much more water was involved, possibly in excess of 1000 km³. This requires a hydrologic cycle and considerable time. This is consistent with the inferences obtained from the sedimentology and stratigraphy revealed by the Curiosity rover [Grotzinger et al., 2014]. In a study of larger equatorial fans on Mars, Moore and Howard [2005] concluded that construction took tens of thousands to hundreds of thousands (perhaps millions) of years, a conclusion recently reinforced by Armitage et al. [2011]. Our smaller fan may fall towards the shorter time period of this estimate, but that depends on how intermittent the favorable climate conditions were.

Post fan deflation occurred across the entire fan, but was greatest in the light toned fractured lower fan. On the western fan, 2.4 m of erosion is needed to elevate the paleo-channels, and if we estimate a younger age of the fan to be 2 Ga, then this indicates an erosion rate of 1 to 5 nm/yr, whereas the eastern upper fan eroded 0.6 m or 0.3 to 1.2 nm/yr. The amount of erosion is highly variable in the lower fan (Figure 4.13a). The average is 5 m, but locally it exceeds 10 m. This corresponds to an erosion rate low of 2.5 nm/yr (over 2 billion years) to a local high to 20 nm/yr (10 m/500 million years). Farley et al. [2014] estimate local wind-driven scarp retreat, based on surface exposure age dating, of 0.75 m/million years, and suggest from this that wind erosion may be episodic in Gale. Other than wind deflation, we have found no evidence of further fan evolution. Others have suggested that fan construction abruptly halted on Mars [i.e., Moore and Howard, 2005]. Our observations do not contradict this inference.

One possible time sequence of events in Gale Crater is: impact, center mound infill and scour, multiple lake formation and associated fan and delta formation (Figure 4.3), and ultimately, the Peace Vallis fan formation. The timing of the numerous incised valleys and downslope paleo-channels in Gale Crater (Figure 4.3) remains uncertain. Further work is needed to establish the relative timing of these valleys and channels. This matters to constraining further the necessary climatic conditions that led to the formation of the Peace Vallis fan.

This analysis of Peace Vallis fan formation differs in some ways from that proposed by Anderson and Bell [2010]. They propose that the light toned fractured deposit (their “high thermal inertia fan shaped unit”) is a relatively thick deposit that underlies the crater floor and

that the “low thermal inertia unit” is a thin overlying mantle [Figure 47, Anderson and Bell, 2010]. They point to fractures in the low thermal inertial unit as evidence of a thin mantle that transmits the fractures. Fractures in the low thermal inertia unit are not strongly developed and crater walls in the upper fan (low thermal inertia unit) do not show the fractured, resistant material. More importantly, Sumner et al. [2013] show the transition from upper to lower fan to be gradational and intermixed.

4.9 Summary

The Peace Vallis fan is defined by convex contours that extend across the lower portion of the northern Gale Crater wall and terminate at the topographically diverse floor at the base of Aeolis Mons. Covering about 80 km², the upper portion of the fan is a relatively fracture-free, mottled surface and is laced on its western margin with frequent ridges interpreted to be paleo-channel beds. The lower portion of the fan has high thermal inertia compared to the upper portion, is light-toned, and generally highly fractured. This lower portion is deflated locally up to 15 m, indicating that it is at least partially composed of finer sediment that is capable of being swept away by wind. The fan also varies laterally from a western side that slopes primarily to the south and contains the paleo-channels, and an eastern side that slopes towards the southeast and has only a few low-relief paleo-channels. The western portion of the fan terminates on a broad, nearly flat ridge that extends across from Aeolis Mons to the crater wall. The eastern portion of the fan may have prograded onto and possibly across the pre-existing Bradbury Rise (the *Curiosity* Rover landing area) and descended to a topographic low to the east of Bradbury rise. The longitudinal profile of the fan progressively declines in slope from about 3% at the fan apex to 1% or less where the fan progrades across pre-existing topography. The eastern portion of the fan enters an enclosed basin, leading to the prediction that the distal end of the eastern fan included lake deposits, which was subsequently demonstrated by observations obtained by the *Curiosity* rover [Grotzinger et al., 2014].

The location of the Peace Vallis fan results from a 15 km setback in the crater rim, which enables drainage from the adjacent plains near the crater rim to spill into a canyon where an incised valley feeds the apex of the fan. A catchment area up to 1021 km² could have fed the valley. The entire incised valley network that fed the fan with water and sediment appears to be excavated into previously deposited debris, likely of colluvial (i.e. mass transport), eolian, fluvial and possibly air-fall origin. Estimates of volume removed to form the incised valley (0.8 km³) match the total volume of the original fan (0.9 km³). This suggests that the fan records a distinct episode of valley incision and fan development. Given the preservation of the sharp incised valley walls and fan topography, no further water-driven processes have modified these features (but wind transport has actively eroded fine grained material and infilled craters).

The gradient of the uplands plains valley networks (~2%) and their considerable distance from steep adjacent slopes in the canyon indicate that incision here was driven by fluvial transport, likely in gravel-bedded channels. The main canyon valley and its adjacent tributaries are much steeper, and locally the mainstem approaches 10%. Here, mass flows may have originated and carried sediment down to the fan. The progressive decline in fan gradient, an apparent reduction in boulder abundance, the preservation of braided paleo-channels in the western portion of the fan, and the transition to actively wind deflated (and thus finer grained)

laterally extensive beds, all point to a predominance of fluvial transport in fan construction. Given the low ratio of sediment volume to water volume that typifies the fluvial transport of sand and coarser materials, fan construction would require at least 600 km^3 and perhaps more than 6000 m of runoff from the 730 km^2 watershed. As others have concluded for similar fans on crater walls on Mars, the construction period must have taken at least 1000s of years if not much longer [Grotzinger et al., 2014].

The source of water for valley incision and Peace Vallis fan development is suggested by the following observations: (1) the valley system, which does form a branching network, is of limited extent (drainage density is about 0.15 km/km^2), (2) the channel tips originate either in broad inclined plains in the uplands or on steep hillslopes, and the tips do not form amphitheater headwalls, (3) a large volume of water is needed to deposit the fan ($>1000 \text{ km}^3$), and (4) valley incision occurred after Gale impact at about 3.6 Ga. The geomorphic features of the valley network, the lack of a large elevated source area, and the requirement of a large volume of water makes groundwater flow an unlikely water source. The lack of channel networks below the source area (that might record runoff from rainfall) and the age of the Gale impact suggest that snow accumulation and melting is a probable explanation for the dominant water source. The large amount of water needed to build the fan requires there to be, however, a hydrologic cycle.

The Peace Vallis fan provides a framework for predicting stratigraphy and potentially habitable environments. With regards to observations by the *Curiosity* Rover, the sediments associated with the fan were derived from previously deposited debris, not directly from erosion of bedrock. This adds an extra step in exploring the controls on sediment composition as influenced by weathering and transport. The enduring activity for 1000s of years for fan development, similar to that reported elsewhere for fans on crater walls, forces an examination of how climate conditions could develop to create a persistent hydrologic cycle of water storage, runoff and erosion. The fan crossed a pre-existing low gradient crater wall and floor, and its topographic form suggests that it may have on-lapped pre-existing topography at its distal margins. The origin of this pre-existing topography, like the accretion of debris on the Gale walls and the generation of local ridges, rises and basins, is not yet explained. Further clues may well come from what lies in the exposures awaiting *Curiosity* at the base of Aeolis Mons.

4.10 Acknowledgements

This research was conducted for the Jet Propulsion Laboratory, California Institute of Technology, under a contract with NASA under the Mars Program Office. We would like to thank Alan Howard, the entire MSL team, but especially John Grotzinger, Bernard Hallet, Ron Sletten, and Mike Malin for insightful discussions along the way. We greatly appreciate the comments and suggestions provided by our reviewers, especially those from Bob Craddock. We also thank the teams responsible for the orbital data, namely the HIRISE, CTX, and HRSC teams. We give a special thanks to Malin Space Science Systems, the USGS at Flagstaff, and Scott Mansfield, Jonathan Joseph, and Thomas Gautier from Cornell for image and data processing.

4.11 Tables

Analysis	Data Source
1. Fan Planform (Contours, Extent, Long Profiles)	HCN DEM
2. Sediment Budget	MSL DEM, CTX
3. Paleo-channels	MSL DEM
4. Catchment Characteristics	MSL DEM, CTX
5. Fan Reconstruction/Mini-Fan	Delta Radii DEM
6. Lake Levels / Gale Fans and Deltas	HRSC DEM, CTX
7. Crater Counts	MSL DEM, CTX DEM, CTX

Table 4.1 Data sources used for each analysis

Property	Value	Range	Units	Data Source
Catchment Drainage Area	730		km ²	CTX
Catchment Relief	2		km	HRSC
Catchment Valley Length	107		km	CTX
Catchment Valley Width	300	80-900	m	HIRISE, CTX
Catchment Valley Eroded Volume	0.77	0.7-0.9	km ³	HIRISE, CTX
Catchment Valley Average Slope	0.05	0.025-0.07	--	CTX
Catchment Drainage Density	0.15	--	km	CTX
Fan Surface Area	80		km ²	HIRISE
Upper	49	--	km ²	
Lower	31		km ²	
Fan Volume	0.9		km ³	HIRISE
Current	0.7	0.4-0.8	km ³	
Eroded	0.2	0.15-0.25	km ³	
Average Fan Depth	7		m	HIRISE
East	6	4-9	m	
West	8	4-11	m	
Overall Fan Gradient				HIRISE
East	1.2		--	
West	1.5		--	
Fan Paleo-channel Width	27	7-110	m	HIRISE
Fan Paleo-channel Deflation Depth				HIRISE
East	0.6	0.3-1	m	
West	2.4	0.5-5	m	
Fan Relief	275	--	m	HIRISE
Fan Apex Angle	60	--	degrees	HIRISE
Dist. Apex to Bradbury Landing	15.2	--	km	HIRISE

Table 4.2 General properties used to describe the Peace Vallis Fan system

4.12 Figures

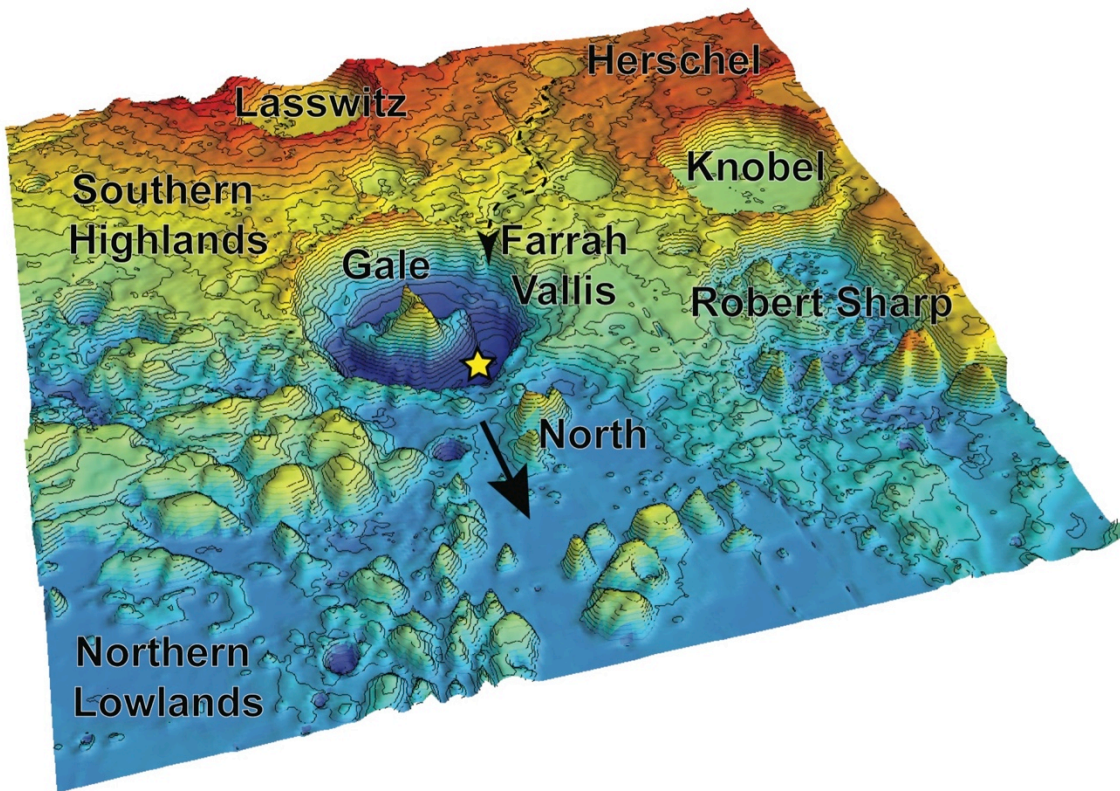


Figure 4.1 MOLA-derived topographic and elevation map of the Gale Crater region overlain with 250 m contours. Gale Crater is located on the crustal dichotomy between the cratered southern highlands and the smoother northern lowlands of Elysium Planitia, and it is this regional slope that contributes to the difference in elevation between the northern and southern rim. The lowest elevations in Gale crater occur on the northwest portion of the crater floor, near the MSL landing site (denoted by a star). Recent crater counts on Gale's ejecta give an age estimate of $\sim 3.6 \pm 0.05$ Ga, corresponding to the Early Hesperian or possibly the Late Noachian [e.g. Le Deit et al., 2013]. The white dashed line between Herschel crater and Gale crater shows a possible paleo-flowpath for water into Gale crater (based on Irwin et al. [2005]).

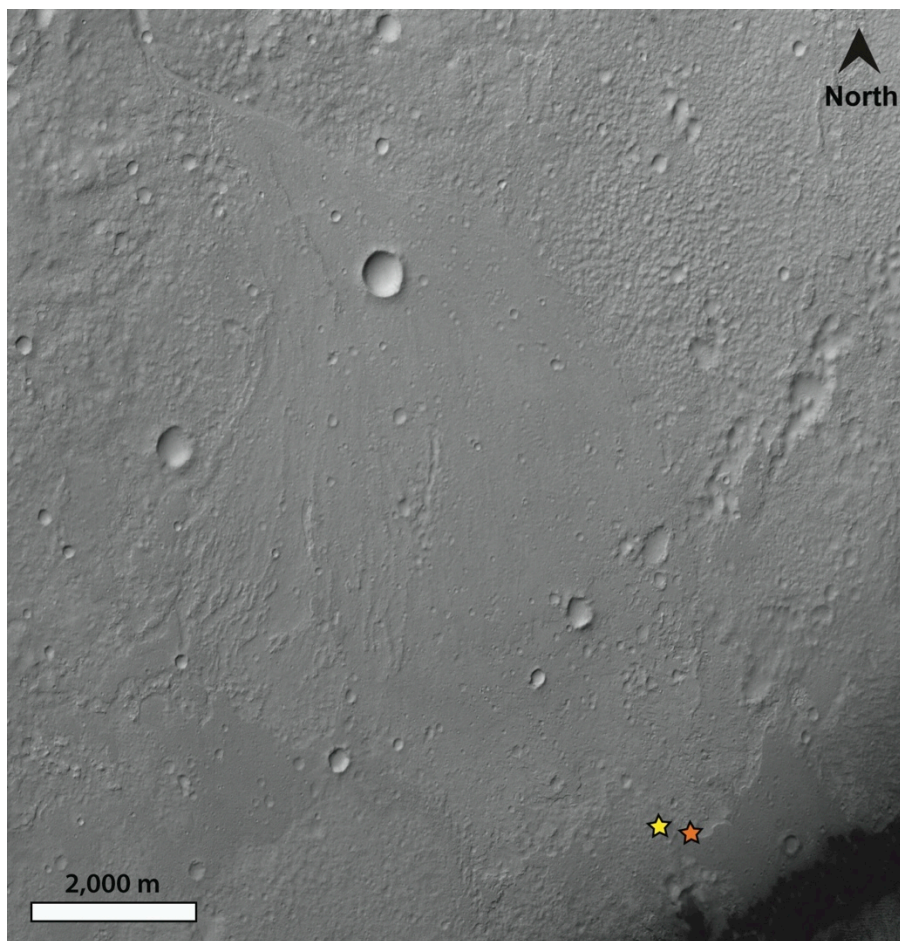


Figure 4.2 CTX imagery (6 m/pixel, image: P22_009571_1756_XI_04S222W) of the Peace Vallis fan. The location of the MSL Curiosity rover landing site (Bradbury Landing) is indicated by the yellow star and Yellowknife Bay is indicated by the orange star. The fan apex is located in the northwest corner of the image and the fan spreads south and southeast.

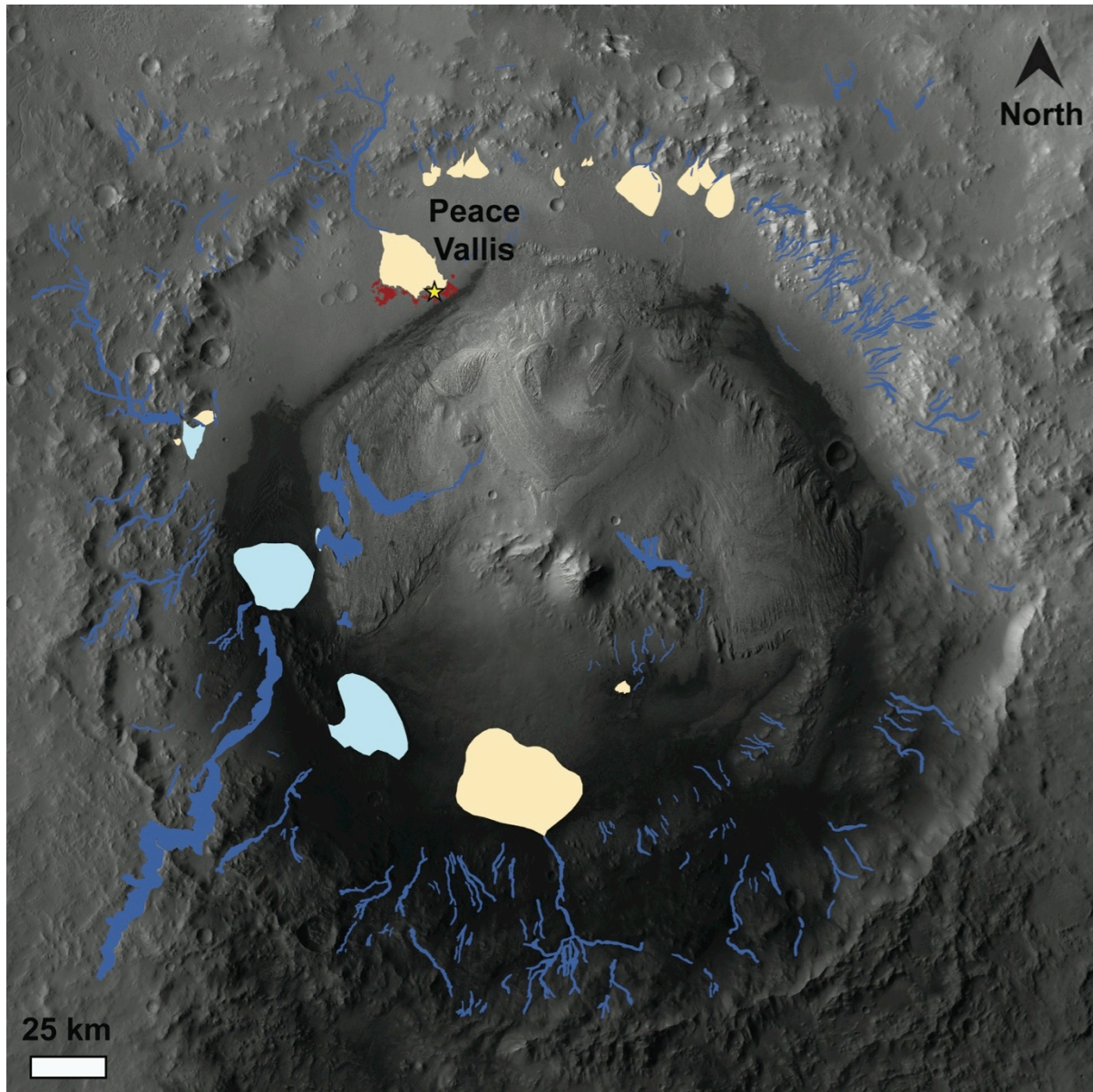


Figure 4.3 A modified version of the geomorphic map from Anderson and Bell [2010] showing their map of “valleys and canyons” (dark blue), “sinuous ridges” (white), and the Peace Vallis fan (labeled) which has cratered surfaces (red) surrounding its distal end. The cratered surfaces are part of Anderson and Bell’s [2010] “mound-skirting unit”. Several large fans (tan) and fan-deltas (light blue), including a large delta feature in the southwestern corner of Gale [Irwin et al., 2005], were added to their map.

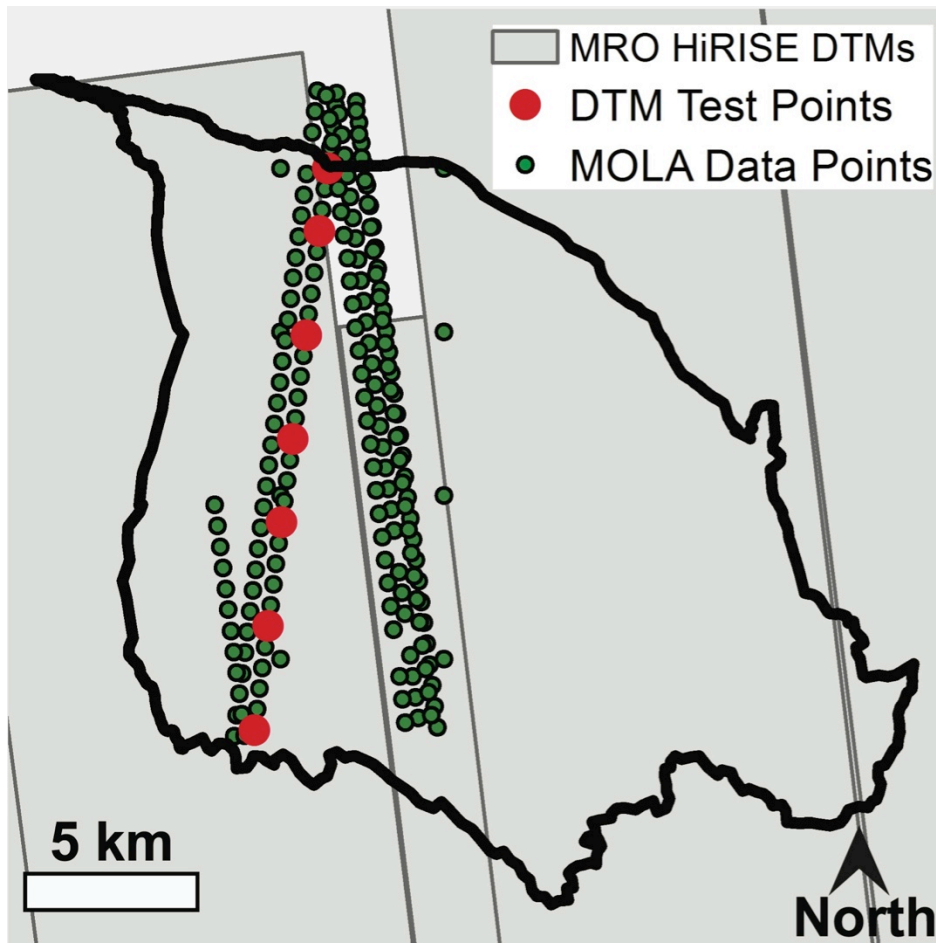


Figure 4.4 The location of MOLA data points (green) in the vicinity of the Peace Vallis fan. DEM test points (red) were used to compare MOLA elevations with those extracted from the MSL DEM and the HCN DEM created for this manuscript. The scarcity of MOLA data along the edges of the HiRISE stereo images (footprints shown in grey) prevents a data project that is wholly free from topographic artifacts.

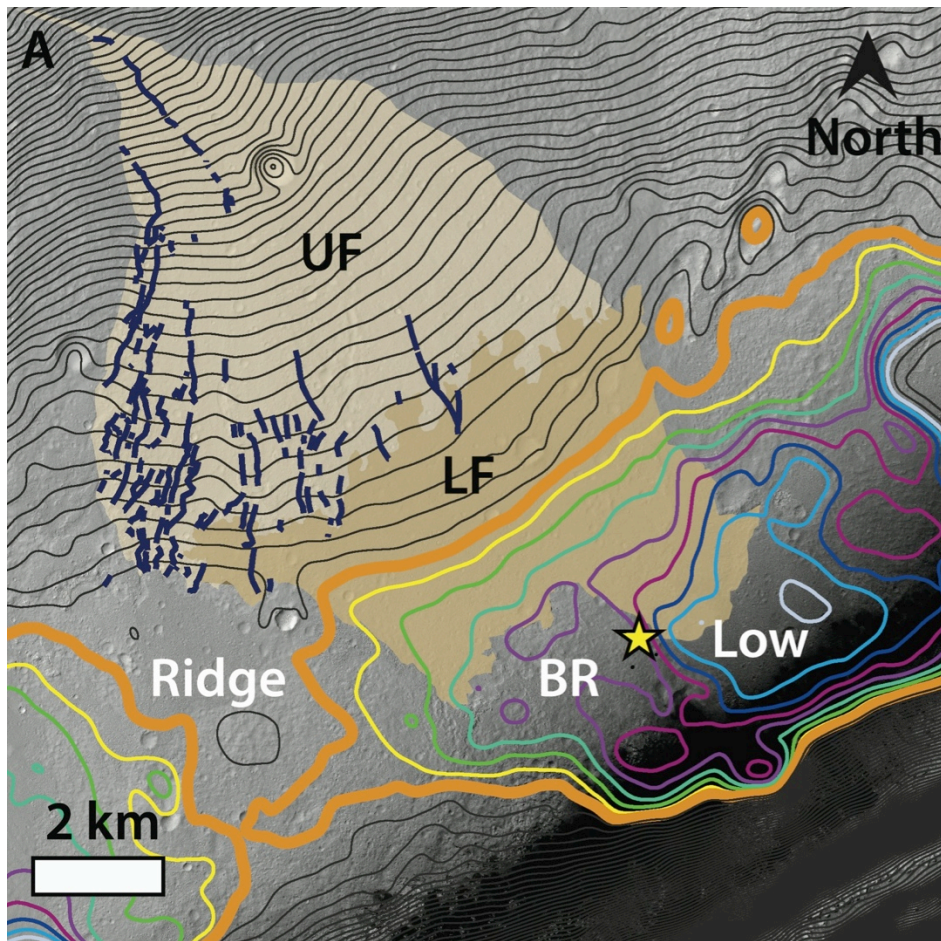


Figure 4.5a

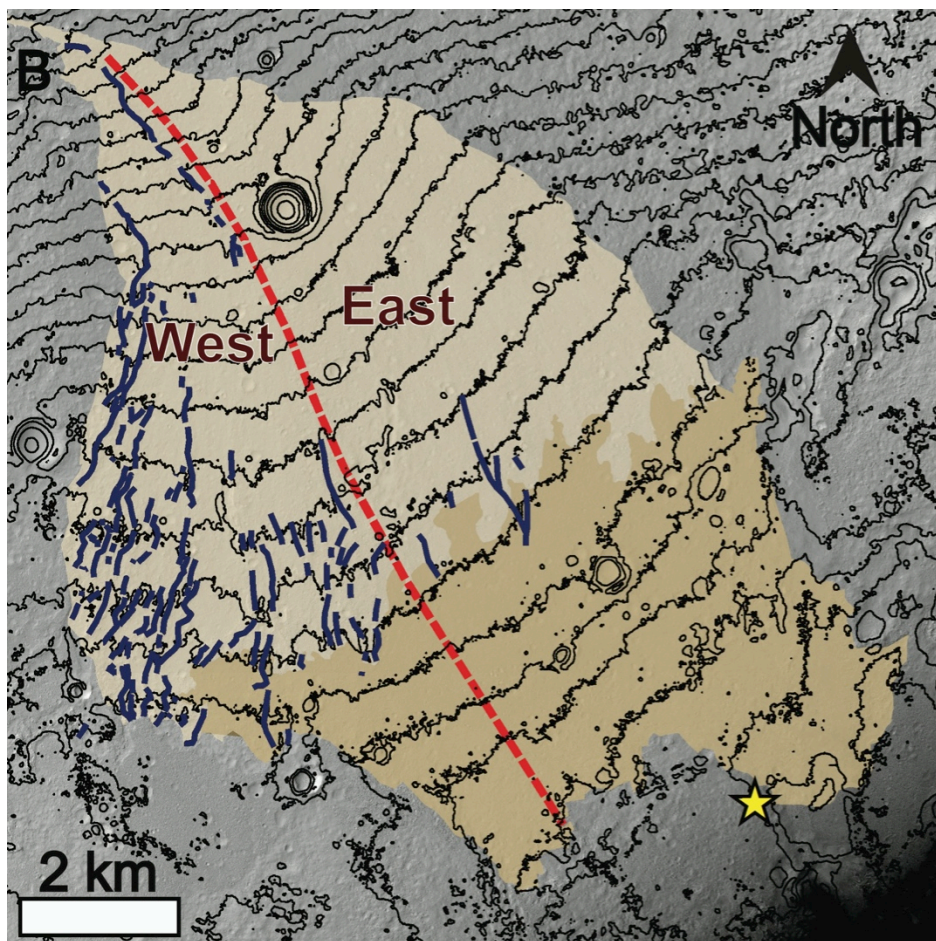


Figure 4.5b

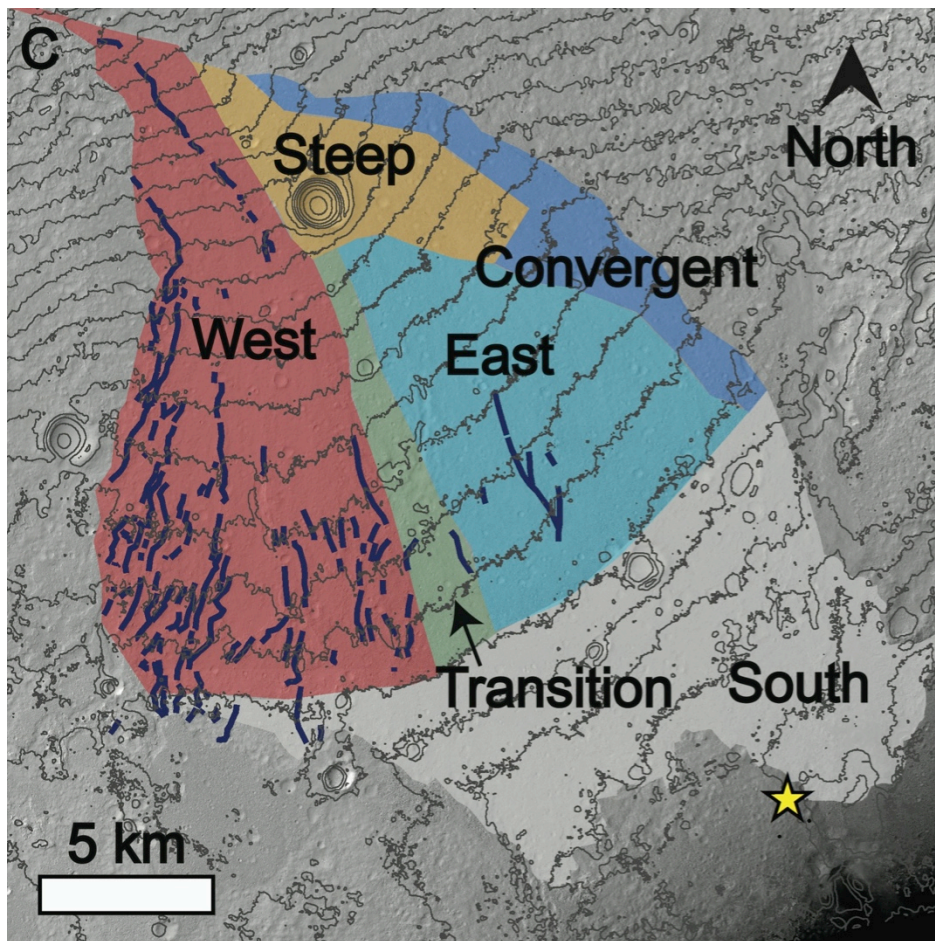


Figure 4.5c

Figure 4.5(a) Anderson and Bell [2010] divided the Peace Vallis fan into an upper “smooth, lower thermal inertia unit” and a lower “high, thermal inertia unit.” Shown here is a revised mapping of their upper/lower fan boundary by the MSL team, where the upper fan (UF) is shown in tan and the lower fan (LF) is shown in light brown. Overlain on the mapping are 5 meter smoothed HiRISE contours, which show the continuation of the fan form from the upper to lower fan. The western edge of the fan terminates on a flat-topped broad ridge, while the eastern fan enters into an enclosed basin (denoted by the thick orange contour) that contains Bradbury Rise and spills into a local topographic low. From HiRISE imagery, 43 distinct ridges were mapped (blue). (b) The 5 meter smoothed contours have been replaced by 10 m contours created from the HCN DEM. A red dashed line denotes a division between the western fan, which slopes to the south and contains most of the mapped ridges, and the eastern fan, which has few ridges and spills to the southeast into the enclosed basin. (c) 10 m contour map with topographic domains highlighted that may record specific phases or process changes in fan development. The distinction is defined by areas that share a common planform curvature, as determined by estimating a focus of radiation and fitting lines of constant arc to successive contour lines.

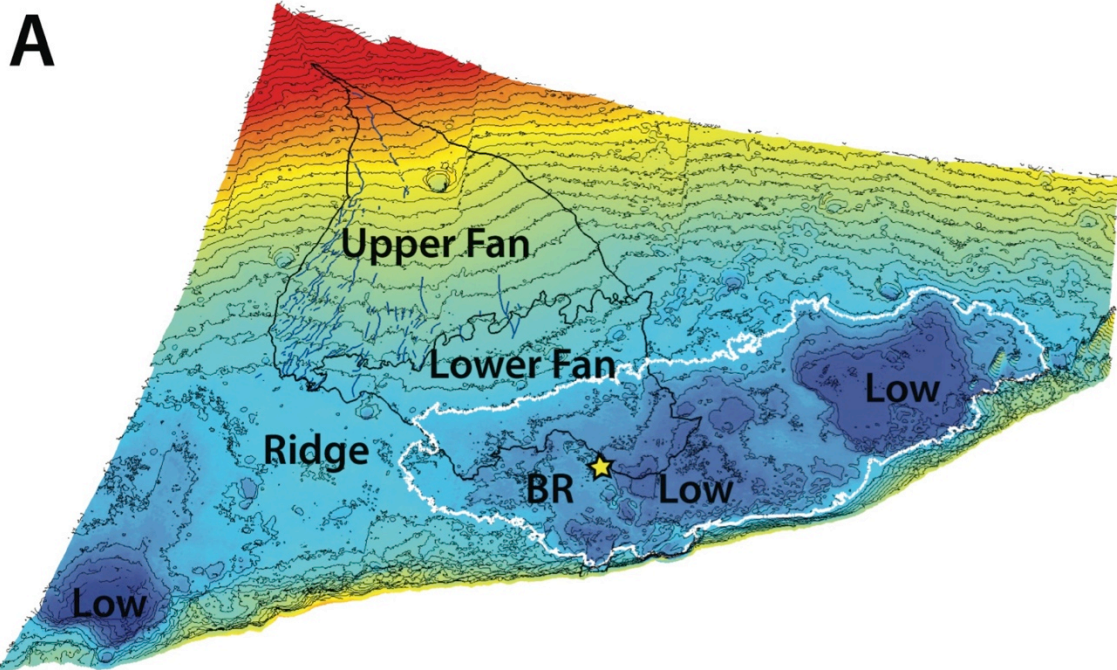


Figure 4.6a

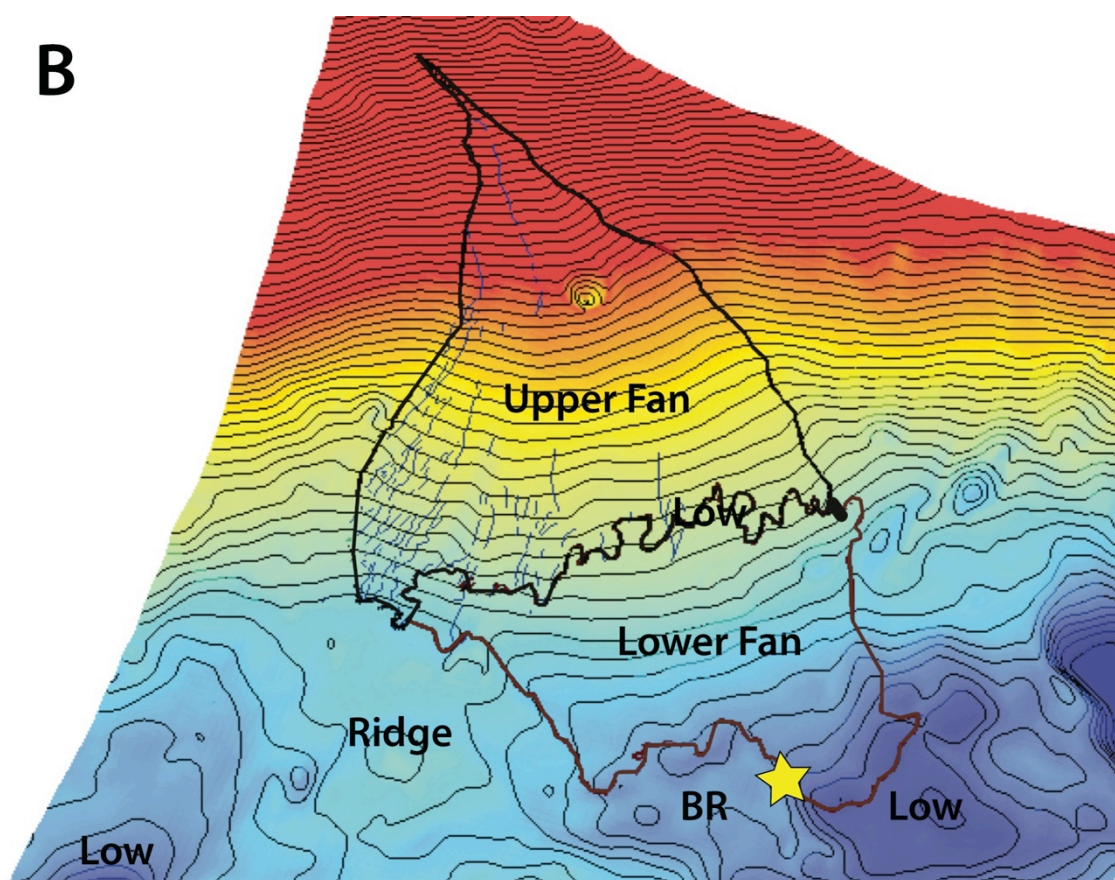


Figure 4.6b

Figure 4.6 (a) A three-dimensional view of the Peace Vallis fan system (vertical exaggeration of 10) with a colorized elevation map and 10 m HiRISE contours. Adjacent to the “ridge” and within the enclosed basin (denoted by the thick white contour) is Bradbury Rise (“BR”), and to the east of Bradbury Rise is a local low into which the lower fan enters. A second, larger low lies adjacent to the eastern border of the enclosed basin. (b) The three boundary conditions discussed in Section 4.3.1 (“ridge”, “Bradbury Rise” and “Low”) are highlighted.

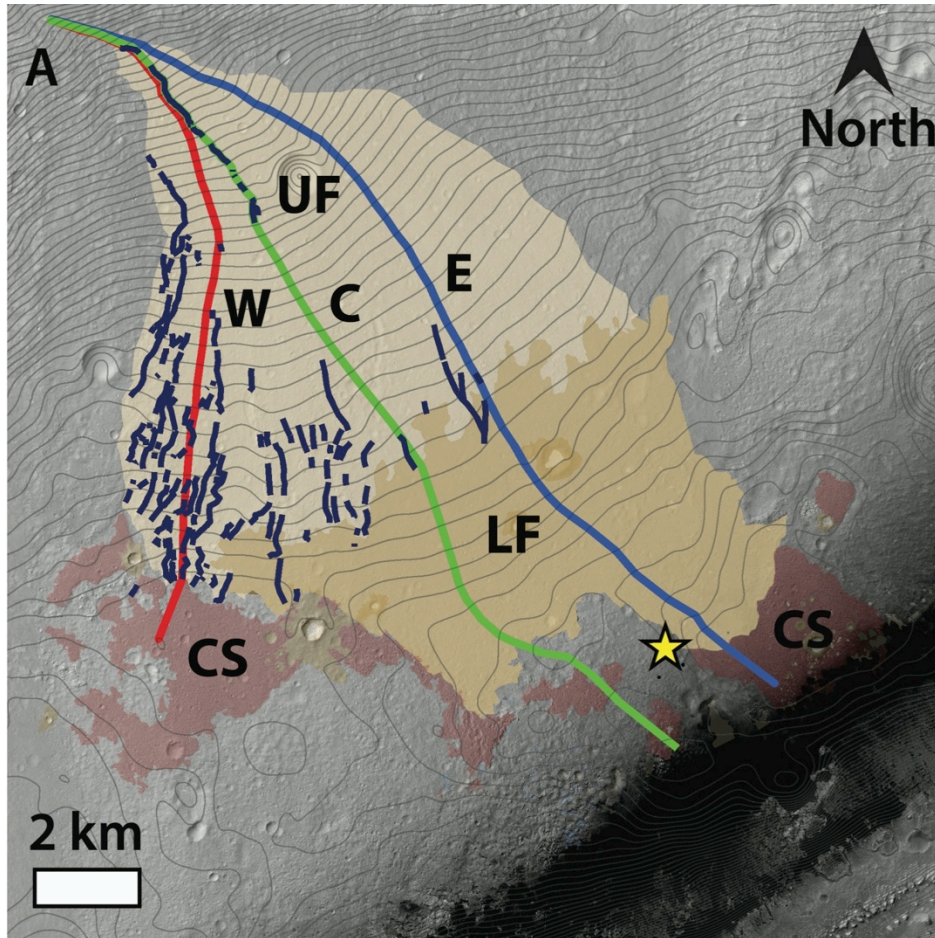


Figure 4.7a

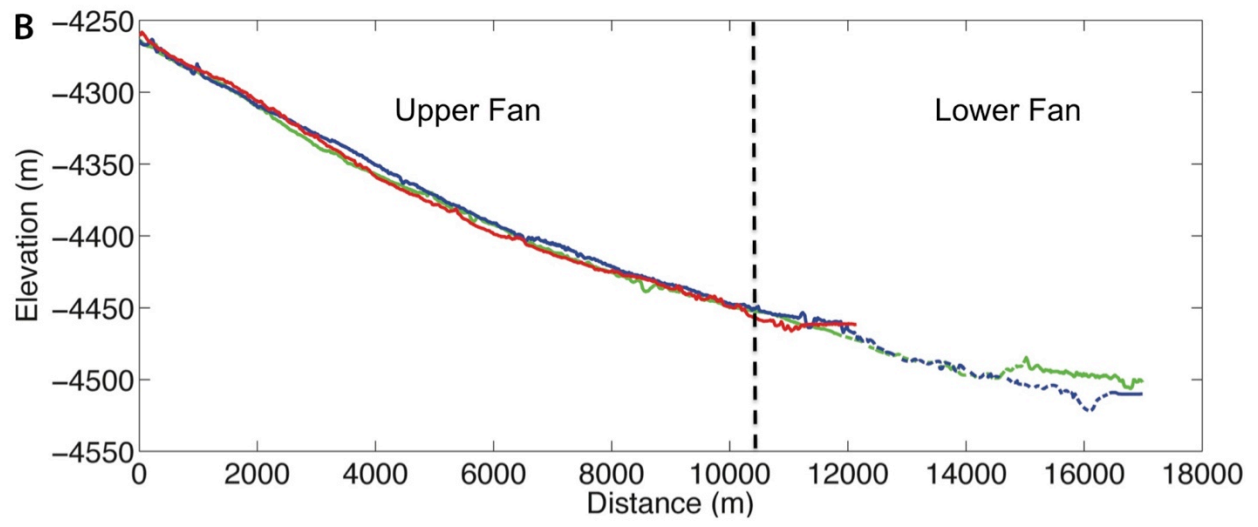


Figure 4.7b

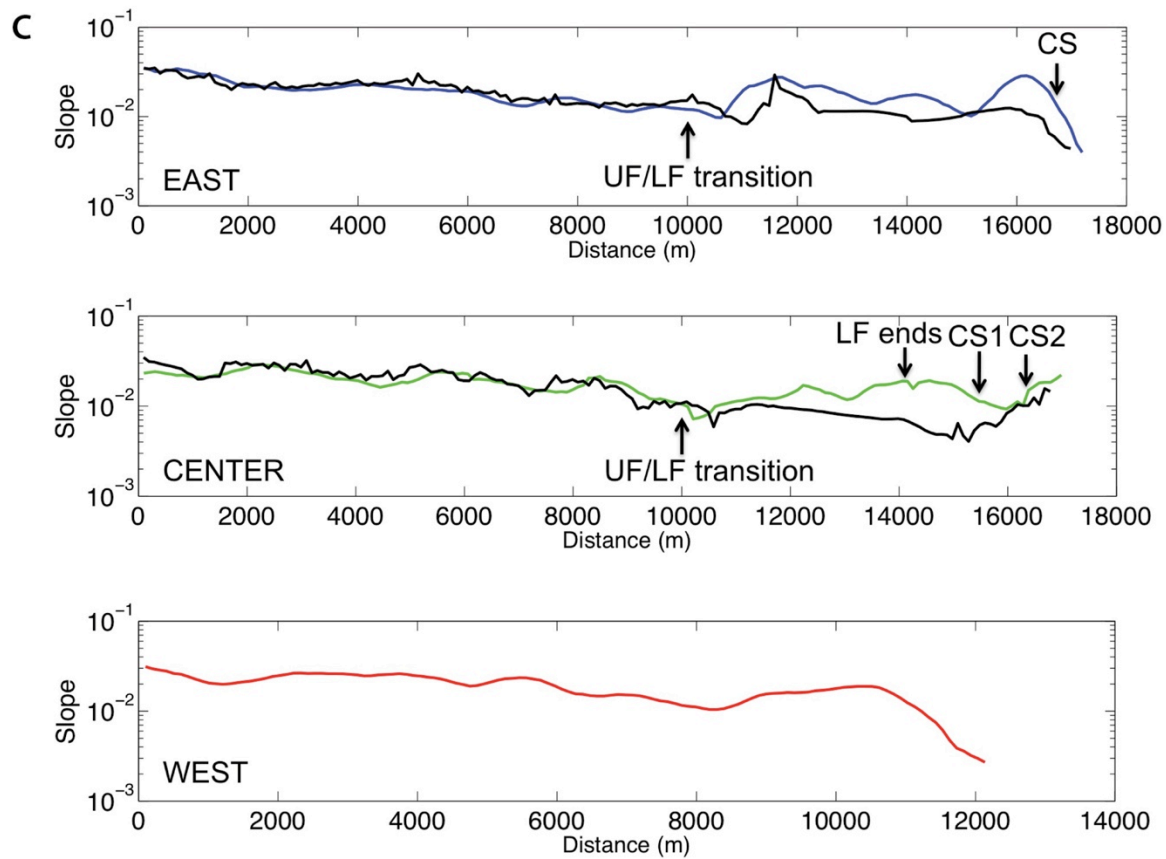


Figure 4.7c

Figure 4.7 (a) The location of three longitudinal profiles down the fan are shown; each profile starts at the fan apex and terminates across one of the lower three boundaries discussed in Section 4.3.1. The western profile (W, red) terminates on a cratered surface to the west (CS), the center profile (C, green) runs onto and across Bradbury Rise, and the eastern profile (E, blue) terminates on a cratered surface to the southeast. (b) Elevation versus distance data (extracted from the HCN DEM) are plotted for the three longitudinal profiles drawn in Figure 4.7a. The dashed line marks the boundary of the upper and lower fan. All three profiles are similar in form, but begin to differ as they approach their lower boundary. For the center and eastern profiles the portion of the profile most heavily influenced by erosion is shown as dashed lines. Ignoring the lower boundaries, all three profiles have relatively low profile concavities of ~ 0.0002 , similar to concavities measured by Moore and Howard [2005] for terrestrial and Martian fans of similar slope (1.5%). (c) The average fan slope was calculated every 100 meters and the 10-point average was plotted as a function of distance for each of the three longitudinal profiles. For the eastern and center profiles, a constructed fan profile, which accounts for post-depositional fan erosion, is also plotted (black). In each case the longitudinal profiles progressively decline from 3% at the fan apex, down to 1%, at which point the fan slope is controlled by pre-existing topography. The transition from the upper fan (UF) to the lower fan (LW) is marked for the center and east profiles.

