

Runoff required to drive postimpact gully development on the walls of Meteor Crater (Arizona, USA)

Marisa C. Palucis^{1,†}, Alan D. Howard², David A. Kring³, Kuni Nishiizumi⁴, Marc Caffee⁵, and William E. Dietrich⁶

¹Department of Earth Science, Dartmouth College, Hanover, New Hampshire 03755, USA

²Planetary Science Institute, Tucson, Arizona 85719, USA

³Lunar and Planetary Institute, Houston, Texas 77058, USA

⁴Space Sciences Laboratory, University of California, Berkeley, Berkeley, California 94720, USA

⁵PRIME Lab, Department of Physics, Purdue University, West Lafayette, Indiana 47907, USA

⁶Department of Earth and Planetary Science, University of California, Berkeley, Berkeley, California 94720, USA

ABSTRACT

Since the impact ~50,000 yr ago, surface runoff has entrained and transported sediment from the walls to the floor of Meteor Crater (Arizona, USA). Previous work interpreted this erosion and deposition to be due to predominantly fluvial (i.e., dilute water transport) processes. However, light detection and ranging (LiDAR)-derived topographic data and field observations indicate that debris flows dominated, which were likely generated by runoff that entrained the talus that borders bedrock cliffs high on the crater walls. The low gradient of the crater floor caused debris flows to stop, leaving lobate deposits, while fluvial processes delivered sediment toward the center of the crater. Cosmogenic radionuclide dating of levee deposits suggests that debris-flow activity ceased in the late Pleistocene, synchronous with regional drying. Assuming a rock-to-water ratio of 0.3 at the time of transport by mass flows, it would have taken $\sim 2 \times 10^6$ m³ of water to transport the estimated $\sim 6.8 \times 10^6$ m³ of debris-flow deposits found at the surface of the crater floor. This extensive erosion would require ~6 m of total runoff over the 0.35 km² upslope source area of the crater, or ~18 mm of runoff per debris-flow event. Much more runoff did occur, as evidenced by crater lake deposits, Holocene fluvial activity (which produced little erosion), and contemporary rainfall rates. Rarely on Earth is the total amount of water that creates and runs through a landscape estimated, yet such calculations are commonly done on Mars. Our analysis suggests that

erosional and depositional landforms may record only a small fraction of the total runoff.

INTRODUCTION

Meteor Crater, a well-preserved impact crater located on the Colorado Plateau in north-central Arizona (35°1.648'N, 111°1.363'E; Fig. 1A), formed ~49,000–61,000 yr ago (Sutton, 1985; Phillips et al., 1991; Nishiizumi et al., 1991; Marrero et al., 2010; Barrows et al., 2019) 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 (Figs. 1B and 1C), which were first discussed in detail by Kumar et al. (2010). Similar features on Mars have gained considerable attention since being recognized by Malin and Edgett (2000), who identified three main attributes to these landforms: a gully head or alcove, channels or chutes, and a fan or apron (Fig. 2). As gullies have been hypothesized by some to have formed from liquid water or brines (Malin and Edgett, 2000; Hartmann et al., 2003; Coleman et al., 2009; see review by Conway et al., 2019), versus dry or gas-supported flows, and have been mapped across geologically young terrains on Mars when liquid water was thought to be highly unstable (Late Amazonian; Harrison et al., 2015), constraining their formation mechanism has been a key goal for understanding recent Martian climate and water reservoirs.

However, the water sources and volumes required to form Martian gullies and erode crater walls through time are uncertain. Models used to assess water amounts on Mars are typically derived from field studies on Earth (e.g., Costard et al., 2002, 2007; Hartmann et al., 2003;

Conway et al., 2015, 2018, 2019), but field studies on Earth rarely ask the question: How much water did it take to make and modify this landscape? Additionally, one important aspect that does not appear to have been considered in previous terrestrial gully studies is the effect of a drying climate on their evolution. This is especially important on Mars, where current conditions are cold and dry, but gully formation may have initiated during periods of high obliquity (Laskar et al., 2004), when there was enhanced potential for precipitation and runoff of water at lower latitudes. Meteor Crater, as first pointed out by Kumar et al. (2010), provides an opportunity to explore how a landscape relevant to Mars 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. Careful mapping of its deposits, using both sedimentology and their surface expression, allows researchers to assess the dominant flow processes (i.e., fluvial vs. debris flow), which is necessary to model the water fluxes and volumes required to form gullies. Furthermore, using a combination of radiometric dating and the fact that there is near-complete conservation of sediment mass in the crater, it is possible to infer changes in the frequency and intensity of runoff events that drove crater evolution.

Observations at Meteor Crater made by Kumar et al. (2010) led them to propose a time sequence of events for the formation and modification of the gullies at Meteor Crater. Immediately postimpact, 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, where 30 m of lake sediments cover allogenic breccias with no intervening alluvium (Shoemaker and Kieffer,

Marisa C. Palucis  <https://orcid.org/0000-0003-0034-5810>

[†]marisa.c.palucis@dartmouth.edu

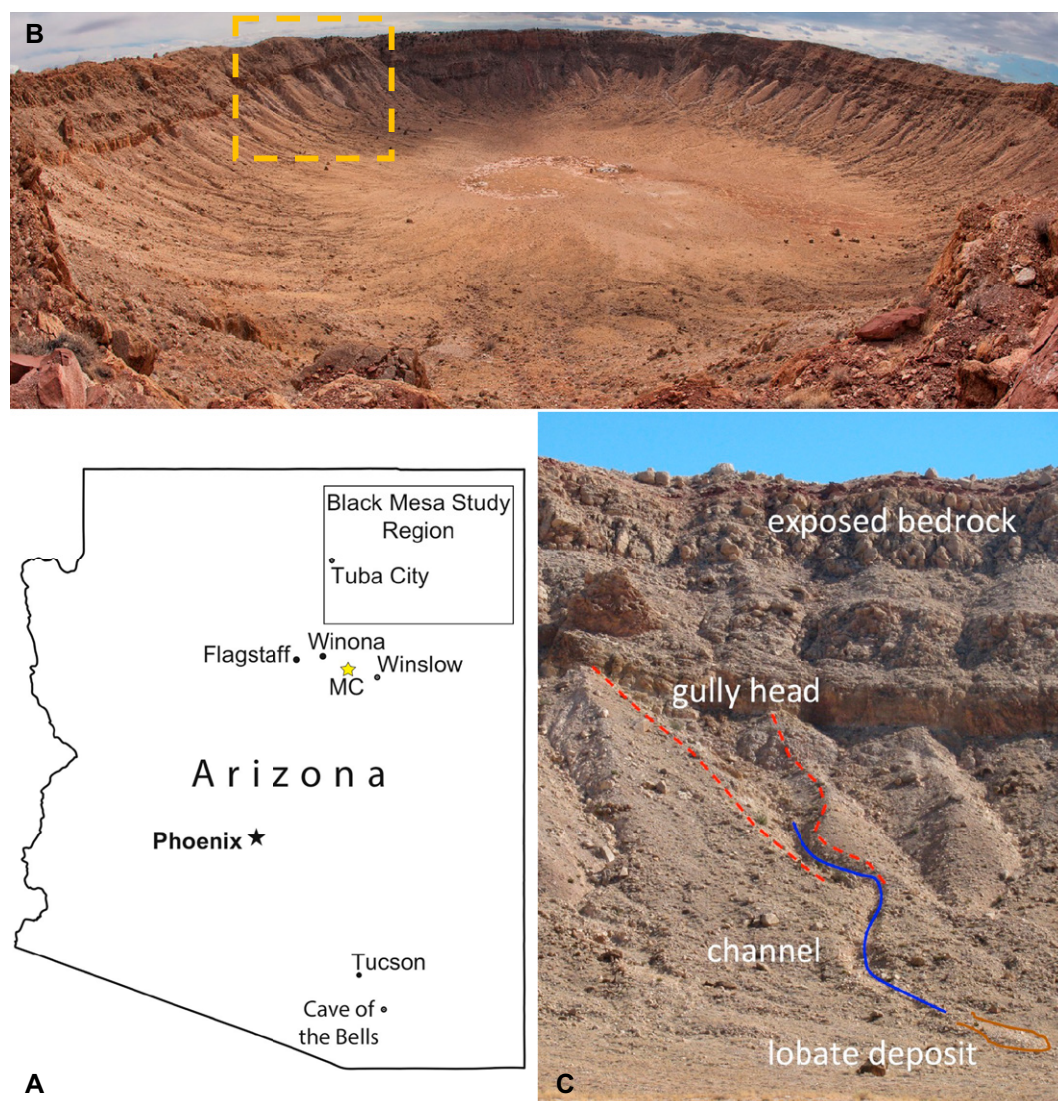


Figure 1. (A) Location of Meteor Crater, Arizona (USA), marked with a yellow star. The location of the Winslow airport, which is 42 km from Meteor Crater, and other relevant locations discussed herein are marked with black dots. Box outlines the Black Mesa study region of Zhu et al. (1998) and Zhu and Kipfer (2010). (B) Image of Meteor Crater (credit: William E. Dietrich, University of California, Berkeley). The crater is ~180 m deep and 1.2 km in diameter with a rim crest that rises ~30 m to 60 m above its surrounding plains. (C) Image of the northeastern wall, which is ~180 m tall, highlighting the main components of a Meteor Crater gully system, which consists of the upper bedrock wall, where runoff is generated during precipitation events, a gully head that is incised into talus/breccia that mantles the crater walls, a gully channel, and a depositional zone (which often consists of coarse-grained lobate-shaped deposits). Latitude, longitude = 35.0278°N, 111.0222°W.

1979). Since the crater is not breached, Kumar et al. (2010) assumed that the source of water for the lake was artesian flooding, groundwater springs, rainfall, or some combination of the three. Surface runoff initially washed crater-wall impact deposits into the lake and then later incised bedrock. Kumar et al. (2010) proposed that lake-level fluctuations would result from varying groundwater seepage exfiltrating from radial fractures and tear faults. These proposed seepage outflow channels would then erode the crater walls and deposit sediment across the lower slopes and crater floor (i.e., the Pleistocene alluvium mapped by Shoemaker and Kieffer, 1979) and interfinger with lake sediments. To explain gullies in the lower slopes of the crater wall, Kumar et al. (2010) proposed that a later, second phase of enhanced precipitation occurred, where material deposited from this post-Pleistocene gully-forming event is now being dissected and overlapped, on the crater

floor, with modern playa deposits. They suggested that 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, 616). No grain-size data, channel slopes, channel geometry, or other quantitative metrics that might support the inference that fluvial processes are the dominant erosive and transporting agent were reported by Kumar et al. (2010).

Our initial visit to Meteor Crater occurred after Kumar et al. (2010) suggested an alternate hypothesis for the processes driving gully formation. The sedimentology and morphology of many of the deposits along the lower walls of the crater, which are dominated by matrix-supported levees and lobate snouts, indicate that many individual debris-flow events occurred in the past. Furthermore, the weathered state of these deposits suggested a possible Pleistocene, rather

than Holocene, origin. Deposits recording fluvial processes appeared to be subsidiary and more recent. These observations warranted another investigation of the erosional processes (and the resulting deposits) at Meteor Crater, with a specific emphasis on the link between sediment transport process and climate and an assessment of the amount of runoff that would have been necessary to erode the crater-wall gullies.