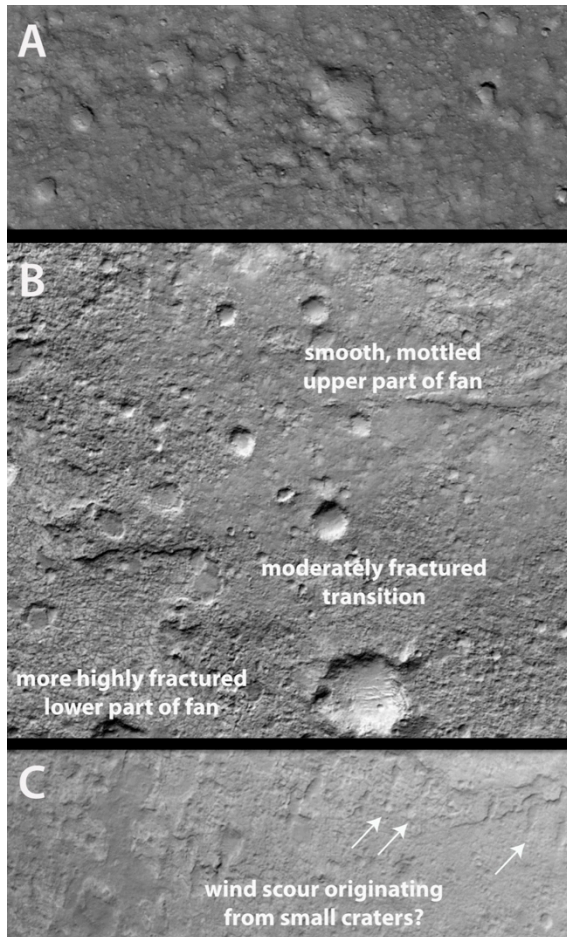


Figure 4.8 HiRISE imagery (0.25 m/pixel) of the Peace Vallis fan. (a) The upper fan is characterized by a relatively smooth, mottled surface that corresponds to the low thermal inertia unit (image: PSP_003453_1750_RED) (b) In the transition zone between the upper and distal fan, the mottled and fractured surfaces are intermixed (image: PSP_003453_1750_RED) and (c) Wind-scoured elongated troughs across the exposed bedrock may originate as small craters (as indicated by the white arrows) (image: ESP_018920_1755).

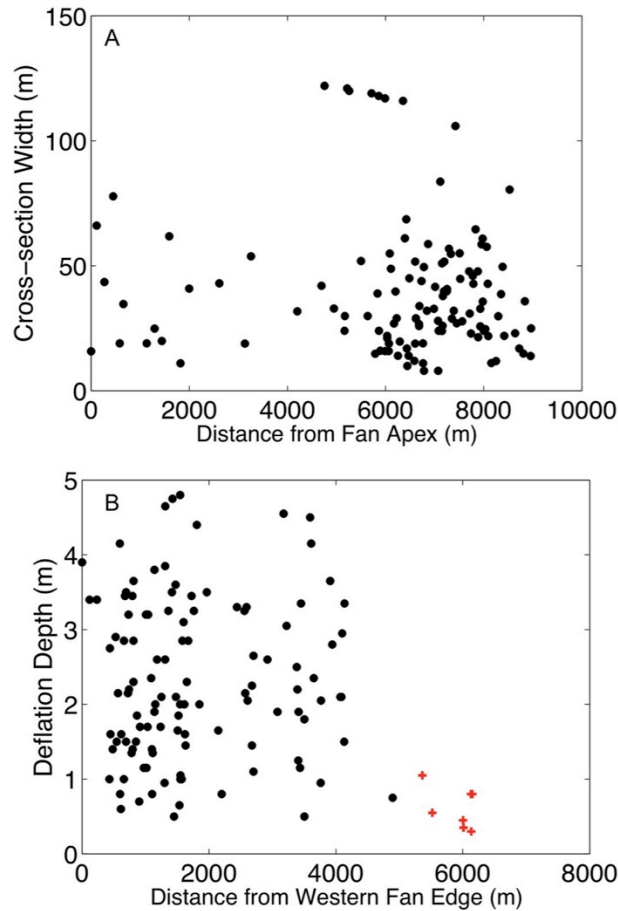


Figure 4.9 Cross-sectional data were extracted at 122 locations across the mapped ridges, interpreted to be paleo-channels, using MSL DEM data to determine the average ridge width and deflation depth (a) Cross-sectional width versus distance from the fan apex shows that there is no systematic change in ridge width as a function of distance downslope (b) deflation depth going from the edge of the western fan to the eastern fan shows a marked difference between the western and eastern fan, with an average depth of 2.4 m and 0.6 m, respectively.

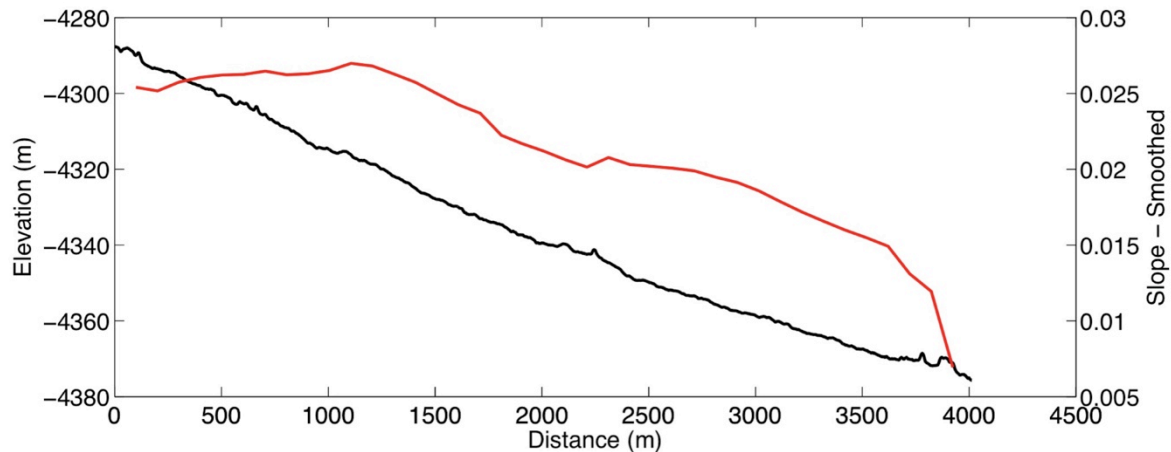


Figure 4.10 A longitudinal profile (black line) for the uppermost paleo-channel on the Peace Vallis fan. Elevation data were extracted from the HCN DEM and plotted as a function of distance on the left y-axis. Slope was calculated every 100 meters and the 10-point average is plotted as a function of distance. The slope declines from 2.5% near the fan apex to 1.5% near where the channel terminates.

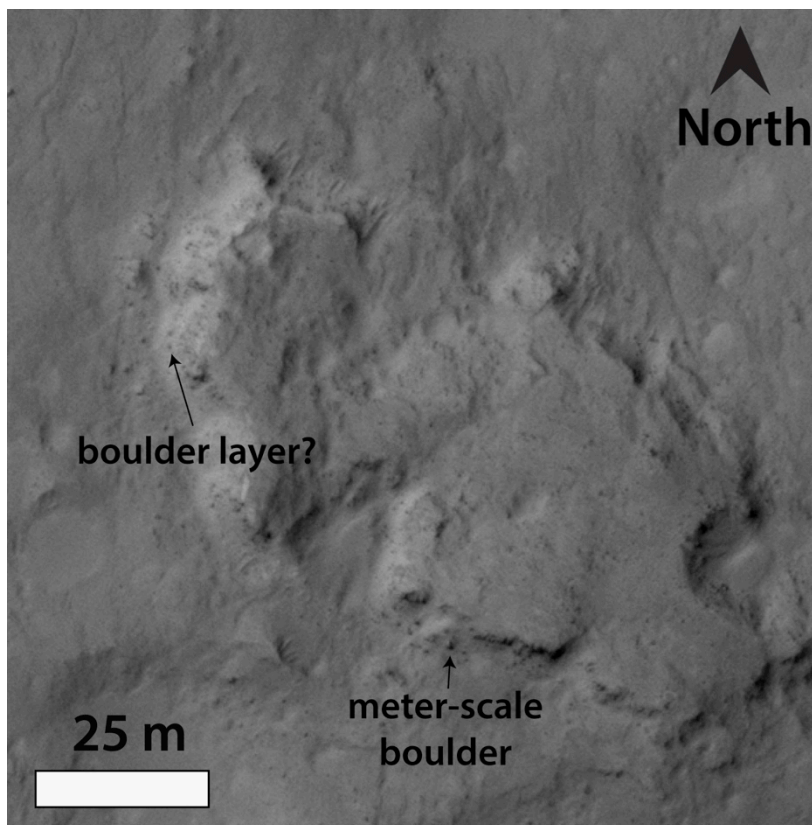
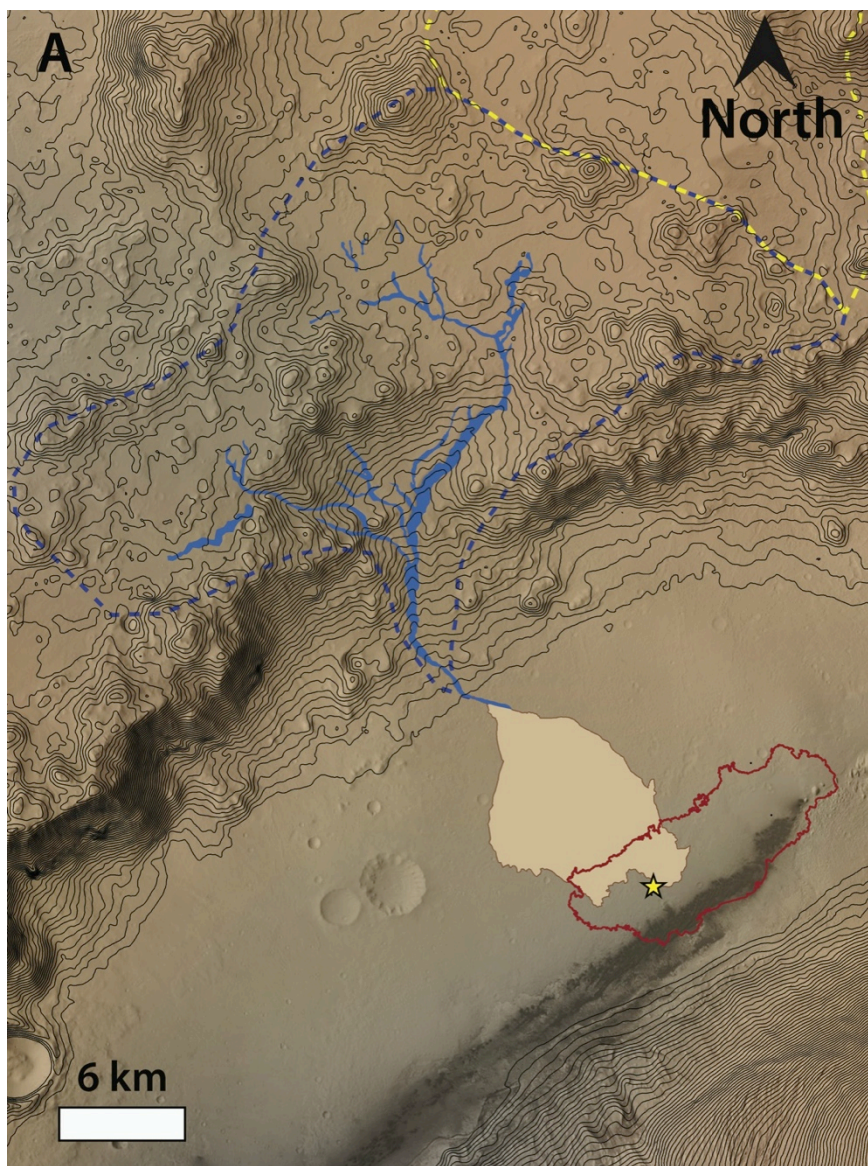
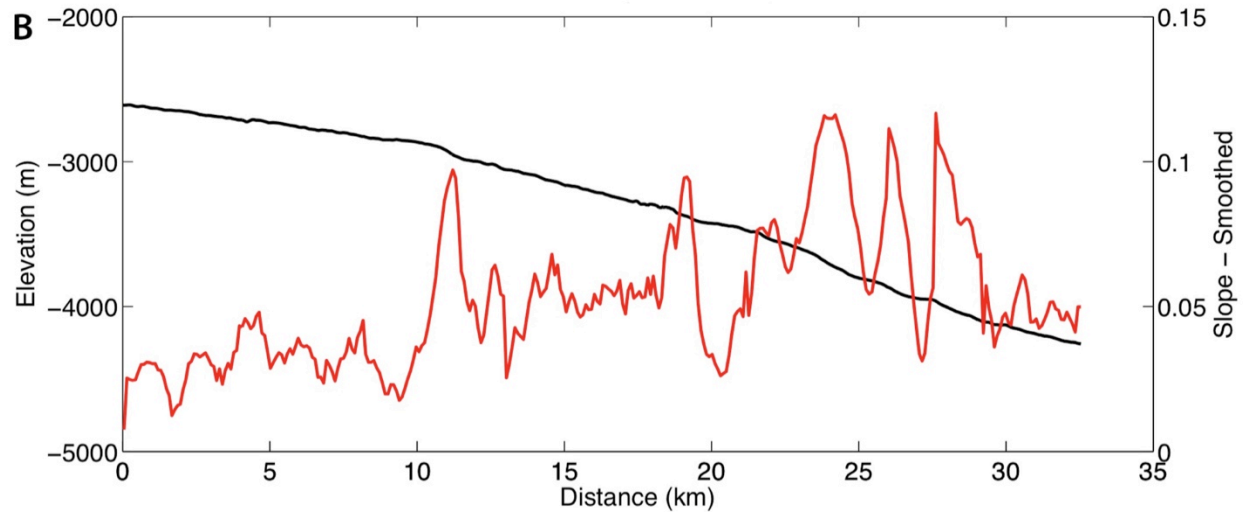


Figure 4.11 HiRISE image PSP_009716_1755 (25 cm/pixel resolution) showing a close-up of the main central ridge located near the fan apex. The ridge is ~50 meters wide and shows alternating layers of light-toned material and darker boulder-rich deposits that are approximately 0.25 to 1 meter thick. The individual boulders exposed in the outcrops in some of these channel deposits are meter to sub-meter scale and they appear to decrease in density with distance from the ridge.

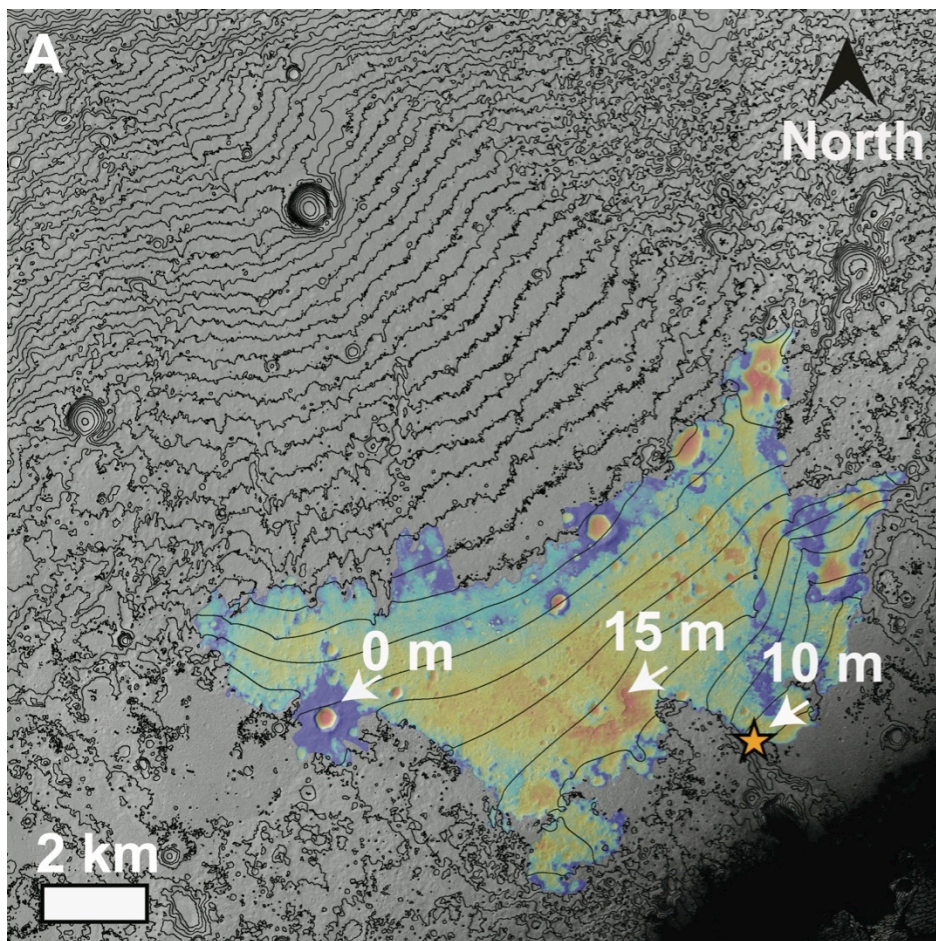


a.

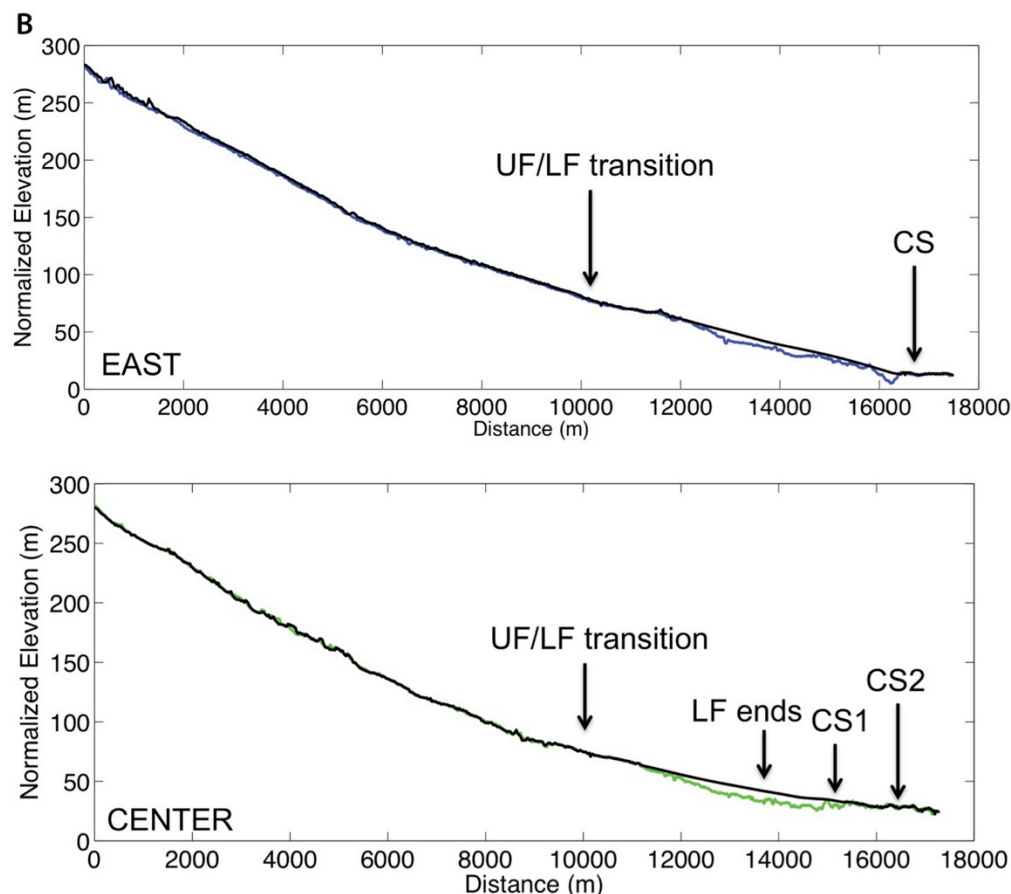


b.

Figure 4.12 (a) Peace Vallis fan emanates from a well-defined incised valley network (light blue) located within a large topographic drainage area. The lower catchment (730 km²) is dashed in blue and the upper catchment (291 km²) is dashed in yellow, such that an area of 1021 km² can drain to the outlet valley. 50 m HRSC contours are shown for the catchment region only (starting above the fan apex) to highlight drainage from the upper plains, where the most distal tips of the valley network originates, and the canyon from which the incised valley network emanates. The large enclosed basin into which the lower eastern portion of the Peace Vallis fan spills is outlined in red. “A” marks the start of the longitudinal profile shown in Figure 4.11b and “B” marks the end. (b) Elevation data was extracted and plotted as a function of distance (from a CTX-derived DEM) on the left y-axis (black line) and on the right y-axis the slope was calculated every 100 meters and the 10-point average is plotted as a function of distance (red line). The average valley slope in the upland plains is roughly 0.02 (from 0 to 10 km), while the mainstem reaches slopes locally of 0.1, which is in the mass flow regime for sediment transport.



a.



b.

Figure 4.13 (a) Shown here are 5 m pre-eroded fan contours that were calculated by projecting a continuation of the current-day upper fan topography to the dark cratered unit to the southeast and other neighboring areas that have experienced minimal erosion. An erosion map was created by differencing the reconstructed fan surface with the current-day fan surface, giving a total erosion volume of 0.14 km^3 from the lower fan and maximum erosion depths of up to 18 meters locally. The location of the John Klein drill site [Grotzinger et al., 2014] is marked with a star. (b) Reconstructed fan longitudinal profiles (black) are plotted on top of the current day longitudinal profiles for the eastern (blue) and center (green) fan to show the extent of erosion on the lower fan. The average reconstructed fan slope was calculated every 100 meters and the 10-point average was plotted as a function of distance for the center and east longitudinal profiles in Figure 4.7c (black).

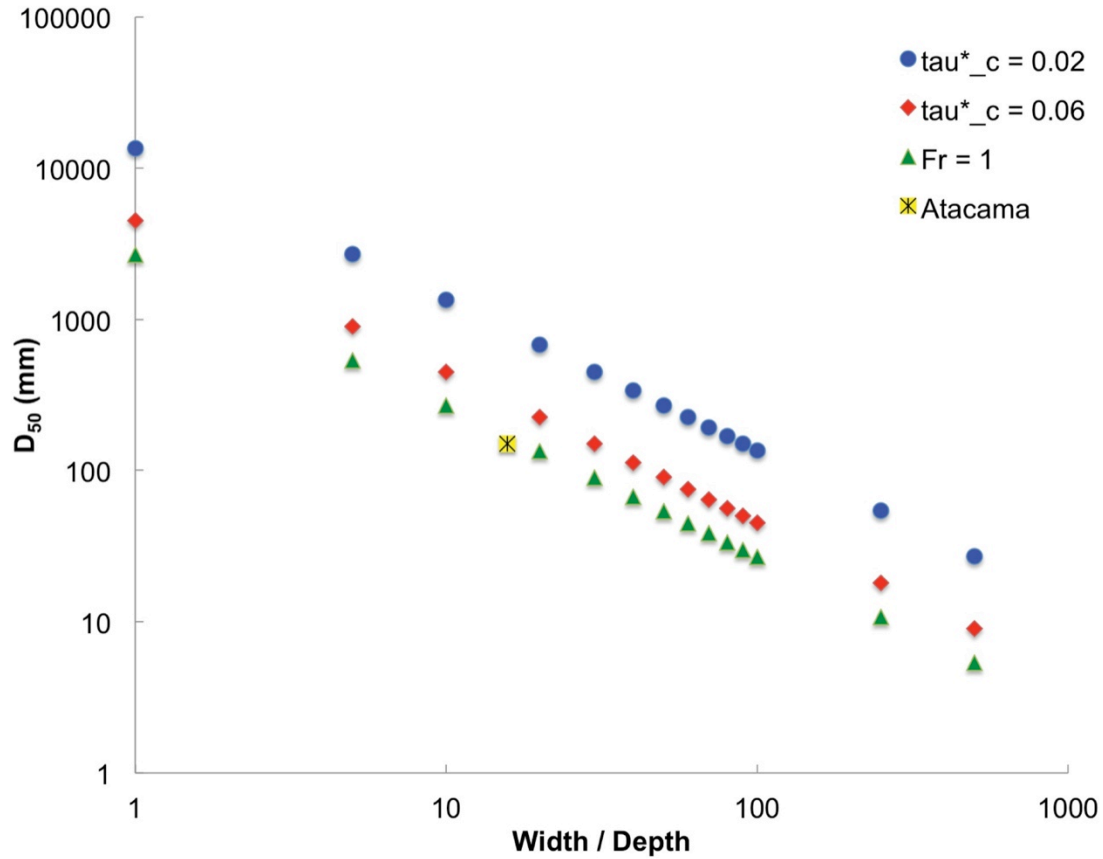


Figure 4.14. Using threshold channel concepts, the median grain size on a 2% channel bed (D_{50}) for given width to depth ratios (W/D) under typical critical shear stresses (defined as the shear stress over the grain weight) is plotted for sandy ($\tau^*_{cr} = 0.02$) to gravelly beds ($\tau^*_{cr} = 0.06$). Using the constraints that terrestrial alluvial channels typically have W/D ratios between 10 and 100, and Froude (Fr) numbers close to one [Stock et al., 2008], Peace Vallis fan would likely have transported gravel to possibly boulder sized material. Pebble count data from a terrestrial analogue fan in the Atacama Desert (Chile), which also has a mean slope of 1.5%, falls within the predicted ranges for Peace Vallis fan.

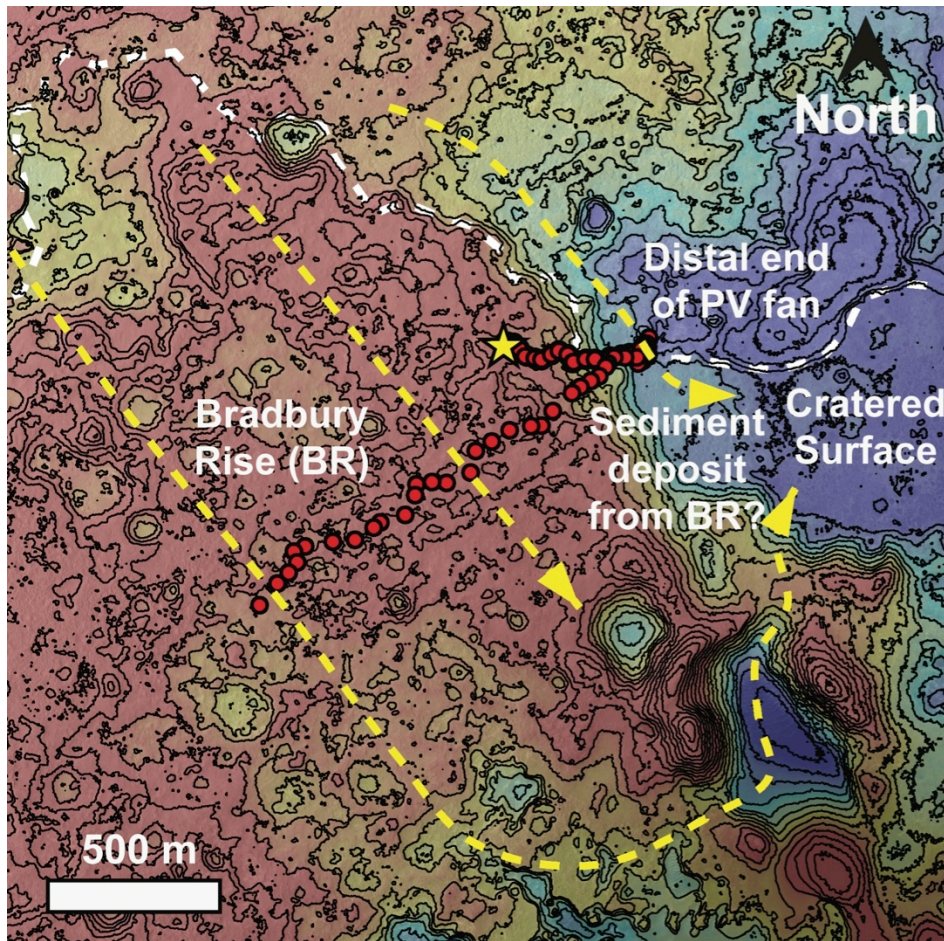


Figure 4.15. Colored elevation map with 2 m delta radii HIRISE contours overlaid. For reference, there is a 17 m difference between the elevation on Bradbury Rise and the eroded portion of the lower fan. The star marks Bradbury Landing and the red dots mark stops along the rover traverse during the first 360 sols. The yellow dashed lines show possible paleo-flow lines across Bradbury Rise and/or onto the cratered surface, creating what is possibly a small fan-like feature as sediment was deposited off of Bradbury Rise.

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Chapter 5: Sequence and Relative Timing of Large Lakes in Gale Crater, Mars

5.1 Abstract

Deltaic deposits across Mars have been identified and used in the past to document the presence of lakes within craters and large basins. The MSL mission to Gale Crater provides us with the opportunity to investigate one such crater where lakes have been hypothesized to exist using improved imagery, topography, and ground-based observations from the *Curiosity* rover. Here we report evidence for at least three large lakes within Gale, all which occurred after the central mound (named Aeolis Mons and referred to informally as Mt. Sharp) reached close to its current topographic form, with a possible time sequence of rising and falling lake levels. From corresponding deltaic deposits off the southern rim of Gale crater and the southern flank of Aeolis Mons, we identified the highest lake level at -3280 meters (based on HRSC-derived contours), which would have had a mean depth of 700 meters. The larger regional context suggests that this lake was derived from regionally sourced water. The next lake level, which was established after a period of drying and rewetting, is defined by four deltaic features, three sourced from sediment from Aeolis Mons and one off of the western rim of Gale, as well as the termination of gullies around the northern rim of Gale. This second lake has corresponds to the -3980 m HRSC contour, has a mean depth of 300 meters, and was likely sourced from water more local to Gale. Lake levels then rose to the third lake level, defined by the -3780 m HRSC contour, with a mean depth of 400 meters. Two deltaic deposits derived from sediment from the rim of Gale and the termination of several gullies define this locally sourced lake. This last lake then declined sufficiently quickly to prevent incision into any of the deltaic forms around the crater, though hydrologic activity continued as evidenced by a time of fan building around the crater. A quantitative topographic analysis of the fan and delta features and their source areas on the crater wall and Aeolis Mons suggest that mass is conserved for all but one system, hence these features have experienced little modification since emplacement. Importantly for the MSL mission, this suggests that most of the sediments the rover has and will encounter has experienced at least a few cycles of drying and wetting.

5.2 Introduction

On August 6, 2012, UTC, the Mars Science Laboratory (MSL) rover, *Curiosity*, landed in Gale Crater (5.3°S 137.7°E , Figure 5.1), a 150-km diameter impact crater located on the crustal dichotomy of Mars. Gale Crater contains a 5 km high sedimentary mound that shows a distinct mineralogical transition from phyllosilicates to sulfates, hypothesized to be the result of planet-wide changes in climate and hydrology [Milliken et al., 2010]. While the main target of the MSL mission is this mound, Aeolis Mons (informally referred to as Mt. Sharp), the final landing ellipse also contained the distal end of the Peace Vallis alluvial fan [Anderson and Bell, 2010; Palucis et al., 2014 - Chapter 4], which the rover traversed to from its original landing site, Bradbury Landing. Here at Yellowknife Bay, fine-grained sedimentary rocks, inferred to represent an ancient lake, were found to preserve evidence of a habitable environment during the post-Noachian period of Mars [Grotzinger et al., 2014]. Gale Crater hosts a number of other morphologic features, however, that are suggestive of past crater-wide fluvial-lacustrine activity whose extent were much greater than the distal region of the Peace Vallis fan. These include (Figure 5.2): incised valleys and valley networks, including Farah Vallis (the large canyon

entering Gale from the south), deltaic deposits, other fan deposits, and channels (both regular and inverted) [Cabrol et al., 1999; Pelkey et al., 2004; Irwin et al., 2005; Anderson and Bell, 2010; Dietrich et al., 2013 and 2014; Le Deit et al., 2013; Chapter 4].

Irwin et al. [2005] concluded that Gale was little modified since impact and that there was just a brief period of reactivation of a channel system from the south of Gale, originating near Hershel Crater and draining an area of 270,000 km². They noted that in the region to the south of Gale Crater, similarly sized craters appear substantially more degraded, are partially to nearly infilled, and have more subtle rim topographies, supporting that these craters likely predate Gale Crater. Ejecta from Gale's impact then partly buried the valley network originating near Hershel Crater (see Figure 4 in Irwin et al., 2005), evidence that the network was active before Gale's formation. This extensive regional drainage system likely provided the water needed to incise Farah Vallis and support lake and delta formation [Irwin et al., 2005]. Irwin et al. [2005] suggested this period of lake building occurred after Aeolis Mons sediments were deposited and perhaps partially removed by aeolian erosion [Malin and Edgett, 2000], likely within the late Noachian to early Hesperian epoch. Recent dating of Gale's ejecta using crater counts puts the age of Gale between 3.6 and 3.8 billion years ago, also in the Late Noachian to Early Hesperian epochs [Thompson et al., 2011; Le Deit et al., 2013], suggesting Aeolis Mons reached its current topographic form early on in Gale's history. This timing of lakes in Gale proposed by Irwin et al. [2005] also coincides with the timing of last activity within large valley networks across the southern cratered highlands before an abrupt cessation of valley network formation and the transition to more episodic and very localized valley network formation [Fasset and Head, 2008]. Others have also proposed large lakes in Gale [see review in Anderson and Bell, 2010; Le Deit et al., 2013; and Dietrich et al., 2013 and 2014], the extent and duration of which has major implications for the sediments being studied along *Curiosity's* traverse, as these lakes would have caused saturation of the deposits in Aeolis Mons and the crater walls and rim.

Despite the fact that climate models cannot simulate atmospheric conditions that could support standing surface water for an appreciable period of time [Haberle, 1998; Segura et al., 2002; Wordsworth et al., 2012], it is evident that at the global scale Mars was once able to support an active hydrological cycle. Evidence includes large deltaic deposits [Moore et al., 2003; Mangold and Ansan, 2006; Dibiase et al., 2013], meandering channels with cutoff loops [Moore et al., 2003; Malin and Edgett, 2003], alluvial fan deposits [Moore and Howard, 2005; Kraal et al., 2008; Morgan et al., 2014; Chapter 4], and well-developed valley networks [Craddock and Howard, 2002; Hynek et al., 2010]. Though lakes were proposed early on [e.g. Carr, 1983; Cabrol and Grin, 1999], higher-resolution imagery has allowed for the recognition of individual basins that may have supported lakes based on deltaic deposits at the mouth of inflowing valleys [Malin and Edgett, 2003; Howard et al., 2005; Lewis and Aharonson, 2006; Kleinhaus, 2010], though often these depositional features are degraded due to impact cratering, aeolian erosion and deposition, and/or mass wasting. Other evidence cited for former Martian lakes is the presence of large incised valleys entering enclosed basins [Irwin and Howard, 2002; Howard et al., 2005; Grant et al., 2011], incised valleys exiting enclosed basins via eroded exit breaches [Howard et al., 2005; Fasset and Head, 2008], clay-rich sedimentary deposits within enclosed basins [Ehlmann et al., 2008; Grant et al., 2011], and morphologic changes from canyons to fan deposits at common elevations within a basin [Dietrich et al., 2013]. Some have

proposed the possibility of a global ocean occupying the northern lowlands [e.g. Clifford and Parker, 2001; Perron et al., 2007; Di Achille and Hynek, 2010], though there is some debate in the literature [Malin and Edgett, 1999; Carr and Head, 2003]. As evidence for lakes across Mars grows, there remains uncertainty on the climatic and hydrologic environment that supported them, including disagreement on whether lakes and their associated features formed from multiple flow events lasting thousands of years or more [Moore et al., 2003; Wood, 2005; Barnhart et al., 2009; Kleinhaus, 2010] or during brief, intense periods of fluvial activity that lasted on the order of days to several years [Jerolmack et al., 2004; Kraal et al., 2008].

Here, we exploit HIRISE and CTX imagery, and photogrammetrically-derived digital elevation data from HIRISE, CTX and HRSC, to build on previous work that identifies several probable lake levels that would have occupied large portions of Gale Crater. We first identify and describe in detail several deltaic deposits sourced either from the rim of Gale Crater or from Aeolis Mons deposits, and for each we compare the eroded volume of material from valley/canyon incision to the volume of the deltaic deposit. These observations indicate at least three distinct lake levels developed within Gale (referred to herein as the Pancake, the Farah Vallis, and the Western lake stands). Observations in and around Gale also offer modest support to the proposal that Gale, lying as it does at the crustal dichotomy, may have been connected to a late Noachian ocean. We then argue that based on the geomorphology and our mass balance results, we can identify a clear sequence of lake level decline followed by lake level rise, as well as changes in the relative contribution of regional versus local water sources. Lastly, we propose a probable hydro-geomorphic sequence of events from the time of Gale's impact to present day.

5.3 Methodology

5.3.1 Topographic Data

As part of the entry, descent, and landing (EDL) modeling effort for the MSL mission to Gale Crater, a crater-wide 50 m/pixel digital elevation model (DEM) was made from Mars Express Orbiter High Resolution Stereo Camera (HRSC) images and was geo-referenced to Mars Orbiter Laser Altimeter data (MOLA) and mosaiced. For the landing ellipse and the proposed traverse up Aeolis Mons, twelve Mars Reconnaissance Orbiter (MRO) High Resolution Imaging Science Experiment (HIRISE) 0.25 cm/pixel stereo pairs (created by the USGS Astrogeology Center in Flagstaff, AZ) and three 6m/pixel stereo pairs from the MRO Context Camera (CTX) were geo-referenced and mosaiced as well. These combined datasets form a “resolution pyramid” to allow mapping within Gale crater at scales all the way down to a few meters or less [Golombek et al., 2012]. These datasets were projected into an Equidistant Cylindrical projection with zero as the center longitude and latitude, on a spherical Mars with a radius of 3396190-m. Most of the analyses presented here were derived either from the HIRISE topographic dataset (1 m/pixel elevation, red box in Figure 5.3) or from the HRSC topographic dataset (50 m/pixel elevation, covers the entire extent of Figure 5.3) (see Table 1).

In addition to the DEMs provided to the MSL team by the USGS (Flagstaff), we created several 18 m resolution topographic dataset using CTX stereo images. The images were controlled using MOLA shots and 16-20 tie points spread throughout the DEM boundary. The tie points were given z-control (elevation) using a MOLA DEM. This DEM was created to improve

upon the MSL-provided HRSC DEM in performing topographic analyses of several of the delta and smaller fan features within Gale Crater (black boxes in Figure 5.3). For consistency throughout the paper, the HRSC derived elevation will be reported for each delta front. Depending on the various datasets used (i.e. HIRISE and CTX), the contour line values are ± 20 m within the HRSC values. Though HRSC is the lowest resolution dataset, it is the only dataset available that spans all of Gale crater.

5.3.2 Volume Calculations

Much of the topographic analysis presented here relies on methods similar to those developed by Warner et al. [2011] for basin volumes and Chaytor et al. [2009] for landslide volumes. The volumes eroded from valley networks in the source regions were calculated by projecting a smooth surface between the bounding sidewalls on either flank of the valley, essentially creating a “lid” over the lower incised valley network. This projected surface was then gridded and subtracted from the original gridded topography. A similar procedure was followed to determine lake volumes for each of the three proposed lake levels. To estimate delta volumes we used two methodologies. When possible we first eliminated the delta form (mapped in planform) and projected the nearby local topography (typically the crater wall or floor) laterally under the area occupied by delta feature to create a pre-delta surface. We then took the difference between the gridded delta surface and this projected pre-delta surface to estimate the volume of the delta. This methodology is described in detail for the Peace Vallis fan in Chapter 4. Due to the complexity of recreating the local topography in some regions, we would draw several long profiles (typically 5-10 profiles depending on the planform area of the delta) down each delta feature and out onto the floor it built upon. For each profile we then projected slope of the floor linearly back upslope under the delta deposit and determined the depth of the deposit. This depth was then applied to the length of the segment (the segment length is one-half the distance to the long profile to the right plus one-half the distance to the long profile to the left) to determine the segment volume. The segment volumes were summed to get the total delta volume. On one delta we tested both methods and got comparable results (1.4 km³ for a complete topographic reconstruction and 1.5 km³ using 5 long profiles). We are assuming that since the source material for the deltaic deposits are mostly granular deposits of colluvium and dust and not bedrock, the bulk density of the eroded and deposited material is approximately the same, such that volumes can be directly compared. Our methodology used ArcGIS tools, though all post-processing was conducted in MATLAB.

Topographic data to determine erosion caused by valley incision in the source areas were often limited and even when CTX DEM data are available, they are too coarse to resolve the smaller channels higher up in the drainage network. We therefore used the area for which we had higher-resolution DEM data and/or resolvable channels to determine width-to-depth relationships for the incised valleys. We then extended this relation to the entire channel network, so that in areas where coverage was limited to visual data alone, or where the elevation was too coarse to discern the presence of channels, measured channel widths were used to estimate a channel depth. By doing this along the entire channel network, the volume of material missing from the headwaters was estimated. Since cessation of channel incision, dust and colluvium has partially infilled many of these channels and dunes are present on the valley floors such that our estimates of sediment produced by channel incision may be somewhat low.

5.3.3 Crater Counting

We generated cratering maps of Gale's lower inner walls and floor (defined as being areas within the crater rim that have slopes less than ~5 degrees – the lower walls typically have slopes between 2 and 5 degrees and the floor is typically <2 degrees - and excluding Aeolis Mons deposits) by mapping onto CTX imagery in ArcGIS using CraterTools software [Kneissl et al., 2011]. In order to compare differences in a region's ability to retain craters, we mapped all craters within Gale's wall and floor, acknowledging that we may have included some secondary craters in our count. In general, the resolution of the CTX data enabled us to map craters down to ~12 m in diameter. We extracted crater area, and hence crater diameter, from our maps to compile crater frequency plots in Craterstats II [Michael and Neukum, 2010] using pseudo-log bins (with no re-surfacing correction) to interpret the age of Gale's floor based on the chronology function of Hartmann and Neukum [2001]. While we also dated the surface of several of the larger depositional features and their source areas within Gale (like the Peace Vallis fan and its source area), the small size of the units (they are approximately 100 km² in planform area) precludes reliable age estimates [Palucis and Dietrich, 2014; Chapter 6].

5.4 Delta, Fan and Channel Deposits within Gale Crater

Across Gale, there are several large depositional features with arcuate forms emanating from or associated with incised valleys. The basic architecture of these features is that they have flat tops that transition to a steeply dipping front that then merges with the topography of the basin into which these features formed. Gilbert [1890] first described in detail the same form on Earth, now referred to as a Gilbert delta, which results from a river encountering a still body of water (e.g. a lake) and thereby causing it to drop its sediment load. Gilbert [1890] then used the elevation of the delta fronts to infer paleo-lake level stands for Lake Bonneville. In the following sections, we describe the Gale delta features and their associated incised valleys in detail, and then use them to infer paleo-lake levels within Gale. Depositional landforms denoted by convex contours relative to their background walls are referred to as alluvial fans and are also documented, as they indicate additional periods of hydrologic activity within Gale.

5.4.1 The Farah Vallis-Pancake Delta and Farah Vallis Delta-Fan Systems

On the southern rim of Gale crater is a large v-shaped incised valley, recently named Farah Vallis (Figure 5.2, 5.4). The term “valley” is used rather than “channel” because the feature is much broader and deeper than the channel that cut it. Farah Vallis is composed of two distinct canyons; one that crosses the rim and cuts into ejecta and one that cuts into deposits from the upslope canyon. In total, it is 74 kilometers long, has an average slope of 6% (Figure 5.5a,b) and varies in width from 500 meters to 3 kilometers, with an average width-to-depth ratio of 9.5 based on 20 cross-sections. The upper portion of the canyon is incised 26 kilometers back from the crater rim into Gale crater ejecta [see Figure 1 in Newsom et al., *submitted*] and terminates in an amphitheater-shaped headwall. There are just two, short, stubby tributary valleys. Although relatively sharp-edged, the floor of Farah Vallis is covered with dunes and colluvium, in some areas with up to boulder-sized material (see HIRISE image ESP_018142_1735) indicating that it is partially infilled. The upper portion of the canyon has an average slope of 6%, calculated every

100 meters using HRSC data, but begins to steepen to 9% as the valley enters Gale crater. Here it is incising into a deposit located ~1000 m below the crater rim, perhaps formed from the collapse of the impact structure [Melosh, 1989] or local slumping processes post-impact (Figure 5.6). Between 26 km and 46 km the canyon takes two 90-degree bends, the first towards the west and the second back towards the north before terminating as the channel at the base canyon crosses the crater wall. From 33 to 46 km, frequent distinct, linear evenly spaced indentations define tracks down the valley walls to large boulders, recording active boulder displacement down slopes of about 45%. Downstream of 46 kilometers, where the channel exists the canyon is a fan-shaped deposit (seen clearly in CTX topographic data, Figure 5.7) with an average slope of 4%. Here Farah Vallis becomes less prominent, the valley width decreases from 1.5 kilometers to 500 meters and valley depth decreases from 200 meters to 60 meters in less than a kilometer and the channel is crossing thick deposits (Figure 5.8). By about 57 km the valley again incises, this time into prior deposits until about 74 km, with Farah Vallis widening again to a maximum of 2.5 kilometers wide and 250 m deep, before transitioning to a distinct fan-delta deposit (Figure 5.6). This will be referred to as the Farah Vallis fan-delta and will be described in detail below. To the east of Farah Vallis, and likely associated with Farah Vallis prior to deposition of the fan-shaped deposit and its later incision, is another delta feature, referred to here as the “Pancake delta” (Figure 5.6).

The Pancake delta (Figure 5.9) is a dark-toned, arcuate-shaped feature that is 117 km² in planform and is approximately 100 meters in depth. Using 10 long profiles and the segment methodology described in Section 5.3.2, the delta volume is ~14 km³. The delta has an average slope of 3% at its head, which decreases to 1.3% on its lower plain (Figure 5.10a,b). At the delta front, which follows exactly the -3280 meter HRSC contour line (orange contour, Figure 5.10a,b), the average slope increases abruptly to 13% (Figure 5.10b). At the time of writing, there is no HiRISE imagery of this delta, but based on CTX imagery, there are no distinct channels on the surface of this feature, normal or inverted. The delta front appears to be well preserved, as it lacks any substantial gully development and maintains a relatively uniform height over 20 kilometers, suggesting perhaps a low wave-energy environment.

The Farah Vallis delta front is located at approximately the -3780 meter HRSC contour (green contour, Figure 5.11a,b), which is 500 meters lower than the Pancake delta front. The delta is 78 km² in planform area, has an average depth of 300 meters, and a volume of 22.5 km³. The volume is based on the analysis of 10 long-profiles and the assumption that the surface the delta aggraded across has the same slope as the regional surface slope for Gale Crater (i.e. 1.4% sloping towards the north). The HRSC contours (Figure 5.11a) near the exit of Farah Vallis are strongly arcuate the first 6 kilometers down the delta, and the long profile in this region does not show a sudden break in slope (the slope does increase from an average of 2% to 3.5% but more gradually than the slope break seen at the delta front which increases abruptly from an average slope of 5% to 12%, Figure 5.11b), suggestive that there is a fan deposit overlaying the Farah Vallis delta. This deposit covers approximately 25 km² and is relatively thin, perhaps 15 to 20 meters on average, based on extending the Farah Vallis delta surface back and underneath the hypothesized fan deposit. The fan deposit, while partially covered in dark sand from the nearby dunes, preserves a large channel running down its center (Figure 5.12). The channel is 5 kilometers long, runs the entire length of the feature, and has an average width of 80 meters $\pm$ 10 meters based on 20 measurements. HRSC data are too coarse to resolve channel depth, and there

is only CTX coverage for the upper portion of the fan. Even where there is CTX DEM coverage, channel depth is difficult to resolve as the limit of resolution in the z direction (elevation) is approximately 18 meters. Since the channel depths are not resolvable, even where the channel is widest and most clearly defined, this suggests that the channel is at most 18 meters deep, but likely between 0.8 and 8 meters if the width-to-depth ratio follows those for typical terrestrial alluvial fan, which typically range from 10 to 100 [e.g. Stock, 2012].

The direct connection of Farah Vallis to the Farah Vallis fan and delta, and its close association with the Pancake delta to the east, suggest that the sediment required to build these depositional features is derived from both the incision of Farah Vallis and from incision of sediment previously deposited from the Farah Vallis canyon. A quantitative test of this hypothesis is to compare estimates of the eroded and the depositional volumes (see Figure 5.13). As mentioned previously, the floor of Farah Vallis is covered with colluvium and dunes, which our volume calculation does not account for, and may underestimate the total removed/eroded volume. This analysis will be repeated for each delta system around Gale, both to constrain the source of sediment (and water) and to place bounds on the extent of preservation of the deposits and the applicability of using these deposits to infer past lake levels.

Based on a combination of CTX and HRSC DEM data, the total eroded volume from Farah Vallis is 40.5 km³ (Figure 5.13, 35.5 km³ in the upper canyon, 5 km³ lower incision). The Pancake delta and the fan-shaped deposit at the canyon mouth (Figure 5.13), which was later incised into by Farah Vallis, are 14 km³ and 5 km³ respectively. The Farah Vallis fan-delta deposit has a combined volume of 23 km³, of which the fan is approximately 0.5 km³. In total, the Pancake delta, Farah Vallis fan-delta, and the fan-shaped deposit emanating from the upper half of Farah Vallis comprise 42 km³ of material, which approximately matches the 40 km³ of material removed to create Farah Vallis.

5.4.2 The Entrance Canyon and Stacked Deposits

The entrance canyon is located on the north side of Aeolis Mons (Figure 5.2, 5.14) and is a destination for the *Curiosity* rover during its ascent of Aeolis Mons. The large canyon within the lower Aeolis Mons unit (referred to as the dark-toned layered yardang unit by Anderson and Bell [2010]) is ~4.5 kilometers long, though it does appear to incise into the light-toned yardang unit [Anderson and Bell, 2010] for another ~6 km. It is on average ~850 meters deep. Within the canyon there is an ~1 kilometer long gully (colored in blue in Figure 5.15 and outlined in blue in Figure 5.16a) with an average depth of 11 meters and an average width of 150 meters (based on 10 cross-sections). The gully incised along an 11% slope. The gully is relatively sharp-edged for most of its extent, but is in-filled with colluvium and meter-scale boulders (Figure 5.16b) and eventually transitions to a boulder-rich and crudely stratified deposit (Figure 5.15, 5.17a - numbered “1” and colored in green), which appears to overlie a series of four distinct deposits (Figures 5.15 and 5.17a). This transition occurs near the -3780 meter HRSC contour interval (green contour, Figure 15), which corresponds to the Farah Vallis delta front. This deposit is relatively narrow with an average width of 140 meters (5 cross-sections) and extends for ~750 meters at an 8% slope (Figure 18a,b). The average depth is 7 meters. This upper deposit transitions to a second deposit (numbered “2” and outlined in red in Figure 15, 17a) that is wider, steeper-sided and spreads radially. The front extends downslope for ~650 meters and has an

average slope of 16%, an average width of 270 (5 cross-sections) and an average depth of 100 meters. There is a clear front to this deposit (red arrow, Figure 18b), which corresponds to the -3927 meter HRSC contour, where it steepens from 16% to 24%. Based on the radial form and steep front we interpret this deposit to be a delta. This deposit extends onto a second distinct deposit (numbered “3” and outlined in orange in Figure 15, 17a), which spreads laterally to the east and appears to have inverted distributary channels (Figure 17b). Again, based on its form and the presence of channels, we interpret this deposit to be a delta as well. This second delta front deposit extends ~260 m, is on average 255 meters wide (10 cross-sections), and 30 meters thick. The delta front is located near the -3980 meter HRSC contour line (yellow contour, Figure 15) and is characterized by a sudden change in slope from 16% to 18% (orange arrow, Figure 18b). The lowest deposit is shown in yellow in Figures 15 and 17a (and is numbered “4”) and extends another 520 meters. This deposit has an abrupt increase in slope (black arrow, Figure 18b) and has an average width of 160 meters (10 cross-sections) and is ~5 meters deep. This may be a third delta deposit, but is the most unconvincing as it is quite narrow and does not spread radially. Based on examination of the HIRISE imagery, the deposits all appear coarse, containing meter-scale boulders, which is consistent with the steep slopes at which they deposited. Combined, the coarse, steep nature of the deposits and the break in slope like that of a delta for at least two, if not three deposits, argues for distinct on-lapping stratigraphy. Underlying the entire sequence is a broad, nearly uniform sloping plain (colored in grey in Figure 15 and numbered “5”), with an average slope of 13%, that is etched with strata that were deposited by well-defined, regularly spaced bedforms (spacing is approximately 20 meters) with lighter toned ridges oriented in a east-west direction (Figure 19). No meter-scale material was observed in the HIRISE imagery on this plain, suggesting a finer-grain size than that observed in the deposits overlying it.

Using 1 meter HIRISE elevation data, the volumes of each deposit were calculated using 3 to 5 long-profiles and are compared to the removed gully volume of 0.0037 km^3 , assuming a constant-slope for the gully bottom where the gully is backfilled. It was found that the uppermost deposit (“1”, green, Figure 15) is 0.0007 km^3 , the first underlying delta deposit (“2”, red, Figure 15) is 0.01 km^3 , the second delta deposit (“3”, orange, Figure 15) is 0.002 km^3 , and the lowest deposit (“4”, yellow, Figure 15) is 0.0005 km^3 . In sum total, the entrance canyon deposits are $\sim 0.013 \text{ km}^3$ in volume, which is ~ 3 times larger than the volume of material excavated from the gully. This suggests, perhaps, that the feeder gully was once larger in extent or depth and has mostly been filled in since it was last active. Evidence for the latter is the presence of large boulders, colluvium, and dunes that are obscuring the head of the gully (see Figure 20). Also, the gully may have not been the only sediment source, as there is very coarse debris lying in the canyon upslope and dark scree and colluvium lying at the base of the east-facing cliff, both of which would be a ready source of debris for runoff from Aeolis Mons.

5.4.3 The Western Valley and Delta System

On the northwestern rim of Gale crater is a third delta system (Figures 5.2, 5.21), referred to here as the “Western delta”, though there are actually three distinct delta deposits. The lower more prominent deposit (W1) has a front that corresponds to the -3980 meter HRSC contour (yellow contour, Figure 5.22). As seen with previous deposits, the slope rapidly increases at this front, in this case increasing from an average slope of 13% to 21% (Figure 5.23). This deposit is

approximately 7.0 km² in planform and is 250 meters thick at its front. A second, smaller delta deposit (W2) is overlaying it in a "stair-step" morphology and its front corresponds to the -3780 meter HRSC contour (green contour, Figure 5.22). Again there is an abrupt increase in slope from 7% to 14%. The deposit has a planform area of 2 km² and is 100 meters deep at its front. To the south is a second, smaller deposit (W3) whose front also corresponds to the -3780 meter HRSC contour (green contour, Figure 5.22). This deposit has an area of 0.3 km² and is 50 meters deep at its front. Directly to the north of these delta deposits is a small fan, based on its fan-like form (Figure 5.21), reflected by convex contours (Figure 5.22). This small fan feature has an area of 6 km² and a slope of 12%. Based on CTX imagery, the deltaic deposits are all light-toned, appear to be minimally eroded into at their fronts (there is no gullying or evidence of wave-cutting at the scale of the imagery), and there are no observable channel-like features.

Upslope of the western delta and fan deposits is an incised valley network. The valley network is 76 kilometers long, has an average slope of 8% (Figure 5.24a,b) and varies in width from 250 meters to 1.6 kilometers, with an average width-to-depth ratio of 10 based on 15 cross-sections. Using the current surface topography from both HRSC and CTX to infer the direction of transport of water and sediment, a watershed area of about 450 km² may drain to the western delta system. The drainage density for the western delta system 0.16 km/km², which is higher than some of the highest valley network drainage densities reported in a global survey by Hynek et al. [2010] of over 82,219 valleys, with most of the highest drainage density valleys forming between the latest Noachian and earliest Hesperian epochs. The source area furthest back from the crater rim consists of gently sloping plains towards Gale Crater (5% slope on average) interspersed with local mounds and ridges, perhaps formed from impact ejecta, which are roughly 50 to 100 meters in relief. There are local bedrock exposures, but only at the steepest portions of the hills, as most of these mounds and ridges are mantled with debris, likely colluvium or aeolian deposits. Nearer to the crater rim the valley incises into a steeper, smaller canyon (average slopes are 12%), which flattens out for several kilometers (7% average slope) before steepening again (Figure 5.24a,b). The overall appearance of the western delta watershed is of disrupted topography associated with impacts and impact ejecta that has been mantled by outfall from airborne sediment, colluvium from retreating slopes, and possibly fluvial reworking. The valley system, with limited typically amphitheater-shaped headwalls, drains to the western delta and is cut into the previously deposited sediments. Although relatively sharp-edged, the incised valleys are also floored with dunes and colluvium, indicating they are partially infilled.

Using 18 meter CTX DEM data, the total volume of material that has been eroded from the western delta valley network is 1.6 km³. Both methodologies for calculating the volume of the deltaic deposits were performed for the Western Delta deposits, resulting in an average net deposit volume of 1.5 km³. We do not have a strong constraint on the volume of the small fan, but by projecting the local topographic slope of the crater floor underneath the fan deposit, its average depth is approximately 15 meters. For a planform area of 6 km², that gives an approximate volume of 0.09 km³. The sum of the two prominent delta volumes and the small fan is 1.59 km³, which is an approximate balance between the volume of material eroded in forming the valley network (1.6 km³).

5.4.4 *Aeolis Mons terrace and delta*

Aeolis Mons is flanked by broad, gently sloping terrace-like features (Figure 5.25) that have previously been interpreted as generated by wind scour [Anderson and Bell, 2010] or cut by shoreline processes [Cabrol et al., 1999]. There appear to be bedform-shaped features on the terraces (visible in HiRISE images) that are consistent with subaqueous deposition. In addition, a 9 km long, roughly 1 km thick exposure of inclined layers just below the terraces, visible in CTX and Mastcam images, may record an ancient delta (Figure 5.26). A deltaic interpretation is problematic, however, in that there is not an obvious sediment source other than reworked Aeolis Mons deposits. If these beds were deposited subaqueously from reworked Aeolis Mons sediments, the lake level would have been well above the northern rim.

5.4.5 Peace Vallis and Fan System

A detailed description of the Peace Vallis fan system is described in Chapter 4, but a brief overview is provided here, as its distal end was the location of the *Curiosity* rover's first drill site and its timing relative to other hydrologic events in Gale Crater is of primary importance to the MSL mission. The Peace Vallis fan is defined by convex contours that extend across the lower portion of the northern Gale Crater wall (Figure 4.5, Chapter 4). The fan covers about 80 km² and is laced on its western margin with frequent ridges interpreted to be paleo-channel beds (Figure 4.5, Chapter 4). The western portion of the fan terminates on a broad, nearly flat ridge that extends across from Aeolis Mons to the crater wall. The eastern portion of the fan may have prograded onto and possibly across the pre-existing Bradbury Rise (the *Curiosity* Rover landing area) and descended to a topographic low to the east of Bradbury rise (Figure 4.6, Chapter 4). The fan progressively declines in slope from about 3% at the fan apex to 1% or less where the fan progrades across pre-existing topography. The fan's apex is located at the -4227 meter HRSC contour, which is 150 meters lower than the lowest western delta front and 100 meters lower than the bottom of the platform associated with the entrance canyon deposit. It is topographically one of the lowest hydrologic features in Gale Crater.

The location of the Peace Vallis fan likely results from a 15-kilometer setback in the crater rim (Figure 4.12, Chapter 4), which enables drainage from the adjacent plains near the crater rim to spill into a canyon where an incised valley feeds the apex of the fan. This incised valley network, which is 107 kilometers long, was fed by a catchment area 730 km² in size, resulting in a drainage density of 0.15 km/km². Similarly to the Western delta incised valley, all of Peace Vallis appears to be excavated into previously deposited debris, likely of colluvial, eolian, fluvial and possibly air-fall origin. Estimates of volume removed to form the incised valley (0.8 km³) match the total volume of the original, pre-eroded fan (0.9 km³).

5.4.6 South Fan System

There is a large fan deposit, referred to here as the South fan, off of the southern rim of Gale (Figure 5.2). The fan covers an area of 240 km² with an average slope of 1.5%. The fan apex is located 200 meters above the -3280 meter HRSC contour line and its distal end is approximately 100 meters lower. CTX imagery did not reveal any shorelines or gullying into the fan deposit at -3280 meters, though it is very possible that these features would not be resolvable at this scale (18 meters/pixel). A dark dune field covers the upper portion of the fan, but the lower portion of the fan is light-toned, preserves some small channel features, and has a pitted

appearance, likely from impact cratering. The low resolution HRSC DEM precludes an accurate volume estimate of this relatively thin feature. This fan emanates from a 97.5 km long incised valley network (that does not cross the crater rim) with an approximate eroded volume of 2 km³ (based measurements on valley width and an assumed width-to-depth ratio of 12 – the average width-to-depth ratio for Farah Vallis, Peace Vallis and the West delta incised valley).

5.4.7 Small fans, small deltas, and channels around Gale Crater

Along the crater walls and from the slopes of Aeolis Mons (Mt. Sharp) are gullies, sinuous ridges (likely inverted channels) and small deposits that based on their form and location also likely record delta and fans. Previous maps have been published that capture many of these features [Anderson and Bell, 2010; Le Deit et al., 2013]. Here, we focus on those features that may provide clues about lake levels and their relative timing. To the east of the Farah Vallis delta-fan system is a small delta-like feature emanating from a large canyon on the south side of Aeolis Mons (Figure 5.27). The apparent front of the delta roughly corresponds to the -3280 meter HRSC contour line, which is the same contour that defines the Pancake delta front, though the HRSC DEM is too coarse to get a long profile and slope to see if there is an obvious break in slope. The feature is approximately 0.5 km² in area and much less than 50 meters thick based on the 50-meter HRSC contour intervals. The regional slope here is approximately 9%. Based on CTX imagery the delta feature does show what may be inverted distributary channels (Figure 5.27), indicating erosion of the deposit has occurred since deposition, but the correspondence of the delta front with the -3280 meter HRSC contour line, which also defines the very prominent Pancake delta, seems unlikely to be due to aeolian erosion. The canyon immediately upslope from the delta has an average depth of 150 meters, an average width of 1.7 kilometers. Since the delta feature is obviously under scale relative to the canyon from which it emerged, it likely records only a small portion of the sediment derived in cutting the upslope canyon, suggesting that the canyon was well established before the formation of the Pancake delta. Importantly, it does indicate that some fluvial activity was occurring off of Aeolis Mons when the Pancake delta developed.

Along the western margin of Aeolis Mons, small (< 3 km²) complex fan-delta shaped deposits lie at the mouth of two of the larger canyons that exit to the Gale floor (Figure 5.28). These roughly arcuate-shaped deposits are strongly wind-shaped and this deflation in the more southern deposit (WAM-2) has exposed subtle crossing ridges that may record channels. There is an elevation difference of 500 m from the top to the bottom of the larger deposit (WAM-1) and a 250 m difference for the smaller deposit (WAM-2) with slopes of 20 to 27%. Collectively these attributes suggest that both deposits record small deltas recording fluvial activity in the upslope canyons but minor amounts of sediment delivery relative to the canyon scales. The -3980 m HRSC contour (yellow) that defines the lower Western Delta (and one of the Entrance canyon deltas) crosses both these small deltas at approximately the elevation where downslope steepening occurs.

Poorly defined fan-shaped deposits are present at the mouth of the largest canyon cutting through Mt. Sharp towards the north (defined as the mass wasting deposits by Le Deit et al. [2013]). The irregular deposits drape the steep-bedded slopes downslope of the canyon and

spread to the northern crater floor. No obvious break in slope is present in the images available that would suggest a delta.

In general, small fan development along the crater walls of Gale is concentrated on the northwest wall (Figure 5.2). These fans were first mapped and discussed in Anderson and Bell [2010], where they were referred to as “fan-shaped mesas on the crater floor” that occur at the termination of branched “valleys/canyons”, usually on the crater rim. The fans are typically 1 to 40 km² in area, and deposit on slopes between 8% and 16% (using CTX slope data). In appearance, the fans are thin, appear to preserve craters thus giving them a pitted texture, and often have ridges, inferred to be paleo-channels by Anderson and Bell [2010], running perpendicular to contour. These somewhat sinuous ridges often continue past where the fan form ends, sometimes continuing to the crater floor. Anderson and Bell [2010] report that the thermal inertia of these fans range from 430 to 780 J m⁻² K⁻² s^{-1/2}, which encompasses the range for the low thermal inertia upper portion (460 J m⁻² K⁻² s^{-1/2}) and high thermal inertia lower portion (620 J m⁻² K⁻² s^{-1/2}) of the Peace Vallis fan.

Incised channels are most numerous along the northwestern wall where the crater rim is least sharply defined. Larger and deeper incised channels are developed along the southern wall, which is considerably longer than that to the north. Most of the incised channels do not terminate in fans, suggesting there is threshold volume of erosion for a fan to be supported. If we assume that the incised valley represents the potential sediment load to a fan, then an eroded volume of 0.5 km³ is needed for a fan to be currently visible at the mouth of an incised valley. In general, most of these fans terminate below the -3780 meter HRSC contour line. As shown in Figure 5.2, however, many of the valley networks (that terminate without fan deposits) end at either the -3780 meter or the -3980 meter HRSC contour line.

5.5. Large Lakes within Gale Crater

Based on the geomorphic evidence presented above, there were at least three large lakes within Gale Crater that endured long enough to leave a geomorphic record of their existence. Each of these lakes will be described and then the case will be made for a possible sequence of lake level decline followed by lake level rise. As noted above, a higher lake may have occurred which would require water above the northern rim, and thus an ocean, but evidence at Gale is not yet sufficient to make a strong case for that occurrence.

5.5.1 Gale-filling Lake (-2450 m HRSC lake level)

Di Achille and Hynek [2010] identified four deltas within 260 km of Gale crater (two to the west and two to the east of Gale) that border the crustal dichotomy and would require an ocean (as the deltas are not confined within craters). The average elevation of the delta front is -2448 m. This local elevation is similar to the global level (i.e. -2540 m) that they proposed for such deposits. Topographic analysis of Gale crater shows that water levels higher than -2500 m would breach a roughly 22 km wide topographic corridor slightly beyond the northern rim. While this suggests the possibility that Gale once was connected to a Martian ocean, there is no evidence of channel inflow or outflow across the topographic breach. This highest lake level remains speculative unless the MSL *Curiosity* imagery eventually enables a definitive

interpretation of the 9 km deposit (Figure 5.26) as a delta deposit. If confirmed as a delta, this would require still even higher water levels.

5.5.2 Pancake Lake (-3280 meter HRSC lake level)

We interpret the highest lake level to have reached about the -3280 meter HRSC contour (orange contour in Figure 5.2) and is defined most clearly by the Pancake delta, though the small delta off of Aeolis Mons (Figure 5.27) also terminates at this elevation. In addition to the delta features, this elevation also corresponds to several small inverted channels off of the southern flank of Aeolis Mons and a topographic bench on the northeastern side of Gale Crater. A lake at this elevation would have covered a surface area of 5832 km² and would have had a mean and maximum depth of 0.7 and 1.4 kilometers, respectively. The lake would have occupied a volume of 3780 km³.

Assuming though that the lake must have remained at this elevation for enough time to form the Pancake delta and its associated features, the steady-state discharge needed to balance out losses to evaporation can be estimated, giving some bounds on the average discharges entering Gale Crater during this period of its hydrologic history. Theoretical and experimental estimates of evaporation rates on present-day Mars typically vary from 0.03 to 0.4 mm/hour, but could perhaps be as high as 2.5 mm/hr [Ingersoll, 1970; Hecht, 2002; Kuznetz and Gan, 2002; Sears and Chittenden, 2005; Sears and Moore, 2005], though the lack of constraints on the temperature and pressure conditions of early Mars make these estimates highly uncertain. While evaporation rates do increase with decreasing pressure, they are even more strongly a function of temperature, so that water at 0°C on Mars will evaporate no faster than water at 60°C on Earth [Hecht, 2002]. A warm, wet early Mars [Craddock and Howard, 2002], therefore, may not have had orders of magnitude differences in evaporation rates as a cold, dry early Mars [Head and Pratt, 2001]. In lieu of better estimates at this time, we will use this range of values (0.03 to 0.4 mm/hour) for the three hypothesized lakes in Gale Crater. At the highest stand, this equates to a steady-state discharge, calculated as the product of lake area and evaporation rate, between 16 and 650 m³/s (or 0.5 to 20 km³/earth year). Wallace and Sagan [1979] showed that for a range of Martian insulations, wind regimes, and pressures, the mean evaporation rates for ice-covered water bodies is 2x10⁻⁹ meters/day. At this evaporation rate, discharges as low as 11 m³/s could maintain this lake level, though it is difficult to discern within the geomorphology whether or not the lake would have had an ice cover (though the apparent lack of wind wave influence on delta development perhaps supports the ice cover hypothesis).

A typical assessment of the runoff needed to build a depositional feature like the Pancake delta is to assume a rock to water ratio [Mangold et al., 2012], or in cases where channel geometry, slope and grain size can be estimated, perform a hydraulic analysis [Wilson et al., 2004]. The topset slope of the Pancake delta is fairly low gradient (~1%), suggestive that the sediment was transported fluvially (as opposed to mass flows, like debris flows or mud flows). While typical rock/water ratios vary by orders of magnitude and depend on grain size and climate characteristics, a survey of terrestrial rivers by Syvitski et al. [2003], who collected runoff and sediment flux data for 340 river basins around the world, shows that the log transformed rock/water ratio yields a roughly Gaussian curve with a mean value of 0.0001±0.00008 (Figure 5.29). For the volume of sediment comprising the Pancake delta (14

km³), this would suggest 140,000 km³ of water runoff would have been needed, roughly 26 times the volume of the proposed lake. Using our calculated discharge of 0.5 to 20 km³/earth-year, and assuming that nearly all the water that kept the lake full came from Farah Vallis, a rock/water ratio of 0.0001 yields 5×10^{-5} to 2×10^{-3} km³/earth-year of sediment transported. For the amount of sediment contained within the Pancake delta plus its associated plains (i.e. 20 km³), it would take 400,000 to 10,000 earth years to build the Pancake delta while holding the water level constant.

5.5.3 Farah Vallis Lake (-3780 meter HRSC lake level)

The second largest lake level is at the -3780 meter HRSC contour (green contour in Figure 5.2) and is defined most clearly by the Farah Vallis delta front, but also corresponds to the upper Western delta (W2) and the termination of approximately half of the gullies on the northern wall of Gale crater. It is also near the elevation where the gully within the entrance canyon transitions to a deposit. A lake at this elevation would have covered a surface area of 3617 km² and would have had a mean and maximum depth of 0.4 and 0.9 kilometers, respectively. The lake would have occupied a volume of 1583 km³. Similarly to the -3780 meter lake, the lack of definable channels on the Farah Vallis delta deposits makes it difficult to determine formative discharges during the time of delta-building, but discharges ranging from 10 to 401 m³/s (0.3 to 13 km³/earth-year) would be needed to maintain a steady-state lake level (assuming losses to evaporation only). For a rock/water ratio of 0.0001, this leads to 3.3×10^{-5} to 1.3×10^{-5} km³/earth-year deposited. To deposit the 23 km³ of material contained within the Farah Vallis delta, would then take 670,000 to 18,000 earth years to deposit.

5.5.4 Western Lake (-3980 meter HRSC lake level)

The third largest lake level would have occurred around the -3980 meter HRSC contour (yellow contour in Figure 5.2) and is defined most clearly by the lower Western delta (W1, Figure 5.22) and the middle delta within the entrance canyon deposits (yellow arrow, Figure 5.18b). This elevation also corresponds to the termination of the other half of the gullies on the northern wall of Gale crater. A lake at this elevation would have covered a surface area of 3008 km² and would have had a mean and maximum depth of 0.3 and 0.7 kilometers, respectively. The lake would have occupied a volume of 915 km³. Discharges ranging from 8 to 334 m³/s (0.25 to 10.5 km³/earth-year) would be needed to maintain a steady-state lake level (assuming losses to evaporation only). The volume of the lower delta is about 1 km³. Assuming a 0.0001 sediment/water ratio, the sediment deposition rate is about 2.5×10^{-5} to 1×10^{-3} km³/earth-year or 40,000 to 950 earth years.

5.6 Sequence and Relative Timing of Lakes

5.6.1 Pre-large Lakes

Thompson et al. [2011] proposed that the impact age for Gale, based on crater counts in the surrounding ejecta, occurred in the Late Noachian/Early Hesperian, though they expressed some uncertainty in this age range due to the occurrence of a few large craters for which it is not clear if they pre or post-date the Gale impact. Le Deit et al. [2013], working independently and

using only crater counts on the ejecta sheet to the south and east, arrive at an age of ~ 3.61 Ga for Gale. Thompson et al. [2001] also counted craters on the crater floor and wall and concluded that the floor/wall were Early Hesperian in age. We independently performed a crater count across the floor and wall of Gale and found that craters ranging from 1 to 4 kilometers in diameter fell approximately on the 3.4 Ga isochron (Figure 5.30; root-2 binning, Hartmann 2004 production function, Hartmann and Neukum 2001 chronology function), though the method of binning altered the mean age from 3.38 (Late Hesperian) to 3.41 (Early Hesperian). These counts within Gale suggest that the Aeolis Mon deposits were formed relatively soon after impact, likely within the first 0.2 Ga. Irwin et al. [2005] proposed that Farah Vallis breached along the southern rim of Gale and formed two small deposits around what must have been the pre-existing mound deposit, and concluded that most deposition and channel erosion into Gale crater occurred in a period of 0.2 to 0.3 Ga post Gale impact.

When the Gale impact occurred, it appears to have disrupted a valley network draining $\sim 270,000$ km² to the south [see Irwin et al., 2005]. It is difficult to discern whether or not this valley network was active at the time of impact. Modeling by Kite et al. [2013], which assumes a pure CO₂ atmosphere, dusty snow and a faint young sun, show that the region near Gale is conducive to both snow accumulation and melting when obliquity is $\geq 40^\circ$, eccentricity is ≥ 0.12 , and perihelion occurs near equinox. They estimate that these conditions are satisfied by 0.01-20% of the probability distribution of Mars' past spin-orbit parameters, and the probability of a melt year any given year is only 0.05%. For the first 0.2 Ga of Gale's existence, this equates to approximately only 100,000 years where runoff might be expected to occur. In addition, we have shown that the volume of sediment comprising the Pancake delta and its associated plains to the west and that in the Farah Vallis fan and delta are approximately equal to the volume of sediment removed to create Farah Vallis (see Section 5.4.1). This suggests that the majority of Farah Vallis' incision occurred during the time of Pancake delta and its associated plains formation. Based on the small delta exiting the canyon off of the south side of Aeolis Mons (Figure 5.27), Aeolis Mons would have already been close to its current topographic extent before this incision occurred. Irwin et al. [2005] argue that the stratigraphy and geomorphology of Gale suggests a mostly dry history with only minor fluvial erosion of the crater wall and mound. In their hypothesized chronology, a hiatus occurred within the erosive epoch of valley network erosion, and it was during this time Aeolis Mons layers were deposited and partially deflated by winds, and then transient lakes formed.

Although Farah Vallis is the only large entrance breach into Gale, it does not mean that Gale was not intermittently wet during its first ~ 0.2 Ga, as Gale's impact formation would have likely resulted in local releases of water. Wang et al. [2005] predict that for a 150 km diameter crater, maximum groundwater volumes ranging from 1 to 10,000 km³ would be released (assuming aquifer depths of 10 to 1,000 meters) due to impact-induced liquefaction, where the release of this pressurized groundwater would be focused in newly formed or pre-existing fractures. Segura et al. [2008] show through modeling that moderate sized impactors (defined as 30-100 kilometers in diameter) may yield 40 cm to 18 meters of global rainfall, depending on the impactor size and background levels of CO₂. For a 30 kilometer impact, which yield's a Gale-sized crater, model results show that the surface temperature will remain above 273 K for 95 to 155 days with average precipitation rates of 0.002 to 0.0009 m/day, depending on whether or not the radiative effect of clouds are considered. Clues about the environmental conditions under

which Aeolis Mons formed are one of the goals of the *Curiosity* rover once it begins its ascent of the mound.

The distinct benches defining the sides of Aeolis Mons have been proposed as evidence of wave-cut erosion at a time of a Martian ocean [Cabrol et al, 1999; Figure 5.25]. The bench level does correspond to the height of the northern rim. If such a water body existed and then evaporated or sublimated (if ice covered), the water saturated Aeolis Mons deposits may generate seepage channels to the lowering water body, which could explain the canyons into mound on its west side. As discussed below, depositional relationships clearly show that the large lakes that formed within Gale post-date canyon formation.

5.6.2 Large Lakes in Gale

At some point, water supplied from the south breached Gale Crater and began to incise Farah Vallis, eventually depositing the Pancake delta. During the time that the Pancake delta was being deposited, the plains to the west were likely being deposited simultaneously. During lake level rise to form the Pancake lake (-3280 m), sediment would have spilled out onto and across the crater floor, a topographic low point within the region, to where the Farah Vallis delta currently resides. The majority of the sediment removed to create Farah Vallis was not deposited in the Pancake delta; it was deposited within the Farah Vallis fan-delta system (Figure 5.13). Later incision into the plains to the west of the Pancake Delta, either from runoff or groundwater seepage, eventually supplied the remaining material comprising the Farah Vallis fan-delta (Farah Vallis Lower Incision, Figure 5.13). This late stage incision appears to have occurred after deposition of the fan on the wall of Gale Crater that blocked the Pancake delta from the inlet of Farah Vallis (Figure 5.7). The switch from sediment delivery resulting in the Pancake delta formation and the early stages of the Farah Vallis delta to the aggradation of the fan on the crater wall indicates lake levels did not rise much higher than the Pancake delta front. The fan maintains a relatively constant slope of 4% and does not show any topographic indications at the resolution of CTX DEM data that its distal end formed sub-aqueously (e.g. rapidly decreasing slopes or decreasing convexity of contours). Fan formation was either the result of decreased runoff (and hence reduced transport capacity), increases in sediment supply [Stock, 2013], or simply the spreading of unconfined flow from Farah Vallis. One possible end member is that all input to the lake ceased during this time, though it seems unlikely, as water and sediment were being delivered to build the fan (Figure 5.7); what is known for sure is that inputs could not outpace evaporative losses and the lake level declined. Evidence for this is later incision into the small fan deposit combined with the -3780 meter Farah Delta front, the latter which records a later, and lower, still stand (Farah Vallis Lake, green contour, Figure 5.2) within Gale crater.

If the Entrance canyon and Western delta deposits record a back-stepping sequence (i.e. rising lake levels), then the -3280 meter lake must have dropped below the Entrance canyon deposits. The lowest deposit from the entrance canyon ends near the -4077 m HRSC contour, suggesting the lake would have dropped below at least this level. This corresponds to an 800 meter elevation drop, and losses of 1034 km³ of water. We have not observed shorelines or strandlines around Gale crater that definitively point to a lake below or near the elevation of the entrance canyon platform, but the lack of wave-cutting on the larger deltas around Gale suggest that this was a low wave energy environment or a shore-lived period of standing water.

Shorelines and strandlines may not have readily formed, or if they did, were subtle and have since been eroded away or are not visible at the resolution of the imagery available.

The small planform area of the Western and Entrance canyon deposits precludes dating the deposits via crater counting [Palucis and Dietrich, 2014; Chapter 6], so the amount of time that passed between the formation of the Pancake delta and its associated deposits and the building of the Entrance and Western canyon deposits is unknown. The common elevation of the steep front of the middle Entrance canyon deposit (deposit outlined in orange, Figure 5.17a,b) with the steep front of the lower Western delta (W1, yellow contour, Figure 5.22), suggest that at some point the lake rose to approximately -3980 meters and remained there long enough to build these deposits. As the Entrance canyon deposits result from incision into Aeolis Mons and the Western delta deposits result from incision into the northern rim of Gale, the Farah Vallis lake must be sourcing some water local to Gale crater. Approximately half of the incised valleys on the northeastern wall of Gale, as mapped by Anderson and Bell [2010], also terminate at this lake level, and therefore would also likely be contributing water to this lake. This does not mean that regional drainage is not contributing water to Farah Vallis, but that the contribution was perhaps waning compared to the time of initial incision of Farah Vallis (during the time of Pancake Lake). As channels are not preserved on the Western delta, or visible at the CTX image scale, back-calculating formative discharges using sediment transport and hydrologic relations is impossible. The topset slopes on the lower delta are within the debris flow regime (~17%) suggesting higher rock to water ratios (e.g. 0.3 to 0.02 [Mangold et al., 2012b]). The -3980 meter delta within the Entrance canyon was also deposited at steeper slopes (~16%) and is composed of coarse-grained material, suggestive of mass flow deposition (i.e. high rock to water ratios), such that 0.007 km³ to 0.1 km³ of water would have been needed (assuming rock to water ratios of 0.3 to 0.02)) to deposit the 0.002 km³ delta.

The upper Western delta deposit (W2) and the small deposit to its southwest (W3, Figure 5.22), the lack of incision and gullying into the lower west delta deposit, the back-stepping sequence of deltas and fans within the entrance canyon, the delta front at Farah Vallis that corresponds to the upper west delta front (W2, green contour, Figure 5.2), and the termination of approximately the other half of the incised valleys mapped by Anderson and Bell [2010] all suggest that there was a lake level rise from the -3980 meter lake to the -3780 meter lake.

Incision of the plains to the west of the Pancake delta would have contributed the remaining sediment (~4.5 km³) to the formation of the Farah Vallis delta. The Farah Vallis delta, which is approximately twice as large as the Pancake delta in volume, has topset slopes indicative of fluvial processes (~5%), though mass flow deposition could occur. For rock to water ratios ranging from 0.01 to 0.001, this corresponds to 450 to 4,500 km³ of water or 1.6 to 16 meters over the entire Farah Vallis catchment (270,000 km²). The upper west delta has a volume of ~0.2 km³ and topset slopes of 7%, which is within the fluvial-mass flow transition regime. For rock to water ratios ranging from 0.01 to 0.001, this corresponds to water volumes of 20 to 200 km³ and depths of 40 to 400 meters of water over the west delta catchment.

5.6.3 Post-lake fan formation

The small fan located on the Farah Vallis delta (Figures 5.11a,b, 5.12) suggests the transition to drier conditions within the Farah Vallis catchment. By performing a hydrologic analysis on the preserved central fan channel similar to that performed on the inverted channels in Chapter 4, we can estimate formative discharges for this feature. This provides an important constraint on the hydrologic conditions that prevailed at the time this fan formed and sets some bounds on the range of discharges coming from the Farah Vallis. These discharges can be compared to those predicted from our evaporation rates and assumptions about evaporative losses to runoff. Standard hydraulic equations and threshold channel concepts are first used along with slope data and measurements of channel width to determine the likely grain sizes that comprised the channel bed (this cannot be determined from orbital imagery). The instantaneous discharges and flow velocities at bankfull conditions for a ‘representative’ cross-section are then calculated.