To obtain a quantitative estimate of the proportion of sediment moved by debris-flow processes, as compared to fluvial processes, we acquired aerial light detection and ranging (LiDAR) data for the crater (~0.25 m per pixel; point density = 5.4 points/m²), performed detailed field mapping and surveying of the debris-flow levee deposits, collected samples for cosmogenic ¹⁰Be dating of several deposits exiting one of the gully systems, and calculated the erosional flux and minimum water runoff associated with sediment deliv-

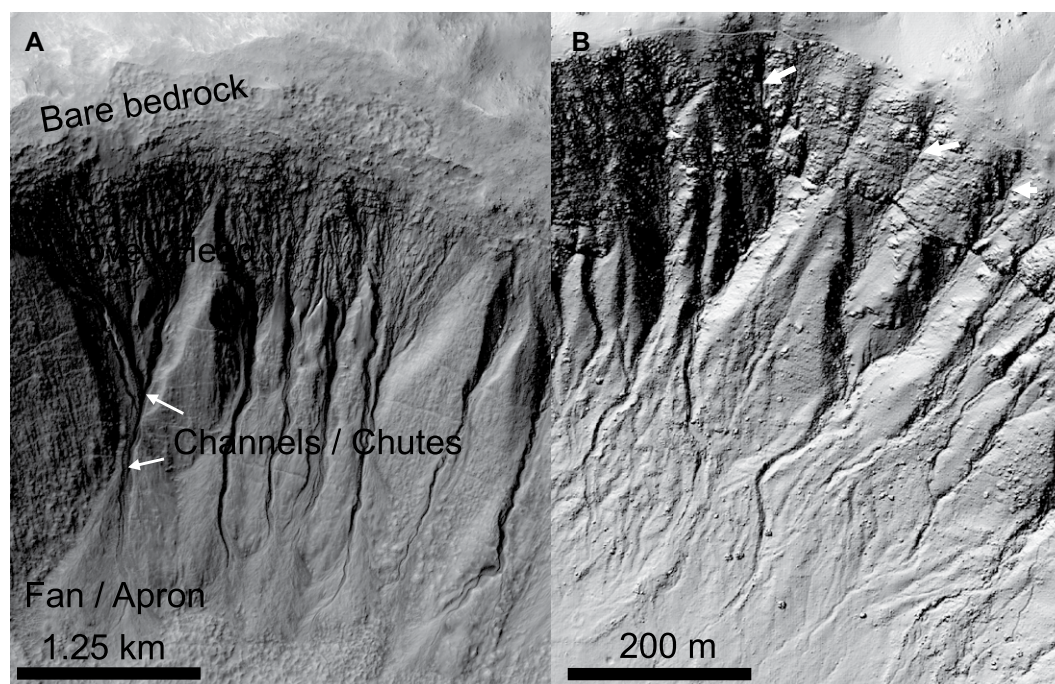


Figure 2. (A) High Resolution Imaging Science Experiment (HiRISE) image ESP_050858_1435 of gullies in a small crater in the Terra Cimmeria region of Mars (credit: National Aeronautics and Space Administration [NASA]/Jet Propulsion Laboratory [JPL]–Caltech/University of Arizona). Latitude, longitude = 28.0145°N, 261.3920°W. (B) Shaded relief image of the southeastern crater wall of Meteor Crater generated with 25 cm/pixel light detection and ranging (LiDAR) data (<https://opentopography.org/>).

ery to the crater floor. Here, we present data that suggest debris-flow processes dominated gully incision up until the Holocene. We then address three central questions: (1) Why did debris flows happen here, and why did they cease? (2) How much water was required to accomplish the observed erosion? (3) Is Meteor Crater an appropriate analog for understanding gullies on Mars? We conclude with a general conceptual model of gully formation processes that we propose is broadly applicable to steep escarpments (on Earth and Mars), where the lower portions are mantled with coarse debris embedded with some finer-grained sediment.

STUDY SITE

Geology

Meteor Crater is a simple, bowl-shaped impact crater, ~180 m deep (from the current-day central floor to the crater-rim top) and 1.2 km in diameter (1.1 km² in area), and it is encompassed by a rim of ejecta that rises 30–60 m above the surrounding plain (Shoemaker, 1959). Independent dating using ¹⁰Be/²⁶Al measurements (Nishiizumi et al., 1991; Barrows et al., 2019), cosmogenic ³⁶Cl measurements (Phillips et al., 1991; Marrero et al., 2010), and thermoluminescence dating of shock-metamorphosed dolomite and quartz (Sutton, 1985) places the age of the impact between ~49,000 to ~61,000 yr ago, corresponding to the Wisconsin interstadial, which was a relatively warm period (Jacobs,

1985). Shoemaker (1959, 1987), Shoemaker and Kieffer (1979), Kring (1997, 2007, 2017), Kumar and Kring (2008), and Kumar et al. (2010) provided detailed descriptions 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 (Permian) to the Moenkopi Formation (Triassic), which are units in the upper portion of the Grand Canyon sequence (Dutton, 1882). The lowest exposed unit in the crater is the Coconino Sandstone, composed of well-sorted quartz eolian sands (McKee, 1947). It is the basal unit excavated by the impact event, but only the upper portions of the formation are exposed in the crater walls. The Coconino Sandstone is overlain by ~1.5 m of Toroweap Formation, composed of sandstone and dolomite. The Kaibab Formation, an ~80-m-thick unit of dolomite, dolomitic limestone, and thin calcareous sandstone horizons, overlies the Toroweap Formation. The Kaibab Formation is exposed along the steep upper wall of the crater. Shoemaker and Kieffer (1974) mapped the Toroweap Formation as being conformable with the Kaibab Formation, although it is recognized as being distinct in outcrops (and unconformable in other areas of northern Arizona; Kring, 2007). Two members of the Moenkopi Formation rest disconformably on the Kaibab Formation. The bottom member is the Wupatki Member, which is 2–6 m thick and composed of very fine sandstone. Atop the Wupatki Member, there is the Moqui Member,

which is 2–10 m thick and composed of fissile siltstone. The overturned rim sequence occurs in this unit, and because of the fissile nature of the Moqui Member, pinpointing the contact between the upright and overturned Moenkopi strata in the rim is difficult. On average, the exposed Moenkopi Formation is ~8.5 m thick (Kring, 2007). The bedrock units in the crater walls are crosscut by faults produced during the impact. Authigenic breccias occur along those faults in near-vertical planes.

Based on the geologic map (Fig. S1¹) and cross section by Shoemaker (1960), the crater

¹Supplemental Material. Figure S1: Geologic map of Shoemaker (1960) that was recolored by Kring (2007) and has been georeferenced to compare with our geomorphic mapping to highlight the correspondence of the debris-flow levees and snouts with material mapped as Pleistocene alluvium. Figure S2: Highly fractured and broken bedrock wall above gully 0. Figure S3: (A) 0.5 m contours draped over shaded relief image of gully 0 to show our definition of gully head area (blue region) and length (black line), (B) long profiles of the floor of the gully (black line in A) vs. unincised talus/breccia (red line in A), where, on average, gully channels tend to be more concave up than their surrounding unincised talus slopes, and (C) cross section of the gully channel (dashed line in A), highlighting its inner steep walls and U-shape. Figure S4: The location and ID of the individual gully alcoves. Figure S5: (A) Example of a line of individual boulders comprising an older levee at gully 16, and (B) corresponding topographic map, with 0.25 cm contours, showing that the boulders are visible within the LiDAR DEM. Please visit <https://doi.org/10.1130/GSAB.S.22362976> to access the supplemental material, and contact editing@geosociety.org with any questions.

floor is brecciated bedrock (where the bottom of the breccia lens is the boundary between the Coconino Formation and the underlying Supai formation), locally over 150 m thick, covered by ~10.5 m of bedrock debris, which is most likely fallout debris from the impact. Atop these postimpact sediments, the crater floor and lower crater walls are covered with Holocene and Pleistocene surficial deposits and breccia. Pleistocene breccia (derived from the impact and fault gouge) borders the lowest portions of the crater walls and is preserved where covered by rockfall derived from erosion of the steep and highly fractured upper crater walls (Fig. S2; Kumar et al., 2010). This talus mantle has been dissected, producing alluvium that interfingers a series of lake beds that are ~30 m thick near the center of the crater. Overlying Pleistocene sediments are ~1.8 m of Holocene alluvium, including the floors of minor stream courses (Shoemaker, 1987), rock levees (Kumar et al., 2010), and playa beds. The uppermost 30 cm of playa bed sediments were deposited above a thin ash layer that was deposited after an eruption from Sunset Crater ~900 yr ago in the nearby San Francisco volcanic field (Kring, 2007, 2017).

Current and Past Climate

The current climate at Meteor Crater is hot and arid. The Winslow station, which is located ~30 km east of Meteor Crater, reports annual rainfall of ~200 mm and mean annual snowfall of ~300 mm (<https://wrcc.dri.edu/>). The wettest month in the region is August, which averages ~35 mm of rain, and the driest month is June with ~5 mm. Over 35% of the yearly precipitation occurs in the fall months, and 15% occurs in the summer months. The maximum daily precipitation on record (over the past 80 yr) was ~99 mm. Based on the regional intensity-duration-frequency (IDF) curves (National Oceanic and Atmospheric Administration [NOAA], 2011), 15 to 30 min storm intensities, which are relevant for generating surface runoff and initiating mass flows in the region (Staley et al., 2020), range from ~8–11 mm (1 yr recurrence interval) to ~17–23 mm (decadal recurrence interval) to ~29–40 mm (100 yr recurrence interval) to ~45–60 mm (1000 yr recurrence interval). July tends to be the hottest month, with average highs of ~94 °F (34.4 °C), and January is the coldest, with lows of ~21 °F (−6.1 °C). The groundwater table is below the observable crater floor, which is dry except for minor ponding after rainstorms (Kumar et al., 2010). Roddy (1978) reported that the water level in the Meteor Crater well, located 1050 m north of the point of impact (35°2.188'N, 111°1.399'E), was 186 m below the ground surface (at 1500 m relative to mean

sea level [MSL]), putting the water table ~60 m below the current crater floor. Pilon et al. (1991) used ground-penetrating radar on both the interior floor of the crater (four transects) and across the ejecta (one transect) 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.

However, interpretations of the crater-floor sedimentology suggest that at the time of impact, wetter conditions may have prevailed. Based on the immediate formation of lake deposits postimpact 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 (1979) 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, thereby forcing water to flow upward into the crater, allowing for a lake to form without the crater floor intersecting the local groundwater table. Mollusk populations within the lake core samples indicate perennial waters, as opposed to fluctuating waters (i.e., bogs, swamps, drying muds; Reger and Batchelder, 1971), suggesting that the lake at Meteor Crater was not short-lived and required a sustained source of water.

Although other local paleoclimatic data are lacking, regional studies strongly indicate a late Pleistocene wet period followed by a warming and drying trend toward current conditions. Cole et al. (2013) collected 60 packrat middens across the southwestern United States that ranged from older than 48,000 yr B.P. to present, where well-preserved pine needles within the middens documented the geographic response of three types of pinyon pines to climate change. Their analysis suggested that in northwestern to central Arizona, during the full glacial Wisconsinan (23.4–14.7 ka), winter precipitation was at least 150% greater than current day. During the subsequent Bølling-Allerød interstadial (14.7–13 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. This was followed by rapid warming and precipitation decline at the start of the Holocene (11.7–9.0 ka), which led to the decline and northward retreat of these populations to their current-day distributions (Cole et al., 2013).

To the north of Meteor Crater in the Black Mesa basin in northeastern Arizona (see region in Fig. 1A), Zhu et al. (1998) and Zhu and Kipfer (2010) performed groundwater dating, numerical modeling, and noble gas analyses of the Navajo Sandstone aquifer. They proposed

that 14,000–17,000 yr ago, the recharge rates to groundwater were three times higher relative to today, and the water level was as much as 60 m higher relative to today. They suggested this pulse of high recharge was due to a northward migration of the southern branch of the split jet stream. At the transition to the Holocene, a shift toward the current drier and warmer conditions has been documented in various studies in the region (Cave of Bells area; Fig. 1A; e.g., Wagner et al., 2010), although the specific timing and magnitude of changes vary among studies (e.g., Ballenger et al., 2011).

METHODOLOGY

LiDAR Data Collection and Processing

On 12 March 2010, the National Center for Airborne Laser Mapping (NCALM) performed an airborne survey of Meteor Crater and its ejecta blanket. To obtain higher-resolution data of the gullies and their meter-scale deposits, two different point densities were used; data at a density of 8 points/m² were collected for the crater walls and rim, and data at a density of 4 points/m² were collected for the surrounding area. All data processing, including generation of the final data product used in this study, was performed by NCALM. All coordinates are relative to the North American Datum of 1983 (NAD 83; Continuously Operating Reference Stations [CORS] 96) reference frame, and no bare earth extraction was required since Meteor Crater 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 product was projected in Universal Transverse Mercator (UTM) zone 12N with units in meters, and heights are North American Vertical Datum of 1988 (NAVD 88) orthometric heights computed from Geodetic Reference System of 1980 (GRS 80) ellipsoid heights using the National Geodetic Survey (NGS) GEOID09 model. The elevation accuracy is 5–30 cm (1 σ). All data can be accessed through the National Science Foundation–supported OpenTopography portal.