The fluid discharge, Q , within a channel can be derived by conservation of mass,

$$Q = uhw \quad (5.1)$$

where u is the average flow velocity, h is the channel depth and w is the channel width. For channels with width-to-depth ratios less than 18, the hydraulic radius, R , which is the ratio of the channel area to its wetted perimeter, is used instead of h . For a turbulent velocity profile (i.e. Reynold’s number, Re_{channel} , is > 1000 [Chow, 1954]), the velocity, u , is found at a given height above the bed, z , using the law of the wall formulation,

$$u = \left(\frac{u_*}{k} \right) \ln \left(\frac{z}{z_o} \right) \quad (5.2)$$

where z_o is the roughness coefficient, k is the dimensionless von Karman’s constant (typically 0.41 [e.g. Furbish, 1997; Garcia, 2007]) and u_* is the shear velocity, defined as

$$u_* = \left(\frac{\tau_b}{\rho} \right)^{0.5} \quad (5.3)$$

In equation 5.3, τ_b is the boundary shear stress, defined as

$$\tau_b = \rho g h S \quad (5.4)$$

where ρ is the fluid density, g is the gravitational constant (3.711 m/s^2 for Mars), and S is water surface slope (assumed to equal measured topographic slope in this analysis). The roughness coefficient for hydraulically rough flow is

$$z_o = \frac{k_s}{30} \quad (5.5)$$

where 30 is the Nikuradse roughness coefficient and k_s is the effective roughness height, typically approximated as $k_s = 3.5D_{84}$, (e.g. Garcia, 2007) giving

$$z_o = \frac{3.5}{30} D_{84} \quad (5.6)$$

The vertically averaged velocity is:

$$\bar{u} = \frac{u_*}{k} \ln \left(\frac{11h}{3.5D_{84}} \right) \quad (5.7)$$

where the 11 in equation 5.7 is the result of integrating the log profile constant and multiplying it by the Nikuradse roughness (i.e. 30). Assuming that the boundary shear stress needed to move the median grain size on the bed (D_{50}) is the critical boundary shear stress [e.g. Buffington and Montgomery, 1997],

$$\tau_{*cr} = \frac{\rho g h S}{(\rho_s - \rho) g D_{50}}, \quad (5.8)$$

(ρ_s is the sediment density) and the shear velocity, u_* is

$$u_* = \left(\tau_{*cr} \frac{(\rho_s - \rho)}{\rho} g D_{50} \right)^{0.5} \quad (5.9)$$

Vertically average velocities reported here are derived from Equation 5.7 and the relationship given for shear velocity given in Equation 5.9. Instantaneous discharge is then calculated from Equation 5.1.

The average channel bed width on the Farah Vallis fan is 80 m and channel slope is ~2%. Martian gravity is 3.711 m/s^2 , the sediment density is 3000 kg/m^3 (typical of basalt) and fluid density of 1000 kg/m^3 (low concentration flow). We assume a range of critical shear stresses, τ_{*cr} , from 0.02 (for sandier beds) to 0.06 (for coarser beds) [e.g., Parker et al., 1998; Parker, 2007; Garcia, 2007] and a range of width-to-depth ratios to determine the mobile median grain size under critical conditions by rearranging Equation 5.8 to solve for D_{50} .

$$D_{50} = \frac{\rho g h S}{(\rho_s - \rho) g \tau_{*cr}} \quad (5.10)$$

We assume width-to-depth ratios between 10 and 100, which are the range of aspect ratios observed for terrestrial alluvial rivers [Knighton, 1998] and for channels on several arid to semi-arid alluvial fans [Stock, 2012]. The roughness length scale D_{84} is set equal $1.5 * D_{50}$ [Morgan et al., 2014]. Also, when the width-depth ratio is less than 20, the hydraulic radius, R , is used in place of channel depth, h . As a separate constraint, flow is also assumed to be critical, i.e. the Froude number is 1, which has been observed on terrestrial fans [see review in Stock, 2012].

The predicted grain size for a 2% slope lies in the gravel to boulder range. For width to depth ratios of 100 for the central fan channel, the predicted flow velocity ranges from 0.6 to 1.3 m/s for low to high critical Shield numbers respectively. This equates to instantaneous discharges

of 31 to 65 m³/s. For the catchment area of Farah Vallis, which assumes the entire catchment is contributing water, this requires runoff supply rates ranging from 0.0004 to 0.03 mm/hr. Kite et al. [2013] suggest that over Mars' history, peak snowmelt rates were likely between 2 and 3 mm/hr, and Williams et al. [2008, 2009] use a snowpack model to show that for the past 5 Ma, latitudes between 70°N and 70°S correspond to significant seasonal snowmelt runoff (>1 mm) on the order of several days a year.

Lake level decline from reduced inflow from Farah Vallis may have been associated with a time of fan building around Gale, not just at Farah Vallis. Runoff from smaller catchments around Gale would have been capable of supplying enough water to transport and deposit sediment, but not enough water to maintain a large lake. The steady state estimates for maintaining lake levels range from 8 to 650 m³/s, but these estimates are likely to be slightly low, as we are ignoring infiltration. The small fan deposits that rim the northern wall of Gale lie on slopes from 8% to 16%, suggesting they are debris flow dominated, with high rock to water ratios. The continuation of the ridges (inferred to be paleo-channels) past the fan deposits down to the crater floor, are perhaps the result of a transition to fluvial processes [Chapter 3]. If the ridge width preserves the original channel width, these channels were ~20 to 50 meters wide. Almost all of the fan and ridge deposits within Gale lie below the -3780 meter HRSC contour line (Farah Vallis Lake, green contour, Figure 5.2), so if they formed before the Pancake lake or between the Pancake lake and Western lake, we might expect to see shorelines or some disruption of the (now) paleo-channels due to re-working. Inspection of CTX imagery does not reveal this, but if Gale was a low-wave energy environment, as we have suggested previously, substantial shoreline development or channel reworking may not have occurred. The timing of the large fan to the south is difficult to constrain. At the resolution of CTX imagery there are no shorelines where the Pancake lake level crosses the fan, which could mean the fan postdates the first large lake in Gale, but better imagery and topography is needed to say for certain. Its incised valley does not cross the rim of the crater, as is the case for the other large fan and delta systems around the walls of Gale. This could mean that there was reduced water availability (i.e. perhaps post-large lakes), or that perhaps the water sourced to the fan was supplied from groundwater draining from the south (i.e. formed during the time of large lakes).

The Peace Vallis fan, whose distal end was located within the MSL landing ellipse, is the most-studied fan within Gale crater and perhaps offers the best clues to the timing of other fans in Gale. Crater counts in the 730 km² source area for the Peace Vallis fan, much of which lies in the ejecta blanket of Gale, yields an age of about 3.6 Ga (Figure 5.31) (though caution should be taken due to the relatively small area counted [Chapter 6]), close to the age of Gale itself. This suggests that the source area has not been substantially re-surfaced since Gale formed and that incision of Peace Vallis to form the fan was the only major incision event. Very early (i.e. shortly after impact) erosion and delivery of sediment to the floor of Gale cannot be ruled out, and perhaps resulted in the setback of the crater rim into which Peace Vallis formed and the creation of the ridge that defines the western edge of the basin into which the Peace Vallis fan flowed [Chapter 4]. While the Peace Vallis fan is likely too small (~80 km²) to obtain reliable crater count age dating directly [Palucis and Dietrich, 2014; Chapter 6], Grant et al. [2014] did attempt to date the fan, concluding that the fan must be Amazonian in age based on craters 90 to 150 meters in diameter. They disregarded a large crater on the upper portion of the fan under the assumption that this crater pre-dated fan formation and is part of a crater fill unit dated to be

middle Hesperian. This particular crater is ~760 meters in diameter and is located near the fan apex (Figure 5.32a). The crater has an average depth of 63 meters, giving it a depth-to-diameter ratio of 0.08. The average depth-to-diameter ratio for post-impact, simple craters is 0.1 to 0.2. The average slope of the crater rim is ~5% and that of the inner crater walls is ~15%. The rim ranges in height from 5 meters to 20 meters, with the highest rim heights located on the eastern side of the crater. This follows expected rim heights (~20 meters) for a crater of this size [Melosh, 1989]. The fan is ~15 to 20 meters thick near this crater based on topographic reconstructions [Chapter 4], such that if the crater were being “buried” by the fan one would expect the crater rim to only be several meters high currently. The rim is not breached (Figure 5.32b) and the inner crater walls do not appear to have been gullied from incoming water during Peace Vallis fan formation or from large lakes in Gale and mapping of fractures near the crater shows a concentric pattern, consistent with fracture formation from the impact (Figure 5.32c). The crater’s ejecta has mostly been removed, presumably from aeolian erosion, as up to 5 to 7 meters of deflation has occurred on the upper portion of the fan as evidenced by the inverted channels [Chapter 4]. The morphologic evidence suggests this crater formed post-Peace Vallis fan formation, and post-lake formation. If it did form around the time of similarly sized craters to the east of it, i.e. ~3.1 to 3.3 Ga [Grant et al., 2014], then the lakes in Gale ended before the late Hesperian, though Grant et al. [2014] base their age on 19 craters over a relatively small area (543 km²).

5.7 Sources of Water

Previous work on the deposition of deltas and large equatorial fans has concluded that they formed during periods of enhanced precipitation, likely as snowfall, through hundreds to thousands of flow events [Armitage et al., 2011; Grant and Wilson, 2011, 2012; Di Achille and Hynek, 2010; Moore and Howard, 2005]. If the lakes in Gale occurred as late as the Late Hesperian to Early Amazonian, the atmospheric pressure and temperature were likely too low to support precipitation that was not snow [Marchant and Head, 2007; Wordsworth et al., 2012]. Groundwater discharge has been considered to be a possible water source for Martian valley networks [Malin and Carr, 1999; Malin and Edgett, 2000; Goldspiel and Squyres, 2011], but within Gale the morphology of the source basins, especially with valley heads often descending from narrow ridges located back from the walls of the crater (e.g. Peace Vallis and the West Canyon), argues for a “spatially distributed surface water source” [Morgan et al., 2014]. In addition, most of the source basins are eroded into the highest parts of the crater rim such that the potential groundwater source areas are negligible. It is possible, however, that lake infilling could have been derived from regional groundwater at depth leading to local lake effects and snow. In either case, it is likely that most of the water driving incision and delta and fan formation is derived from the melting of ice and snow at the surface. A terrestrial analog might be the McMurdo Dry Valleys, in which most precipitation falls as snowfall and channelized ephemeral streams are sourced completely from meltwater [Conovitz et al., 1998; Fountain et al., 1999].

Global climate models show that snow accumulation and melting is possible in the equatorial highlands of Mars, assuming a faint young sun, high obliquity, and a denser CO₂ atmosphere [e.g., Wordsworth et al., 2012; Kite et al., 2013], where snowmelt rates up to 2 and 3 mm/hr are possible [Kite et al., 2013]. These melt rates are adequate to drive even our highest

estimated discharges. Furthermore, the relatively well-preserved rim of Gale [Irwin et al., 2005] may have helped to create a microclimate that could trigger appreciable snow accumulation around Gale [e.g. Moore and Howard, 2005].

If our hypothesis is correct, and Gale shifted from hosting crater-wide lakes to more localized fan-building events, then water was perhaps supplied regionally early on, from the area draining Farah Vallis to the south, and then more locally as the climate became drier. A survey around Gale (within about 600 km of the crater) shows that there are 16 locations where relatively short deep canyons breach a crater rim or the edge of a block of the dichotomy, and in 15 of these cases the channels enter from the south to southwest- in the direction of mean topographic gradient, just as Farah Vallis does. In most of the cases, the canyon tips thin out to the south towards wide, shallower valley systems often strongly degraded but that in some cases can be traced more than 100 km to the south. In 13 of these cases, the steep canyons terminate in arcuate shaped deposits with steepened edges suggesting deltas. This suggests that at least at the time of Farah Vallis incision there was a regional hydrologic event resulting in large volumes of surface water flowing downslope from the south towards the dichotomy.

Fasset and Head [2011] compiled the work of others to show that prior to the mid-Hesperian, Mars had a higher impact flux, was wetter, had more volcanic resurfacing, a pH-neutral alteration environment, an intense magnetic dynamo and perhaps a denser atmosphere. These processes likely did not all happen synchronously, but based on their Figure 5.2 [Fasset and Head, 2011], all besides large basin formation (e.g. Hellas and Argyre) and their baseline core dynamo scenario persisted at least into the end of the Noachian (~3.5 to 3.75 Ga) and often into the Early Hesperian. Volcanism continued at a relatively high average rate throughout the Hesperian, resulting in the resurfacing of at least 30% of the planet, and large water floods formed episodically [Carr and Head, 2010]. Observations suggest that by the end of the Noachian most aqueous activity at the surface of Mars was suppressed, resulting in the growth of a thick cryosphere, though the occasional presence of Hesperian valley networks indicates that water activity did not decline to zero [Fasset and Head, 2008; Carr and Head, 2010]. By the Amazonian, the average rate of volcanism was approximately a factor of ten lower than in the Hesperian and small floods occurred only episodically, likely driven from sudden groundwater release or the drainage of lakes [Ghatan et al., 2005; Harrison and Chapman, 2008]. The formation of features attributed to the presence, accumulation, and movement of ice dominated during this time, and changes in obliquity play a prominent role in ice stability relations [Head et al., 2003].

Taken together, this suggests that runoff flowing from the south to the north could have supplied water both for a global ocean and for Farah Vallis formation, as well as 16 other incised valleys that breach crater rims. Gale is unique in that its internal dynamics, as recorded by multiple lake levels and smaller incised valleys (as mapped by Anderson and Bell [2010]), are not seen in any other crater in the region. The nearest craters of comparable size or larger are older and more infilled, but they still expose ~1 km of inner crater wall, on which no relatively fresh incised valleys are observed. This suggests that local hydrologic processes were important at Gale, in addition to regional water inputs. Kite et al. [2011] propose, based on modeling, that “intense localized precipitation will occur for a lake size $\geq 10^3 \text{ km}^2$ ”. Due to Mars’ low-density atmosphere, deep convection can be triggered for even just small amounts of latent heat released,

and since the temperatures are below 273 K, condensation will occur at low altitudes above the lake, leading to continuous precipitation as snow. As all of the large lakes within Gale are on the order of $3 \times 10^3 \text{ km}^2$ or greater, it is possible that the presence of these lakes are responsible for the relatively fresh-looking incised valleys rimming the crater. The Peace Vallis fan however is located below these major lakes, on the crater floor, meaning that local precipitation occurred independently of lakes at Gale as well.

5.8 Proposed history of hydro-geomorphic events with Gale Crater

Gale crater formed after a period of ubiquitous crater degradation, which terminated during the late Noachian [Irwin and Howard, 2002], and its ejecta appears to have disrupted a valley network draining $270,000 \text{ km}^2$ to its south (Figure 5.33a-5.33c), whose tributaries begin on the rims of craters that overlie the larger Herschel crater rim [Irwin et al., 2005]. Crater counts on the wall and floor of Gale (Figure 5.30 and Thompson et al., 2011) suggest that the mound reached (generally) its current topographic form before the deltas were deposited (Figure 5.33d). Very shortly after impact, erosion and delivery of sediment to the floor of Gale likely occurred, as large Gale-forming impacts are predicted to result in the release of groundwater due to impact-induced liquefaction and increases in surface temperature above 273 K on the order of ~ 100 days. The setback of the crater rim into which Peace Vallis formed may have occurred then, as crater counts of the sediment source area catchment date it to be $\sim 3.6 \text{ Ga}$ (Figure 5.31). Some of the sediment delivered from the crater rim post-impact likely comprises the ridge that borders the western edge of the local topographic depression into which the Peace Vallis fan eventually flowed (Figure 5.33e, Figure 4.6 [Chapter 4]).

Surface runoff from the valley network to the south of Gale appears to have flowed into the subsurface and seeped out near the crater rim, eventually incising Farah Vallis. Though the exact timing of incision has not been established, it occurred after Aeolis Mons reached its current extent. Runoff and sediment outflow spilled into Gale to form the Pancake delta and the adjacent plains to the west, while maintaining the 700 m deep Pancake Lake (Figure 5.33f). During this time, it is also possible that much (up to 80%) of the sediment comprising the Farah Vallis delta was also deposited on the western floor of Gale, which would have been the lowest point topographically. Runoff from Farah Vallis eventually decreased or ceased altogether, resulting in a considerable drop in the Pancake lake level, at least below the entrance canyon deposits (Figure 5.33g). At a later time, increased in local precipitation (or perhaps groundwater seepage) resulted in lake level rise that lasted long enough to form the lower Western canyon delta (W1) and the middle Entrance canyon delta (Figure 5.33h). Farah Vallis likely contributed much of the water needed for lake level rise, though groundwater seepage from the south may also have supplied water to Gale. Evidence of the former is incision into the plains to the west of the Pancake delta (Figure 5.7). The lake level rose again, approximately 200 meters, to the -3780 meter lake level, resulting in the upper Western canyon delta (W2) and an associated deposit to its southwest (W3) and the Farah Vallis delta front (Figure 5.33i). This hydrologic system must have shut down sufficiently abruptly that the deltas did not cut down as the lake evaporated (or if frozen, as they sublimated). The time gap between these two major lake forming events has not been established, but is perhaps driven by different hydrologic systems (distal via Farah Vallis/runoff from the south and local via precipitation/lake effects). As lake levels declined a second time, local precipitation (again, likely snowmelt) may be associated with a time of fan

building around Gale (Figure 5.33j). Large fans around Mars have been dated to the late Hesperian to the early Amazonian [Grant and Wilson, 2011; Morgan et al, 2014], and it is possible that the much smaller fans in Gale were contemporaneous. The lack of shorelines on these features can be used to argue that they formed after these large lakes, but as none of the deltas in Gale show evidence of wave-cutting, it is possible that shoreline development was weak or non-existent and hence the absence of shorelines is not strong evidence for the timing of fans relative to deltas.

Several other alternative lake histories for Gale have been proposed in the literature, which differ some from the timeline presented here. Pelkey and Jakosky [2002], using THEMIS data, claim that the valleys rimming Gale Crater would post-date any large lake in Gale since they extend to the crater floor, but our analysis shows that many of these valleys likely formed when the lakes were present. Irwin et al. [2005] suggests that post-impact there was a brief re-activation of the highland valley networks resulting in the breach of Gale's rim, the dissection of Aeolis Mons, the formation of gullies, and the deposition of fans and deltas. Aeolis Mons is posited to have formed prior to this time during a hiatus in erosive valley network activity. While we are in general agreement with this model, improved imagery and topography allows for a more precise timeline of events. Anderson and Bell [2010] suggest a lacustrine origin for the rocks comprising Aeolis Mons, following Malin and Edgett [2000], but do not make mention of any large lakes after Aeolis Mons reached its current topographic form. They also examined the Entrance Canyon and its deposits (which they call the "fan-shaped unit") and suggest that material from the mound eroded and deposited as an alluvial fan atop what they refer to as the "mound-skirting material", which they describe as being pitted, mesa-forming, and having parallel ridges at times. This is mapped in Figure 5.15 (numbered "5") as our broad, nearly uniform sloping plain. Figure 35 in Anderson and Bell [2010] shows the fan eventually eroding back, as well as the mound-skirting unit/sloping plain, leaving the current-day extent of the deposit (Figure 5.15). The clear transitions between different units within the Entrance Canyon deposits, as well as the correspondence between the elevation of deposit 3's (orange) front with the Western W1 delta front and the coarse nature of the deposits (which would be difficult to erode by aeolian processes), points to multiple stages of deposition, some into a standing body of water (as opposed to a single deposit that formed sub-aerially and eroded back). Le Deit et al. [2013] interpreted the lower Western delta (W1) to be a delta as well, suggesting a paleo-lake occurred at -4200 m (the elevation of the delta's apex). They also infer that the terraces on the western flank of Aeolis Mons were wave-cut terraces, the highest located at -3700 m (though they note this terrace is not horizontal) and therefore posit a second (highest) lake level at -3700 m. The fan shaped deposits on the western and southern slopes of Aeolis Mons are assumed to be retrogressive thaw slumps that may have partially been deposited in standing water. The Pancake delta is referred to as a "fan-like" deposit and the Farah Vallis delta is not mentioned. They cite as a water source for these lakes to be the incised valleys and sinuous ridges rimming Gale crater. In their first model, they assume that the Gale region was characterized by continuous permafrost overlain by an active layer that was subject to freeze-thaw cycles. During suitable obliquity and orbital conditions, snow was both stable at the surface and able to melt, though a shallow frost table prevented infiltration thereby causing overland flow and the formation of the incised valleys rimming Gale and the deep canyons on the flanks of Aeolis Mons. The lakes would have been ice-covered and recharged primarily by snowmelt. In a second model, they assume discontinuous permafrost, such that the lakes would have been supplied by snowmelt percolating

through the mound via fractures and groundwater flow. While possible that permafrost played some role at Gale, we do observe strong evidence for features one might expect (i.e. pingos, gelifluction or solifluction lobes, rock glaciers, eskers or stony patterned ground). On their timeline (see Figure 22 in Le Deit et al. [2013]) they place lake formation after the eastern and western flanks of Aeolis Mons developed (whose elevation corresponds to the northern rim), but coeval with the yardang units stratigraphically above these flanks and their crater floor units 1 through 4, which they say may have continued depositing until 1.1 Ga, long after the lakes had disappeared. Crater floor unit 3 encompasses our deposits associated with the Pancake delta, which we hypothesize deposited early on during the incision of Farah Vallis and crater floor unit 4 contains the Western delta deposits. The crater-wide lakes and most fluvial activity occurred from about 3.46 Ga into the Late Hesperian, with episodic fluvial activity and continued deposition of crater floor units 1 through 4 up to 1.1 Ga. Figure 5.34 shows our updated timeline, which shows a sequence of lake level rise (Pancake Lake) around 3.4 Ga, an abrupt decline, followed by a rise to the Western Lake and then a second rise to the Farah Vallis lake. A decline in regional water supply, either from reduced runoff from Farah Vallis or reduced groundwater influx from the south led to the disappearance of lakes. Many of the smaller fans around Gale's rim may have formed when the lakes were present, due perhaps to lake effects [Kite et al., 2011], but the Farah Vallis fan atop the Farah Vallis delta and the low elevation of the Peace Vallis fan on the northern crater floor suggests a later-stage episode of fan-building as the water supply became more local to Gale.

5.9 Conclusions

Detailed examination of delta and fan forms around Gale crater support past work that Gale once hosted large, crater-wide lakes and with the use of CTX, HiRISE and HRSC topography, we are able to document the most probable sequence of lake levels, thereby providing a likely relative timeline of the hydrologic events within Gale crater. The largest lake within Gale (besides a crater-filling lake if Gale were connected to a northern ocean) is best defined by the Pancake delta off of Gale's southern rim. This Pancake lake would have been ~700 meters deep and could have persisted for 400,000 to 10,000 earth years (based on upper end estimates of the time needed to build the Pancake delta). The lake level eventually dropped below the elevation of the Entrance Canyon deposits, before rising again to the Western lake stand, defined best by the Western delta W1 deposit. The Western lake was ~300 meters deep and would have lasted up to 40,000 to 950 earth years before rising again to the Farah Vallis stand. The Farah Vallis lake, defined best by the Farah Vallis delta, was 400 meters deep and would have persisted for up to 670,000 to 18,000 earth years. The Peace Vallis fan, which is located near the MSL *Curiosity* landing site, is located on the crater floor, topographically under all three of these large lakes. If it formed contemporaneously with the Farah Vallis fan and the Western fan-like feature, then it post-dates these large lakes. If it formed during a period of lake decline, then it is possible that its sediments were at one point saturated. The lack of a shoreline on the fan suggests the former, but shoreline detection within Gale is difficult, perhaps due to limited shoreline development, erosion of shorelines from aeolian processes, or too low image resolution across the crater.

In any case, these three high lake levels within Gale crater would have caused saturation of the deposits in Mt. Sharp and in the crater walls. As lake levels in Gale decreased, the low

area immediately adjacent to the *Curiosity* landing site would have been one of the last areas to dry out, possibly contributing to shallow localized lake formation along the floor of the crater. This provides context for the geology and sedimentology along the current *Curiosity* rover traverse from Bradbury Landing to the base of Mt. Sharp, as all of the deposits that *Curiosity* will examine would have experienced one or more periods of saturation-desaturation cycles.

We posit that the reason Gale shifted from hosting crater-wide lakes to more localized fan-building events was that water was perhaps supplied regionally early on from the south, and then more locally as the climate became drier. Our survey around Gale shows 16 locations where relatively short deep canyons breach a crater rim or the edge of a block of the dichotomy, all in the direction of the mean topographic gradient, suggesting that at least at the time of Farah Vallis incision there was a regional hydrologic event resulting in large volumes of surface water flowing downslope from the south towards the dichotomy. This has important implications for understanding regional paleo-climates on Mars in the post-Noachian.

5.10 Acknowledgements

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5.11 Tables

Analysis	Data Source
1. Delta / Fan Planform (Contours, Extent, Long Profiles)	CTX and HRSC DEM
2. Sediment Budgets	MSL, CTX, and HRSC DEM
3. Paleo-channels on FV and PV fans	MSL and CTX DEM
4. Gale Catchment Areas	CTX and HRSC DEM
5. PV Fan Reconstruction	MSL DEM
6. Lake Levels, Depths and Volumes	HRSC DEM
7. Crater Count of Gale Walls and Floors	CTX

Table 5.1 Data Sources used for analyses (DEM = Digital Elevation Model, CTX = MRO Context Camera derived DEM, HRSC = Mars Express Orbiter High Resolution Stereo Camera, MSL = Mars Science Laboratory provided HIRISE-based DEM)

Name	Value	Units	Defined By:
<i>Western Lake</i>			<i>-3980 meter HRSC</i>
Area	3008	km ²	1. Western Delta (W1) - CR
Mean Depth	0.3	km	2. Entrance Canyon Delta ("3") - AM
Max Depth	0.7	km	Western AM deposits (WAM)
Volume	915	km ³	
<i>Farah Vallis Lake</i>			<i>-3780 meter HRSC</i>
Area	3616	km ²	1. Farah Vallis Delta – CR
Mean Depth	0.4	km	2. Western Delta (W2, W3) - CR
Max Depth	0.9	km	
Volume	1582	km ³	
<i>Pancake Lake</i>			<i>-3280 meter HRSC</i>
Area	5832	km ²	1. Pancake Delta – CR
Mean Depth	0.7	km	2. South Delta - AM
Max Depth	1.4	km	
Volume	3780	km ³	

Table 5.2. General properties for the three major lake stands within Gale Crater. AM = Aeolis Mons, CR = Crater Rim

Name	Value	Units	Source
<i>Pancake delta</i>			CTX
Area	117	km ²	
Mean Depth	0.1	km	
Volume	14.0	km ³	
<i>FV Fan/Delta</i>			CTX / HRSC
Area	78.2	km ²	
Mean Depth	0.3	km	
Volume	22.9	km ³	
<i>Farah Vallis</i>			CTX / HRSC
Length	84	km	
Drainage Area	270,000	km ²	
Removed Volume	40.2	km ³	
<i>West Delta</i>			CTX
Area	7.0	km ²	
Mean Depth	0.1	km	
Volume	1.4 / 1.5	km ³	
<i>West Delta Valley</i>			CTX
Length	76	km	
Drainage Area	450	km ²	
Removed Volume	1.6	km ³	
<i>Entrance Canyon</i>			HIRISE
<i>Deposits</i>			
Mean Depth	0.17	km	
Volume	0.016	km ³	
<i>Entrance Gully</i>			HIRISE
Length	1.7	km	
Drainage Area	68	km ²	
Removed Volume	0.004	km ³	
<i>Peace Vallis Fan</i>			HIRISE
Area	80	km ²	
Mean Depth	7	m	
Volume	0.9	km ³	
<i>Peace Vallis</i>			HIRISE / CTX
Length	107	km	
Drainage Area	730	km ²	
Removed Volume	0.8	km ³	

Table 5.3. General properties for the major valleys, deltas and fans within Gale Crater

5.12 Figures

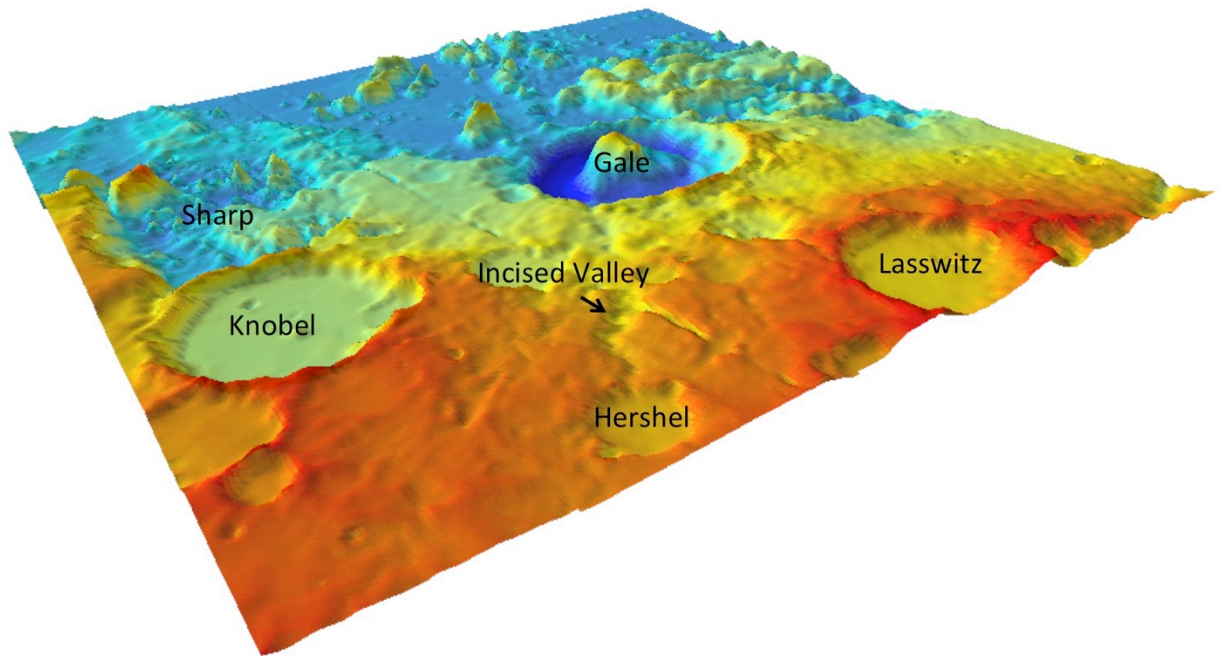


Figure 5.1 MOLA-derived topographic map of the Gale Crater region. Gale Crater is located on the crustal dichotomy between the cratered southern highlands and the smoother northern lowlands of Elysium Planitia, and it is this regional slope that contributes to the difference in elevation between the northern and southern rim. The lowest elevations in Gale crater occur on the northwest portion of the crater floor, near the MSL landing site. Gale's impact disrupted a pre-existing valley network originating near Herschel crater; this is likely one of the main sources of water into Gale crater (based on Irwin et al. [2005]).

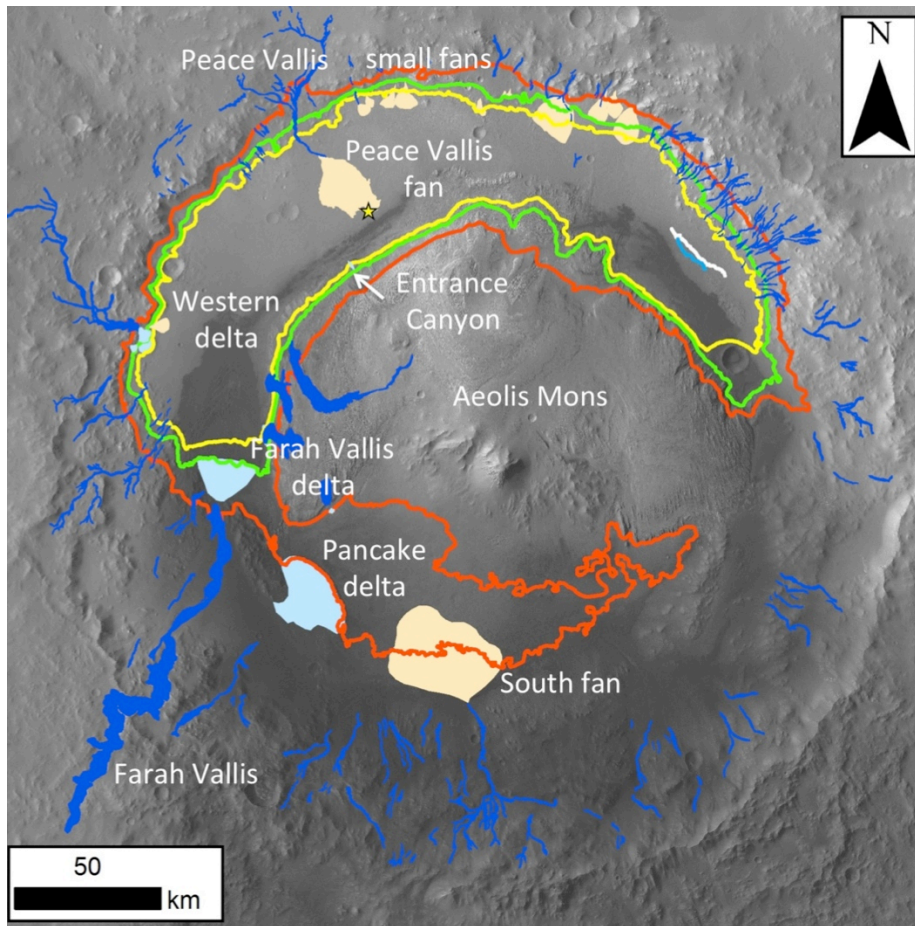


Figure 5.2 A modified version of the geomorphic map from Anderson and Bell [2010] showing their map of “valleys and canyons” (dark blue) and the Peace Vallis fan, where the MSL landing site has been marked with a star. Several large fans (tan) and deltas (light blue), were added to their map, and are discussed in detail within the text. The inferred paleo-lake levels are shown, which correspond to the -3280 meter HRSC contour (Pancake lake, orange), the -3780 meter HRSC contour (Farah Vallis lake, green) and the -3980 meter HRSC contour (Western lake, yellow).

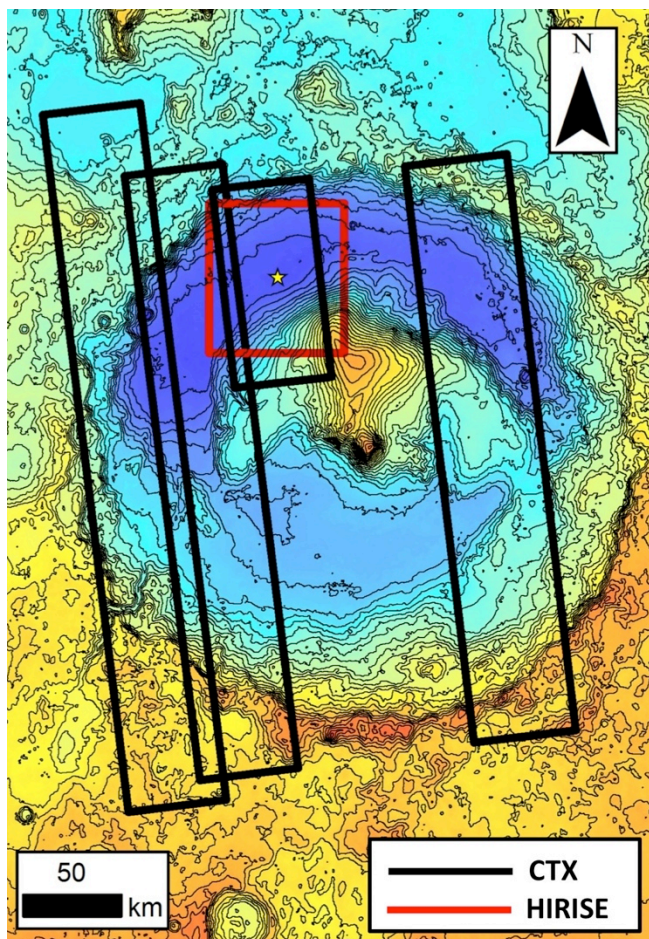


Figure 5.3 An HRSC elevation map of Gale crater with 200 meter HRSC contours. The areas for which we have CTX DEMs are outlined in black. The region where HIRISE coverage is available is outlined in red. The CTX DEM near the landing site (marked with a star) was provided by the MSL project, while the remaining CTX DEMs were created using CTX stereo pairs and the commercial software SOCET SET.

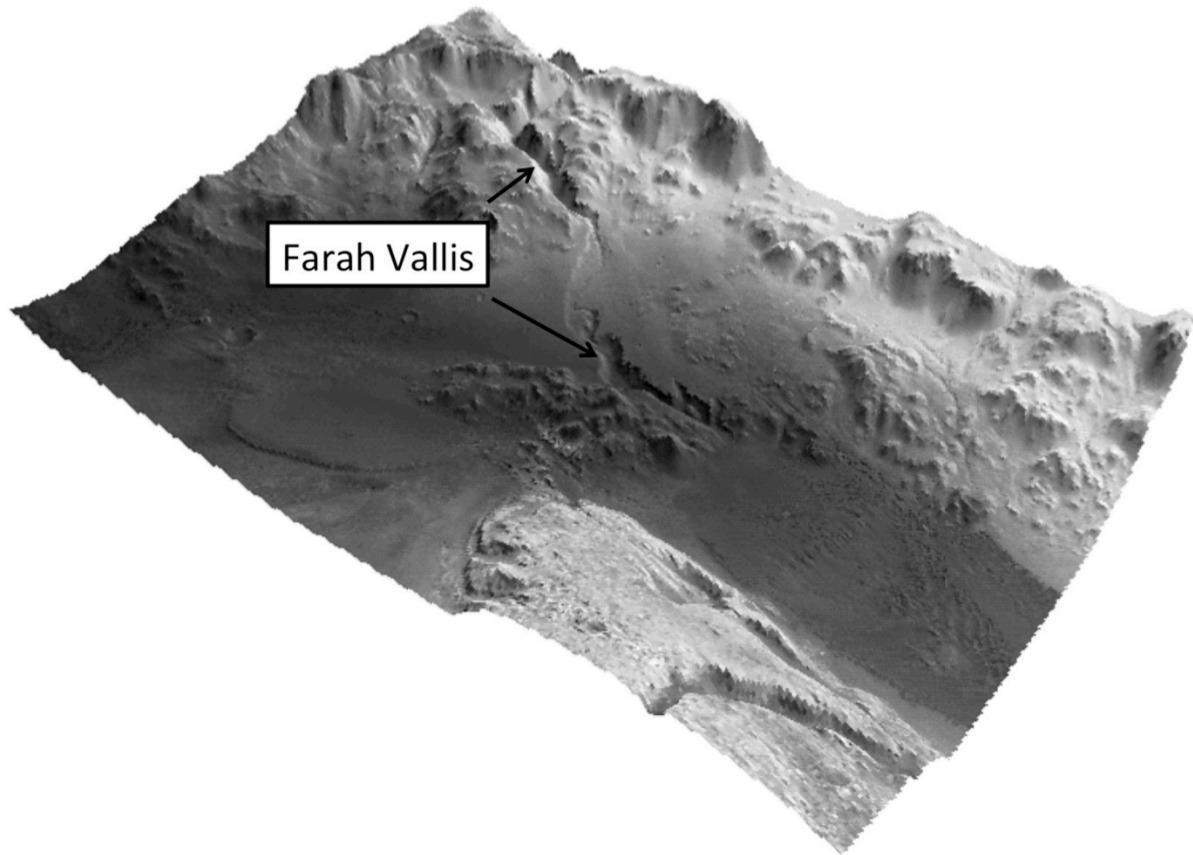
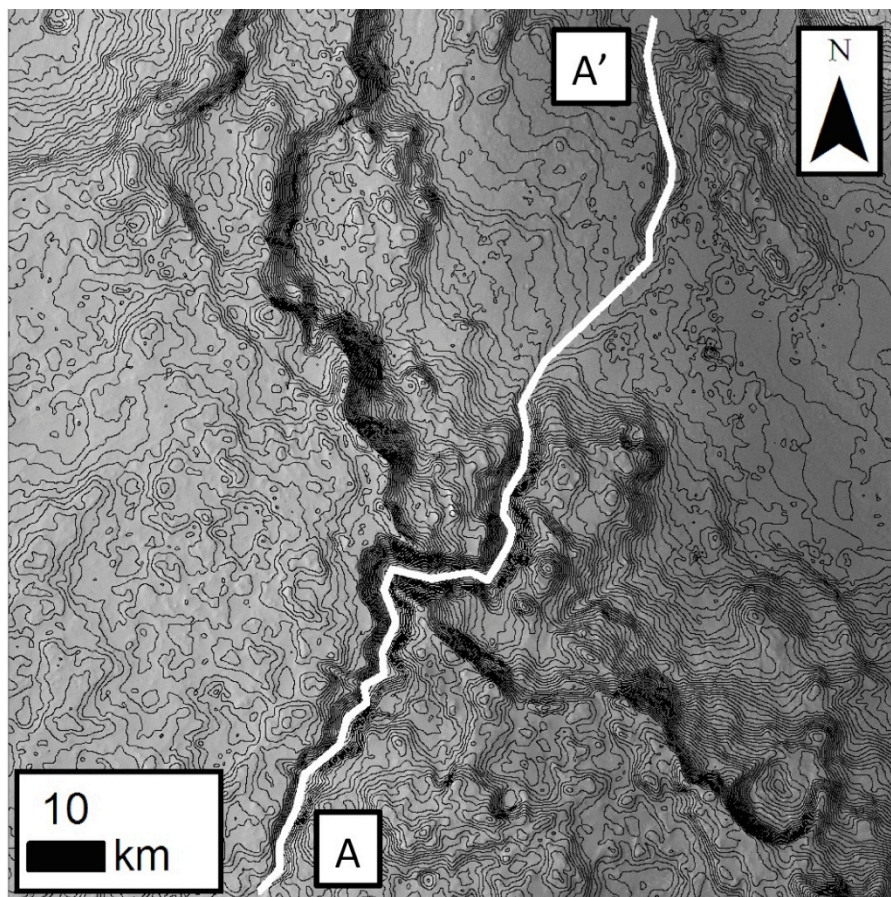
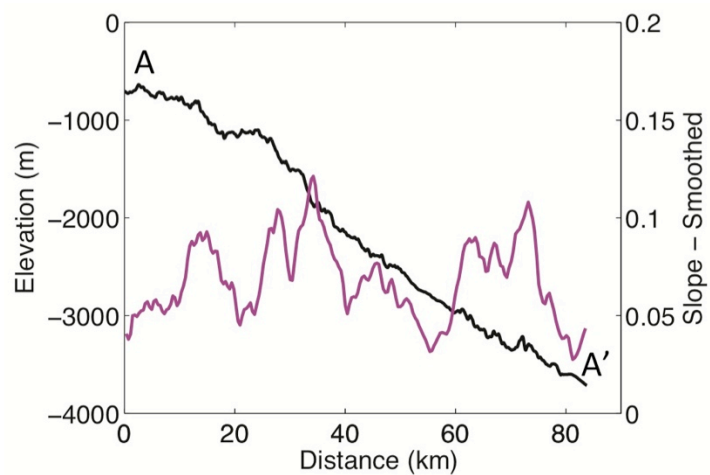


Figure 5.4 A 3D regional view of Farah Vallis on the southern rim of Gale crater with a CTX mosaic (courtesy of M. Malin) overlain. Farah Vallis drains an area of 270,000 km² and is over 74 kilometers long and on average 200 meters deep. It is the only entrance breach into Gale crater associated with an extensive external source area.



a.



b.

Figure 5.5 (a) 50-meter HRSC contours for the Farah Vallis region are shown, as is the location of the longitudinal profile (A is upslope of A'). (b) A longitudinal profile (black line) and slope profile (pink line) of Farah Vallis. Elevation data were extracted from the HRSC DEM and plotted as a function of distance on the left y-axis.

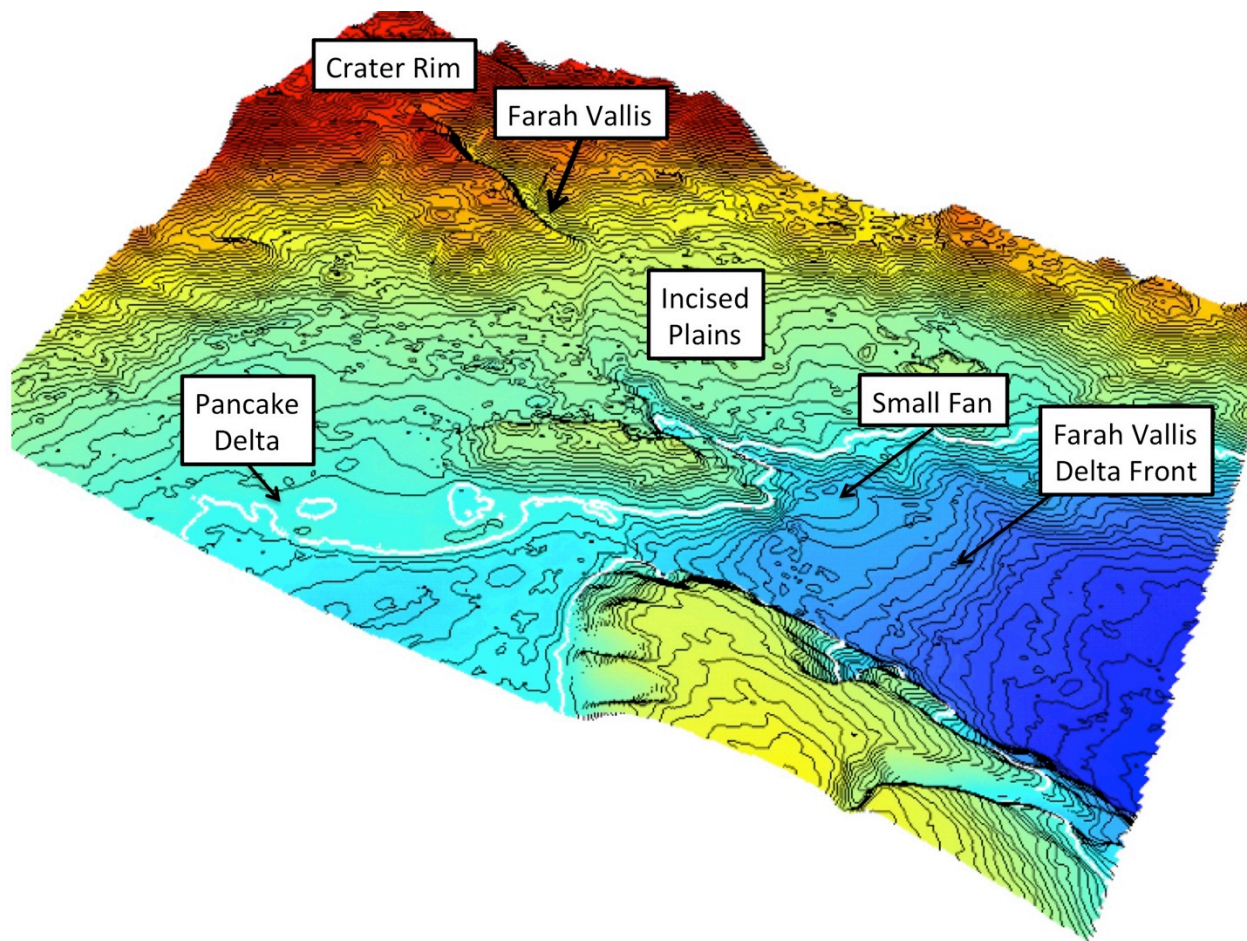


Figure 5.6 A three-dimensional view of the Farah Vallis fan system (vertical exaggeration of 1.5) with a colorized elevation map and 100 m HRSC contours. The contour line that corresponds to the front of the Pancake delta is shown in white to highlight where associated deposits to the west of the Pancake delta would have been deposited.

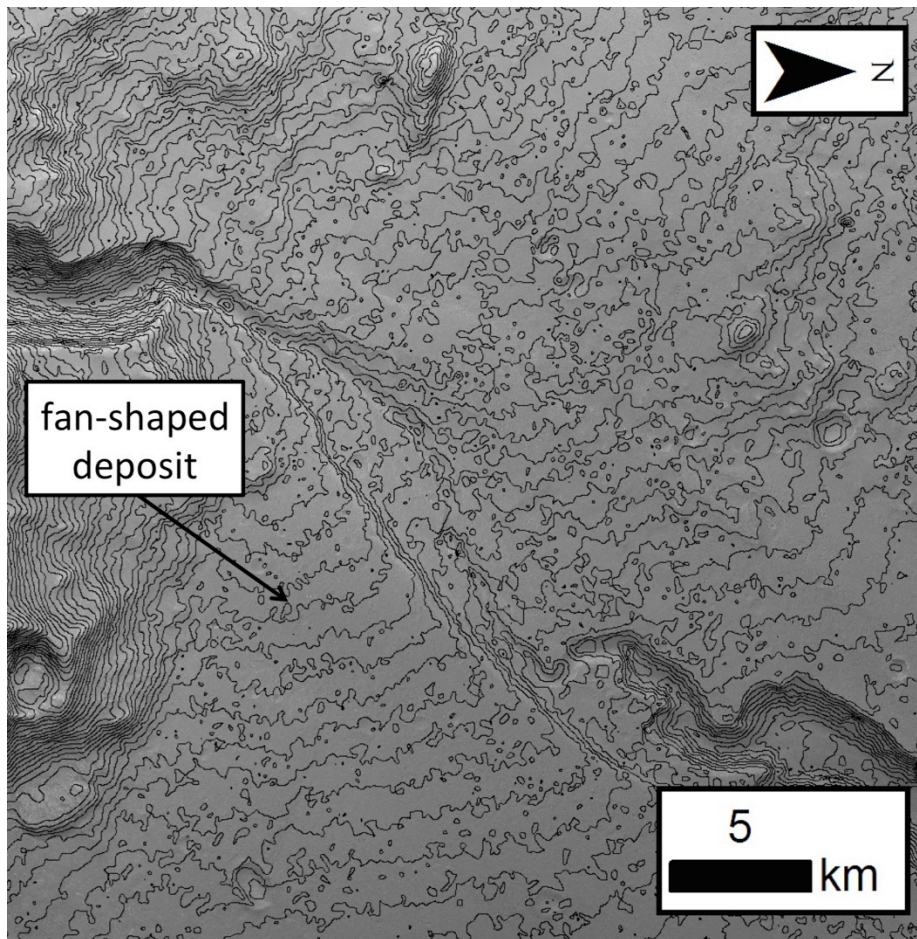


Figure 5.7 A close up view of the “incised plains” region from Figure 5.6, with 25 meter CTX contours overlain. This fan deposit is defined by distinct convex contours that extend across the southern Gale Crater wall and terminate a few hundred meters above the -3280 meter lake.

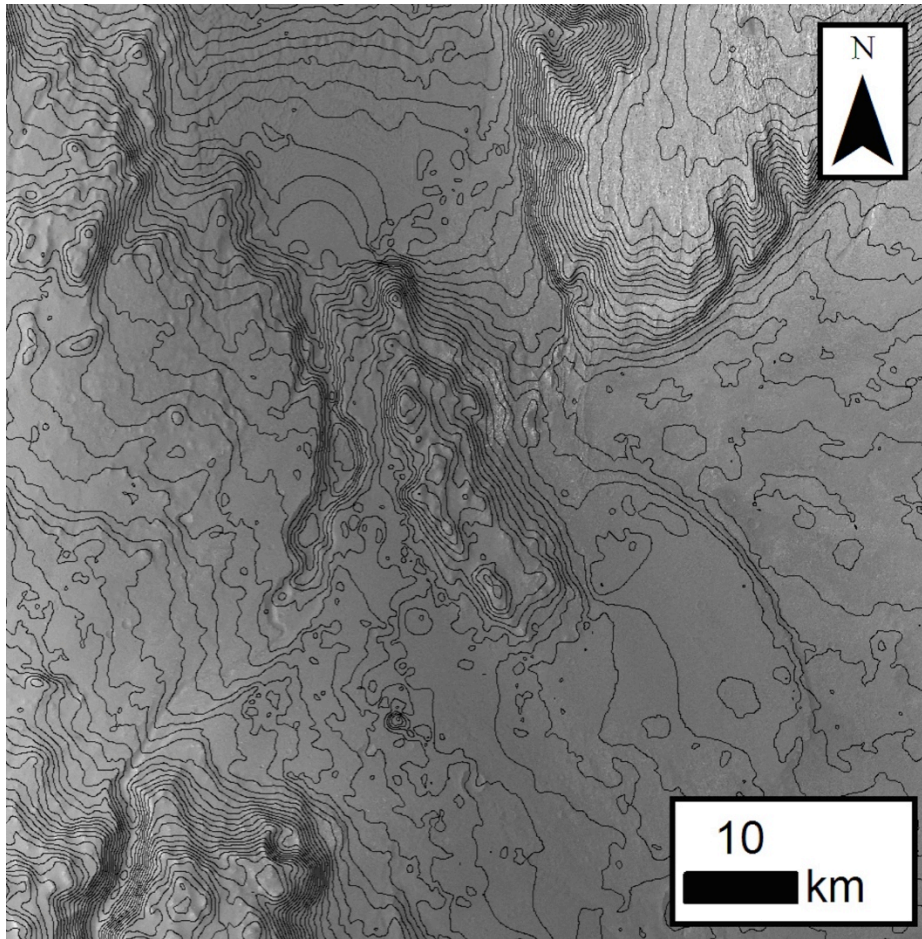
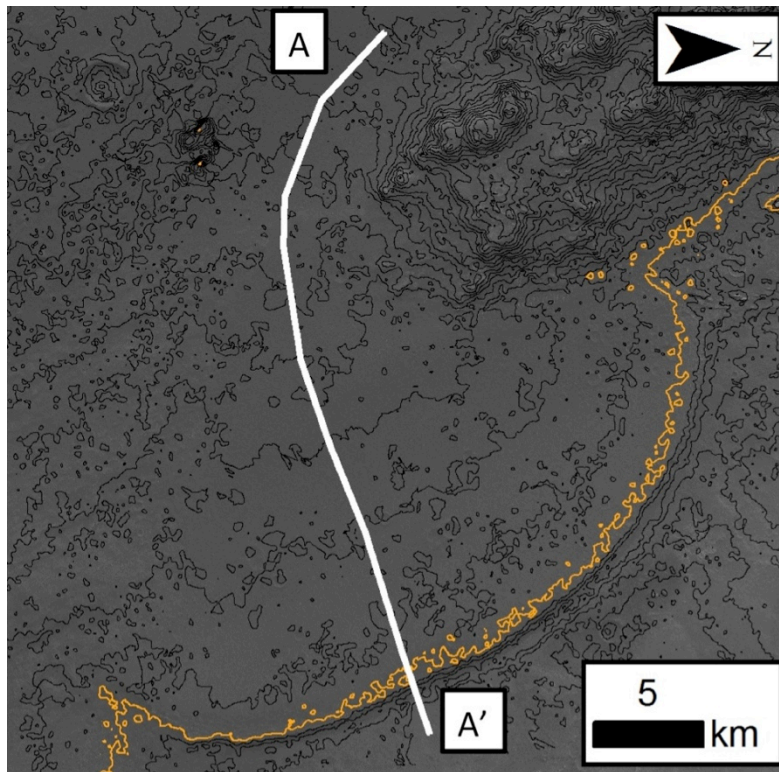


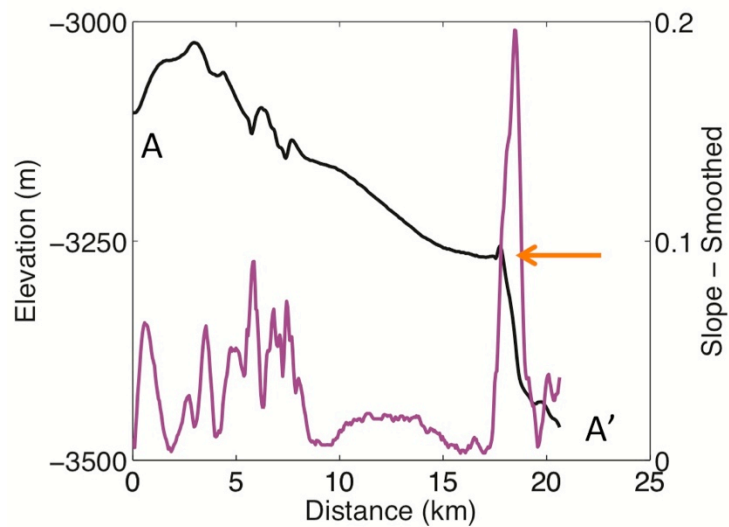
Figure 5.8 An overview of the lower Farah Vallis region, showing incision into the “Pancake delta associated deposits” and the fan highlighted in Figure 5.7. In the lower left, Farah Vallis becomes less prominent, decreasing in valley width and depth. At its lower end, Farah Vallis widens again before transitioning to a distinct fan-delta deposit.



Figure 5.9 CTX image B21_017786_1746_XN_05S222W of the Pancake delta, which is a dark-toned, arcuate-shaped feature that is best described as a classic gilbert-type delta, typical of freshwater/lake system here on Earth [Gilbert, 1890].

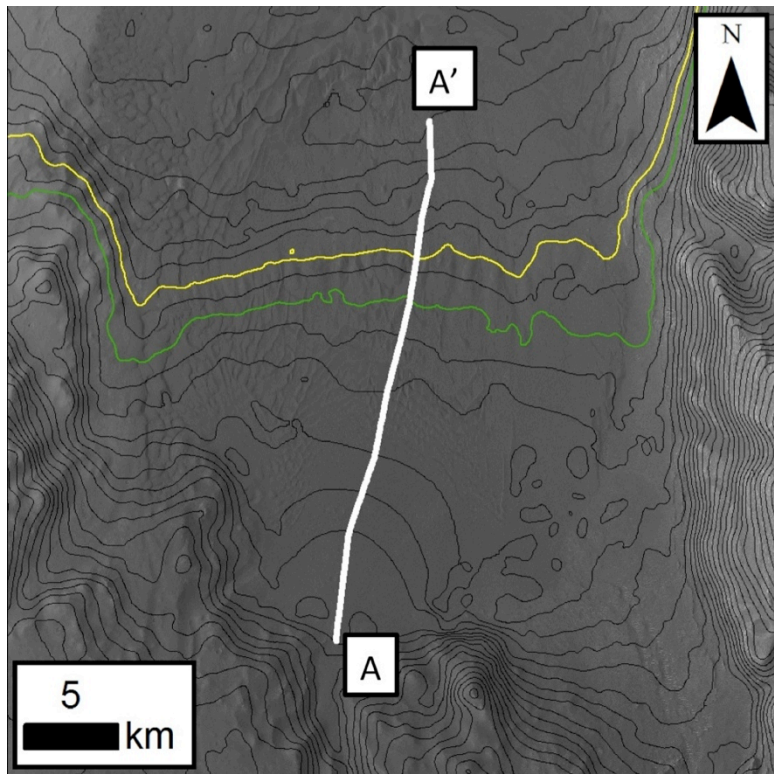


a.

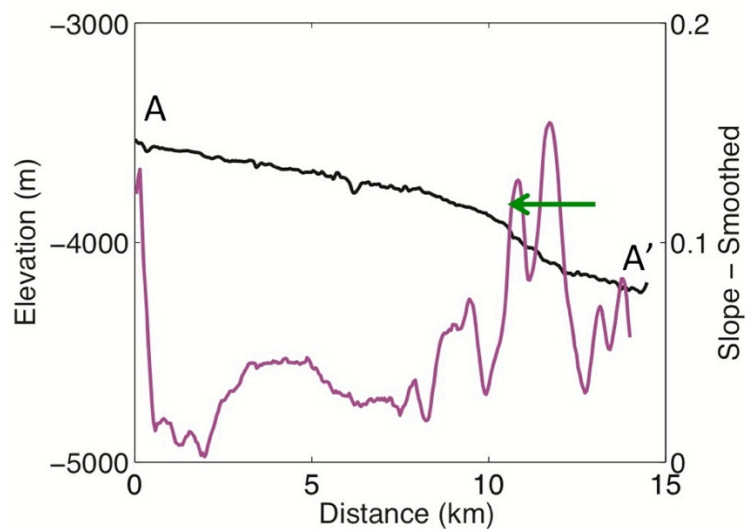


b.

Figure 5.10 (a) 25-meter CTX contours of the Pancake delta are shown, as is the location of the longitudinal profile (A is upslope of A') (b) A longitudinal profile (black line) and slope profile (pink line) of the Pancake delta. Elevation data were extracted from the CTX DEM and plotted as a function of distance on the left y-axis. Slope was calculated every 100 meters and the 10-point average is plotted as a function of distance. The -3280 meter contour corresponding to the delta front is shown in orange, and an orange arrow points out the delta front within the longitudinal profile and slope data.



a.



b.

Figure 5.11 (a) 100-meter HRSC contours of the Farah Vallis fan and delta are shown, as is the location of the longitudinal profile (A is upslope of A') (b) A longitudinal profile (black line) and slope profile (pink line) of the Farah Vallis fan-delta. Elevation data were extracted from the HRSC DEM and plotted as a function of distance on the left y-axis. Slope was calculated every 100 meters and the 10-point average is plotted as a function of distance. The -3780 meter contour corresponding to the delta front is shown in green, with a green arrow pointing out the delta front within the longitudinal profile and slope data. For reference, the -3980 m contour is shown in yellow.

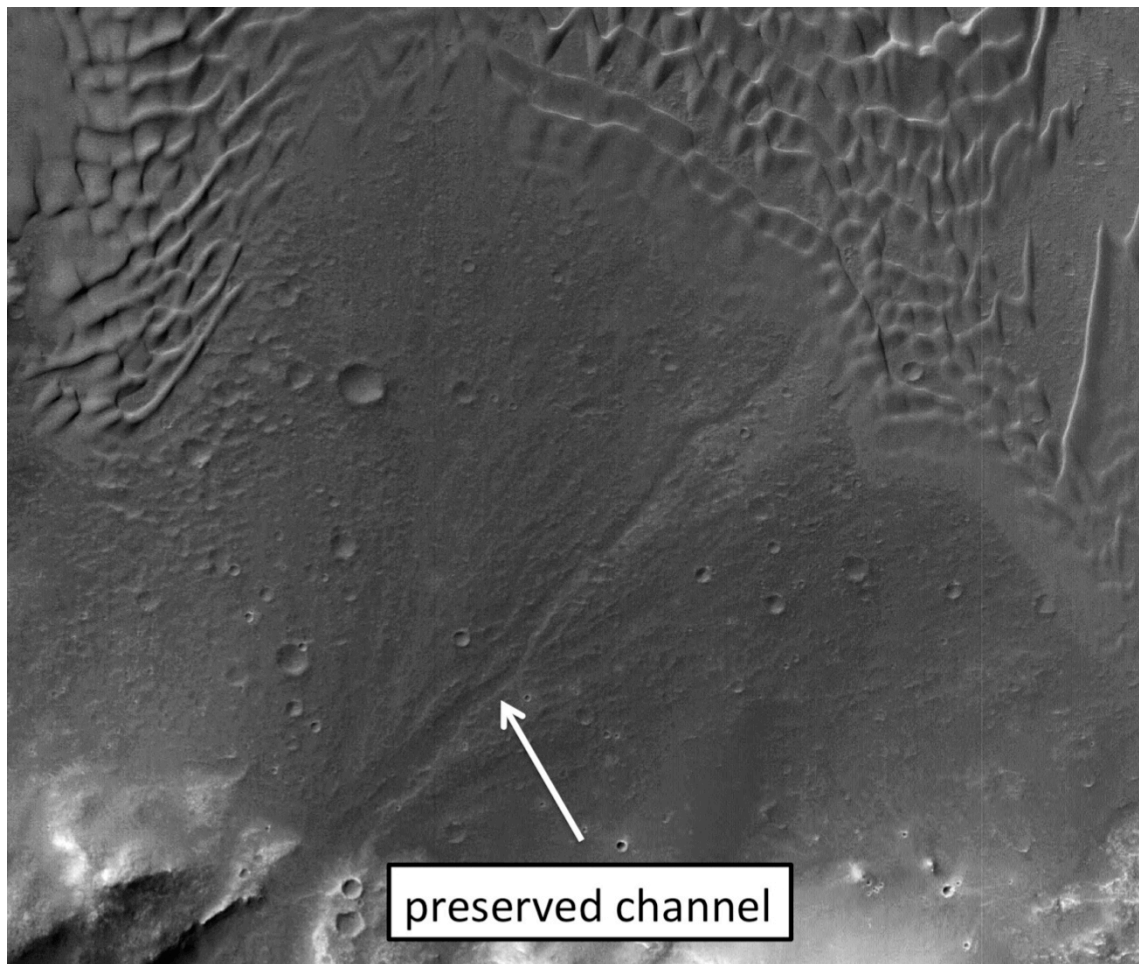


Figure 5.12 CTX image P13_006143_1745_XN_05S223W of the fan deposit, based on its arcuate convex contours (see Figure 5.11), overlying the Farah Vallis delta deposit. This deposit covers approximately 25 km² and is relatively thin. It preserves a large channel running down its center that is nearly 5 kilometers long.

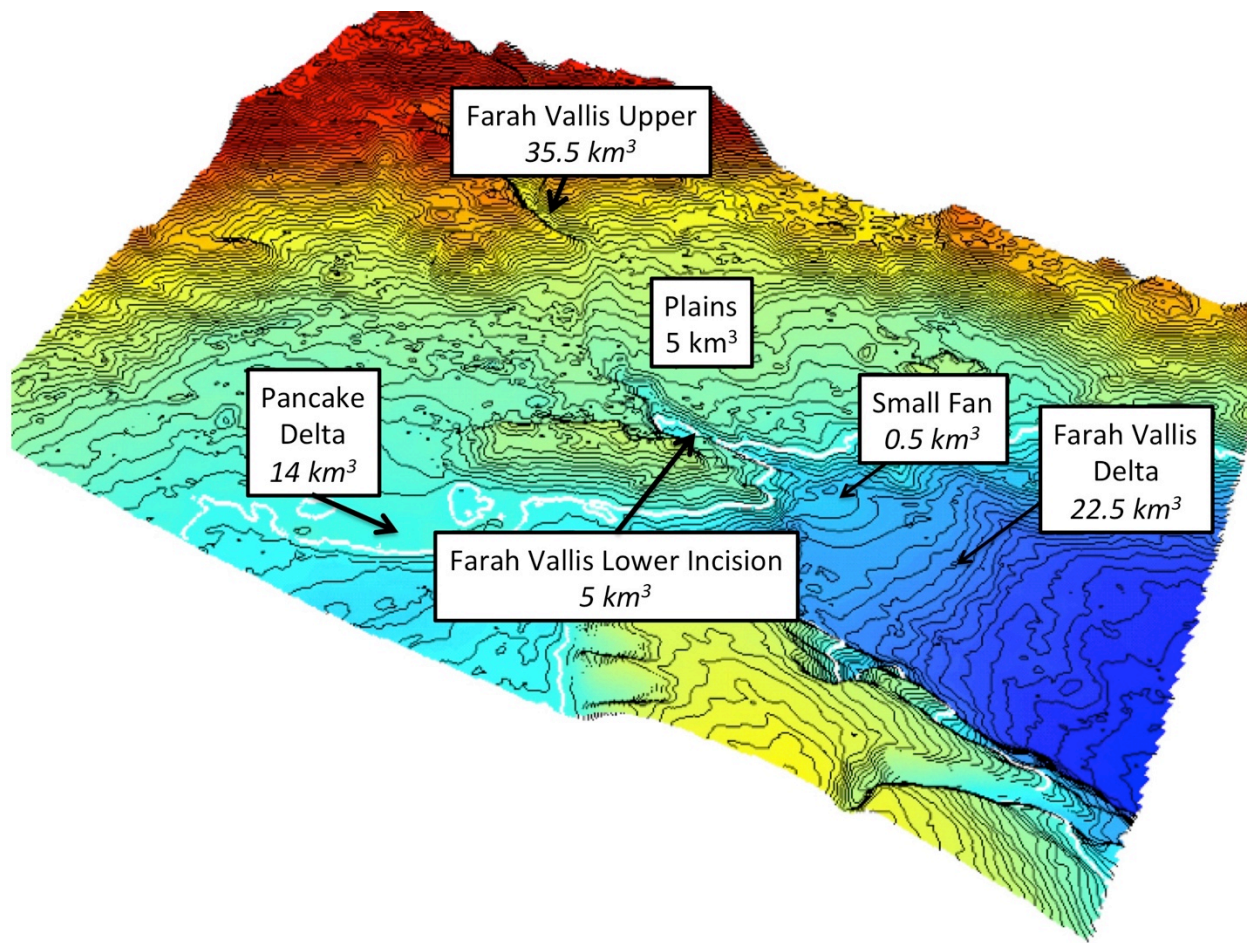


Figure 5.13 This is the same overview of the region shown in Figure 5.6, though the volumes of sediment associated with depositional features and the removed volumes of sediment from erosional features are shown. The Pancake delta and its associated deposits, the Farah Vallis fan-delta, and the fan-shaped deposit emanating from the upper half of Farah Vallis comprise 42 km^3 of material, which approximately matches the 40 km^3 of material removed to create Farah Vallis.

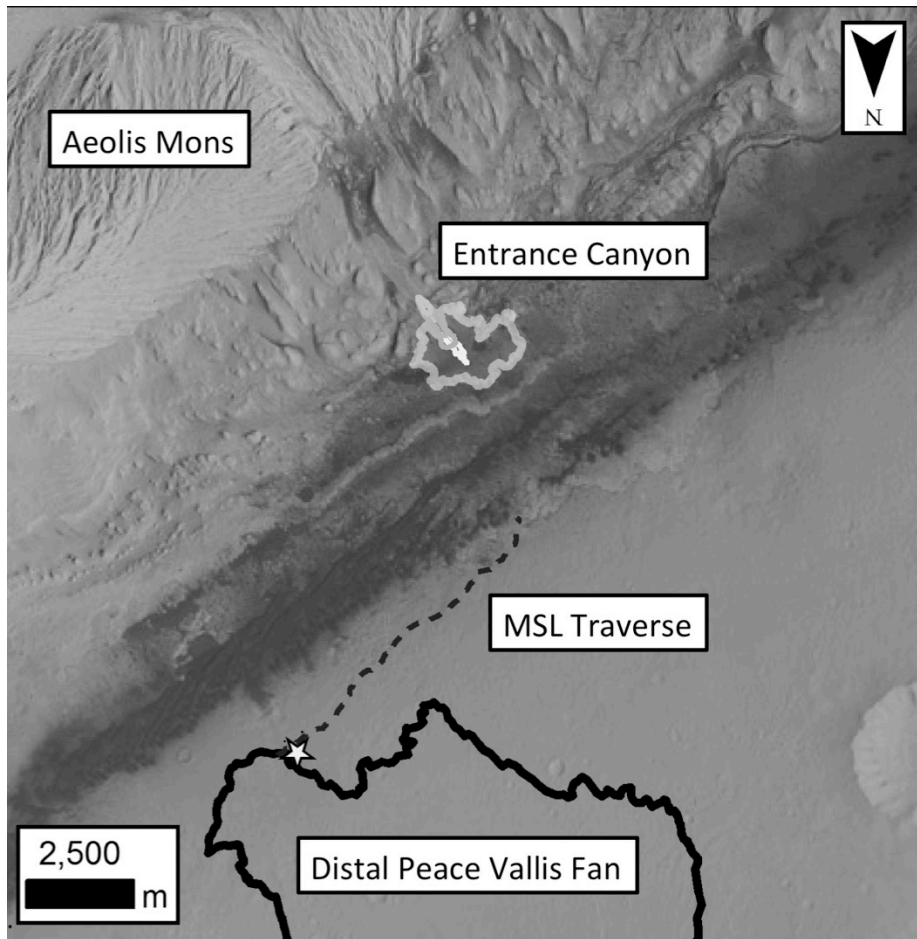


Figure 5.14 The entrance canyon is located on the north side of Aeolis Mons and is a destination for the *Curiosity* rover during its ascent of Aeolis Mon. Its relation to the distal end of the Peace Vallis fan and the MSL landing site is shown.

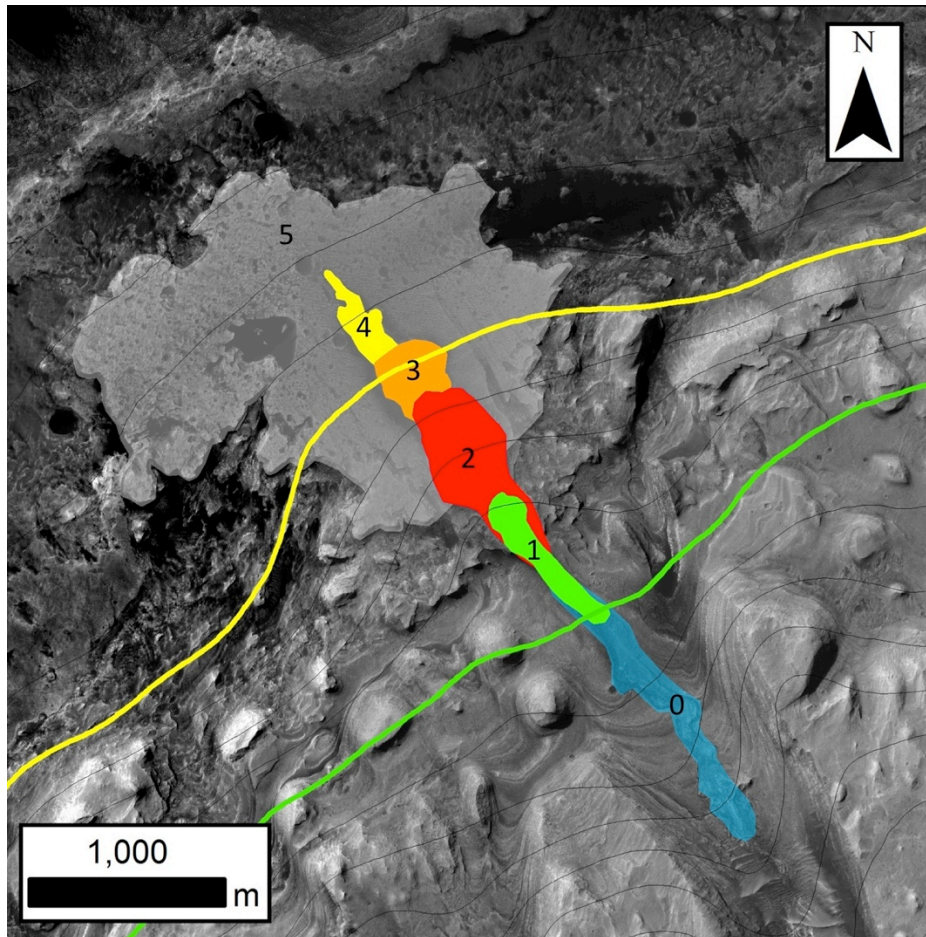
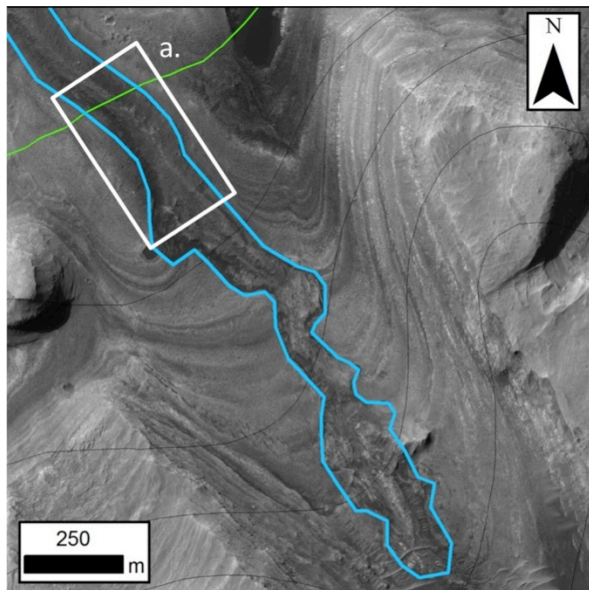
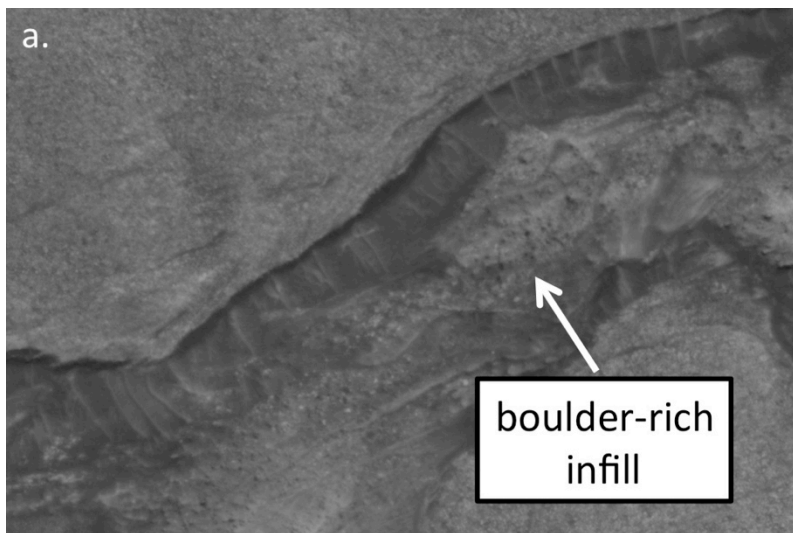


Figure 5.15 Within the entrance canyon there is an approximately ~1 kilometer long gully (blue) that eventually transitions to a boulder-rich, relatively narrow, and crudely stratified deposit (green). This upper deposit transitions to a second, likely deltaic, deposit (red) that is wider, steeper-sided and spreads radially. This deposit extends onto a second distinct deltaic deposit (orange), which spreads laterally to the east and appears to have inverted distributary channels. The lowest deposit is shown in yellow and may preserve strata deposited by large-scale bedforms. Underlying the entire sequence is a broad, nearly uniform sloping plain (grey). HIRISE images ESP_019988_1750 and ESP_019698_1750.

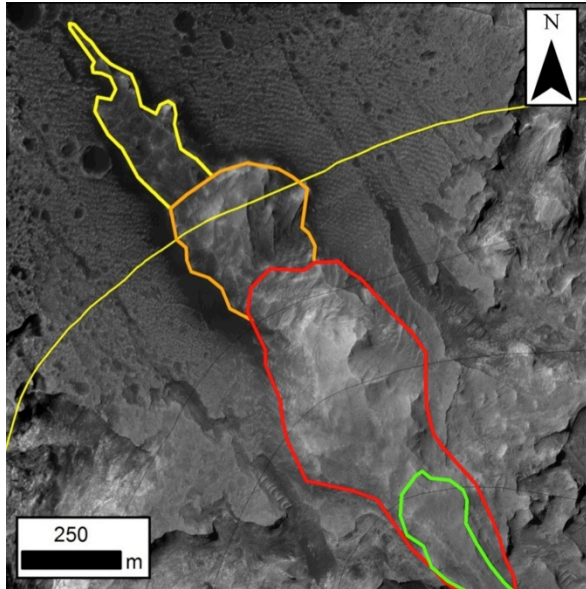


a.

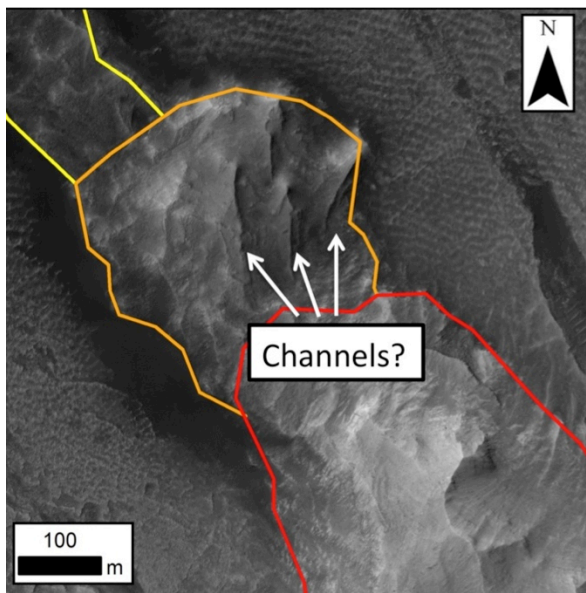


b.

Figure 5.16 (a) A view of the gully (outlined in blue) within the entrance canyon and (b) a close up view of the region outlined that white in (a) showing the boulder-rich infill.

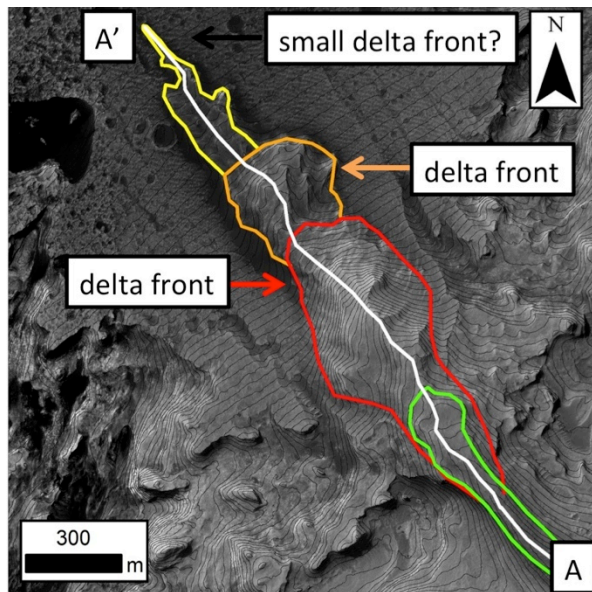


a.

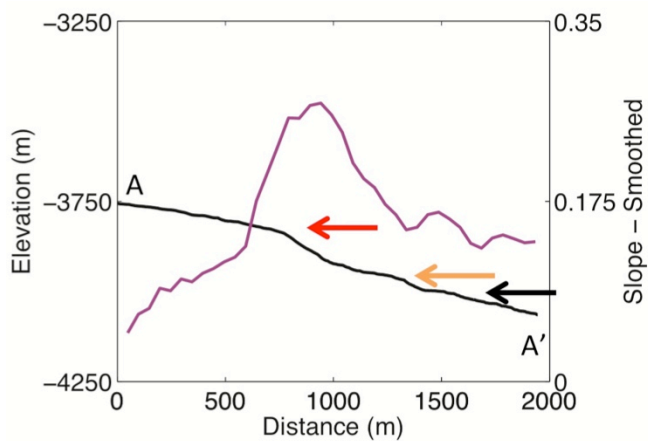


b.

Figure 5.17 (a) A close-up view of the entrance canyon deposits, colored in the same color scheme as Figure 5.15 and (b) an even closer view of the second distinct deltaic deposit, which spreads laterally to the east and appears to have inverted distributary channels.



a.



b.

Figure 5.18 (a) 5-meter HIRISE contours of the entrance canyon deposits are shown, as is the location of the longitudinal profile (A is upslope of A') (b) A longitudinal profile (black line) and slope profile (pink line) of the entrance canyon deposits. Elevation data were extracted from the HIRISE DEM and plotted as a function of distance on the left y-axis. Slope was calculated every 100 meters and the 10-point average is plotted as a function of distance. A yellow arrow is pointing out the -3980 meter delta front within the longitudinal profile and slope data.

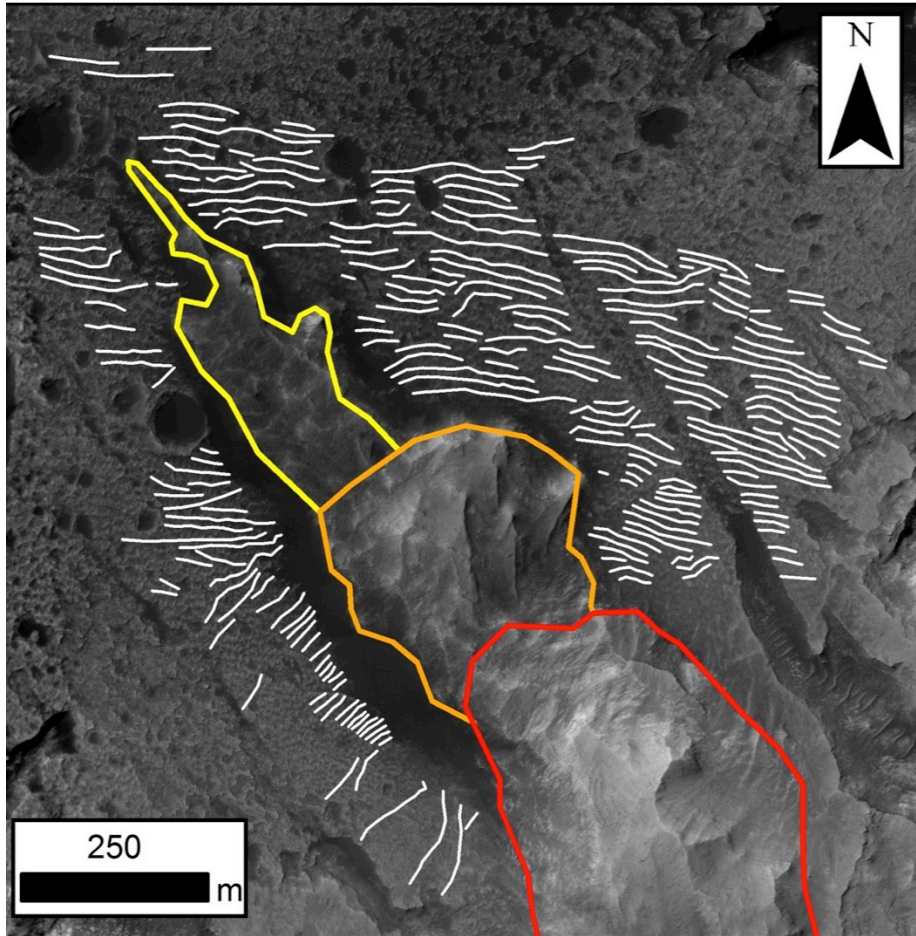


Figure 5.19 Well-defined, regularly spaced bedforms (spacing of approximately 20 meters) are located near the entrance canyon deposits on a broad, nearly uniform sloping plain.

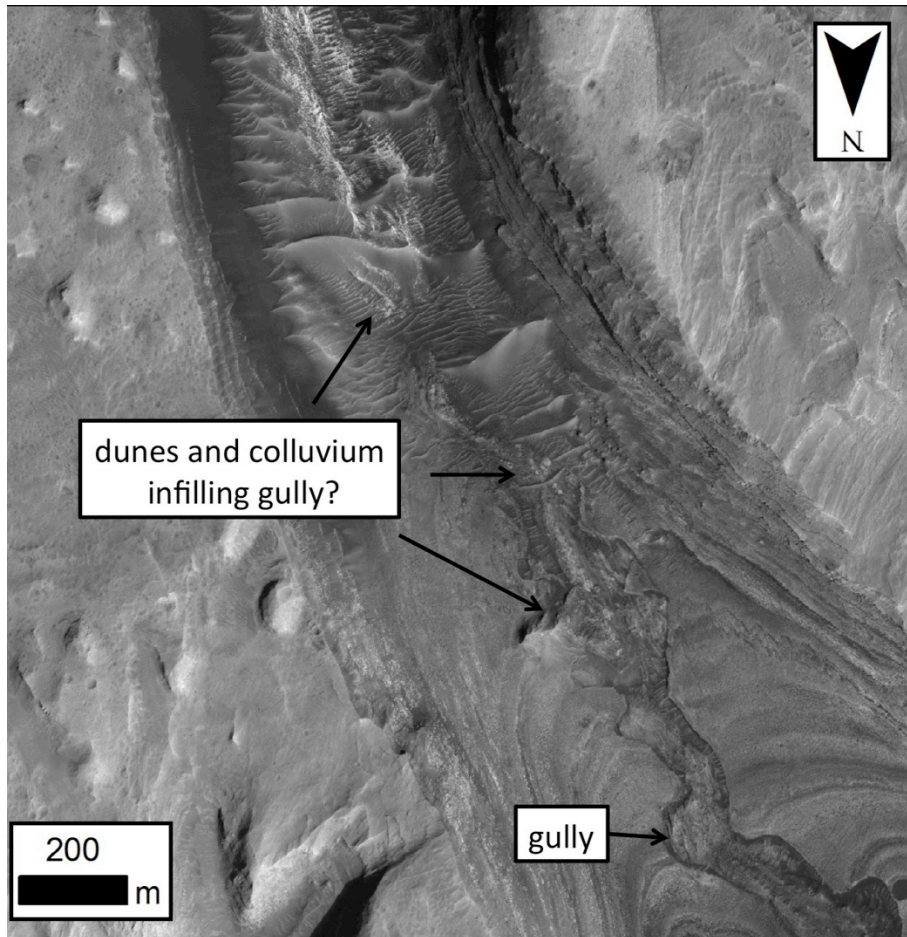


Figure 5.20 Evidence that the feeder gully was perhaps once larger in extent or depth and has been filled in since it was last active is the presence of large boulders, colluvium, and dunes that are obscuring the head of the gully.