Sediment Volume Calculations

Much of the topographic analysis presented herein (i.e., calculation of the volume of material eroded from talus and levee and lobate deposit volumes) relied on methods like those developed by Warner et al. (2011). To estimate the volume of sediment that has eroded

from the talus slopes that border the bedrock cliffs within the crater, we mapped regions of incision and then projected a smooth surface between the bounding sidewalls, essentially recreating the pre-eroded talus slope topography. This projected surface was then gridded and subtracted from the original gridded topography. To estimate the levee and terminal deposit volumes on the surface of the crater floor (i.e., nonburied deposits), 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 predeposit surface. We then took the difference between the gridded current-day deposit surface and the projected predeposit surface to estimate the volume of the deposit. To estimate the volume of sediment deposited on the crater floor as lake sediment, we used the contour at the maximum extent of the lake (at 1570 m [dark blue contour in Fig. 3], the top of “Silica Hill”; Kring, 2007) to estimate its areal extent and multiplied that by the average depth of the lake deposits based on drill cores (~30 m). To estimate the volume of Holocene alluvium

and playa deposits, we used the geologic map from Shoemaker and Kieffer (1974), which we georeferenced into ArcMap (Fig. S1), to map their areal extents and then multiplied those by 1.8 m, i.e., the depth reported in Shoemaker (1987). To estimate the volume of Pleistocene sediment on the crater floor, we used two methods. The first was to simply map the area these deposits currently occupy on the crater floor and walls (using the geologic map from Shoemaker and Kieffer, 1974) and multiply that by their average depths based on the cross section by Shoemaker and Kieffer (1974). The second method involved fitting second-order polynomials to the boundaries defining the interfaces between the breccia and Pleistocene talus/lake sediment/alluvium 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 revolved around the y axis and integrated along the x axis). As the “volume” above the breccia had Pleistocene talus, alluvium, and lake sediments, we estimated the percentage of each (Pleistocene talus ~7%, alluvium ~28%, and lake sediment ~65%) from the Shoemaker and

Kieffer (1974) cross section and then applied that to the total volume. These methods inherently assume the crater is a perfect bowl-shape and that deposition of each unit was uniform around the crater; as neither is the case, these are rough estimates of the volumes of each depositional unit.

Geomorphic Field Mapping and Surveys

The relative roles of debris-flow versus fluvial processes at Meteor Crater were determined from examination of surficial deposits and exposures across the crater, and detailed measurements were made in one individual gully system, referred to herein as “gully 0.” Criteria used to distinguish past debris-flow events included the presence of boulder-rich lobate snouts, debris-flow tracks (i.e., identified by a single or paired narrow train of boulders or boulders suspended in a fine-grained matrix), poorly sorted and matrix-supported sediments, and the absence of stratification, sorting, or rounding of clasts (e.g., Bull, 1977). Fluvial processes were identified by V-shaped channels with defined banks, sorting and rounding of clasts, clast imbrication, and the development of bed or

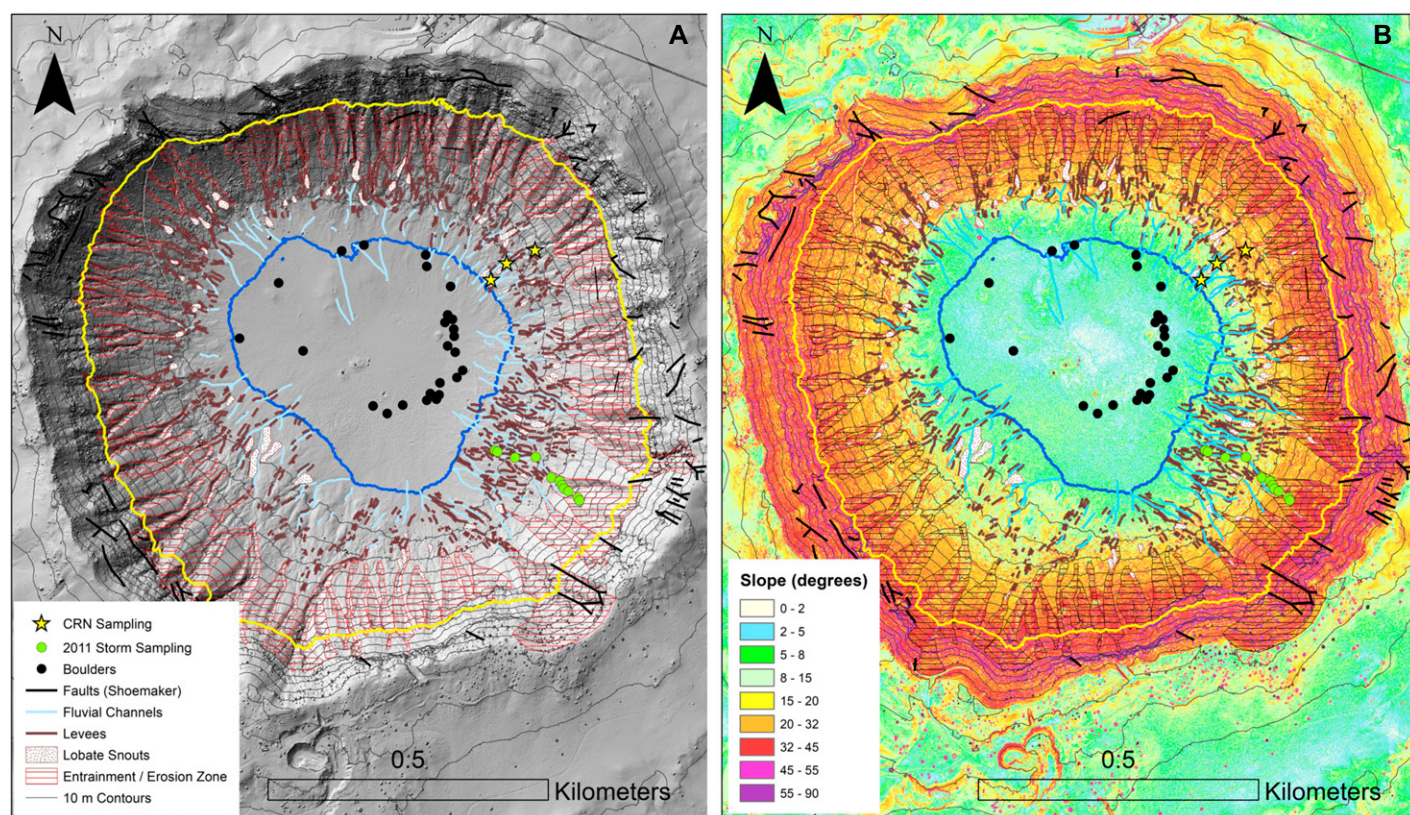


Figure 3. (A) Geomorphic map of Meteor Crater overlain on shaded relief image. Blue contour shows the highest mapped extent of the Pleistocene lake, and yellow contour shows the average elevation where gully incision begins. (B) Same geomorphic map as in panel A, except with a slopeshade image underlain to highlight the slope dependence of different erosional and depositional landforms. CRN—cosmogenic radionuclide.

bar forms within a channel. The spatial distributions of channels, boulder snouts, and levees and the crosscutting relationships among deposits were identified and mapped onto 25 cm contour data in the field, which informed subsequent mapping. Using a handheld global positioning system (GPS) and LiDAR data, we also identified the lowest position of boulders on the crater floor. Fluvial channel cross sections and debris-flow tracks were surveyed in the field; these surveys were used to estimate levee volumes for comparison with the volume estimates performed using the LiDAR data alone.

Following a monsoonal storm in early September 2011, we noted the occurrence of fluvial channel development on the floor of many gullies (i.e., where runoff incised V-shaped, slightly meandering channels, often with small gravel bars, on the floor of previous debris-flow tracks). Within “gully 16” (Fig. 3A), where a mostly continuous fluvial channel developed from the headwall to the crater floor, we collected data on bankfull channel dimensions, channel slope, and grain size at each of the points (green dots) marked in Figure 3A. We measured fluvial channel width (w) and bankfull depth (h) in the field, and the local slope was calculated over 10 m (5 m upstream and downstream of the channel cross section). The upslope drainage area from each of these sites was calculated using LiDAR data. For grain size, we conducted Wolman pebble counts (e.g., Wolman, 1954; Bunte and Abt, 2001) on bars that developed in the fluvial channels to determine the critical discharge for fluvial sediment motion and the source area runoff rate. To conduct these counts, we set up grids and measured the intermediate axis of the particle at each grid node, with the particle being chosen using the tip of a mechanical pencil, making sure to count at least 100 grains for the determination of the median grain size (D_{50}) and D_{84} (i.e., the grain size at which 84% of the grains are smaller) within tolerable limits (Wolman, 1954).

Fluvial Runoff and Sediment Transport Formulation

Instantaneous storm discharge (Q) and corresponding runoff rates (i.e., discharge divided by the upslope drainage area) for the September 2011 storm event were estimated at each channel cross section (green dots in Fig. 3A) using principles of continuity:

$$Q = a_{xs}U, \quad (1)$$

where a_{xs} is the flow cross-sectional area (estimated as the channel width, w , multiplied by the bankfull flow depth, h), and U is the average flow

velocity. The average flow velocity was calculated using the Darcy-Weisbach equation:

$$U = \sqrt{\frac{8ghS}{f}}, \quad (2)$$

where g is gravity (9.8 m/s²), S is the fluid surface slope, and f is the Darcy-Weisbach friction factor, a quantification of flow resistance. For channels with width-to-depth ratios less than 20, the hydraulic radius, R_h , which is the ratio of the channel area (wh) to its wetted perimeter ($2h + w$), was used instead of h in Equation 2. We assumed uniform flow such that the fluid surface slope equaled the local tangent of the bed slope angle, θ . Local bed slope was extracted from our LiDAR-derived digital elevation models (DEMs) from long profiles that extended 5 m upstream and 5 m downstream of each cross section. Ferguson (2007) derived an empirical expression for f that was intended to work for a range of flow conditions and channel slopes:

$$\sqrt{\frac{8}{f}} = 17.7 \frac{h}{D_{84}} \frac{1}{\sqrt{52.3 + 5.57 \left(\frac{h}{D_{84}} \right)^{5/3}}}. \quad (3)$$

In order to estimate modern-day sediment fluxes transported by fluvial processes from the crater wall to its floor, we used an empirical model by Schneider et al. (2015). This model was developed using field data from sites that covered a wide range of channel bed slopes, grain sizes, and water discharges, and it accounts for bed forms and other sources of macroscale roughness that can affect flow resistance and resulting sediment fluxes. This model was found to best predict sediment motion in a fine gravel-bed channel on an alluvial fan with slopes similar to those at Meteor Crater (Palucis et al., 2023). In this model, the nondimensional transport rate (W^*) is a function of both (1) the median grain size (D_{50}) of the coarse grain-size distribution (grains >4 mm) and (2) the ratio of the total dimensionless boundary shear stress on the bed (τ^*) to the reference shear stress (τ_r^*), where the reference shear stress τ_r^* is defined as

$$\tau_r^* = 0.56S^{0.5}, \quad (4)$$

and the boundary shear stress on the bed is

$$\tau^* = \frac{\tau_b}{(\rho_s - \rho)gD_{50}} = \frac{\rho g R_h S}{(\rho_s - \rho)gD_{50}}. \quad (5)$$

The nondimensional bed-load transport model is

$$W_{tot}^* = \begin{cases} 0.002(\tau^*/\tau_r^*)^{16.1} & \text{for } \tau^*/\tau_r^* < 1.2 \text{ and } D_{50} > 4 \text{ mm} \\ 14 \left(1 - \frac{0.85}{(\tau^*/\tau_r^*)^{0.7}} \right)^{4.5} & \text{for } \tau^*/\tau_r^* \geq 1.2 \text{ and } D_{50} > 4 \text{ mm} \end{cases} \quad (6)$$

We then converted dimensionless transport rates to volumetric bed-load transport rates (q_{vol}) per unit width (m³ s⁻¹ m⁻¹) using

$$q_{vol} = \frac{W_{tot}^* u_*^3}{\Re g}, \quad (7)$$

where $u_* = \sqrt{\tau_b/\rho}$ is the shear velocity, and $\Re$ is the submerged density of quartz (1.65).

¹⁰Be/²⁶Al Exposure Dating

To determine the timing of debris-flow activity at Meteor Crater (via levee emplacement), we collected four samples from levee deposits associated with gully 0, labeled MC-11-01 through MC-11-04 (Fig. 4). The samples were derived from sandy dolomite blocks (Kaibab Formation), which consisted chiefly of dolomite with various amounts of detrital and secondary quartz. They were chosen because they were expected to be the most resistant to weathering and have the most continuous exposure ages. Maximum sample depth from the exposed surfaces was 1.5 cm. All samples, except one (MC-11-04), were collected from relatively flat horizontal surfaces; MC-11-04 was collected on a vertical face to test for shielding effects on the same boulder as MC-11-03. The sampled deposits were located downslope from a weathering-resistant, thumb-shaped knob rising above the oldest Pleistocene talus near the base of the beta member of the Kaibab Formation; this knob was also sampled by Nishiizumi et al. (1991) (M-01, Fig. 4).