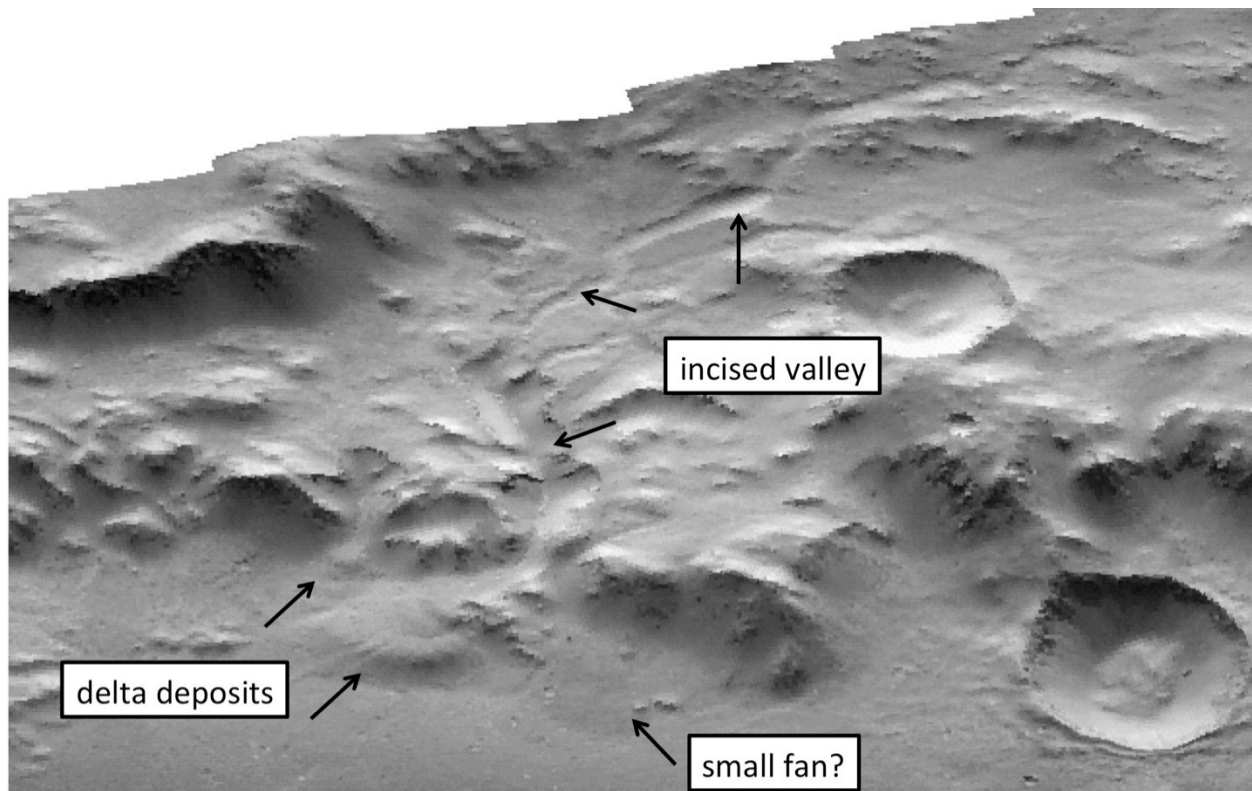


Figure 5.21 A three-dimensional view of the west canyon delta system (vertical exaggeration of 1.5) that is located on the northwestern rim of Gale crater (overlain with a black and white orthorectified 6-meter CTX image, G05_020265_1746_XI_05S223W). In addition to the delta deposits, there is a small depositional feature to their north that may be a small fan deposit.

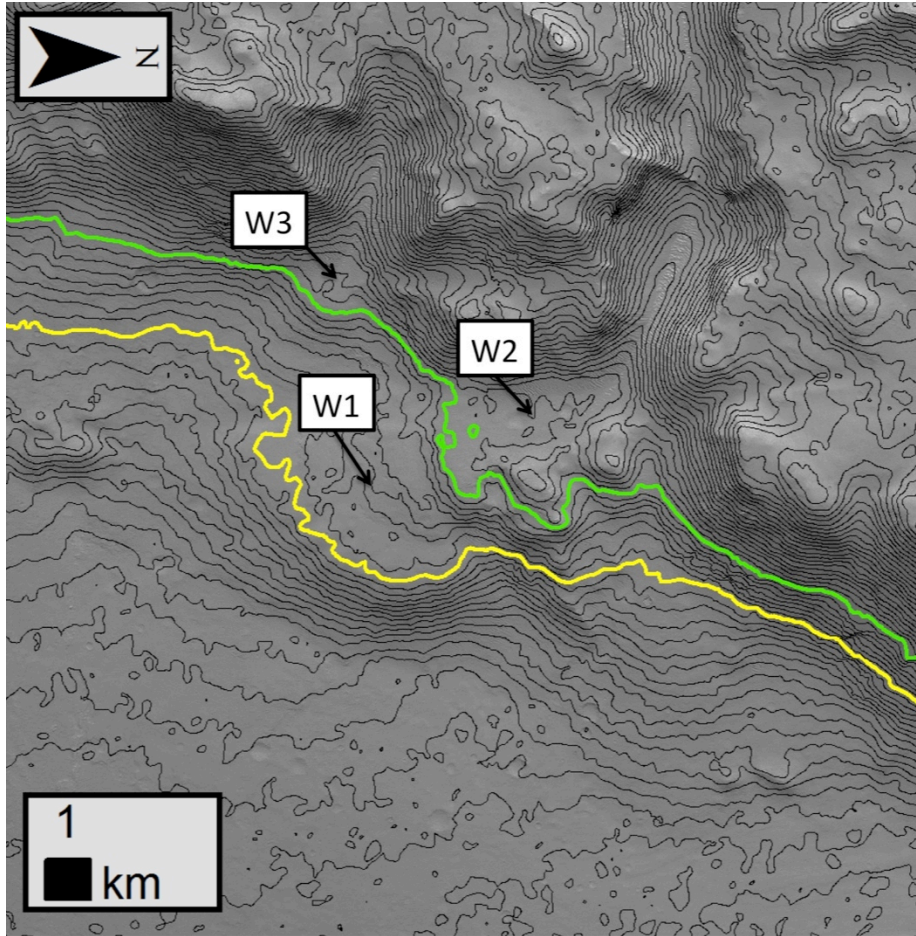
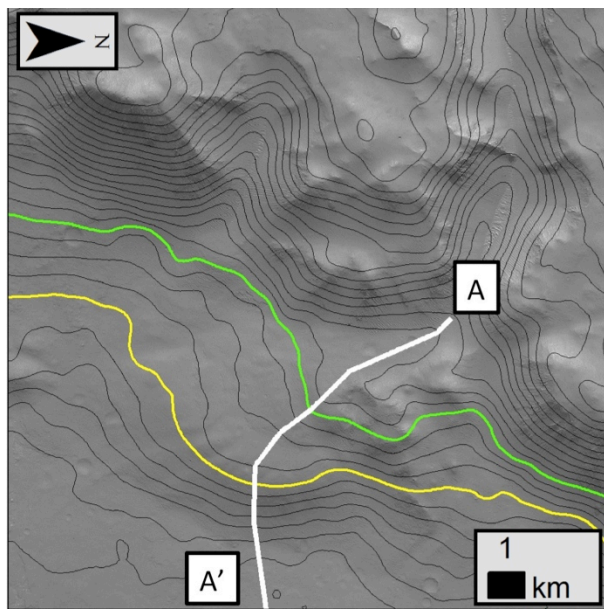
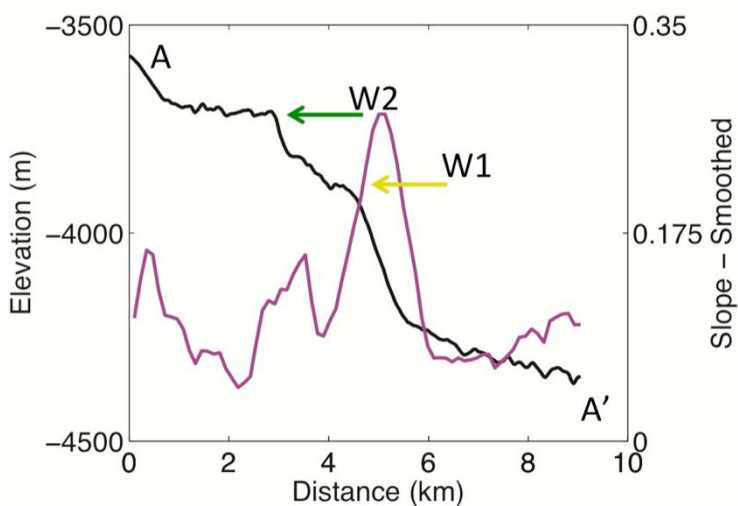


Figure 5.22 The lower more prominent deposit (W1) has a front that corresponds to the -3980 meter HRSC contour (yellow contour). A second, smaller delta deposit (W2) is overlaying it with a “stair-step morphology” and its front corresponds to the -3780 meter HRSC contour (green contour). To the south is a second, smaller deposit (W3) whose front also corresponds to the -3780 meter HRSC contour (green contour).

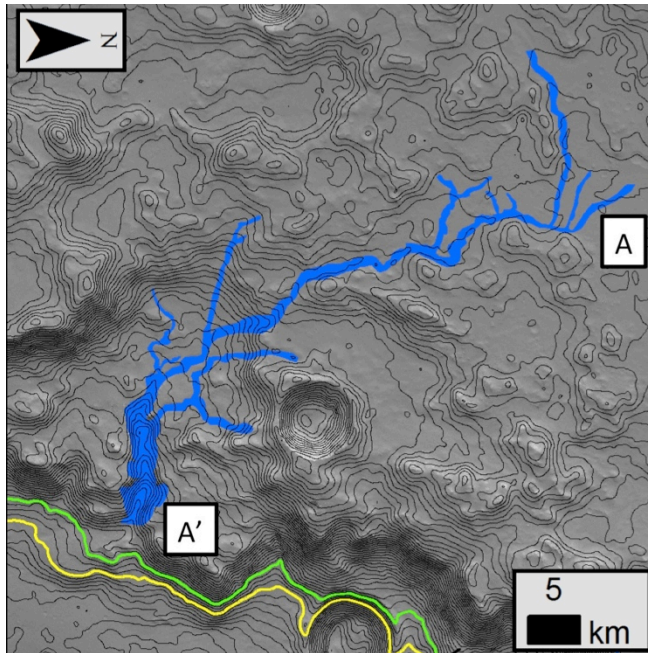


a.

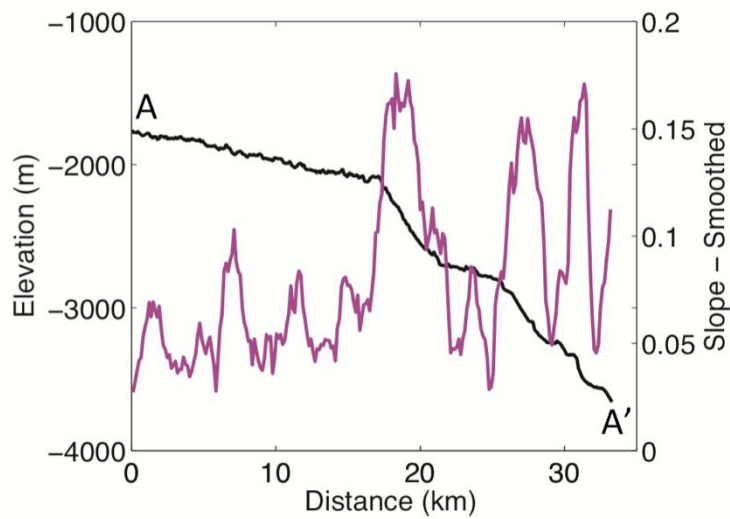


b.

Figure 5.23 (a) 25-meter CTX contours of the west canyon deposits are shown, as is the location of the longitudinal profile (A is upslope of A') (b) A longitudinal profile (black line) and slope profile (pink line) of the west canyon deposits. Elevation data were extracted from the CTX DEM and plotted as a function of distance on the left y-axis. Slope was calculated every 100 meters and the 10-point average is plotted as a function of distance. A yellow arrow is pointing out the -3980 meter delta front and a green arrow is pointing out the -3780 meter delta front within the longitudinal profile and slope data.



a.



b.

Figure 5.24 (a) 25-meter CTX contours of the west canyon incised valley is shown, as is the location of the longitudinal profile (A is upslope of A') (b) A longitudinal profile (black line) and slope profile (pink line) of the west canyon incised valley. Elevation data were extracted from the CTX DEM and plotted as a function of distance on the left y-axis. Slope was calculated every 100 meters and the 10-point average is plotted as a function of distance. The west canyon incised valley is located within a 450 km² drainage area.

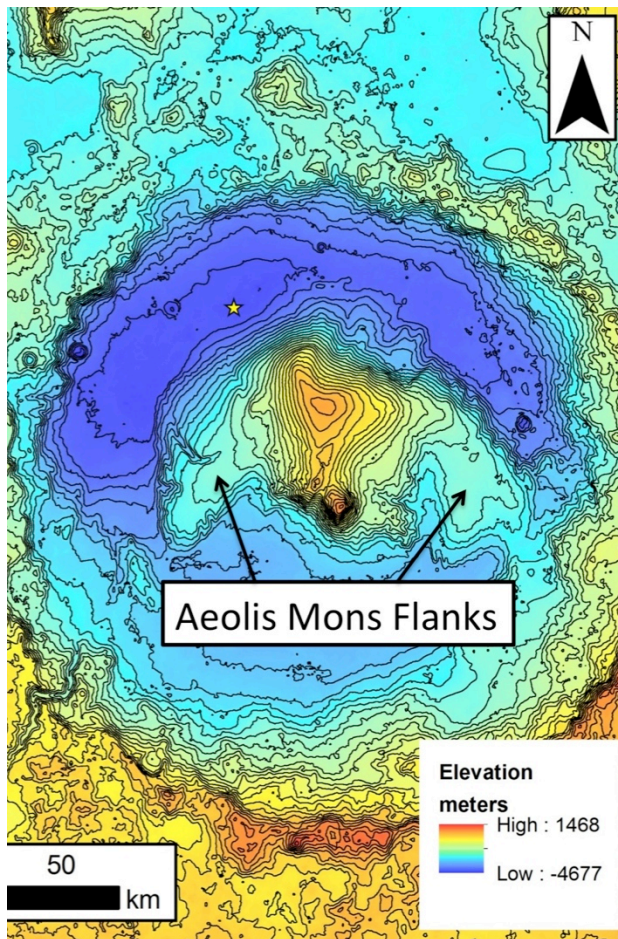


Figure 5.25 An HRSC-derived elevation map of Gale crater with 200 meter HRSC contours. The star marks the location of *Curiosity*'s landing (i.e. Bradbury Rise). The elevation of the eastern and western "flanks" of Aeolis Mons matches that of the northern crater rim, leading to the suggestion that the flanks were wave-cut when Gale was connected to a northern ocean.

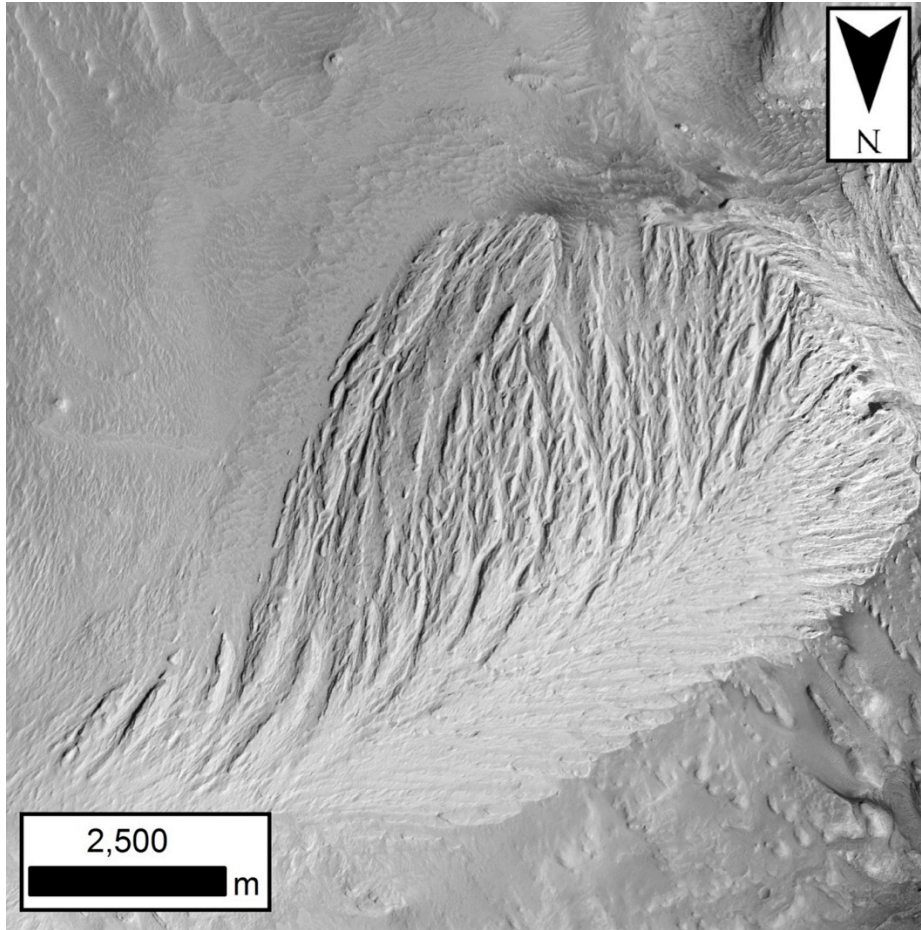


Figure 5.26 A 9 km long and ~1 km thick exposure of inclined layers, located just below the flanks pointed out in Figure 5.25, may record a delta. This interpretation would require a lake level well-above the northern rim of Gale crater.

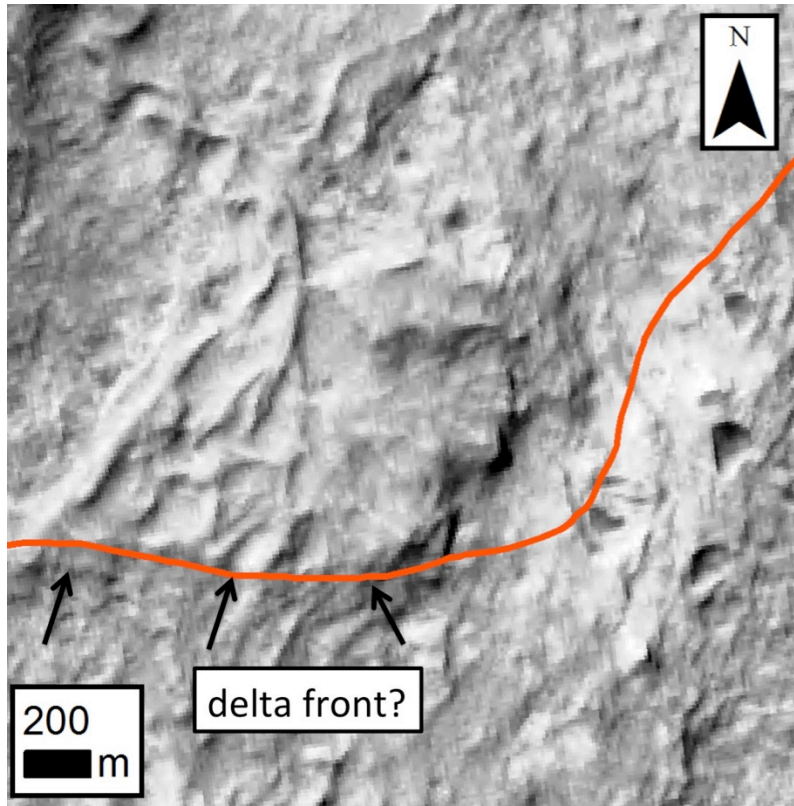


Figure 5.27 CTX image of a small delta-like feature that lies to the east of the Farah Vallis delta-fan system and emanates from a large canyon on the south side of Aeolis Mons. The apparent front of the delta roughly corresponds to the -3280 meter HRSC contour line (orange), which is the same contour that defines the Pancake delta front. The delta has what may be inverted distributary channels.

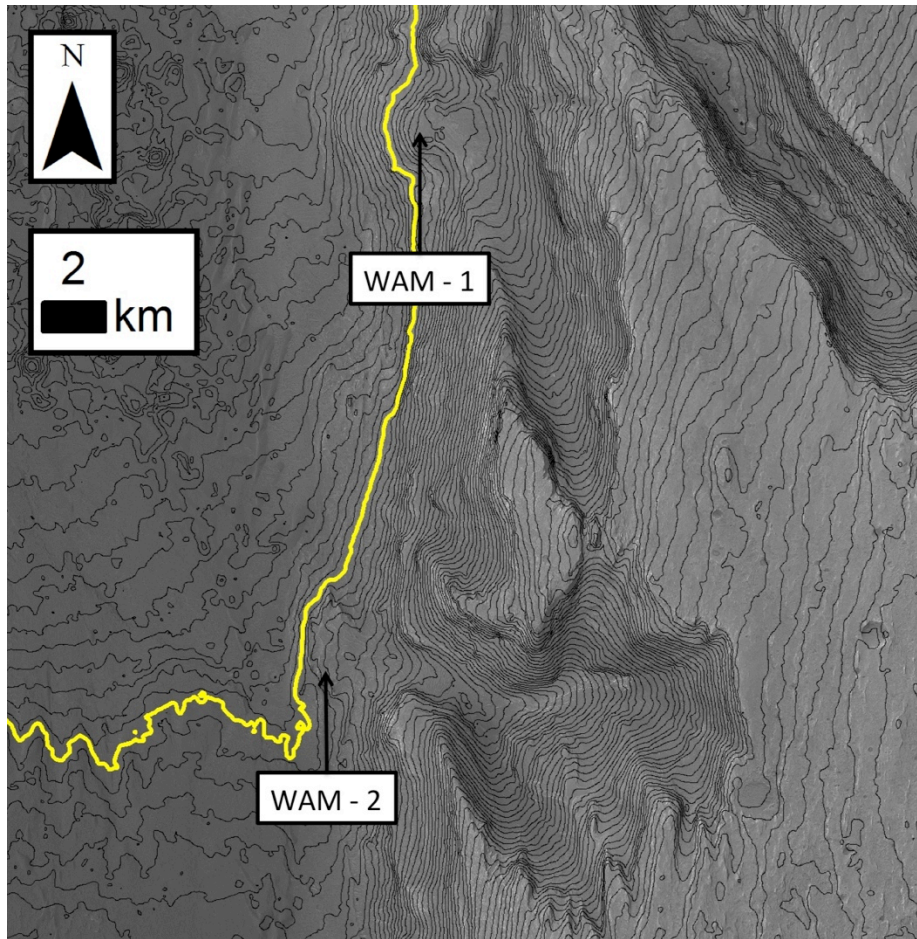


Figure 5.28 Along the western margin of Aeolis Mons, small ($< 3 \text{ km}^2$) complex fan-delta shaped deposits lie at the mouth of two of the larger canyons that exit to the Gale floor. These deposits may be small deltas recording fluvial activity in the upslope canyons, but there is only a minor amount of sediment delivery relative to the scale of the canyons. The -3980 m HRSC contour (Western lake, yellow) that defines the lower Western Delta (W1) crosses both these small deltas at approximately the elevation where downslope steepening occurs.

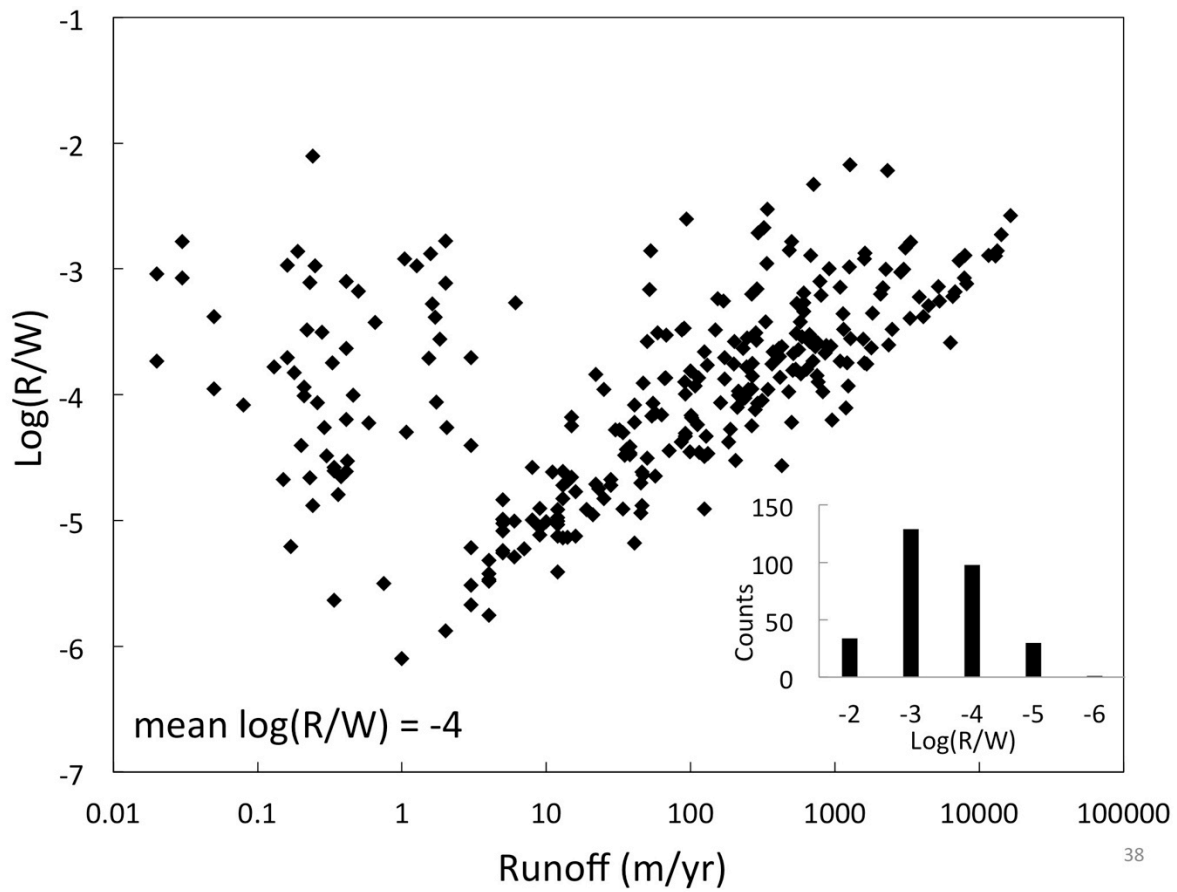


Figure 5.29 A survey of terrestrial rivers by Syvitski et al. [2003], who collected runoff and sediment flux data for 340 river basins around the world, shows that the log transformed rock/water ratio yields a roughly Gaussian curve (inset plot) with a mean value of 0.0001.

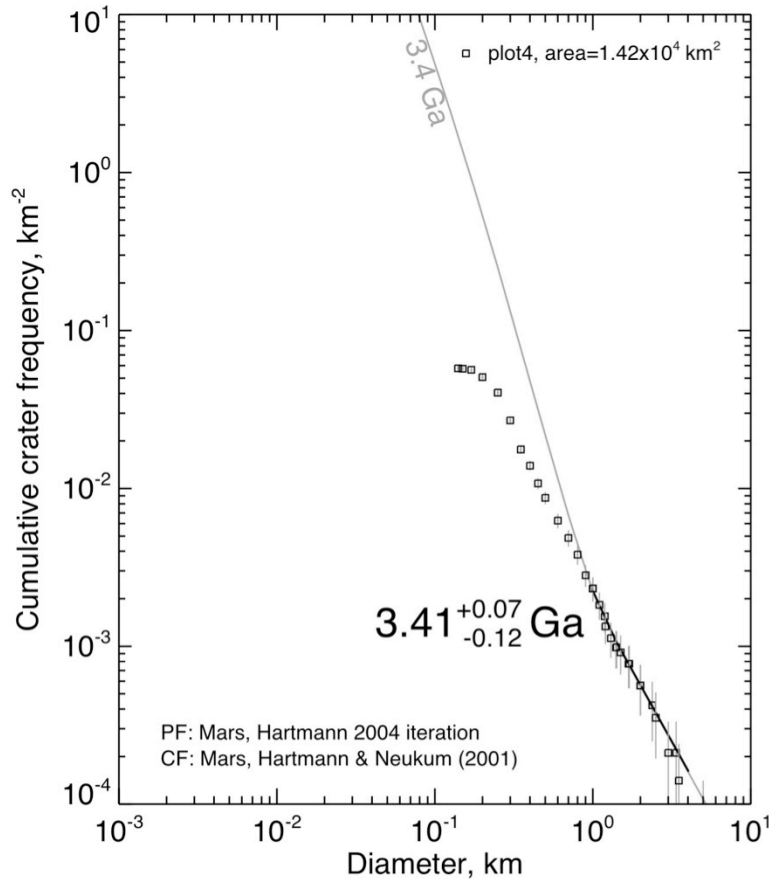


Figure 5.30 Crater count data for the wall and floor for all of Gale (area $\sim 14,000 \text{ km}^2$) are shown on a cumulative crater frequency plot. All mapped crater sizes are plotted using root-2 binning and there have been no re-surfacing corrections. The production function (PF) is Mars Hartmann 2004 and the cratering function (CF) is Hartmann and Neukum (2001). Craters ranging from 1 to 4 kilometers in diameter fell approximately on the 3.4 Ga isochron.

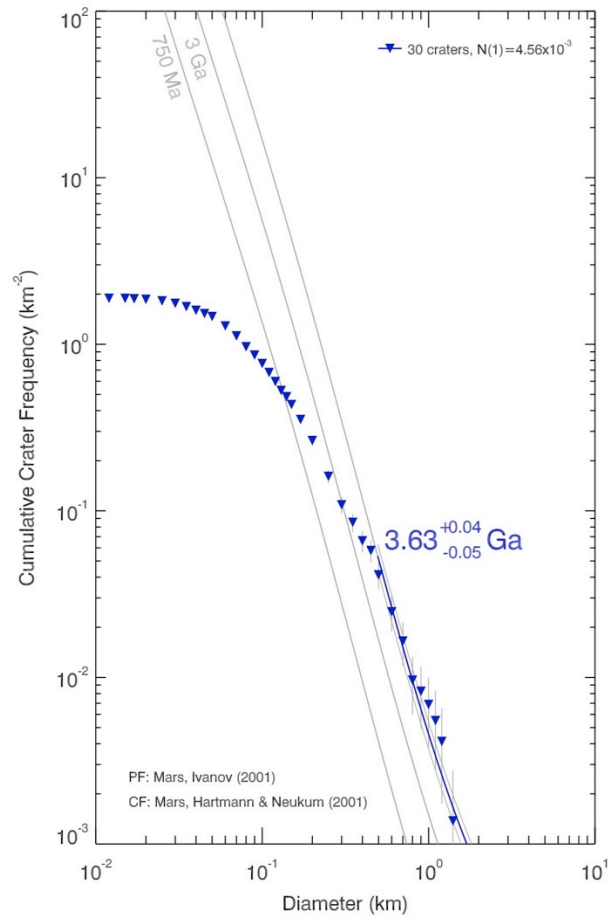


Figure 5.31 Crater count data for the Peace Vallis catchment (area $\sim 730 \text{ km}^2$) are shown on a cumulative crater frequency plot. All mapped crater sizes are plotted using root-2 binning and there have been no re-surfacing corrections. The production function (PF) is Mars Hartmann 2004 and the cratering function (CF) is Hartmann and Neukum (2001).

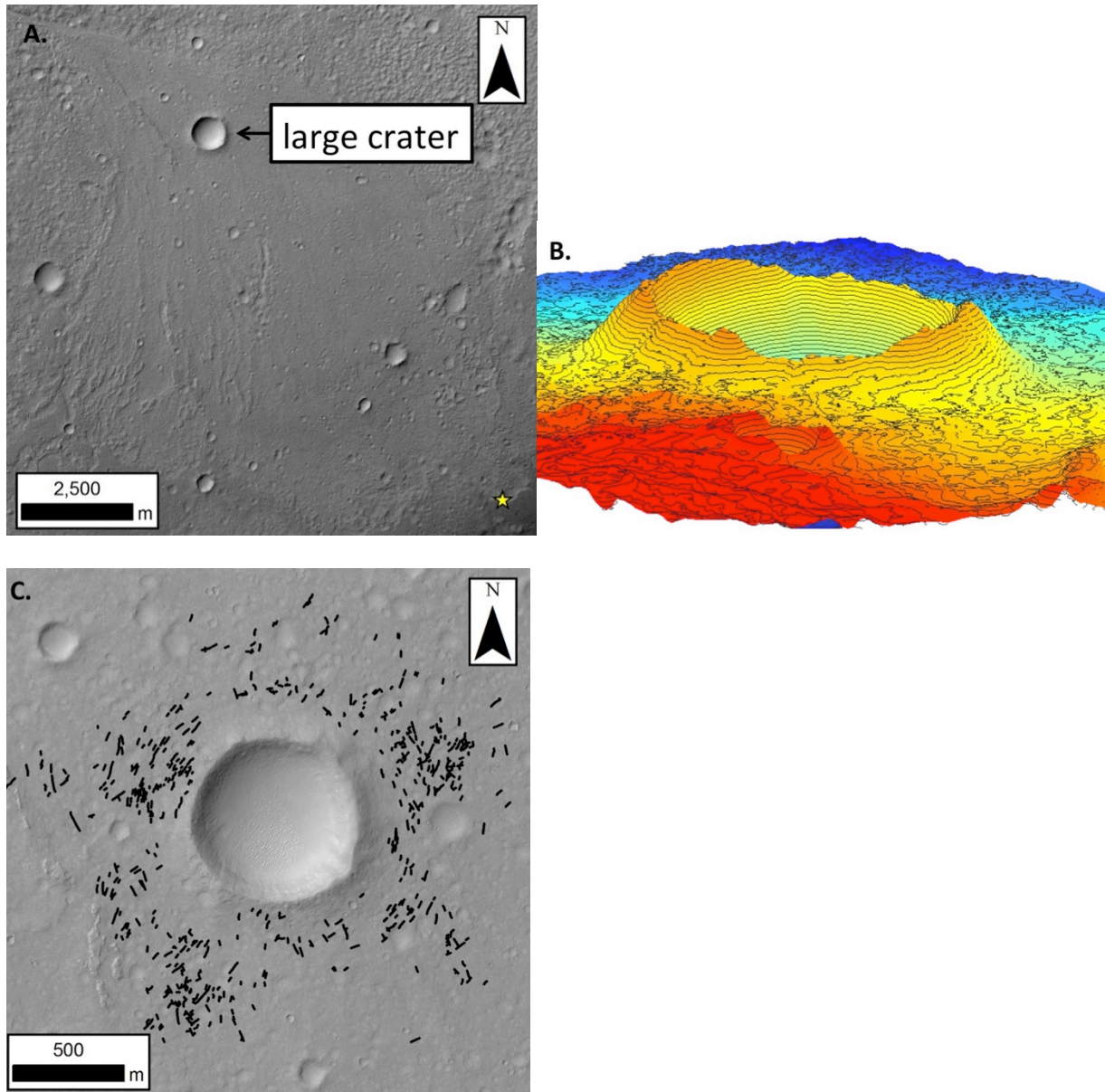
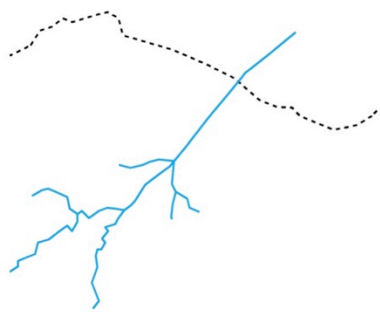


Figure 5.32 (a) CTX imagery (6 m/pixel, image: P22_009571_1756_XI_04S222W) of the Peace Vallis fan. The yellow star indicates the location of the Bradbury landing site. The fan apex is located in the northwest corner of the image and the fan spreads south and southeast. The large 760-meter diameter crater is highlighted. (b) A three-dimensional view looking south of the Peace Vallis fan 760-meter crater (vertical exaggeration of 10) with a colorized elevation map and 10 m HiRISE contours. The crater has a continuous rim and is highest to the east and south. (left and top of image respectively). (c) Mapping of concentric fractures around the Peace Vallis fan crater indicates that the crater post-dates fan formation.



a.



b.



c.



d.

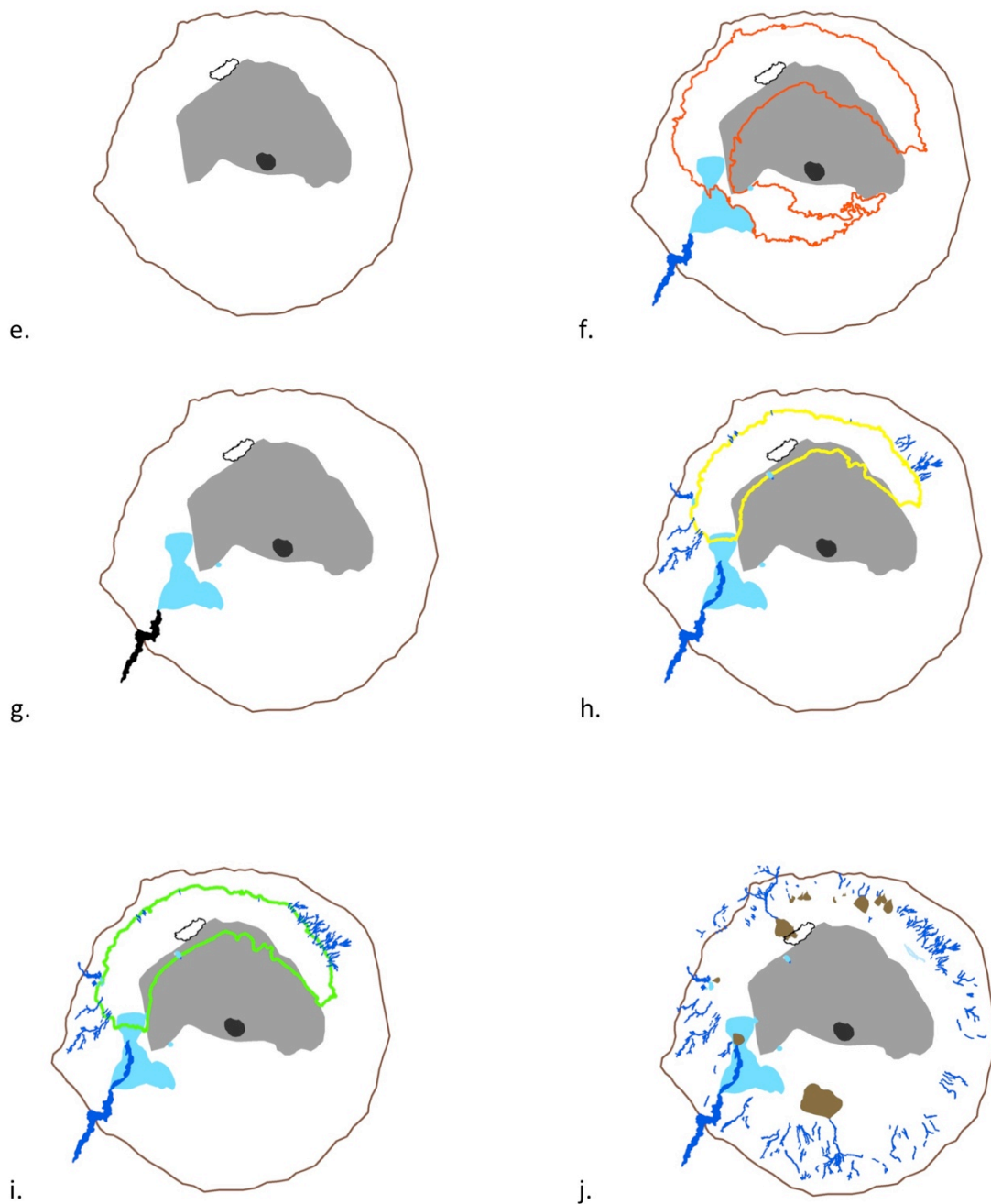


Figure 5.33 A cartoon representation of a plausible hydrologic sequence within Gale crater since its impact: (a) development of an incised valley network that drained the Noachian terrain near Hershel crater (b) Gale's crater-forming impact occurred during the late Noachian / early Hesperian (c) Gale crater's ejecta blanket (outer brown line) disrupted the valley network to its south; prior to Aeolis Mons development Gale likely had a central peak (black dot) (d) growth of Aeolis Mons (grey), perhaps due to infilling and subsequent excavation of Gale crater [Malin and Edgett, 2000] or from slope enhanced aeolian transport and deposition [Kite et al, 2013] (e) perhaps coeval with Aeolis Mons' growth is the deposition of sediment from the northern crater

rim into Gale, forming the 15 km setback in the rim into which Peace Vallis eventually formed (and may have resulted in the complex topographic boundary conditions that the Peace Vallis fan encountered at its distal end, including a 64 km² basin [Figure 4.6, Chapter 4]) (f) incision of Farah Vallis due to renewed runoff from the south and the subsequent deposition of the Pancake delta and the plains to its west, which record one of the largest lakes to have occupied Gale (-3280 meter HRSC contour) (g) decreasing runoff resulted in the decline of the -3280 meter lake; perhaps completely, but at least to a level lower than the entrance canyon deposits (h) renewed runoff from Farah Vallis, combined with local precipitation that drove incision of the crater rim and Aeolis Mons, lead to a second large lake in Gale, which had a still stand at -3980 meters (i) further runoff, as evidenced by the incision of more gullies around Gale's rim and the formation of the Farah Vallis delta front, lead to the second highest stand in Gale at -3780 meters (j) the -3780 meter lake rapidly declines and disappears, based on the lack of gullying of the delta deposits, leading to a period of fan (and gully) formation around Gale.

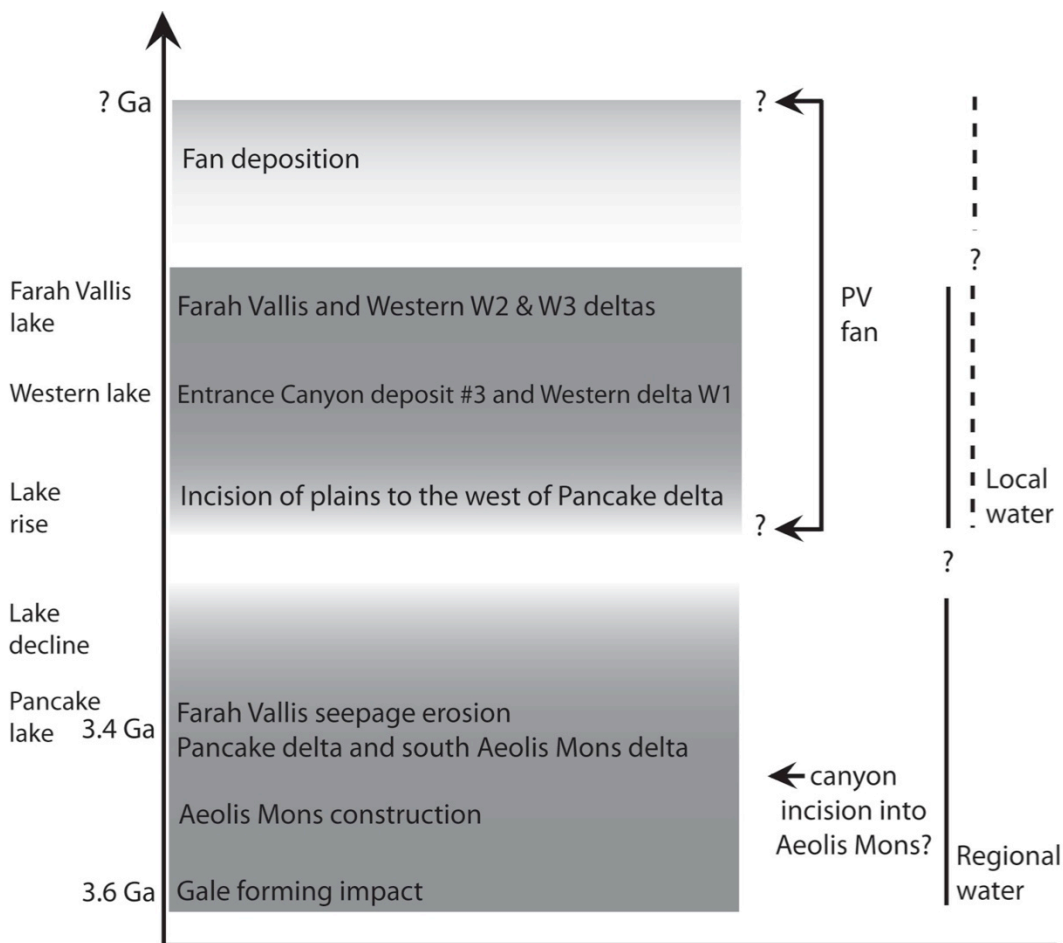


Figure 5.34 A second representation of the sequence and relative timing of large lakes in Gale; The Peace Vallis fan, similarly to the Farah Vallis fan, likely formed post-lake, but the possibility remains that it formed between the two lakes, hence its location on the timeline could be inserted in two places (denoted with question marks). The likely water sources (distal via Farah Vallis versus local via precipitation) and amounts of water (dark grey shading corresponds to a wet Gale, while light grey to white indicates drier periods) are also indicated.