Prior to the impact, the ~250-m.y.-old Kaibab dolomite was shielded by ~8–10 m of Moenkopi sandstone, such that production rates of ¹⁰Be would be on the order of 1×10^{-6} atoms/g/yr (surface rates based on geographic latitude). At this production rate, any inherited ¹⁰Be would be ~0.1% of the total ¹⁰Be measured in each sample (Table 1). Once the impact occurred, the rocks we sampled were likely exposed to cosmogenic radiation, but presumably under different shielding conditions. When choosing our samples, we looked for rocks that appeared to be in place (i.e., had not recently been overturned or moved) and 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.

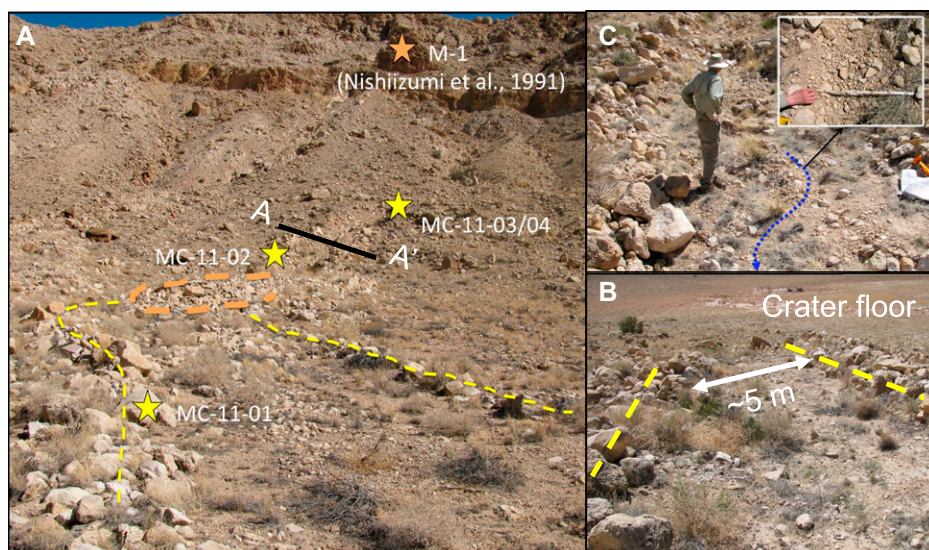


Figure 4. (A) Sampling locations (yellow stars) within gully 0 for ^{10}Be cosmogenic dating of coarse-grained levee deposits (yellow dashed lines). A lobate deposit is traced in orange. Cross section A–A' is shown in Figure 5B. (B) An ~5-m-wide, debris flow-created channel between debris-flow levees. (C) Location where grain-size data were collected in gully 0; the low-flow fluvial channel is highlighted with a dashed blue line, and the low-flow fluvial channel bed is shown in the inset panel.

TABLE 1. RESULTS FROM ^{10}Be AND ^{26}Al COSMOGENIC NUCLIDE DATING OF THE LEVEE DEPOSITS AT GULLY 1, METEOR CRATER, ARIZONA

Sample name	Elevation (m)	Mass (g)	$^{10}\text{Be}/\text{Be}$ (E-15)	$^{26}\text{Al}/\text{Al}$ (E-15)	Be-10 (atoms/g)	Al-26 (atoms/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	0.104 E + 06	0	7.42 ± 0.31

The samples were processed to obtain a pure quartz phase, and then the Be and Al were chemically separated and purified (Kohl and Nishi-

izumi, 1992). The ^{10}Be and ^{26}Al measurements were obtained using accelerator mass spectrometry (AMS) at the Purdue Rare Isotope Measure-

ment (PRIME) Laboratory. 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 atoms/g/yr of SiO_2 , respectively (Nishiizumi et al., 1989), and correcting for sample altitude, latitude, and exposure geometry (Lal and Peters, 1967). We used the geographic latitude to calculate exposure ages since the present geomagnetic latitude is not the same as it was during the last 50,000 yr. We estimated an overall uncertainty of about $\pm 10\%$ in the production rates of these nuclides (Nishiizumi et al., 1989). Exposure ages were 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 the present time but could possibly have been important in the past.

OBSERVATIONS

Gully 0: Model Gully System

Gully 0, located near the northeastern corner of the crater (Figs. 4 and 5; Fig. S3), was chosen to map in detail because of its well-preserved deposits (which we also sampled for cosmogenic radionuclide [CRN] dating) and its location under a prominent bedrock outcrop that was dated by Nishiizumi et al. (1991). The relief at this site is ~169 m (from the crater floor to its rim), of which the top ~76 m section is exposed bedrock with an average slope of 35° . Below

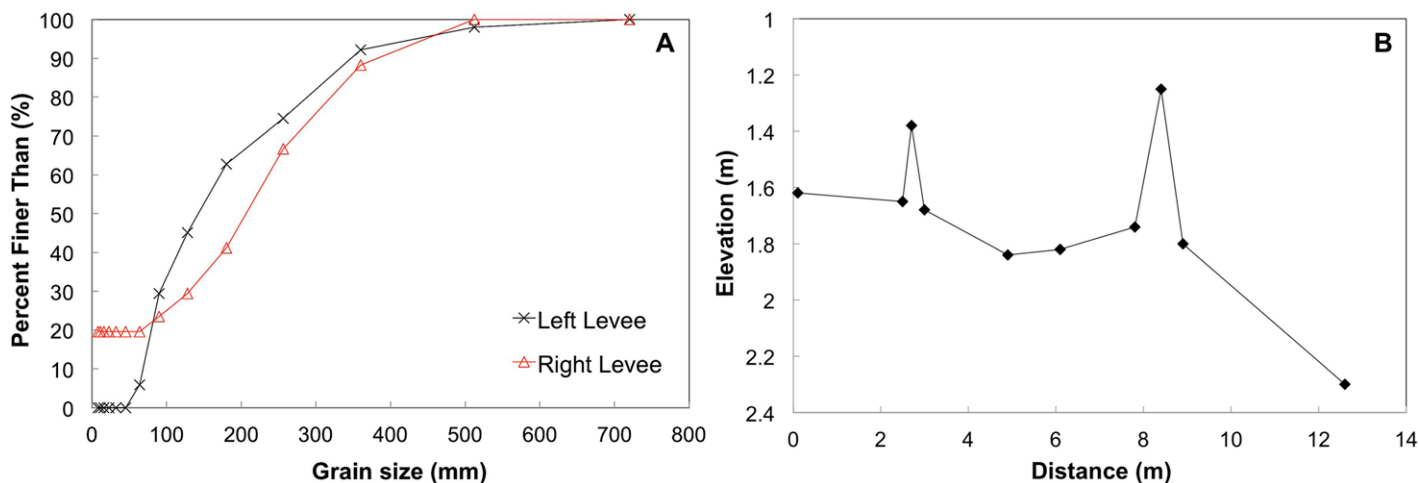


Figure 5. (A) Grain-size distribution of the levee deposits, located near MC-11-02 in gully 0. Right levee (looking downslope) distribution is shown in red, and left levee is shown in black. (B) Cross section, performed in the field, of a levee deposit in gully 0. Location of the cross section is shown in Figure 4A (A–A', black line).

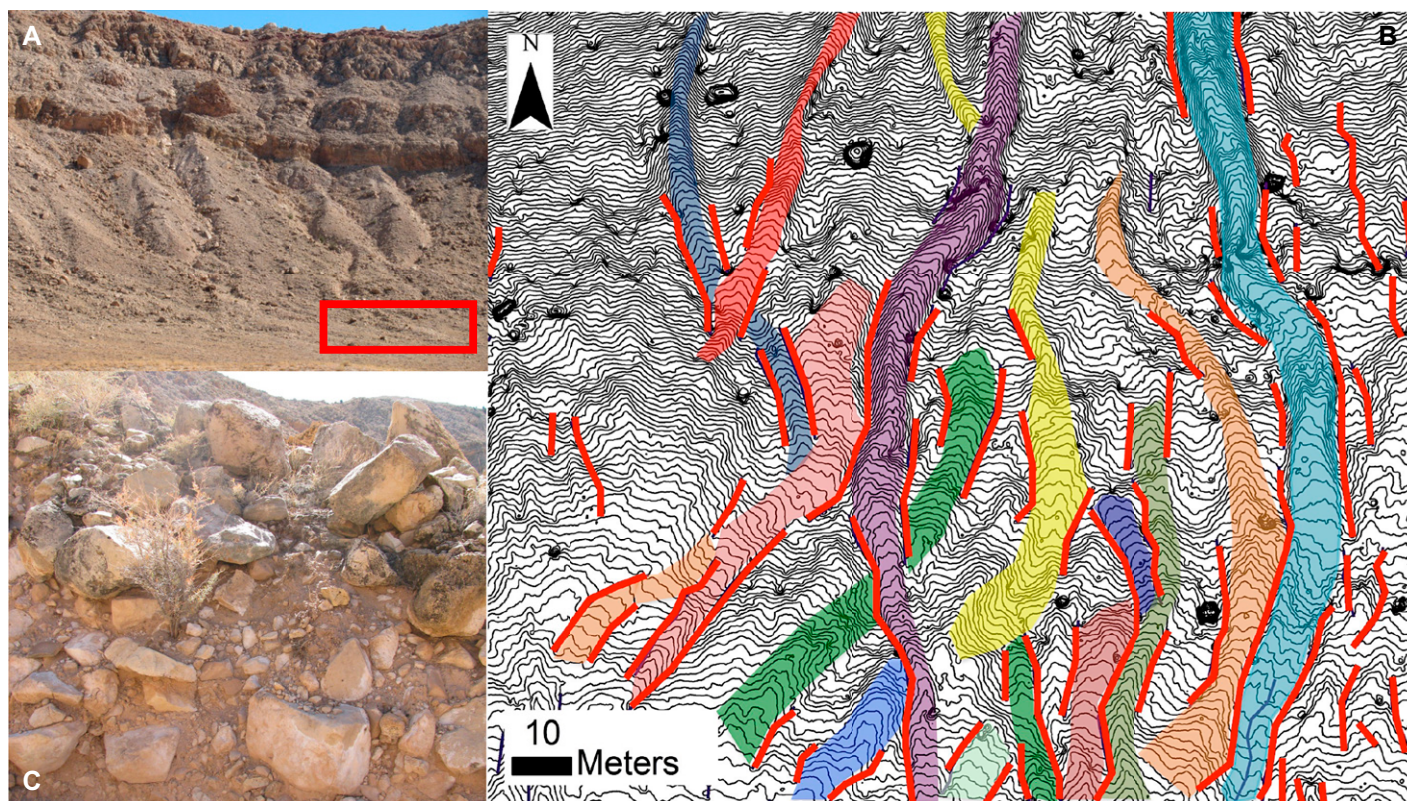


Figure 6. (A) Gully systems on north-northeastern side of the crater; red box shows mapping location in B. The crater wall is ~175 m tall. (B) Topographic map (0.5 m contours) of levee-lined channels (channels are shown in solid colors, and levee deposits are shown in red lines), where each channel is color-coded differently to show the sinuous, crossing, “noodle-like” nature of the debris-flow channels within Meteor Crater. The flow direction is approximately south. (C) Lateral deposit found in gully 0 showing poorly sorted, matrix-supported clasts indicative of mass-flow processes. The largest clasts are ~30 to 50 cm in length.

the bedrock, there is an ~53 m section of talus, composed mostly of impact breccia and rock-fall from the bedrock cliffs above. There is an amphitheater-shaped alcove (gully head) where the talus is incised that has steep (~33°–35°) sidewall slopes. The average depth of the alcove is ~3.7 m. Downslope of the talus, there is an undulating surface (~10 m in relief) with an average slope of ~7°, marked by crisscrossing boulder-laced, lobate snouts and debris-flow tracks (Fig. 5B). Some of the boulders within the levees of the debris-flow tracks showed evidence of having a calcium carbonate coating on them before being entrained in the flow, which allowed us to trace the source of these coated boulders to a local knickpoint in the gully. Exposures in the walls of the levees (Fig. 5C) show them to be predominantly matrix-supported, though careful digging into the deposits revealed isolated regions of stratified sand and gravel, suggesting minor fluvial deposition. The matrix was predominantly sandy, though finer-grained black material, possibly ash from eruptions in the nearby San Francisco volcanic field (Conway et al., 1998), was observed.

We surveyed an especially well-preserved pair of levees leading to a lobate deposit that likely defined a single debris-flow event; this event crossed a preexisting, visually older pair of levees created by another debris flow that had run farther out (see Fig. 4). In both flows, the levee deposits are ~5 m apart and ~0.4–0.7 m high (Fig. 6) and are detectable in the LiDAR data. The local depositional slope for both flows was ~5°. In addition to surficial evidence for debris flows, we observed an ~4-m-wide and 15-cm-deep channel with a gravel bar (D_{50} ~50 mm) between the levee deposits on the gully floor. The V-shaped contours of this channel were detectable in the 25 cm LiDAR data. Incised into this “high-flow” channel, there was a smaller “low-flow,” 0.4-m-wide channel (Fig. 4C); a pebble count on a bar in the smaller channel gave a median grain size of 22 mm. 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).

Kumar and Kring (2008) had proposed that groundwater seepage was the main source of

water driving gully incision in Meteor Crater. Our field observations at this gully system (and elsewhere across the crater) do not support this hypothesis. At the headwall of the gully, there was no evidence of staining or coarse gravel lag deposits, and no evidence for sustained fluvial wash in the channel. We also did not see any caves or large fractures from which groundwater would have discharged, though smaller fractures and joints with ~1–3 m spacing occur regularly in the Kaibab Formation due to uplift of the Colorado Plateau (Roddy, 1978; Shoemaker, 1987; Kring, 2007). We did observe channel development within the bedrock cliffs above the talus (white arrows, Fig. 2B), suggesting that during storm events, water from precipitation (rain or snowmelt) concentrates and flows over the steep bedrock cliffs onto the talus below, where it entrains sediment. In the neighboring nongullied talus, we did not observe evidence of landsliding processes (e.g., landslide scars, headscarps, rotational blocks, etc.), nor did we observe more than minor rilling within the talus, both suggestive that gully incision was predominately from runoff off bedrock and sediment entrainment.

Morphometric Analysis of Gully Systems at Meteor Crater

Using our observations from gully 0, as well as several neighboring gullies, we were able to correlate features on the ground (i.e., levees, small fluvial channels, and lobate debris-flow

deposits) with their topography on the LiDAR map. We characterized the morphology of 71 individual gully systems within Meteor Crater (Fig. S4) using a combination of field mapping and LiDAR analysis, as reported in Table 2. Almost all the gullies at Meteor Crater begin at approximately the same elevation around

the crater wall, ~1660 m (see yellow contour in Fig. 3; “Elevation of alcove top” in Table 2), which generally corresponds to the transition from the cliff-forming Kaibab limestone to the underlying talus slopes into which the gullies are incised. The gullies on the southern portion of the crater wall, however, begin at a slightly higher elevation (~1680 m) where uplift of the Kaibab limestone is higher (Kring, 2007). The water source areas contributing to the gullies ranged from 900 to 14,000 m², with an average of 2460 m². The individual alcoves, which result from the removal of breccia and talus, range from ~0.7 to 7.6 m in depth (average ~3 m) and ~500 to 95,000 m³ by volume (average ~12,000 m³). The gully channels are “U-shaped” in cross section and typically maintain a constant width, being on average ~7 m wide and ~1 m deep (based on five cross sections per gully channel). They are commonly found to either terminate abruptly without fans or end in a lobate deposit of debris (“lobate snouts” in Fig. 3A). Features suggesting a transition into standing water (i.e., deltaic deposits) were not observed. Incised into the gully channels, there are smaller fluvial channels, as was observed at gully 0. These channels tend to be straight, not sinuous, and are sometimes observed to flow around or through mapped lobate debris deposits.

Like the levee deposits found in gully 0, 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 (e.g., Fig. 5). Cross sections of 100 levees (obtained from LiDAR data) show that they tend to be less than 0.6 m high and ~2 m wide. 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 m. The volume of material comprising an individual debris-flow event was calculated by identifying levees linked to specific lobate snout deposits. We identified 20 well-defined flow events (i.e., a lobate snout connected to paired levees), and from these, we found flow volumes to range from 10 to 1000 m³, with an average of 200 m³.

In some cases, levee tracks extend onto the crater floor (Fig. 3A). These visibly older tracks are detectable at the resolution of the elevation model, but field examination showed these deposits to often consist of a single line of individual boulders, where boulders are defined as having a *b* axis greater than 256 mm (Fig. S5). In addition to boulder levees, we also mapped the location of individual boulders/large cobbles (average grain size of 250 mm) found along the

TABLE 2. MORPHOMETRIC PARAMETERS OF GULLIES WITHIN METEOR CRATER, ARIZONA

ORIG_FID	Head area (m ²)	Depth (m)	Volume (m ³)	Length (m)	Gully floor slope (°)	Elevation of alcove top (m)
0*	2759*	3.71*	10,230*	153*	279*	1658*
1	2968	3.60	10,700	116	25.6	1660
2	4517	3.45	16,894	104	26.1	1659
3	2141	3.16	6774	108	24.1	1650
4	1867	3.90	7276	56	34.3	1690
5	739	0.85	998	85	25.3	1635
6	2317	3.32	7701	89	26.1	1643
7	1914	2.68	5133	104	20.6	1632
8	1910	3.01	5742	102	25.1	1642
9	1274	2.99	3804	106	19.8	1633
10	3307	4.37	14,443	82	25.1	1641
11	2278	1.88	4281	121	28.3	1677
12	1945	0.96	1874	101	30.3	1661
13	916	2.40	1400	116	25.5	1650
14	4514	1.78	8021	192	22.9	1671
15	6329	4.67	27,562	123	21.6	1660
16†	9736†	3.70†	35,985†	193†	23.8†	1683†
17	3260	5.02	16,347	108	22.6	1641
18	4924	5.25	35,848	159	24.8	1670
19	14,087	6.74	94,900	237	19.4	1697
20	2741	2.88	7883	136	24.7	1659
21	1065	1.22	1294	110	23.2	1658
22	3028	1.94	5874	112	24.8	1647
23	4175	4.94	20,611	113	29.4	1670
24	2842	3.82	10,845	116	28.2	1665
25	1941	2.51	4871	99	24.2	1658
26	2023	4.78	9678	97	24.4	1659
27	1521	0.85	1290	163	27.8	1661
28	2210	1.01	13,003	112	26.9	1658
29	4781	2.72	13,003	158	23.4	1661
30	14,404	6.23	92,637	220	23.6	1662
31	7578	7.39	37,563	171	25	1663
32	5072	7.50	38,018	121	29.1	1670
33	4981	7.64	76,113	151	24.6	1672
34	1576	2.61	4116	68	20.1	1619
35	6887	2.47	17,005	150	24.5	1662
36	3608	6.84	24,697	123	21.8	1657
37	3337	1.73	5770	156	24.8	1672
38	4980	3.57	17,762	132	22.9	1658
39	726	0.67	489	99	26.8	1660
40	2130	3.25	6917	103	23.7	1658
41	3780	4.49	14,864	133	25.7	1668
42	1409	2.60	4200	70	29.5	1640
43	1146	1.16	1326	93	33.1	1661
44	563	1.33	751	54	31.6	1656
45	1614	1.31	2112	92	26.2	1654
46	1524	3.24	4932	92	29.8	1659
47	1842	2.94	5413	101	31.1	1658
49	903	1.34	1210	87	27.3	1653
50	4405	2.73	11,111	153	26	1653
51	2942	2.81	8262	162	18.5	1654
52	2319	6.19	14,344	90	25.5	1661
53	2857	4.98	14,229	102	25.9	1648
54	1331	1.89	2510	93	28.3	1660
55	1899	0.74	1407	122	24.7	1676
56	3147	4.20	11,011	101	31.5	1657
57	1376	2.22	3055	78	27.5	1648
58	1213	3.02	3661	74	22	1628
59	2502	3.42	8552	130	22.4	1660
60	1340	1.72	2308	83	29.9	1651
61	917	1.50	1372	46	27.5	1643
62	1203	2.17	2615	56	32.7	1660
63	2786	5.27	10,679	86	29.6	1660
64	1381	0.98	1357	98	27.8	1659
65	566	1.55	874	49	30.5	1630
66	200	2.52	800	62	28	1645
67	322	2.52	810	85	26.8	1668
68	693	3.31	2293	59	35.7	1685
69	640	0.95	609	78	24.9	1645
70	975	2.24	2184	69	33.7	1669
71	581	1.17	679	41	30.2	1633

*Cosmogenic radionuclide (CRN) sampling and mapping.

†September 2011 storm survey site.

floor of the crater (black dots, Fig. 3A). 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 ~ 7 m below the highest mapped lake sediments (i.e., Silica Hill) on the crater floor (Kring, 2007).

In Figures 3A and 3B, we show our geomorphic map, as well as a slope map generated from LiDAR data. Slopes steeper than $\sim 32^\circ$ are either in bedrock or lining the channels cutting through the talus, while the talus itself lies almost entirely on slopes between 20° and 32° . The debris-flow levees crisscross slopes down to $\sim 5^\circ$ – 8° , with localized boulder snouts sometimes found stopped in the tracks of previous leveed debris flows (as was the case for gully 0). The boulder snouts typically were deposited on slopes above 8° , and most are concentrated between 8° and 15° . 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 – 10 m in elevation above the highest mapped lake level (dark blue line, Fig. 3A). The lowest-elevation boulders we mapped were found on slopes of 2° – 5° ; if they came down as debris flows, the fine-grained matrix has likely been removed by eolian erosion. Fluvial channels extend past the lobate snout and levee deposits onto slopes less than $\sim 5^\circ$, and some of them extend all the way to the center of the crater (slopes $< 2^\circ$).

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 and intensity-duration patterns

of precipitation that produced runoff inside the crater are not known, as it is not gauged around its rim. However, we do have daily precipitation data from the nearby NOAA Meteor Crater station (35.036°N , 111.023°W). Here, a maximum of 22 mm fell over 24 h, and 85 mm in total fell over 10 d (Fig. 7A). The nearby Winslow airport (~ 35 km to the east of Meteor Crater) has a daily average rainfall record for the past 109 yr, as well as the daily record maximums, which have been as high as ~ 100 mm. After the storm (~ 1 wk after final precipitation event), 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. Rilling and sheetwash were evident across the gully floors of Meteor Crater, as well as on the outer crater rim. For the gully system we surveyed (i.e., gully 16;