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Chapter 6: How small is too small? Assessing the uncertainties in crater retention ages for small surface areas on Mars

6.1 Abstract

As there becomes increasingly more coverage of Mars' surface with high-resolution imagery, smaller features like alluvial fans and deltas (typically $<500 \text{ km}^2$) are being discovered that may provide clues to the timing of hydrologic events over the planet's history. An obvious question to ask is: can crater counting, which has been used to date much of Mars' surface, be used on such small surfaces and still provide accurate age dating? Here we introduce a simple simulation model to quantify the effects of sample area size and crater obliteration effects on age estimates derived from crater size-frequency distributions. Our results show, quantitatively, that crater counting to estimate age of small surfaces ($<1000 \text{ km}^2$) has unavoidable large uncertainties. This arises because the small craters are eliminated due to erosion and infilling and there is a low probability of counting large craters on small sample areas. These effects on both the small and large craters leads to a narrow range of crater sizes in which the correct age may be reflected in the crater density functions that are used to date surfaces. This range decreases with decreasing surface area and increasing erosion and infilling. Steps in the isochron data, in which crater density data are shifted downward for some crater sizes, are often cited as indicators of resurfacing events. We find, however, that such steps occurred randomly in $\sim 5\%$ of the crater size-frequency distributions we generated. This can lead to large errors when determining which tangential isochron is used to assign a surface age. Our modeling, which accounts for obliteration of craters, suggests that in general the least reliable ages occur between 1 and 3 Ga years. Younger surfaces preserve the smaller craters and older surfaces collect larger ones, both improving the probability of obtaining the correct age. For areas less than 1000 km^2 and true ages of 1 to 3 Ga, there is only a 20 to 40% probability of the calculated age lying within 0.25 Ga of the true value, where as for surfaces smaller than 500 km^2 the corresponding probabilities drop to about 10 to 20%. To highlight the constraints on dating small deposits, we applied this model to cratering data from the Peace Vallis fan ($\sim 80 \text{ km}^2$) and one of the larger deltaic deposits within Gale crater (117 km^2), which were deposited at different elevations and likely record different hydrologic events. Even though portions of the observed crater count data in each case follow the proposed crater density isochrones, the uncertainties for such small areas are so large as to make the age assignment unreliable.

6.2 Introduction

For the past several decades, methods for determining absolute ages of planetary surfaces using craters have been developed and refined [e.g. Hartmann et al., 1981; Hartmann and Neukum, 2001; Ivanov, 2001; Hartmann, 2005], including using craters down to several meters in size to determine the current cratering rate [Daubar et al., 2013]. The relative abundances of craters can provide a minimum age for major geologic units, assuming a spatially random impact flux, a calibrated size distribution of impactors, and a randomized flux over short timescales and a "known" variation over longer timescales (the theoretical model and mathematical framework for surface age-dating is summarized by the Crater Analysis Techniques Working Group [1979]). Isochrons, which are defined as the crater size-frequency distribution for a surface of a specified age, assuming the absence of erosional processes or obliteration, were initially derived from

crater size-frequency distributions on the lunar maria that were calibrated with radiometric dating of lunar rock samples before being converted for use on the Martian surface, where differences in gravity and impact velocity [e.g. Ivanov, 2001] and atmospheric corrections [Hartmann, 2005] are taken into account. The result of decades of work is a set of cratering chronology models that can be used to obtain absolute ages for Martian landscapes [e.g. Hartmann and Neukum, 2001; Ivanov, 2001; Neukum et al., 2001; Hartmann, 2005 and previous iterations]. As our sample of surface dates using cosmogenic dating on Mars is currently small and limited to Gale Crater [Farley et al., 2014], these cratering models combined with superposition relations provide the only estimates of age needed for the interpreting the geologic and thus climatic history of Mars [e.g. Tanaka et al., 2014].

Due to volcanic and geomorphic processes occurring over time, erosional or resurfacing events will change the crater population, often preferentially removing craters from the small-diameter end of the distribution [Chapman et al., 1969; Hartmann, 1971; Smith et al., 2008; Michael and Neukum, 2010]. Michael and Neukum [2010] describe a procedure for deriving surface ages for regions that have experienced partial resurfacing events that have preferentially removed a limited range of crater sizes (typically the smaller crater sizes). They note that the shape of the fall-off curve can vary with time and resurfacing process, making it difficult to discern the crater diameter range associated with the re-surfacing events versus roll-off due to image resolution and/or erosional effects. Michael and Neukum [2010] also state that the deviation of the crater data above the “best-fit” isochron for larger diameters is an indication of clustering in the area. Palucis and Dietrich [2014] show, however, that for small areas, the presence of just one crater can cause this deviation as well. Combined, these small and coarse tail effects can cause large errors when attempting to date smaller surface areas, as the smallest craters are more likely to be influenced by geomorphic and volcanic processes and are susceptible to image resolution limitations, while the presence of just a few large craters can lead to erroneously old ages. As such, the standard practice [e.g. Platz and Michael, 2012] for determining a surface age from cratering data has been to basically ignore smaller craters that deviate systematically from the Hartmann curves due to obliteration effects and to ignore the few big craters due to poor statistics. From the remaining crater size bins, one then looks for several bins falling on a single Hartmann isochron. That isochron is then used to assign an age.

With the increase in high-resolution imagery of the Martian surface due to the High-Resolution Imaging Science Experiment (HIRISE) and the Context Camera (CTX), smaller scale features such as alluvial fans [e.g. Moore and Howard, 2005; Kraal et al., 2008a; Anderson and Bell, 2010; Grant and Wilson, 2011; Morgan et al., 2014], deltas [e.g. Kraal et al., 2008b; Di Achille and Hynes, 2010; Dibiase et al., 2013; Dietrich et al., 2014] and young valley networks [e.g. Fasset and Head, 2006; Fasset and Head, 2008; Dickson et al., 2009] are being discovered, and a natural question that arises is when did these features form, which has major implications for understanding the hydrologic and climatic history of Mars. In a study of alluvial fans in the southern Margaritifer Terra region of Mars, Grant and Wilson [2011] looked at eight fan-hosting craters, all of which had well-developed alcoves, incised walls and etched deposits inferred to be distal alluvial or lacustrine in origin. The fan deposits were typically 1000 km² in size and over 100 craters were counted, though most craters in their counts were smaller than 1 to 2 km in diameter and up to 20 m of erosion was known to have occurred based on the height of inverted distributary channels. Four of the crater fans (E, P, S, R; Figure 2A incremental in Grant and

Wilson [2011]) had craters greater than 1 km, which fell on isochrons dating to the mid-Hesperian epoch or earlier, while crater frequencies for all of the fans (except E) for crater diameters ranging from ~250 m to ~800 m fell within the Amazonian epoch. For these relatively small features, does the presence of these larger craters retain any information about the surfaces' true age or were they simply rare events? Secondly, is there a range of crater sizes that is most likely to preserve the actual age of the surface or can the natural variability in the distribution of craters hitting the surface, combined with losses due to erosion and infilling, affect the crater size-frequency for a larger range of craters than might be expected based on erosion depths from inverted relief or crater depth-diameter relations alone?

To address these questions, and to test and quantify the standard assumptions made when assigning an age to a surface using cratering data, we introduce here a simple simulation model to quantify the effects of sample area size and crater obliteration effects on age estimates derived from crater size-frequency distributions. Michael and Neukum [2010] argue that the area over which craters are counted is not important, but rather, the number of craters counted. Their Poisson analysis suggests that for very young and very old surfaces, relatively few craters need to be counted (order of 10 craters) to achieve a narrow probability curve centered on the true age of the surface. For moderately aged surfaces (i.e. 2 to 3 Ga), closer to 100 craters need to be counted to get approximately the same narrow distribution of ages around the true surface age. The difficulty in using this model is that one needs to know the age of the surface *a priori* to know the appropriate number of craters to count and there is no indication of which crater-size class is best to count. In our model, we specifically focused on determining the errors in ages derived from areas ranging from 10,000 km² down to 100 km². We first address the case in which we have an stable cratered surface, such that all volcanic and geomorphic processes are neglected, before considering how low to moderate long-term rates of erosion and crater infilling, based on the model of Smith et al. [2008], affects surface age uncertainty (i.e. our obliterated crater surface). From this, we present a methodology for determining the level of confidence one has in assigning a crater retention age to a given surface based on its area. This work is partially based partly on the growing interest in dating small-scale features such as alluvial fans and deltas [e.g. Grant et al., 2014], and as such, we explore the application of this methodology using depositional features (i.e. an alluvial fan and a deltaic deposit) within Gale Crater (5.4° S 137.8° E), the landing site of the Mars Science Laboratory *Curiosity* rover.

6.3 Methodology

6.3.1 Model Description

In both our stable and obliterated cratered surfaces version of the model, 200,000 km² surfaces (referred to as ideal surfaces) were generated with crater populations ranging between 60 m and 64 km as defined by the Hartmann isochrons for Mars for 3.5 Ga (Hesperian) down to 0.5 Ga (Late Amazonian) (2004 iteration, Hartmann [2005]). The center coordinates of each crater placed on the surface were determined using a random number generator, with the only constraints being that the craters must not extend past the 200,000 km² area and that all of the craters for a given age, as defined by Hartmann [2005], be present (i.e. craters cannot obliterate one another). For example, for a 3.5 Ga surface that is 200,000 km², the Hartmann isochron dictates that there be 407 craters that are 1 km in diameter and 4 craters that are 10 km in

diameter, and as such, if a 10 km diameter crater is placed “over” a 1 km crater, the 1 km diameter crater must remain to preserve the Hartmann size-frequency relationship. The model is not a landscape evolution model, and as such it does not evolve over time or take into account regional topography. This model also does not try to reproduce any true physical processes, like fragmentation mechanics to generate secondaries, rather it simply assumes that the isochrons as defined by Hartmann [2005] are the theoretical size-frequency distributions for well-preserved surfaces of a known age.

These ideal cratered surfaces are then randomly subsampled over smaller and smaller areas (10,000 km² down to 100 km²), the number of subsamples, n , chosen such that 200,000 km² was sampled (in total) for the smallest sub-area (i.e. $n=2000$ for 100 km²). The x and y coordinates for the upper left-hand corner of the subsampled area was chosen using a random number generator, again the only condition being that the entire subsample area must fall within the 200,000 km² area. The crater populations contained within each sampled area were tallied (total number of craters with a given diameter divided by the sub-sampled area) and the results were compared to Hartmann’s 0.1 to 4.0 Ga isochrons. Normalized root mean square error (RMSE) was used to determine the best fit between the subsampled data and the known isochrons (at 0.1 Ga intervals) and a sample age was assigned. Depending on the experiment, all crater bin sizes were used, a sub-set of crater bin sizes were used, or individual crater size bins were used to determine the best-fit age.

For the obliterated cratered surface model, the obliteration model of Smith et al. [2008] was used, which accounts for erosion and infilling. The use of this model allows for the modification of the Hartman [2005] isochrons in a realistic way (i.e. creating the pronounced crater roll-off for smaller diameter craters). Smith et al. [2008] demonstrated the ability of the model to reproduce roll-off curves for crater size-frequency data at Gusev Crater, Meridiani Planum, and a landslide deposit and the floor of Coprates Chasma. The effect of this model on the crater diameter distributions (i.e. number of craters/area versus crater diameter) is shown in Figure 6.1 for a combined rate of erosion and infilling of 100 nm per year on the 3.5 Hartmann isochron. The Smith et al. [2008] model is based on an equation typically used for modeling populations of radionuclides produced by cosmic-ray exposure

$$N(t,d) = \frac{p(d)}{\lambda(d)} (1 - e^{-\lambda(d)t}) \quad (6.1)$$

where N is the abundance of the measured craters at time, t , p is the crater production rate (number of craters per year) and λ is the crater-loss function (1/year), which is itself a function of β , the combined rate of erosion and infilling (nm/year). N , p , and λ are all functions of crater diameter, d . The production function, $p(d)$, is defined as

$$p(d) = 0.029 * \left\{ \begin{array}{l} 0.0035(0.13 \ln(d) + 0.83) / d^{3.3} \\ 10^{-1.81 \log(d) - 2.59} \\ 10^{-2.2 \log(d) - 1.89} \end{array} \right\} \quad (6.2)$$

where the first expression in Equation 6.2 is for crater diameters between 0.001 and 1.4 km, the second is for craters greater than 1.4 and less than 48.1 km, and the third is for craters greater than 48.1 km. The expression for λ is the ratio between b and the depth-diameter relationship for primary craters (ξ),

$$\begin{aligned}\lambda &= \frac{\beta}{1000\xi_x}; \\ \xi_1 &= 0.2 * d \\ \xi_2 &= 0.42 * \ln(d) - 0.01\end{aligned}\tag{6.3}$$

where ξ_x is based on a best-fit function of crater depths for fresh primary craters from satellite images by Pike [1980]. ξ_1 is for craters less than 5.8 km in diameter and ξ_2 is for craters greater or equal to 5.8 km in diameter. There is also a correction factor for incorporating secondaries, based on the modeling work of McEwan et al. [2005], though Smith et al. [2008] acknowledge that the influence of secondary cratering is controversial, especially its use with the Hartmann production function, hence Smith et al. [2008] acknowledge that their model results should be taken as lower limits for a given β (i.e. there may actually be more small craters on a given landform due to spatial heterogeneity of secondary production and greater secondary production on Mars compared to the Moon).

For the purpose of this paper, three erosion rates were chosen, with β equaling 0 nm per year (no erosion), 10 nm per year and 100 nm per year. According to Smith et al. [2008], erosion rates an order of magnitude larger (i.e. $\beta = 1000$ nm per year) were likely to have occurred mainly in the Noachian (>3.7 Ga [Hartmann and Neukum, 2001]), though isotopic data from the MSL Curiosity rover suggest that scarp retreat rates of mudstones at the distal end of the Peace Vallis fan over the past ~ 80 Ma averaged 750 nm/year [Farley et al., 2014]. Global erosion rates across Mars, however, likely vary greatly depending on location and lithology. Smith et al. [2008] calculated erosion rates in Gusev Crater to be 4.7 nm per year using their model, while ground-based observations of b ranged from 0.08 to 5 nm per year [Grant et al., 2006]. Meridiani Planum, likely owing to its sulfate-rich composition [Arvidson et al., 2006] lead to observed β values of 13.5 to 108 nm per year [Grant et al., 2006]. The same procedure for assigning ages to individual subsampled areas was used for both the stable and obliterated cratered surface scenarios.

6.3.2 Crater Counting at Gale Crater

Cratering maps were generated of Gale Crater's entire wall [Chapter 5] by mapping onto CTX imagery in ArcGIS using CraterTools software [Kneissl et al., 2011]. Some secondary craters may have been included in the counts, but careful attention was made to avoid including obvious rays or clusters of craters. In general, the resolution of the CTX data enabled craters down to ~ 12 m in diameter to be mapped. The crater area, and hence crater diameter, was then extracted for various depositional features around Gale Crater (i.e. alluvial fans and deltas), and from these maps crater frequency plots were compiled in Craterstats II [Michael and Neukum, 2010] using pseudo-log bins (with no re-surfacing correction). These data were compared to our modeled data as discussed in Section 6.5.

6.4 Model Results

6.4.1 Stable Cratered Surfaces

In the first set of model results, the effects of erosion and infilling were not taken into account (i.e. $\beta = 0$), such that each subsampled area had a random distribution of crater sizes derived from an unmodified Hartmann isochron. Figures 6.2a,b and 6.3a,b compare random subsampling for two different areas (i.e. 10,000 km² (a's) and 100 km² (b's)) for two surfaces of known ages, 3.5 Ga (Figure 6.2) and 1 Ga (Figure 6.3). The open circles represent the average crater frequency per area for a given crater diameter for 2000 subsamples. In both age and sample area cases, these average points most closely follow the expected curves for the smaller, and more numerous, craters. Detectable deviation from the expected curves occurred for the largest crater sizes counted even for the large subsample area (10,000 km²). The smaller subsurface area (100 km²) and younger surfaces (1 Ga) had smaller average maximum crater diameters. Because the model does not count fractions of craters within a given subsampled area, the crater frequency data for the larger crater sizes becomes strongly modal. For example, for 8 km diameter craters, when the subsample area is 100 km², either 0 or 1 crater was counted such that the crater frequency was either 0 or 0.01. As this distribution is bimodal, the use of Gaussian statistics is not technically valid, so we also report the mode of the crater frequency per area for a given crater diameter (plotted in Figures 6.2 and 6.3 with an 'X'). For both the old (3.5 Ga) and young (1.0 Ga) surface, when the surface area is reduced to 100 km², crater sizes above 0.25 to 0.5 km in diameter are likely more to be missed (based on the mode), effectively making them 'rare events'. We also included examples of typical crater size-frequency data extracted from individual subsampled areas (represented by orange filled triangles and blue filled circles in Figures 6.2 and 6.3). In all examples, in both Figures 2 and 3, since most of the crater size bins counted fall on a single Hartmann isochron (i.e. the dashed isochron that represents the true age of the surface), that isochron would be used to assign an age. In Figure 6.3b, the effect of counting one large crater over a small area is obvious (see arrow) as the crater-frequency data for these craters falls above the 3.5 Ga isochron. Michael and Neukum [2010] attribute this pattern to clustering, but here it is due to a single crater being counted over a small area. The general practice when dating real surfaces is to avoid counting these large deviants since they skew the surface age towards being older [e.g. Michael and Neukum, 2010; Grant et al., 2014]. In general, all of these plots show that without erosional processes, the smallest craters best retain information on the true age of the surface, even with subsampling areas as small as 100 km². When the subsampled surface area is large (i.e. 10,000 km²) the crater frequency for diameters up to 1.4 km for a 3.5 Ga surface and diameters up to 0.5 km for a 1 Ga surface are not greatly affected, though these crater sizes are still relatively small compared to the crater sizes used to date much of Mars [Tanaka et al., 2014].

Figure 6.4 shows an example of the effect that sampling progressively smaller and smaller areas has on the surface age that would be calculated based on a single crater diameter bin (e.g. using the crater frequency for 1 km diameter craters to determine a surface age). While it is not common to base surface ages on one crater size alone, areas with significant erosion and small areas often lead to ages based on limited crater size classes [e.g. Grant and Wilson, 2011]. The true surface age in this case was 3.5 Ga. The ages plotted are the average model-based age

from 2000 subsamples for a given area (10,000 km² down to 100 km²). In some cases, especially for craters larger than several kilometers, a given subsampled surface will not contain any craters within a given size class, resulting in a model age of 0. Those cases were excluded from the averaging. For all surface area considered, there is an initial dip in the data, where the average surface age appears younger than the true age. When larger crater diameters are considered (large being relative to the area over which counting occurred), a single crater typically results in much older surface ages. When the surface area sampled is only 100 km², only the smallest crater bin, 62.5 m, reflects the true surface age.

Figure 6.5 shows the results from a set of experiments in which we took 2000 subsamples of 10,000, 5,000, 1,000, 500, and 100 km² areas over surfaces with ages ranging from 0.5 Ga to 3.5 Ga. The age for each subsampled area was determined using all crater sizes. To check that our model was assigning ages comparable to those one would get using CraterStats [Michael and Neukum, 2010], a standard tool used for determining surface ages from cratering data, we looked at every 100th model output (20 outputs in total), and determined the age using CraterStats. This was done partly to ensure that larger crater sizes were not biasing the results towards being older, as was discussed in earlier in this section and shown in Figure 6.3b. The CraterStats age was compared to the model output age to make sure that the two methods were within 0.2 Ga of each other. The probabilities of determining the true surface age within 0.25 Ga for a given area were then calculated (number of times age is within ± 0.25 Ga of true age / number of subsamples). Figure 6.5 shows curves of constant area where the probability of being ± 0.25 Ga of the true surface age is on the y-axis and true surface exposure age (in Ga) is on the x-axis. In general, the curves have a pronounced parabolic shape, less so for the 500 km² case, where the lowest probabilities are found for surfaces that are closest to 1.5 Ga in age. This parabolic shape with a minimum at 1.5 Ga arises from upper and lower limits on possible crater sizes for very old and young areas.

Figure 6.6 shows an example of the distribution of ages for the case of 10,000 (light) and 1,000 (dark) km² subsampling areas for true surface ages of 0.5 Ga (blue), 1.5 Ga (red), and 3.5 Ga (black). The distribution of ages that result from subsampling the 0.5 Ga surface and 3.5 Ga surface are right-skewed and left-skewed, respectively, as there is a natural limit preventing outcomes below 0 Ga and above 4 Ga. For the case of a 1.5 Ga surface, the distributions of ages is more normal, as the true age of the surface is further away from the natural limits imposed by the range of ages of the Martian surface. In Figure 6.5, when a very old or very young surface is sampled, the probability of determining it is indeed very old or very young is fairly high. For surfaces that are more “middle-aged”, the distribution is broader as some subsamples will contain more large craters per unit area, making those surfaces appear older, while others will have less large craters per unit area, making them appear younger. For subsampled areas that were 100 km² (grey line, Figure 6.5), the probability of determining the true surface age ranged from 0.2 to 0.4 for all ages investigated. This is likely due to only the smallest crater bin size (i.e. 62.5 m) remaining unaffected by the small subsample area.

6.4.2 Obliterated Crater Surfaces

In the second set of model results, the effects of erosion and infilling (using the obliteration model of Smith et al. [2008]) were taken into account by setting β equal to 100

nm/yr. Similar to Figures 6.2 and 6.3, Figures 6.7a,b and 6.8a,b compare random subsampling for two different areas (i.e. 10,000 km² (a's) and 100 km² (b's)) for two surfaces of known ages, 3.5 Ga (Figure 6.7) and 1 Ga (Figure 6.8). The open circles represent the average crater frequency per area for a given crater diameter for 2000 subsamples. For both age and sample area cases, these average points deviate greatly from the expected Hartmann isochrons for the smaller crater bins, which were more affected by erosion and infilling processes. Like the case of stable cratered surfaces, there was also detectable deviation from the expected Hartmann curves for the larger crater size bins as well. When either the age of the surface or the subsample area was decreased, the maximum crater size bin counted decreased. As described for the stable cratered surface case, as the crater sizes become larger, the resulting crater frequency per area distributions become modal (since the model does not count fractions of craters within a subsampling area). We, therefore, have also plotted the mode (X's) for each crater bin size on Figures 7 and 8. In each case the mode and average are very similar for the smaller crater size bins, but the mode goes to 0 more quickly (i.e. you are more often not going to count a crater of a given size) when the age of the surface decreases or the subsurface area decreases. The deviation between crater size-frequency results for individual subsamples are shown in Figure 6.7 and 6.8 by the orange filled triangles and the blue filled circles. In all cases, the crater frequencies for smaller crater sizes fall below the true age of the original surface, denoted by the dashed isochron, while the larger crater sizes, when counted, fall way above the dashed line. In Figures 7b and 8b, when the area is reduced to 100 km², none of the individual data points fall on the true (dashed) isochron. As larger craters are usually ignored, and craters that are “rolling-off” are ignored, for both cases in Figure 6.7a, craters ranging between 0.75 and 3 km would typically be used to fit an isochron. As such, younger ages than the known age of 3.5 Ga would be assigned (3.2 for orange triangles and 3.1 for blue circles).

Figures 6.9 and 6.10 show the range of craters diameters that have a greater than 90% probability of being within ± 0.25 Ga of the true surface age, for a 3.5 Ga and 1.0 Ga surface respectively. These data are only looking at individual crater size bins and comparing the number of craters counted within that size bin per area as a function of erosion. The crater frequency results are then compared to that expected by the 3.5 Ga and 1.0 Ga Hartmann isochrons. While there is a range of crater sizes that preserve information about the true surface age when erosion and infilling is absent, moderate erosion rates between 10 and 100 nm/year effectively lead to a loss of information at all crater sizes likely to be encountered on smaller sample areas. This occurs even for areas as large as 10,000 km². We have not yet investigated at what size one does not need to worry about obliteration effects, but for reference, the recent dating of the ejecta blanket from the Gale-forming impact (the landing site of the *Curiosity* rover) had an area of ~10,000 km² [Le Deit et al., 2013].

As most surfaces on Mars have experienced obliteration processes (i.e. wind, volcanism, the flow of water and ice), we created a plot similar to Figure 6.5 that takes into account erosion. Unlike the case for a non-eroded surface, where the probability of obtaining the true surface age was reasonably high for very old (>0.96 for 3.5 Ga) and very young surfaces (>0.94 for 0.5 Ga) for all areas over 500 km², Figure 6.11 shows that the sampling area has a much greater effect on the probability of determining the correct surface age. For all age surfaces, there is a progressive drop in the probability of determining the true surface age as area decreases (e.g. for a 3.5 Ga

10,000 km² surface, the probability is 0.94 and for a 500 km² surface the probability is 0.42). For 100 km² areas, probabilities range from 0.06 to 0.13.

6.4.3 “False” Resurfacing

On active planetary surfaces, erosion or deposition can occur as an event rather than as a continuous process and can lead to a partial “resurfacing” in which craters below a certain diameter are preferentially removed, hence restarting the “crater counting clock” [Michael and Neukum, 2010]. Due to the resurfacing event, the surface effectively has two surface ages, the original surface age, and the age of the re-surfaced region, which appears as a step in the distribution between two segments that have different tangential isochrons (see Figure 2a in Michael and Neukum [2010]). In looking at the individual distributions that result from our model runs, however, we found that such steps can occur randomly, as shown in Figure 6.12, which was an individual subsample taken from a 10,000 km² area with an initial surface age of 3.5 Ga and zero obliteration rate. Using a simple automated code to cull out patterns like these in the model data, we found that these “false resurfacing” events occurred 150 times out of 2000 subsamples for the case of subsampling areas of 10,000 km² for a 3.5 Ga surface (~8% of the time), and occurred 130 times for the case of subsampling areas of 10,000 km² for a 1.0 Ga surface (~7% of the time). While the probability of having a “false resurfacing” event is low, it is non-zero based on these simple model results. For the case presented in Figure 6.12, it could result in incorrectly assuming there was an erosion and/or mantling event and would lead to either an erroneously young or old surface exposure age, depending on which set of data (upper or lower red dashed line) one used.

6.5 Case Study: Gale Crater

The modeling presented in Section 6.5 is motivated in part by a growing interest in dating small-scale features such as valley networks, alluvial fans and deltas. In Chapter 5 we showed that there are several depositional features (i.e. alluvial fans and deltas) around the rim and wall of Gale Crater, as well as around Gale’s central mound, Aeolis Mons, that indicate Gale once hosted several large lakes. All of these depositional features are small and most of the fans and deltas occur at different elevations, suggesting they may record different depositional events, but the relative timing between events is unknown. Can crater counting provide accurate age dating on such small features?

Previously, Grant et al. [2014] used cratering counting methods to date one of these features, the Peace Vallis fan, whose distal end was located near the *Curiosity* landing site (see Figures 6.3 and 5 in Chapter 4). Grant et al. [2014] took the fan to be composed of two “units”, the lower fan (BF) and the upper fan (AF). Based mainly on crater counting statistics, they claim that the lower fan is older, likely 3.2 to 3.3 Ga, with a younger lobe (AF) that was emplaced much later, perhaps within the Amazonian. They estimate ~1.2 Ga based on a cumulative analysis that combines the Peace Vallis Fan with deltaic deposits on the southern rim of Gale, which we refer to as the Pancake delta and its associated deposits in Chapter 5. While this interpretation of fan evolution is not supported by the fan geomorphology [Chapter 4], it does lead to the question of how much can one say about the timing between two surfaces that are quite small (the AF unit is ~49 km² and the BF unit is ~31 km²) and have experienced up to 20 m

of erosion [Chapter 4]. Grant et al. [2014] also point out the presence of an ~750 m crater on the upper fan (AF), which they exclude from their statistics because it both appears to predate the upper fan, based on the lack of ejecta and a smaller than expected rim on its NW side, and because “interpretation of statistics from the AF unit is complicated” [Grant et al., 2014; page 1146]. This leads to a second question of what is the probability of having a young fan with a “large” crater on it.

In Figure 6.13 our own crater frequency measurement for the upper (red circles) and lower (black squares) portion of the Peace Vallis fan are plotted on the Hartmann plot. The two data fields are nearly identical. We discounted the craters that appear to be “rolling-off”, either due to erosion and infilling or the limits of crater detection based on the imagery resolution, which results in a surface age for both units of approximately $670 \text{ Ma} \pm 380 \text{ Ma}$. Figure 6.14 shows the combined crater data for the upper and lower fan, as geomorphically it is a single fan [Chapter 4], and the best fit results in a surface age of 848 Ma. According to our model analysis, however, for an 80 km^2 surface that is dated to be between 0.5 and 1 Ga, the probability that the distribution observed matches the actual age is less than 10%.

As to whether it is likely to have a ~750 m crater on a 80 km^2 surface for a specific surface age, Figure 6.15 shows the probabilities that at least one crater of size i (where i ranges from 500 to 1 km) will be found on a feature that is 80 km^2 in size (the area of the Peace Vallis fan) for a range of surface ages. There is a 50% chance of having one or more 750 m diameter craters if the surface is 3.5 Ga, but if the surface is 1 Ga, there is still a 1 in 10 chance. Hence, our analysis suggests that accounting for erosion, the Peace Vallis fan is more likely much older than the standard analysis of just using a minimum crater size cutoff would predict, and that a single relative large crater (750 m) does not indicate a pre-fan formation. Such a crater is highly probably at 3 Ga and remains at least a 10% if the fan were just 1 Ga (which according to our model seems highly unlikely). Fracture mapping around the crater [Chapter 5] supports the notion that the crater impacted after the Peace Vallis fan was emplaced.

On the southwestern wall of Gale crater is a pancake-shaped delta (with an area of 117 km^2) that is described in detail in Dietrich et al. [2013] and Chapter 5. Figure 6.16 shows the crater data for this feature, which has a step in the data, a feature often associated with a resurfacing event [Michael and Neukum, 2010]. Due to the lack of HiRISE imagery of this delta (only CTX imagery is available), it is difficult to determine if there is an actual geomorphic process that may have caused a resurfacing event, or, as mentioned in Section 4.3, this step is a statistical artifact. Using CraterStats, we fit crater data for craters ranging from 250 to 500 m and 63 to 200 m. The larger craters give a surface age of 2.96 Ga and the latter give a surface age of 673 Ma (Figure 6.16). Figure 6.11 shows that with the small size of the feature, the probability of arriving at the true age of the surface based on the measured crater data is low.

6.6 Discussion

Based on our modeling results, assuming that obliteration and sample area size is accounted for, there remains only a probabilistic condition that the crater count data yields the true age. Due to obliteration at the smaller crater sizes and the small number of larger craters counted (even for the oldest surfaces, which have time to build up greater large crater

populations) the greatest uncertainty in ages occurs for surfaces that have true ages of 1.5 Ga. This minimum at 1.5 Ga arises from upper and lower limits on possible crater size-frequencies for very old and young areas. For areas of 1000 km² or less, the probability of crater counts of an individual deposit yielding the correct age drops below 30% at these intermediate ages and for deposits smaller than 100 km² does not exceed 15%. This means that there is an irresolvable significant uncertainty in the dating of any modest to small size depositional feature. Importantly, even when the data appear to line exactly along the reference Hartmann isochrons, the true age may be greatly different due to the effects of obliteration and rare impacts operating on small areas.

The application to the two Gale Crater deposits illustrates these issues. On the Peace Vallis fan, standard practice in analysis of crater count data leads to selecting a threshold diameter below which it is assumed obliteration has altered the crater densities and another threshold diameter above which it is assumed the statistics are poor, or perhaps such craters record events prior to deposition. Our analysis shows that obliteration effects may sweep across the entire population of crater sizes on such a small area, such that what could be assigned an age of less than 1 Ga is perhaps more likely to record 3 Ga since fan deposition. And even with this improved model, the uncertainty remains very large. It can be said definitively that a young age cannot be assigned with any level of confidence to the Peace Vallis fan based on the available crater densities and estimates of erosion.

The simplicity of this model makes it easily adaptable to surfaces of different areas and different planets where isochron data exists. For the Mars isochrons, the isochron data for 1 Ga is the standard (in the model), and all other isochrons are just a known multiple [Hartmann, 2005] of the 1 Ga crater size-frequencies (for example, the crater frequencies for 0.5 Ga are 0.5 times the crater frequencies for 1 Ga; the crater frequencies for 3.5 Ga are 5.72 times the crater frequencies for 1 Ga). The only model inputs are for the area and age of the “true” surface, the subsample area, the number of subsamples, and the amount of erosion and infilling (i.e. b in nm per year). The output can be specified as either the number of craters counted for each crater size for each subsample area, the ages assigned to each subsample area (there are several options for choosing how the crater age is determined and what crater bins should be used), or the probabilities of being plus or minus some age from the true surface age. Here we chose areas mainly ranging from 100 km² to 10,000 km², but probabilities for any size and age combination can be determined. To use Figure 6.5 and 6.11, we recommended plotting crater count data in the Hartmann style if using CraterStats [Michael and Neukum, 2010] and using the roll-off curve of the smaller craters to determine the extent of erosion. If there is no roll-off in the data, or it is known to be due to image resolution effects, Figure 6.5 can be used, otherwise Figure 6.11 should be used. As a first estimate on the age of the surface, it is advised to ignore larger crater bins that contain only a single or a few craters and to ignore craters that are rolling-off (as those populations are likely being highly affected by erosion and infilling). If the moderately sized craters are observed to fall on a single isochron, that isochron is likely the best first estimate of the surface’s age. The surfaces we investigated showed obliteration through a large range of crater sizes though, at times the entire range counted, so the age estimate using moderately sized craters may still be underestimating the true age of the surface. With a best estimate for the surface age, use that surface age on the x-axis of either Figure 6.5 or 6.11 and find the nearest

corresponding area curve. The probability that the surface dated is actually the age found on the Hartmann plot is found on the y-axis.

6.7 Conclusions

Due to volcanic and geomorphic processes occurring over time on Mars, erosional or mantling resurfacing events will change the crater population on the surface of the planet, typically removing craters from the small-diameter end of the distribution. At the large-diameter end of the distribution, craters are more likely to survive, but counting a large crater on a small surface area leads to erroneously old ages. Here we have developed a simple model to quantify the effects of sample area size and crater obliteration effects on age estimates derived from crater counting. Results from this model suggest several general outcomes when using crater counting to ascertain Martian surface ages for small landforms. First, crater counting on small surfaces cannot give an exposure age with certainty due to loss of information at the lower end (due to erosion and infilling) and upper end of the distribution (due to a low probability of counting large craters). This was already known generally, but here we show this quantitatively. Secondly, for small areas, there is a constraint on the range of crater diameters that best retain information about the surface's true age, and this range decreases with decreasing surface area and increasing erosion and infilling. Thirdly, when looking at the individual distributions that were output from the model, we found that steps in the data, which are often indicators of resurfacing events, can occur randomly. While the probability of having a "false resurfacing" event was found to be low, less than 5% of the distributions, it could result in assigning either an erroneously young or old surface exposure age, depending on which tangential isochron is used. When accounting for the obliteration of craters, the model suggests that, in general, the least reliable ages occur between 1 and 3 Ga years. Younger surfaces preserve the smaller craters and older surfaces have had enough time to acquire larger craters, both improving the probability of obtaining the correct age. For areas less than 1000 km² and true ages of 1 to 3 Ga, there is only a 20 to 40% probability of the calculated age lying within 0.25 Ga of the true value, where as for surfaces smaller than 500 km² the corresponding probabilities drop to about 10 to 20%.

Crater-counting data for two small (80 km² for the Peace Vallis fan and 117 km² for the Pancake delta) depositional features within Gale Crater were presented to illustrate how the model provides further constraints on the likely ages of these deposits. The analysis presented shows that based on the small size of the fan, there is an ~10% chance that the crater distribution observed on the Peace Vallis fan is actually from a 0.5 to 1.0 Ga surface, as was suggested by Grant et al. [2014], and furthermore, the fan could plausibly be three times older and due to erosion and infilling now has a very similar crater distribution to a ~1.0 Ga surface. Similar results were found for the Pancake delta. Even though portions of the observed crater count data for each feature follow crater density isochrons, the uncertainties are large leading to an unreliable age assignment.

6.8 Acknowledgements

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6.9 Figures

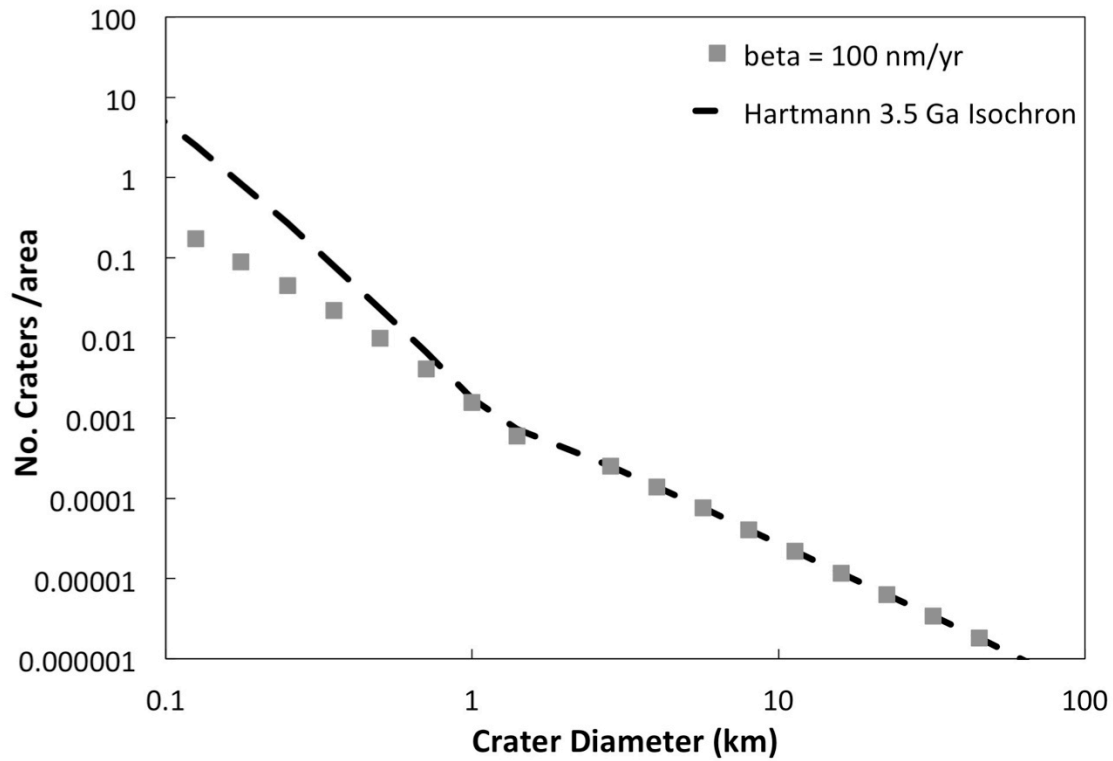


Figure 6.1 Crater diameter distribution as defined by the Hartmann 2004 iteration [Hartmann, 2005] shown by dashed line (for 3.5 Ga) and the effect that 100 nm/yr of erosion and infilling (beta=100) would have on that distribution [Smith et al., 2008] shown by grey squares.

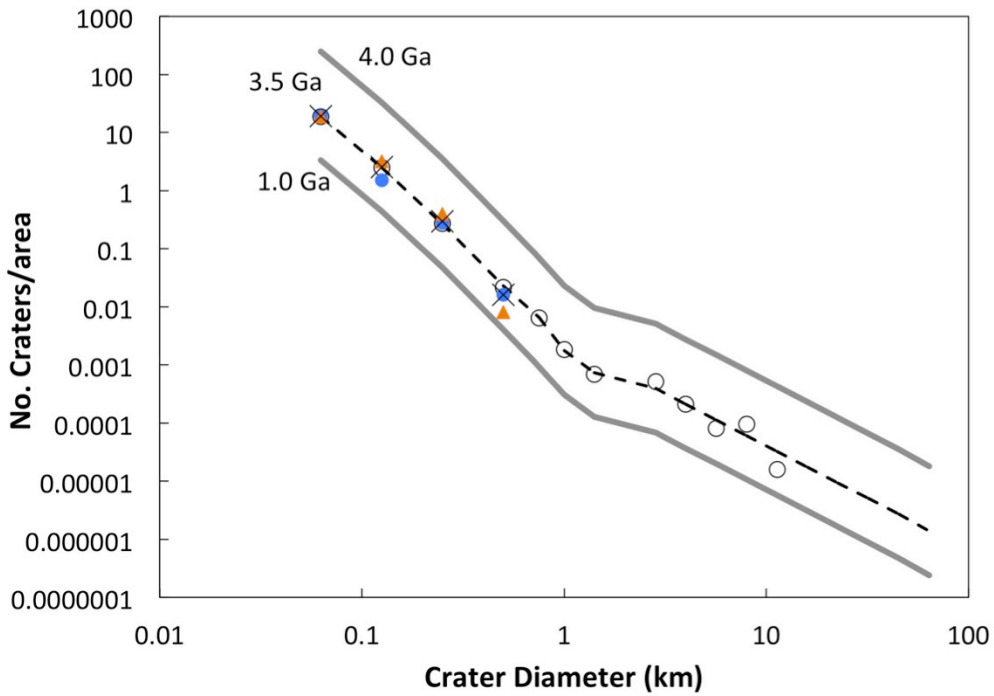
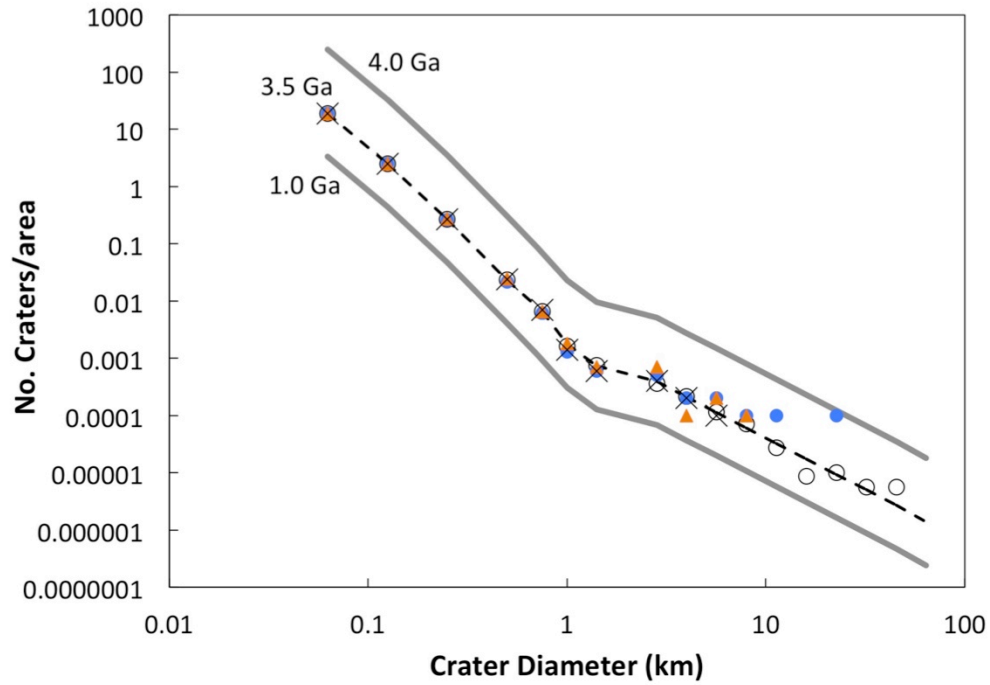


Figure 6.2 (a) Hartmann plot where the thick grey lines represent the 4.0 Ga (upper) and 1.0 Ga (lower) isochrons, respectively, and the dashed line represents the 3.5 Ga isochron, which is the age of the “true” surface. The orange triangles and blue circles show the results for two randomly selected individual crater counts taken from a 10,000 km² subsampling area. The open circles represent the average crater frequency for all 2000 subsamples and the X’s represent the mode. (b) The results are now shown for two individual crater counts taken from a 100 km² subsampling area.