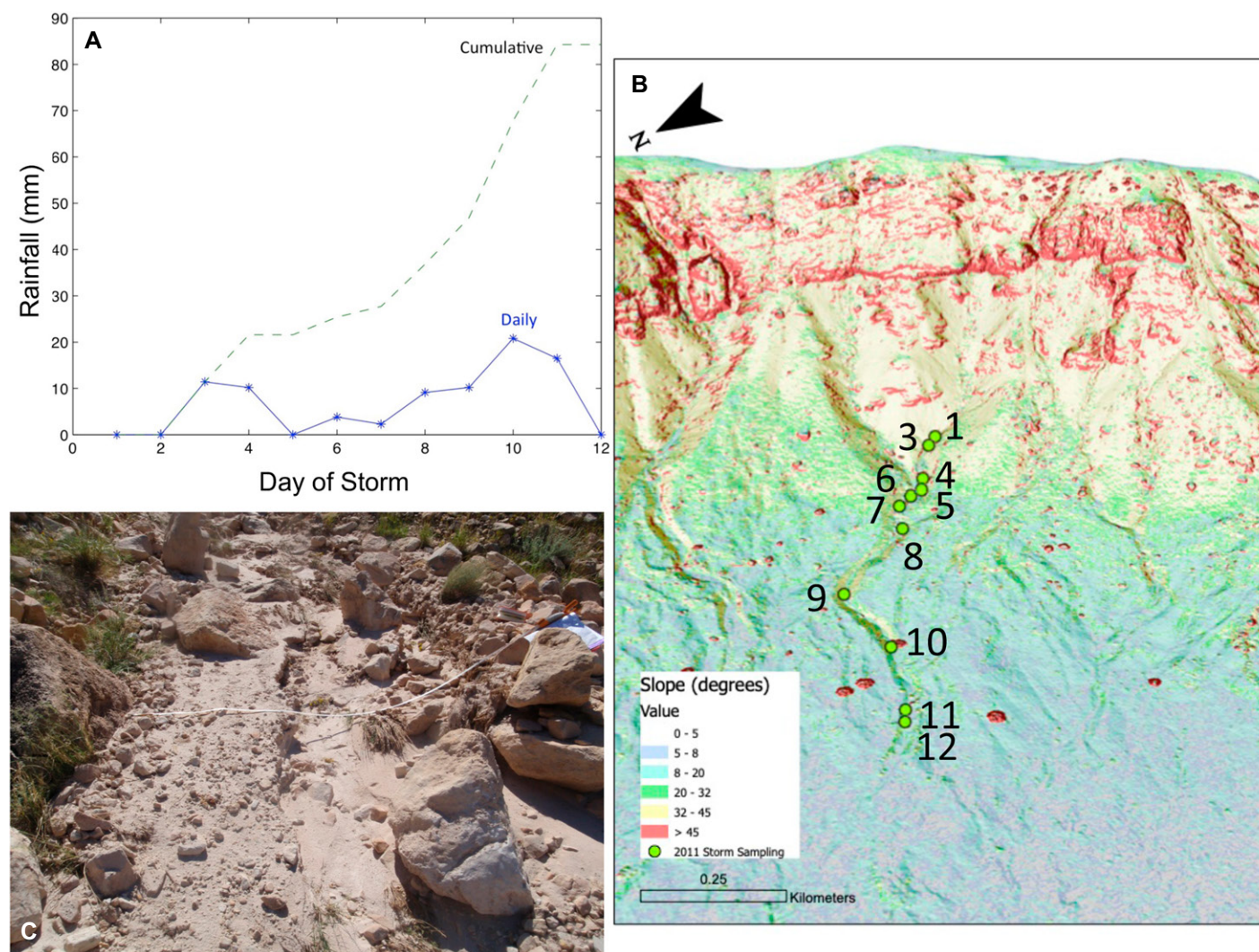


Figure 7. (A) Precipitation data from the National Oceanic and Atmospheric Administration (NOAA) Meteor Crater gauge for the September 2011 storm. Cumulative rainfall amount was ~ 85 mm, while the maximum daily rainfall was ~ 22 mm. (B) Sampling locations within gully 16 after 2011 storm event. (C) Fluvial erosion that occurred near site 10.

TABLE 3. DEPTHS, AREAS, AND VOLUMES OF FEATURES AND DEPOSITS WITHIN METEOR CRATER, ARIZONA

Feature/unit	Surface area (km ²)	Source	Depth (average) (m)	Source	Volume (m ³)	Source	Notes
Depositional							
Breccia	N.D.	N.D.	175	Shaft data; Kring (2007)	N.D.	N.D.	Deposited immediately postimpact
Pleistocene lake	0.15	1570 m contour (Silica Hills deposit)	20	Merrill (1908); Hager (1953)	3.0E + 06	Calculated	Highest level of lake sediments observed in Silica Hills
Pleistocene alluvium	0.34	Shoemaker (1960); geologic map	20	Shoemaker (1960); cross section	6.8E + 06	Calculated	6.0E + 06 m ³ estimated to be debris-flow deposits
Pleistocene talus	0.23	Shoemaker (1960); geologic map	15	Shoemaker (1960); cross section	3.5E + 06	Calculated	
Holocene alluvium	0.21	Shoemaker (1960); geologic map	1.8	Shoemaker (1987)	3.8E + 05	Calculated	
Holocene playa	0.026	Shoemaker (1960); geologic map	1.8	Shoemaker (1987)	4.7E + 04	Calculated	Sediments are composed of pink eolian silt that blows in from outside the crater
Meteoritic material	N.D.	N.D.	N.D.	N.D.	1.0E + 04	Rinehart (1958)	Within alluvium, lag, and colluvium
Erosional							
Gully erosion	0.2	This study	3	This study	8.6E + 05	This study	
Bedrock walls	0.35	Shoemaker (1960); geologic map	9.3	Based on erosion rates from Nishiizumi et al. (1991)	3.2E + 06	Calculated	
				N.D.	5.3E + 06	Calculated	Corrected for changes in porosity to compare with talus/deposits
N.D.	N.D.	N.D.	N.D.	Shoemaker and Kieffer (1974)	7.0E + 06	Calculated	
				N.D.	1.2E + 07	Calculated	Corrected for changes in porosity to compare with talus/deposits

Note: N.D.—no data.

survey locations shown in Fig. 7B), the upslope drainage area at the top of the gully (where the incised talus encounters exposed bedrock) is 3867 m², and the main gully channel is on average ~8 m wide and ~200 m long. There was no clear pattern in changes to the fluvial channel depth going downslope, but the fluvial channel width increased from ~0.4 m (slopes of ~25°) to 1 m (slopes of ~5°). It was also clear that the fluvial channel path was greatly affected by the vegetation and microscale 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 survey 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 fluvial channels, like the gullies themselves, was runoff from the exposed bedrock slopes above the gully (i.e., via Horton overland flow).

ANALYSIS

Timing, Frequency, and Magnitude of Debris-Flow Events

Samples from debris-flow levees were collected for CRN dating from several of the best-preserved (and presumably most recent) mass-flow events within the crater. The exposure age obtained on the stratigraphically older flow (sample MC-11-01), based on crosscutting relationships, was found to be 13,910 ± 670 yr (late

Pleistocene; Table 1). The exposure ages taken from the stratigraphically younger flow (MC-11-02 and MC-11-03) were 18,600 ± 1230 yr (late Pleistocene) and 10,050 ± 380 yr (early Holocene), respectively (Table 1). The oldest exposure age, which was found in the younger flow levee, is suggestive that the boulder we dated was exposed at the surface prior to its entrainment and transport within this flow event.

Assuming debris-flow activity ended by the late Pleistocene to early Holocene, we estimated the average frequency of debris-flow events at Meteor Crater by taking the total volume of material deposited by debris-flow activity and dividing it by the average volume of a debris-flow event (based on the size of surficial deposits) and the time over which debris flows were active within the crater. This assumes one flow event occurs at a time within the crater, but it is possible multiple gullies could be activated at once; our estimate is thus the highest expected frequency. From our geomorphic mapping and slope map (Fig. 3), we assumed that any material deposited between the lowest elevation where debris-flow levees were observed on the crater floor (~5°), which generally coincides with the elevation of the highest mapped lake deposits, and the transition to talus (20°) was from debris flows. Material deposited below ~5° (except for the large, isolated boulders on the crater floor) was assumed to have been emplaced via fluvial processes (and therefore was not included in our estimate). We calculated that there is ~6.0 × 10⁶ m³ of debris flow–derived sediment stored on

the crater floor (Table 3), which would equate to ~30,000 debris-flow events, assuming the average event size is ~200 m³ (as discussed above). Based on our CRN dating, we assumed that debris-flow activity was active from the time of impact to the late Pleistocene, which is ~40,000 yr. This equates to ~1 debris-flow event every 1.5 yr (i.e., 6.0 × 10⁶ m³ debris flow–derived sediment ÷ 200 m³ per debris-flow event ÷ 40,000 yr with ~0.75 events/yr or 1 event every 1.5 yr) at Meteor Crater.

Timing, Frequency, and Magnitude of Fluvial Events

Based on our field data from the September 2011 storm event at Meteor Crater (Table 4), we estimated that the instantaneous water discharge increased from ~0.03 m³/s near the head of gully 16 to ~0.12 m³/s near the floor of the crater. This corresponds to peak runoff rates of ~16–21 mm/h down to ~6–10 mm/h (average ~14 mm/h). Due to the small size of the gully, its steepness, and the relatively impermeable bedrock over which water concentrates, we assumed minor losses to infiltration and evaporation and that the time of concentration is much less than the storm duration (<60 min). As such, we assumed that the system reaches a steady state and that we can directly compare rainfall rates to runoff rates. Based on IDF curves for the region (NOAA, 2011), storms capable of delivering 14 mm of rain in 1 h occur approximately every 1–2 yr, and storms capable

TABLE 4. CHANNEL PARAMETERS AND RUNOFF ESTIMATES

Site number	Drainage area from sample site (m ²)	Bedrock drainage area (m ²)	Slope (°)	Depth (m)	Width (m)	Width/depth (m)	Hydraulic radius (m)	D ₅₀ (mm)	D ₈₄ (mm)	H/D ₈₄ (m)	Velocity (m/s)	Q storm (m ³ /s)	Runoff storm (mm/h)	Schneider model tau* (m/s)	tau*/tau* _{cr} (m/s)	W* _{tot} (m ³ /s)	W (m ³ /h)
September 2011 poststorm (gully 52)																	
1	6475	3867	0.4	0.06	0.46	8	0.05	25	51	1.18	1.17	0.032	179	0.46	1.3	0.06	0.00028
3	7575	N.D.	0.475	0.065	0.4	6	0.05	30	54	1.20	1.32	0.034	16.3	0.47	1.2	0.03	0.00022
4	8450	N.D.	0.3	0.06	0.6	10	0.05	21	37	1.63	1.38	0.050	21.1	0.43	1.4	0.10	0.00034
5	17225	N.D.	0.35	0.065	0.55	8	0.05	22	42	1.56	1.46	0.052	10.9	0.51	1.5	0.16	0.00075
6	17800	N.D.	0.225	0.07	0.75	11	0.06	25	50	1.40	1.14	0.060	12.1	0.32	1.2	0.03	0.00009
7	18,775	N.D.	0.325	0.08	0.68	9	0.06	25	53	1.52	1.54	0.084	16.0	0.51	1.6	0.20	0.00114
8	19,400	N.D.	0.225	0.075	0.7	9	0.06	27	46	1.63	1.33	0.070	12.9	0.31	1.2	0.02	0.00007
9	24,650	N.D.	0.175	0.08	0.9	11	0.07	17	34	2.35	1.64	0.118	17.2	0.42	1.8	0.34	0.00084
10	28,675	N.D.	0.1	0.06	1.1	18	0.05	15	29	2.11	1.01	0.067	8.4	0.22	1.2	0.04	0.00003
11	31,900	N.D.	0.1	0.06	0.95	16	0.05	14	25	2.38	1.11	0.063	7.1	0.23	1.3	0.06	0.00004
12	35,775	N.D.	0.125	0.07	1.1	16	0.06	17	27	2.57	1.41	0.109	10.9	0.28	1.4	0.09	0.00012
Pre-May 2011 channel																	
Gully 67	10,783	4700	0.14	0.15	4	27	0.14	50	870	1.72	1.64	0.986	329.1	0.24	0.210	0.02	0.00008
Prancevic model for debris-flow initiation																	
	2460	N.D.	0.36	0.056	7	125	0.06	31	65.0	0.86	0.89	0.350	512.8	N.D.	N.D.	N.D.	N.D.
	2460	N.D.	0.62	0.0099	7	707	0.01	31	65.0	0.15	0.09	0.006	9.2	N.D.	N.D.	N.D.	N.D.

Note: D₅₀—median grain size; D₈₄—grain size at which 84% of the grains are smaller; H—depth; Q—storm discharge (asterisk indicates non-dimensional value); cr—critical; W*_{tot}—nondimensional bed-load transport rate; W—bed-load (sediment) transport rate; N.D.—no data. Prancevic model is from Prancevic et al. (2014). Schneider model is from Schneider et al. (2015).

of delivering 21 mm of rain in 1 h occur every 2–5 yr. If infiltration and evaporation reduce runoff by 50% (such that ~28 mm of rain is required to get an average of ~14 mm of runoff), storms of this magnitude occur approximately every 10 yr. To estimate the volume of sediment transported in a yearly to decadal storm event at Meteor Crater, we compared the nondimensional bed shear stresses during the storm (τ^*) to the critical Shields stress (τ^*_{cr}), i.e., the threshold at which sediment begins to move. When the ratio $\tau^*/\tau^*_{cr} > 1$, sediment transport theoretically occurs. Using the Schneider et al. (2015) model, we found that τ^*/τ^*_{cr} ranged from 1.2 to 1.8 at all sites, which translates to sediment fluxes between 0.1 and 2.8 m³/h (average ~0.9 m³/h). If similar transport rates occurred in every gully around the crater, then a storm event of this magnitude would transport ~66 m³/h of sediment. Assuming all the Holocene alluvium deposited on the crater floor (Table 3) is fluvial in origin (as it has been deposited on slopes <5°), then ~5700 transport events likely occurred over the last ~10,000 yr (or 1 event every 2 yr on average). Larger fluvial runoff events have occurred at Meteor Crater, based on the 4-m-wide fluvial channel observed in gully 0 that formed prior to the September 2011 storm event.

Sediment and Water Budget

Our field observations and CRN dating results suggest that debris-flow activity was mostly confined to the Pleistocene (with events as often as every 1 to 2 yr), whereas in the Holocene, most sediment has likely been transported via fluvial events. In terms of a sediment budget, the sediment available for transport within Meteor Crater is largely postimpact material (i.e., the breccia lens and postimpact talus) and debris from erosion of the rim (see fig. 6.4 in Kring, 2007). Nishiizumi et al. (1991) estimated rim retreat rates of ~9 m on the west side of the crater using CRN (i.e., 30 cm/1000 yr for the first ~27,000 yr and 5 cm/1000 yr in the last ~23,000 yr), while Shoemaker and Kieffer (1974) estimated ~12 m of erosion on the NE crater flank, which they assumed was a minimum number, and so they estimated that closer to a total of 15–20 m of erosion has occurred. Taking this range of estimates into account and applying them to the inner crater bedrock wall area (~0.35 km²) results in ~5.3 × 10⁶ to 1.2 × 10⁷ m³ of sediment delivered to the crater walls and floors (Table 3). Based on our estimates of the volume of material deposited on the crater floor as lake sediment (~3 × 10⁶ m³) and Pleistocene alluvium (~6.8 × 10⁶ m³), and assuming negligible material has been blown in from outside the crater, this would argue for rim

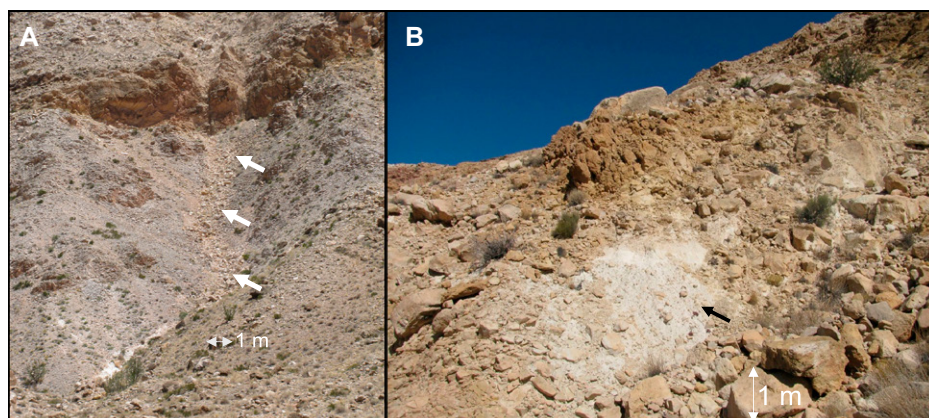


Figure 8. (A) Debris from upslope rockfall filling the floor of a large gully located on the southern wall of the crater. (B) White fine-grained material supporting a coarser gravel layer within the wall of gully 0.

erosion rates closer to those suggested by Shoemaker and Kieffer (1974), or considerable erosion into the impact-generated breccia and talus. However, we calculated that the volume of material removed from areas mapped by Shoemaker (1960) as Pleistocene talus to form the gullies we observe today is $\sim 8.6 \times 10^5 \text{ m}^3$. This accounts for only $\sim 13\%$ of the material mapped as Pleistocene alluvium by Shoemaker (1960), which we suggest has been emplaced by mostly debris-flow processes on the crater floor, and only $\sim 9\%$ of the total sediment on the crater floor. This suggests that once formed, these gullies have repeatedly been filled with sediment from rockfall and erosion of the upslope bedrock and then scoured by debris-flow and fluvial events. Evidence of this infilling process is observed today, where we observed $\sim 1 \text{ m}$ of infill in the gully channel in several gullies on the eastern wall (Fig. 8A).

Similar to calculations done on Mars (e.g., de Haas et al., 2015a), we can use our volumes of fluvial- and debris flow-deposited sediment on the lower crater slopes and crater floor to estimate minimum water amounts that were needed to transport these sediments (i.e., not accounting for infiltration and evaporation). We first determined the minimum amount of water needed to transport sediment as a debris-flow event for comparison with modern precipitation inputs to see if drier conditions in the Holocene can explain the lack of mass-flow events. We took our estimate of the volume of sediment within the terminal snout and levees at gully 0 (283 m^3) and then assumed a rock-to-water ratio (by volume) of 0.3 based on experimental work by Kaitna et al. (2016), who showed that this ratio is the transition from debris-flow to fluvial behavior. This requires at least 85 m^3 of water to initiate and transport sediment from the crater wall to the floor (assuming no losses to infiltration and

evaporation). The water source area for gully 0 is $\sim 4700 \text{ m}^2$, which equates to 18 mm of runoff to generate a flow. If we use a rock-to-water ratio of 0.3 and apply it to the entire volume of debris-flow deposits on the crater floor, then it would have taken a total of $\sim 2 \times 10^6 \text{ m}^3$ of water to transport the observed sediment, or $\sim 6 \text{ m}$ of runoff (for a bedrock area of 0.35 km^2). Using rainfall data from the Winslow Airport (Arizona) (wrcc.dri.edu), a total of almost 20 m of rainfall has fallen over the past 109 yr of record, and daily records show that moderate storms at the crater routinely deliver $>18 \text{ mm}$ of precipitation. This would suggest that even under Holocene climate, enough *total* water for debris-flow transport has been delivered to the crater. However, over the past 100 yr 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 (Fairchild, 1907; Kring, 2007). Furthermore, our CRN ages of the youngest boulders and those reported by Nishizumi et al. (1991) date to the late Pleistocene or early Holocene. This implies that (1) much more water than just that needed to fill the pore space of a pile of sediment is needed for debris flows to occur at Meteor Crater and (2) the rate of water delivery and ability to generate runoff need to be considered. IDF curves for the region (NOAA, 2011) suggest that rainfall can be delivered relatively rapidly, with $\sim 18 \text{ mm}$ of rainfall falling in under 30 min at a frequency of every 5–10 yr. If infiltration and evaporation use 50% of the water needed for initiation (such that a total of 36 mm of rain is required to get 18 mm of runoff), storms of this intensity lasting $<30 \text{ min}$ occur every 50–75 yr. Either much higher intensities are needed than can be achieved by the current climate, or the ability to generate sufficient runoff has declined (as will be discussed below).

In terms of sediment delivery by fluvial processes, we can also take a rock-to-water approach. We first assumed that most of the sediment deposited within the Pleistocene-aged lake was transported fluvially (though we acknowledge that some fraction may have been blown into the crater, similar to the modern-day playa deposits). If we take our estimated volume of lake sediment and use a rock-to-water ratio of 0.0002 (based on a survey of desert rivers by Pepin et al., 2010), it would have required $\sim 15 \text{ km}^3$ of water to transport the observed sediment, or $\sim 13.6 \text{ km}$ of runoff (for a crater area of 1.1 km^2). Applying modern precipitation rates ($\sim 20 \text{ m}$ per 109 yr) over the possible duration of the lake during the Pleistocene ($\sim 40,000 \text{ yr}$), this would equate to only $\sim 7 \text{ km}$ of runoff, or half of the water needed. This would imply that during the Pleistocene, the Meteor Crater region would have been at least $\sim 2\times$ wetter to transport all the crater lake sediments.

DISCUSSION

Controls on Debris-Flow and Fluvial Processes at Meteor Crater

A key motivation for investigating the deposits and erosional features at Meteor Crater was to quantify the role of fluvial versus debris-flow processes in the crater's evolution, and how (and whether) the occurrence and frequency of these processes are functions of climate change. As discussed above, prior work by Kumar et al. (2010), following Shoemaker and Kieffer (1974) and Grant (1999), suggested that postimpact, fluvial processes from seepage eroded the crater wall and subsequently deposited the material now mapped as Pleistocene talus. They hypothesized that further incision of the talus was due to a later period of enhanced erosion, perhaps via rainfall, and pointed to their mapping of a branched network of channels that originate on talus (see fig. 7 in Shoemaker and Kieffer, 1974) as evidence for this second erosion event. We propose an alternative erosion history, where postimpact (immediately after which there was breccia and talus emplacement) runoff over upslope bare bedrock (from rainfall or snowmelt) entrained sediment and incised the gullies predominately via debris-flow processes. Based on our field and LiDAR mapping, the branched network of channels on the talus interpreted by Kumar et al. (2010) to be fluvial in origin is instead from debris-flow tracks that formed as debris flows were deposited/terminated (and fluvial channels form within these tracks and can transport sediment further out onto the crater floor). Debris flows are self-bounding, such that they create levees as coarse debris is advected laterally (e.g., Whipple and Dunne, 1992), and net accumulation of sediment occurs

as successive flows cross and switch canyons (Fig. 5). As such, a second period of enhanced precipitation is not necessarily required in our model. When water inputs are not sufficient to generate a debris flow at Meteor Crater, as was observed during the September 2011 event, fluvial activity, also driven by overland flow processes, leads to minor modification of the gully floor and the transport of finer sediment within the debris-flow tracks. CRN dating of the visually youngest pairs of debris-flow levees in the crater suggests that debris-flow activity occurred mostly in the Pleistocene, while the Holocene has seen mainly fluvial activity. We discuss the possible sources and roles of water, sediment supply, and crater evolution on controlling this process transition in the following.

Both debris-flow and fluvial processes, then, appear to be driven by surface runoff, where water is concentrated on the exposed bedrock walls during storms (or perhaps snowmelt events in the past) and is delivered to the talus slopes below. However, due to the highly fractured nature of the Kaibab plateau surrounding Meteor Crater (Kumar and Kring, 2008), fractures may act as conduits for transferring water from the surface to the subsurface. Huntoon (2000) showed that these fractures are so efficient at dewatering the Kaibab plateau in the vicinity of the Grand Canyon that there are no perennial streams in that region. It is therefore possible that a secondary source of water into the crater (after Horton overland flow) is from infiltrated surface water (Kumar et al., 2010). However, while water can rapidly drain downward through joints in the Moenkopi and Kaibab Formations, when it hits the less permeable Toroweap Formation, it will likely flow radially away from the crater, as the Toroweap surface is uplifted and tilted back from the crater wall. Also, the lack of staining and/or gravel lag at the head of each gully and the fact that erosion from seepage has only ever been validated in cohesionless sediment (Howard and McLane, 1988; Schumm et al., 1995), or very weakly cemented sandstones (Lamb et al., 2006), support the interpretation that this is likely not the primary source of water leading to gully formation.

In regard to the amount of runoff delivered to Meteor Crater, our simple water and sediment budget suggests that perhaps twice as much runoff was needed in the Pleistocene (compared to contemporary amounts) to deliver fine sediment from the crater walls into the lake. Furthermore, sufficient runoff and/or groundwater inputs would have been needed to support a sustained lake. For sediment delivery by mass flows, however, the rock-to-water approach, which is commonly used on Mars, would suggest relatively little water is needed and that modern-day

precipitation and runoff are sufficient to trigger debris flows. This method only considers the total volume of water entrained within a debris flow, however, and it does not consider the rainfall-runoff conditions needed to initiate a flow. Other mass-flow events observed in this region typically initiate under high-intensity rainfall that occurs over 15 to 30 min intervals (Staley et al., 2020). Unfortunately, paleoenvironmental data from the late Pleistocene and early Holocene in the region cannot constrain precipitation or runoff rates, especially short-duration, high-intensity events, but evidence generally points to a cooler, wetter period, with precipitation amounts being ~120% to 150% greater than current day (Cole et al., 2013). At a broader spatial scale, modeling by Matsubara and Howard (2009) for the western United States showed that runoff depths in the late Pleistocene would have ranged ~1.7–4.1 × modern values to reproduce paleolake basin distributions (e.g., Lakes Bonneville, Lahontan, and Manly). It is not clear, however, whether the cooler and wetter climate from the late Pleistocene would have had more frequent high-intensity rainfall events than the current climate, which is dry but subject to monsoons. While more monitoring and modeling are needed to determine the rainfall intensities that generate runoff at Meteor Crater, there is value in understanding the possible mechanisms by which mass flows initiate within the crater, as it provides clues on why debris flows ceased ~10,000 yr ago.

Mechanisms of Debris-Flow Initiation Processes

Here, we discuss in more detail several mechanisms proposed at other study sites for how concentrated runoff can lead to debris-flow initiation, downslope entrainment, and deposition.