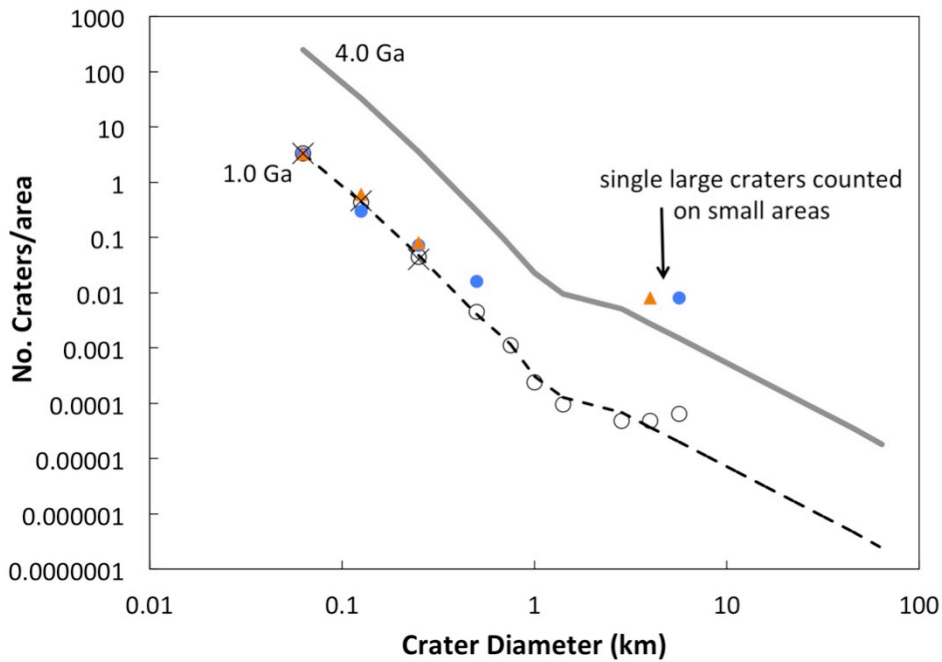
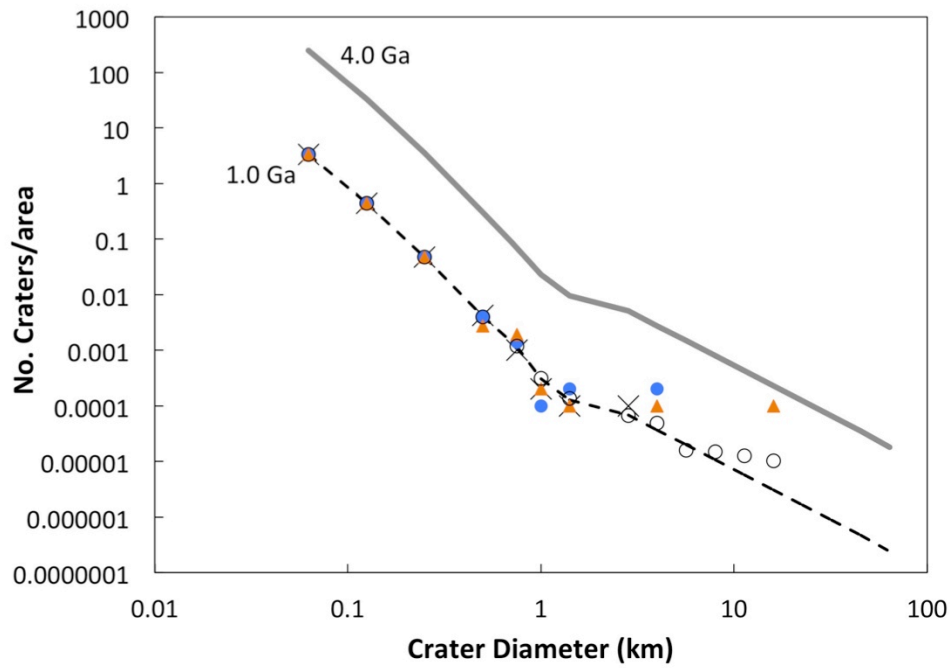


Figure 6.3 (a) Hartmann plot where the thick grey line represents the 4.0 Ga isochron and the dashed line represents the 1.0 Ga isochron, which is the age of the “true” surface. The orange triangles and blue circles show the results for two individual crater counts taken from a 10,000 km² subsampling area. The open circles represent the average crater frequency for all 2000 subsamples and the X’s represent the mode. (b) The results are now shown for two individual crater counts taken from a 100 km² subsampling area.

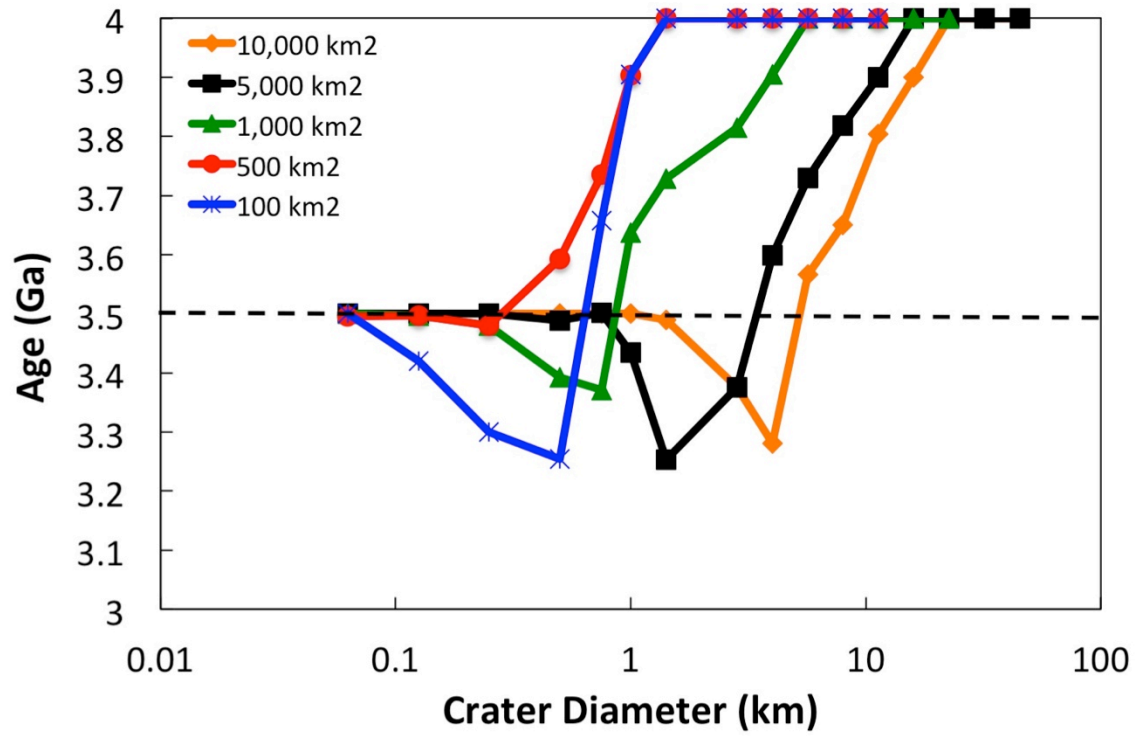


Figure 6.4 The effect of sampling progressively smaller and smaller areas on the surface age calculated from a single crater diameter bin is shown. The “true” surface age in this case was 3.5 Ga. The ages plotted are the average age based on 2000 subsamples.

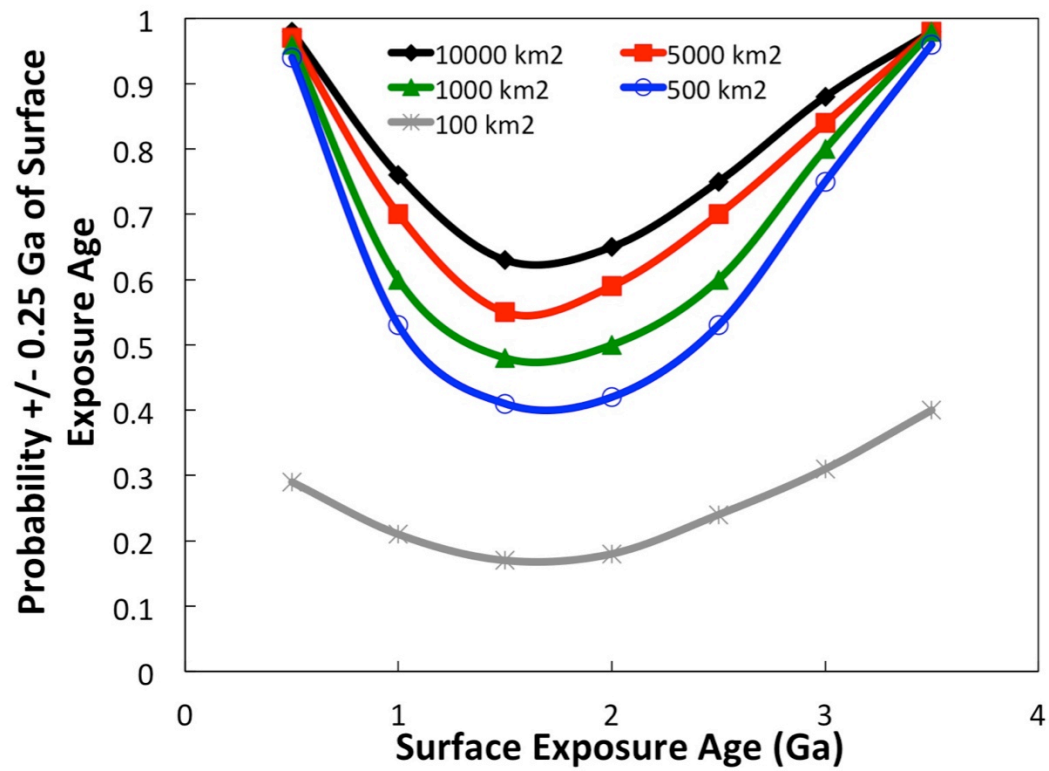


Figure 6.5 Curves of constant area are shown where the probability of being ± 0.25 Ga within the true surface age is on the y-axis and the true surface age (in Ga) is on the x-axis.

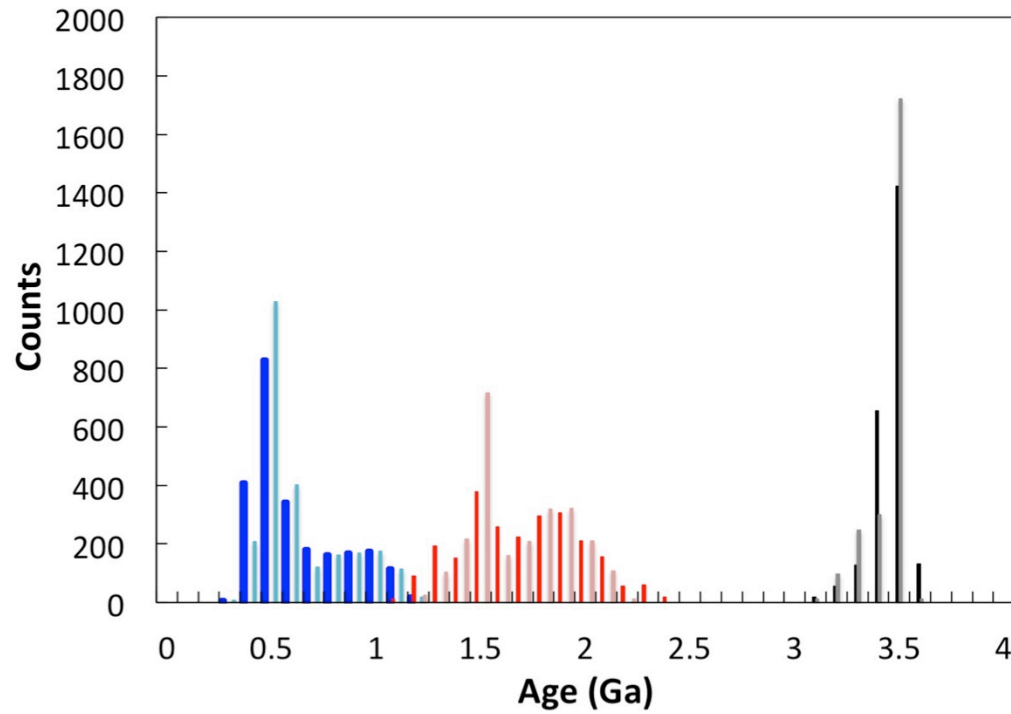


Figure 6.6 The distribution of ages from the “non-eroding” model for the case of 10,000 (light) and 1,000 (dark) km² subsampling areas for true surface ages of 0.5 Ga (blue), 1.5 Ga (red), and 3.5 Ga (black). The distribution of ages that result from subsampling the 0.5 Ga surface and 3.5 Ga surface are right-skewed and left-skewed, respectively, while the 1.5 Ga distribution is more normal.

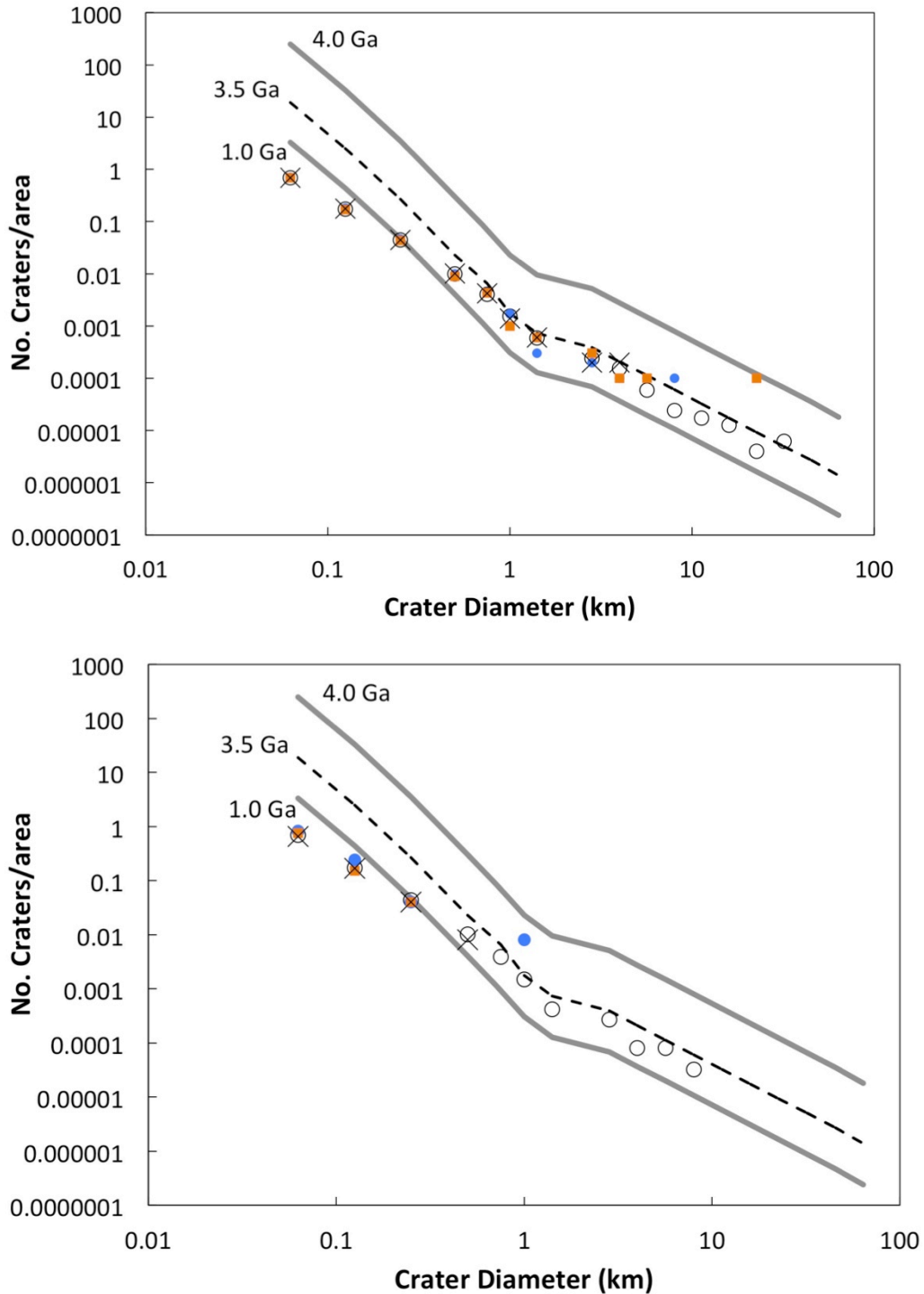


Figure 6.7 (a) Hartmann plot where the thick grey lines represent the 4.0 Ga (upper) and 1.0 Ga (lower) isochrons, respectively, and the dashed line represents the 3.5 Ga isochron, which is the age of the “true” surface. The orange triangles and blue circles show the results for two individual crater counts taken from a 10,000 km² subsampling area, where 100 nm/yr of erosion and infilling has been taken into account. The open circles represent the average crater frequency for all 2000 subsamples and the X’s represent the mode. (b) The results are now shown for two individual crater counts taken from a 100 km² subsampling area.

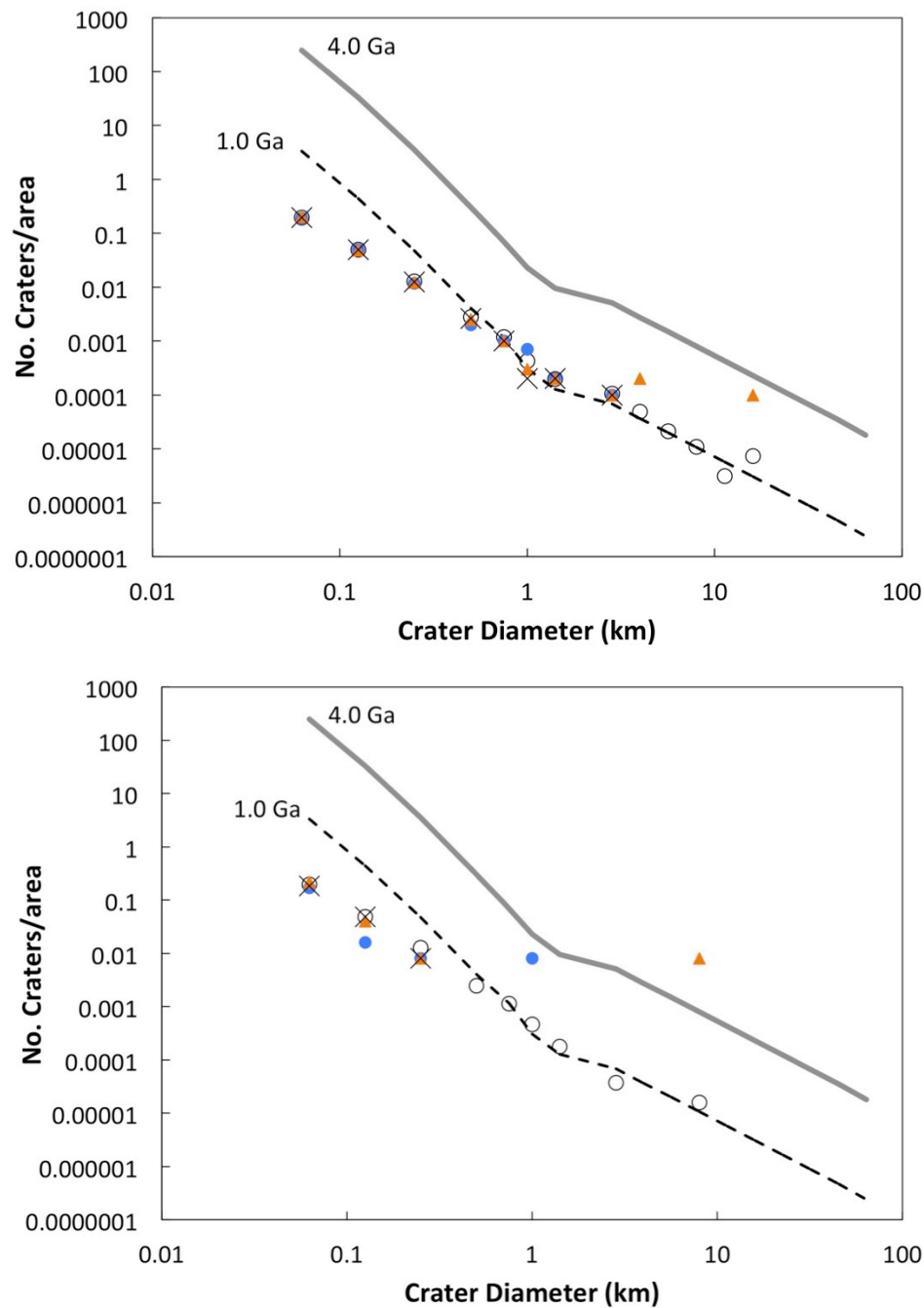


Figure 6.8 (a) Hartmann plot where the thick grey line represents the 4.0 Ga isochron and the dashed line represents the 1.0 Ga isochron, which is the age of the “true” surface. The orange triangles and blue circles show the results for two individual crater counts taken from a 10,000 km² subsampling area, where 100 nm/yr of erosion and infilling has been taken into account. The open circles represent the average crater frequency for all 2000 subsamples and the X’s represent the mode. (b) The results are now shown for two individual crater counts taken from a 100 km² subsampling area.

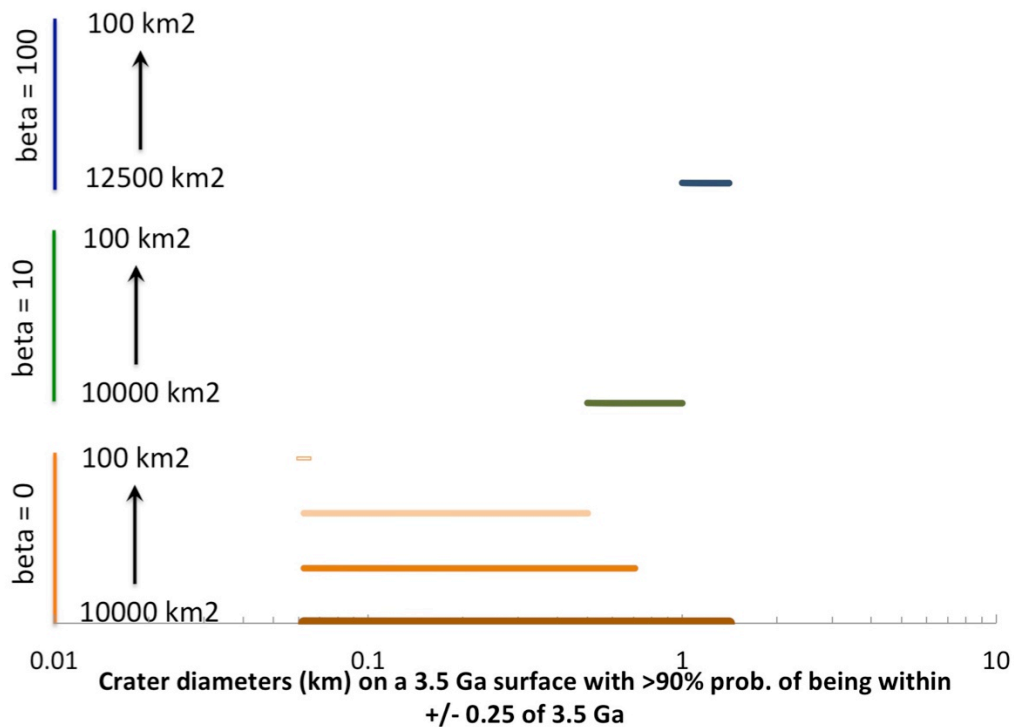


Figure 6.9 Plotted are the range of craters on a 3.5 Ga surface that have >90% probability of being within ± 0.25 Ga of true surface age. For small areas, there is a constraint on the range of crater diameters that may retain information about the surface's true exposure age. This range decreases with decreasing surface area (dark to light colors) and increasing erosion and infilling (bottom to top on y-axis). For example, craters in the range of 500-750 m will accurately reflect the age of a 10,000 km² 3.5 Ga surface that has experienced moderate erosion, but if this surface is only 100 km², there is a loss of information at all crater sizes.

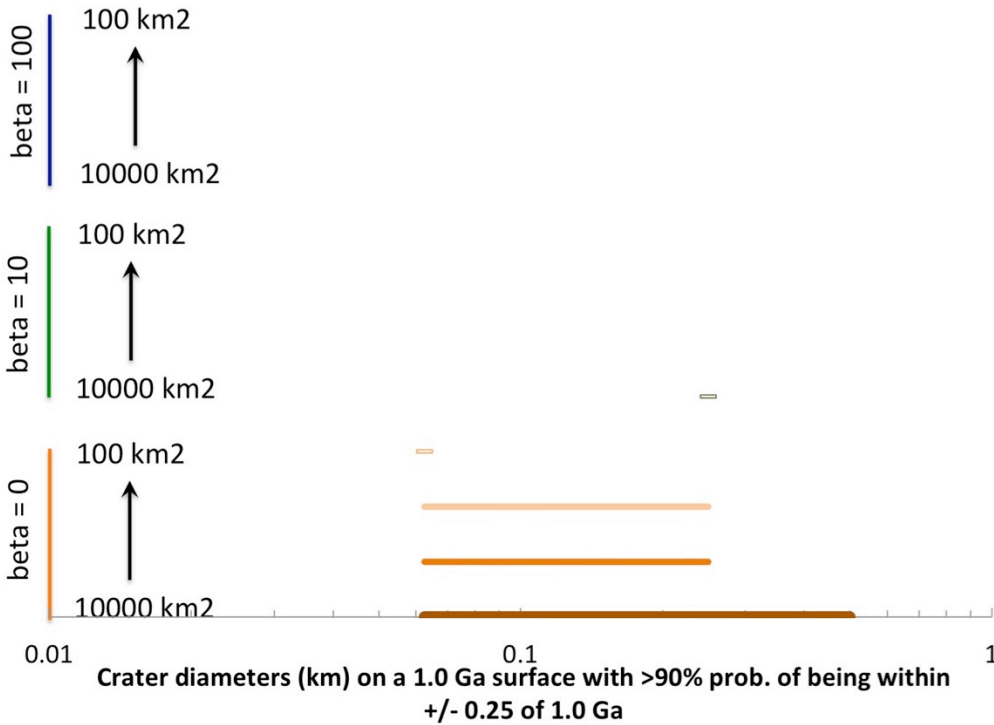


Figure 6.10 Plotted are the range of craters on a 1.0 Ga surface that have >90% probability of being within ± 0.25 Ga of true surface age.

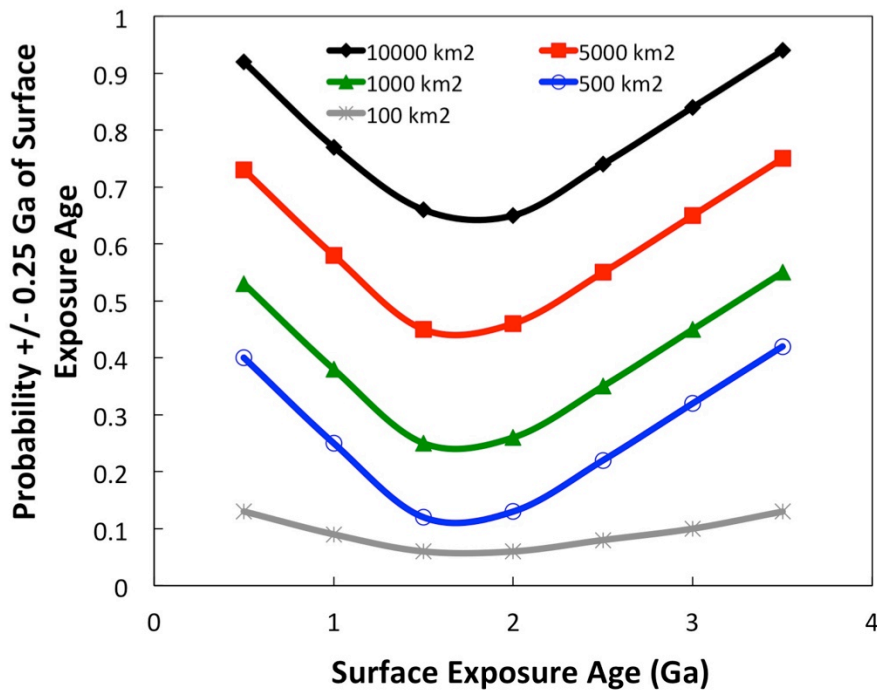


Figure 6.11 Curves of constant area are shown where the probability of being ± 0.25 Ga within the true surface age, for a moderately eroded surface ($\beta = 100$ nm/yr), is on the y-axis and the true surface age (in Ga) is on the x-axis.

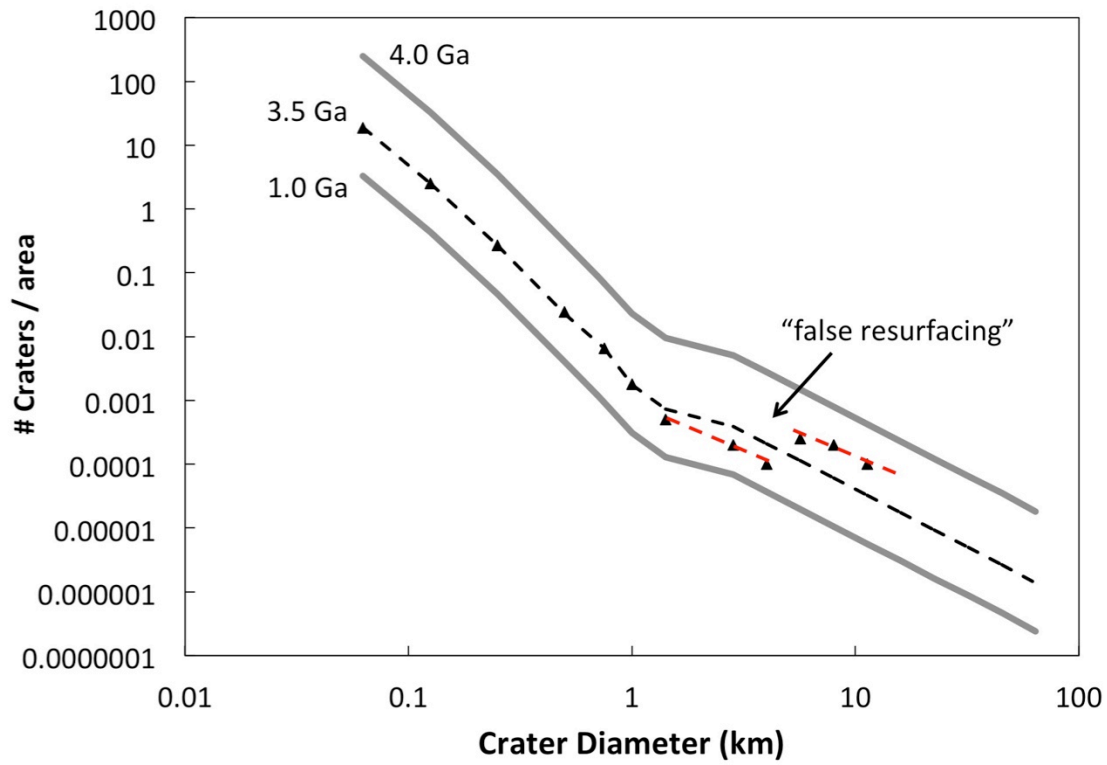


Figure 6.12 An example of a “false resurfacing” event; individual subsample taken from a $10,000 \text{ km}^2$ area with an initial surface age of 3.5 Ga (no erosion).

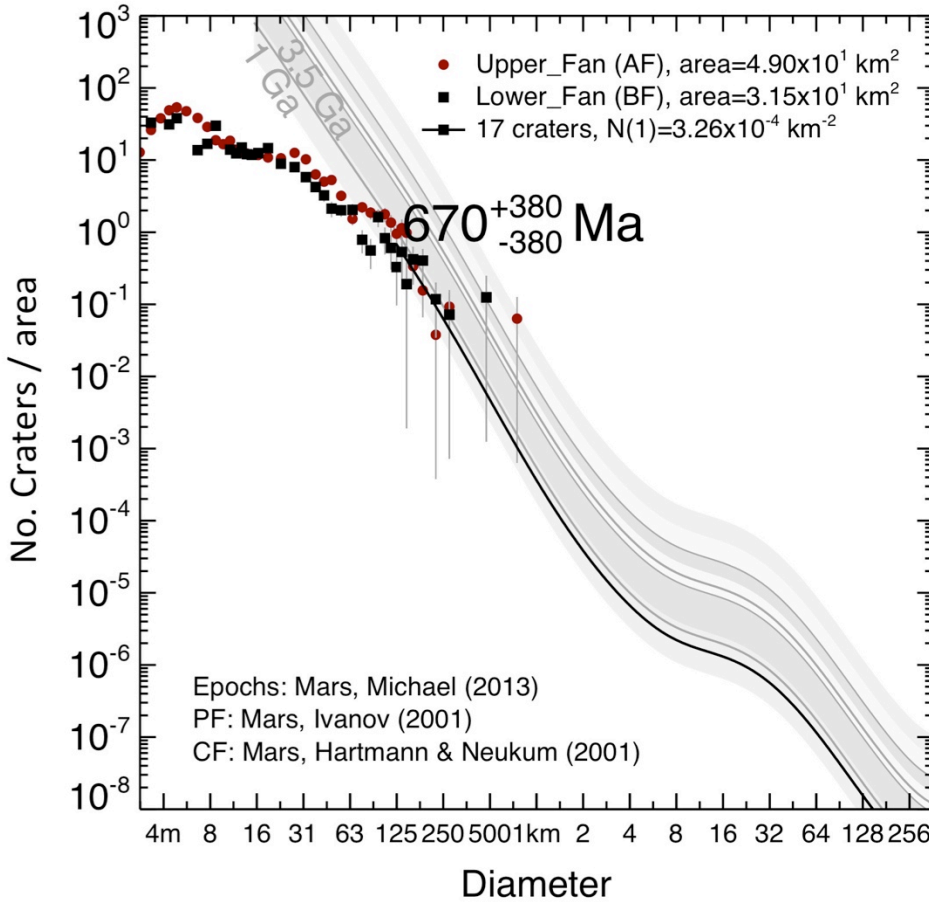


Figure 6.13 Crater counting data from the Peace Vallis fan, where data from the upper fan (AF) in Grant et al. [2014] is represented by red circles and the lower fan data (BF) is represented by black squares, is shown on a typical Hartmann plot. An isochron is fit through craters larger than 125 m.

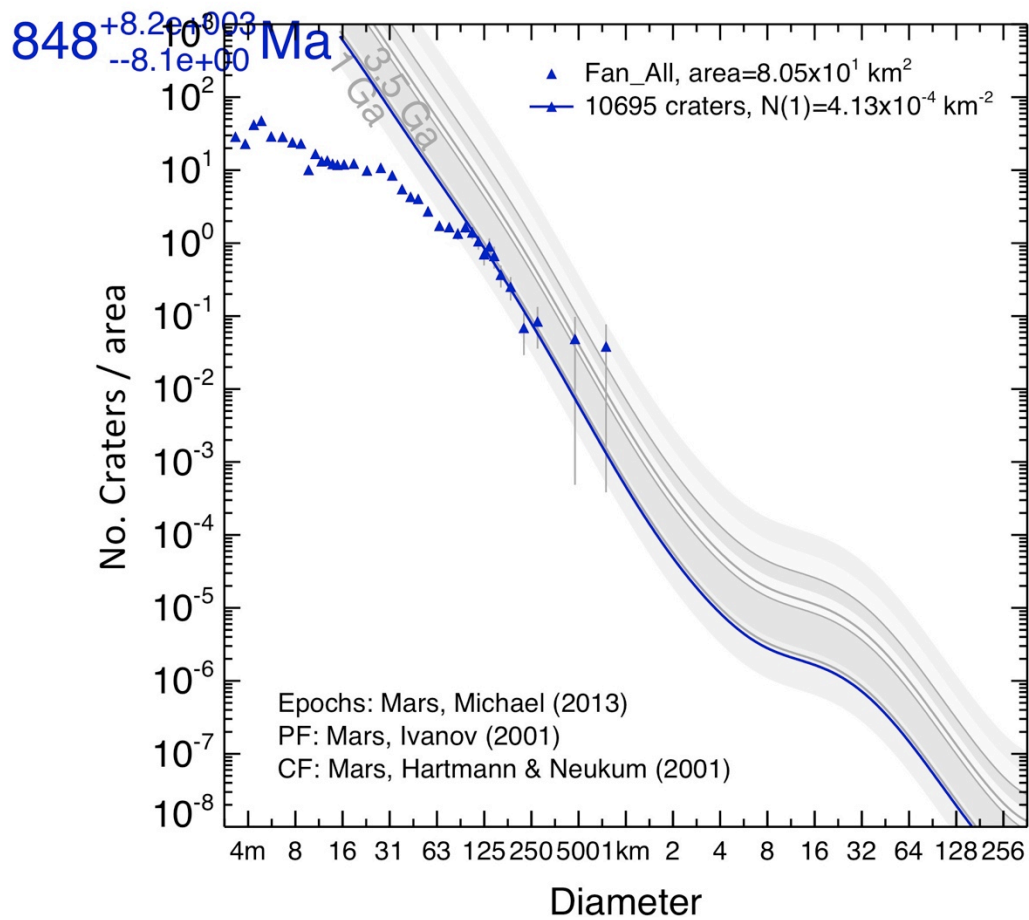


Figure 6.14 Crater counting data from the entire Peace Vallis fan, which is a single geomorphic unit [Palucis et al., 2014], is shown on a typical Hartmann plot. An isochron is fit through all craters diameters.

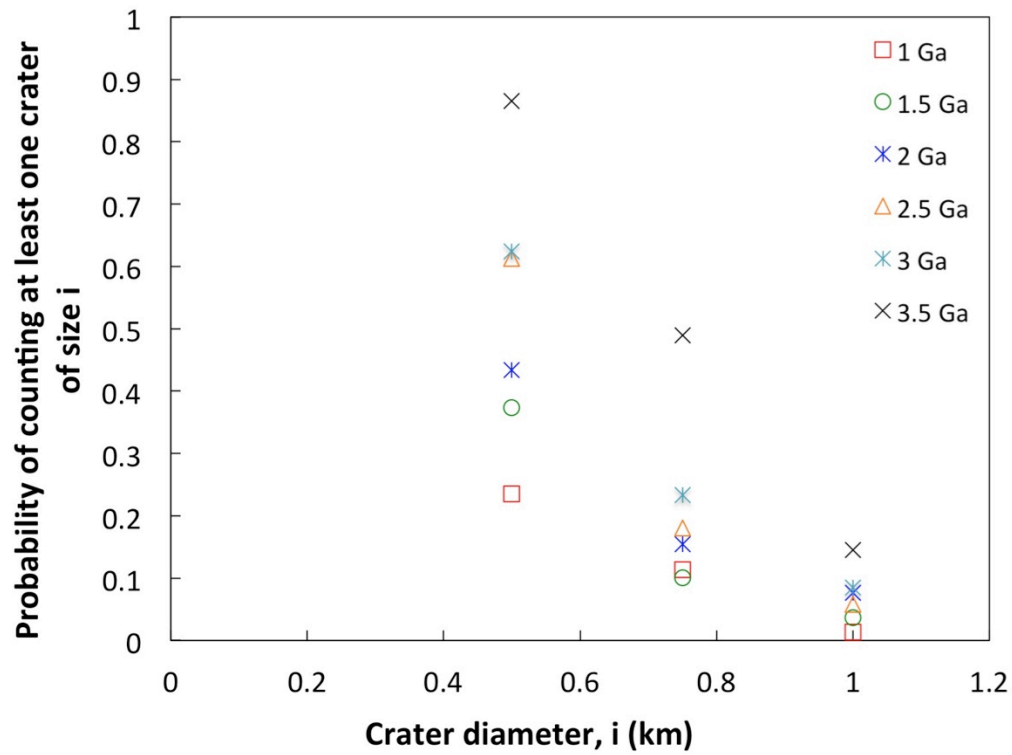


Figure 6.15 Results from model assuming that no erosion and infilling has occurred (i.e. $\beta = 0$). Plotted are the probabilities that at least one crater of size i (where i ranges from 500 m to 1 km) will be found on a feature that is 80 km^2 in size.

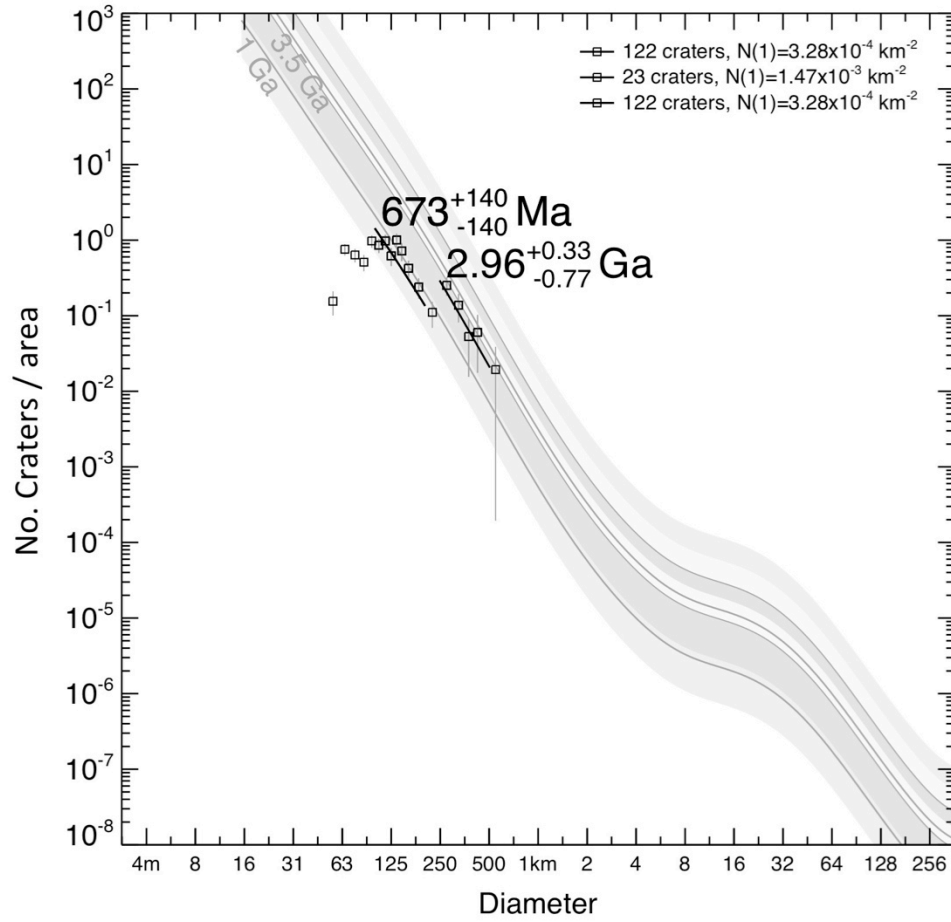


Figure 6.16 Crater counting data from the pancake delta [Dietrich et al., 2013], is shown on a typical Hartmann plot. We fit isochrons for craters ranging from 250 to 500 m and 63 to 200 m.

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