One mechanism proposes failure of a fluvial channel bed by surface runoff. Takahashi (1978) put forth a quantitative model that assumes that failure of a sediment bed with slope-parallel seepage occurs when surface runoff exerts stresses that can exceed the resistance of the granular bed. In addition, if the bed failure depth is equal to or greater than 0.5 times the flow depth, then sediment can disperse uniformly through the flow and move as a debris flow. Prancevic et al. (2014) recast the Takahashi (1978) model in terms of critical Shields stress for debris-flow initiation ($\tau_{c,df}^*$):

$$\tau_{c,df}^* = \frac{\rho h S}{(\rho_s - \rho) D_{50}} = (1 - \eta)(\tan \phi - S) - \frac{\rho}{\rho_s - \rho} S, \quad (8)$$

where η is porosity, and ϕ is the internal friction angle of the sediment. For the range of slopes (S) where debris-flow entrainment likely occurred at Meteor Crater, ~20° to ~32°, the experimental results from Prancevic et al. (2014) suggest that the critical Shields stress for debris-flow initiation would range from 0.4 down to 0.12 (for $\phi = 55^\circ$), respectively. Rearranging Equation 8 and using a D_{50} of 31 mm (based on our pebble count of the gully floor in gully 16 near site 1), we calculated that the runoff depth needed to cause failure of the bed is $h \sim 10$ –57 mm. We used Equations 2 and 3 to estimate mean velocity, but as the Ferguson (2007) formulation (Eq. 3) has only been validated up to slopes of ~20°, and fluvial channels were not observed at 32°, if we take our runoff depth estimate of 56 mm (for a slope of 20°), we find a flow velocity of 0.8 m/s and a discharge of 0.02 m³/s (for a 0.5-m-wide fluvial channel). For a crater average upslope bedrock area of 2460 m², this equates to a runoff rate of ~33 mm/h (though this does not account for the water load needed to saturate the gully bed material). While the Takahashi (1978) and Prancevic et al. (2014) model has not been tested on slopes as steep as those at Meteor Crater (i.e., > 20°), or with such coarse grain sizes, it does illustrate that while the body of a debris flow contains relatively little water for its sediment load, appreciable amounts of water (both surface runoff and subsurface flow) might be required for debris-flow initiation.

While the Takahashi (1978) model proposes that debris flows can form within fluvial channels under relatively steady flow conditions, others have proposed that debris flows initiate in a more chaotic fashion, often termed the “fire-hose effect.” In this model, sediment is mobilized when a concentrated flow of water (normally from intense rainfall over steep bedrock channels) hits unconsolidated sediment, such as debris dams that form in steep channels or talus material at the base of bedrock slopes (e.g., Johnson and Rodine, 1984; Berti et al., 1999; Berti and Simoni, 2005; Coe et al., 2008; Gregoretti and Fontana, 2008). The amount of water needed to initiate debris flows by this mechanism has not been well quantified. Field observations suggest that some amount of water is needed to saturate the sediment/talus prior to its entrainment (e.g., Curry, 1966), but the common assumption of steady, uniform flow for determining shear stress is likely not appropriate, as observations suggest the flow is highly nonuniform (e.g., jets; Costa, 1984; Berti et al., 1999; Cannon et al., 2003; Kean et al., 2013), and part of the sediment entrainment process is likely due to the force of the fluid momentum and the fact that the rapid delivery of water outpaces infiltration. Looking at other modern-day debris-flow events sug-

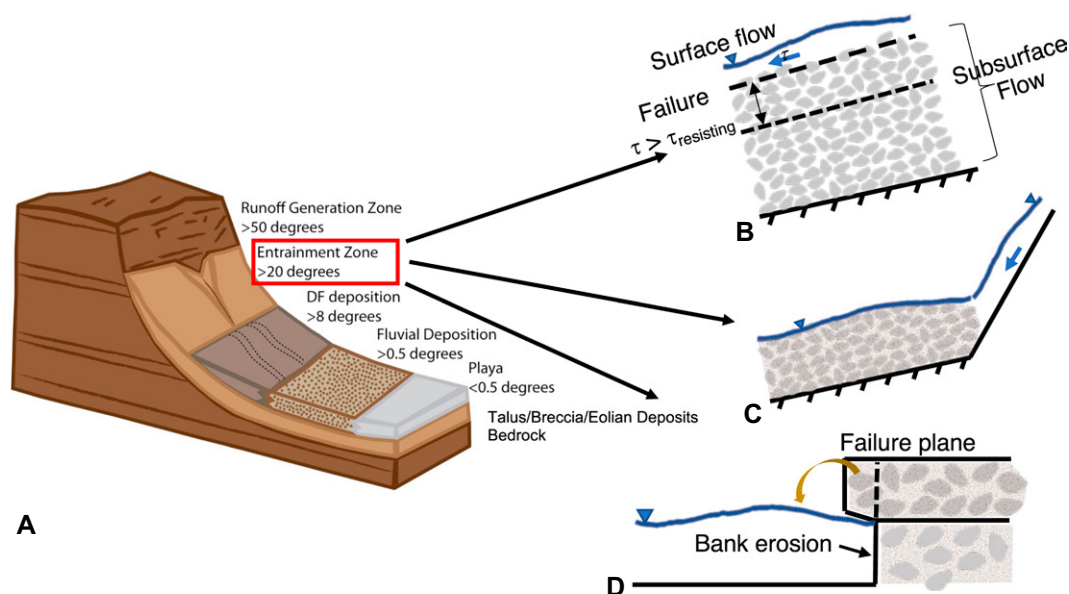


Figure 9. (A) Conceptual model for gully formation. (B) Entrainment into a debris flow (DF) can result from surface runoff that leads to shear entrainment of the sediment bed. (C) “Fire-hose” effect, which requires rapid delivery of water from the upper crater bedrock walls to the talus below. (D) Alternatively, fluvial runoff may undermine the channel floor and adjacent banks, which collapse onto the flow and transform it into a debris flow.

gested to be initiated in this way in the western United States, Godt and Coe (2007) found that debris flows were triggered in talus in response to a thunderstorm that produced ~ 43 mm in 4 h, most of which fell in the first 2 h, 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. Regional IDF curves suggest that similar short-duration and high-intensity storms occur at Meteor Crater at approximately decadal to 50 yr frequencies, but the grain-size distribution and talus properties likely matter when predicting the hydrologic conditions under which they will be entrained and mobilized.

For example, fine-grained sediment (i.e., clay- and silt-sized material) is often cited as a necessary ingredient for producing debris flows (e.g., Major and Pierson, 1992; Griffiths et al., 1996; Meyer and Well, 1997). The fines are thought to mix with water to form the highly viscous pore fluid that mediates intergranular collisions, prevents separation of the fluid and solid components of the flow, and reduces resistance to flow, enabling transport on gentle slopes (Pierson, 1981; Iverson, 1997; Kaitna et al., 2016). At Meteor Crater, near the gully headwalls, a layer of finer-grained white material supports many of the larger clasts (see Fig. 8B); these fines, perhaps generated during the impact, are likely necessary for causing entrainment into a mass flow. However, winnowing of this material has occurred (both by fluvial processes and wind erosion), as evidenced by rock flour mapped on the crater floor and within borehole sediments (Kring, 2007). Loss of this material over time will both increase the saturated conductivity of the talus, which in boulder-rich talus is set by

the conductivity properties of the fines filling the pore space (Bouwer and Rice, 1984), and decrease the viscosity of the pore fluid, both of which make it more difficult to generate the conditions needed to transport coarse grains as a debris flow (e.g., Kaitna et al., 2016).

A third mechanism for debris-flow formation at Meteor Crater is from collapse of gully walls/banks due to fluvial erosion and undercutting, such that runoff is suddenly charged with a high sediment load and transitions into a debris flow (i.e., bulking; Takahashi, 1978; Gregoretti and Fontana, 2008). Under this model, boundary shear stresses along the wetted perimeter of the gully channel and sidewalls must be sufficient to overcome bank cohesion, where the local boundary shear stress is proportional to the downstream component of the weight of the fluid (which is a function of the hydraulic radius in narrow channels; Nagata et al., 2000). Thus, as gully channels evolve and widen with time due to undercutting and bank failure, it would take higher discharges to maintain the same local boundary shear stresses (due to decreasing flow depths) following conservation of mass (i.e., Eq. 1). It is therefore possible that even if Pleistocene climate conditions persisted, debris-flow activity would have decreased.

As originally suggested by Kumar et al. (2010), Meteor Crater provides a unique environment in which to probe whether there is a direct relationship between erosional processes and climate change. The simplest analysis we performed, using sediment-to-water ratios and deposit volumes, suggests much greater runoff during the Pleistocene. However, this method would also suggest that very little water is needed to transport sediment via debris-flow

processes, so even modern-day runoffs would be sufficient. As discussed above, the ways in which debris flows initiate also matter to the water budget (Fig. 9). Fluid shear entrainment of the crater talus and gully floors likely requires large volumes of water to both saturate the sediment on the gully floors and generate sufficient overland flow (~ 1 – 5 cm on steep slopes with small drainage areas) to cause bed failure. It is possible that if this mechanism dominated at Meteor Crater, then the climate transition from the Pleistocene to the Holocene was the main driver in debris-flow cessation. The second mechanism is the “fire-hose” effect, which also requires rapid delivery of water from the upper bedrock walls of the crater to the talus below (i.e., a wetter climate), and enough fine-grained sediment to generate interstitial fluid capable of mobilizing the coarser talus grains. Debris flows suggested to have initiated by this mechanism in the southwestern United States did so under storm conditions that currently occur at Meteor Crater (based on IDF curves), suggesting that winnowing of fines from the talus over time (by wind or fluvial erosion) might also play a key role in the transition from debris flow-dominated to more fluvial-dominated activity at the crater. Last, the growth and evolution of the gullies themselves might lead to a shift in process; if sediment bulking from bank collapse is a potential debris-flow trigger, then widening of the gullies over time would require greater discharge to generate the shear stresses capable of scouring the gully walls and floor and causing bank failure. In this scenario, even if climate change does not occur, debris-flow activity may ultimately diminish due to gully evolution. Of course, all these mechanisms might have occurred at different times in

the evolution of the crater walls. This points to the importance of climate in the rate and type of erosional processes acting on the crater wall at Meteor Crater, but also the importance of environmental factors, such as grain-size distribution, sediment supply, and gully morphology, which might lead to spatially variable responses across the crater for a given storm event.

Conceptual Model for Gully Formation in Talus in Steep Bedrock Landscapes

Our topographic analysis of Meteor Crater suggests a general morphodynamic model for gully incision into talus and subsequent deposition by debris-flow and fluvial processes. Examples of environments where large talus deposits develop at the base of steep bedrock slopes, besides impact craters, are below cirque heads and sidewalls after glacier retreat (e.g., in the Colorado Rocky Mountains; Caine, 1986; Godt and Coe, 2007), in periglacial basins affected by frost weathering processes (Hales and Roering, 2007; Palucis et al., 2023), and steep postfire landscapes (e.g., San Gabriel Mountains; Lamb et al., 2011). Generally, three primary process zones can be identified: a runoff generation zone (typically off bedrock), a sediment entrainment zone (a bedrock slope mantled with colluvium, impact breccia [in the case of a crater], and/or eolian deposits), and a depositional zone (Fig. 9). Empirical studies elsewhere, as well as at Meteor Crater, suggest that each of these zones can be distinguished by characteristic slopes (Fig. 3B), in which the water source region tends to be steeper than 32° , the entrainment zone extends from $\sim 32^\circ$ to $\sim 20^\circ$, and the deposition zone progressively declines from debris flow–dominated (down to about $\sim 8^\circ$ – 5°) to, if present, fluvial (commonly $< 1^\circ$) and flat lake sediments.

When runoff (either from snowmelt or rainfall) is generated over impermeable bedrock, it crosses onto a lower, sediment-mantled zone, and entrainment occurs. The mechanism by which sediment is entrained into a debris flow via runoff can vary, but, in general, some combination of sediment entrainment, bed failure, and destabilization of gully walls causes a mixture of colluvium (i.e., boulders, gravel, sand, and mud) to rapidly charge the flow, leading to a debris-flow surge. This sediment-rich mass may continue to entrain colluvium and increase in size until it reaches lower unconfined slopes, where deposition ensues.

Runoff into a gully lacking sufficient duration or intensity to cause entrainment to mass flows may be able to transport finer 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 nonlinear, in that initial incision will focus subsequent runoff, entrainment (if the shear stress exceeds the threshold for sediment transport), 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, 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.

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 documented recent Martian gullied craters because it is much smaller (the gullies identified by Malin and Edgett, 2000, have gully heads that are several hundreds of meters wide, as opposed to tens of meters at Meteor Crater), it is not within a basaltic lithology, erosion is driven by rainfall (possibly rapid snowmelt as well), and the terrain into which Martian gullies are observed to form, which likely contains a large fraction of fine sediment and possibly appreciable ice (e.g., Mustard et al., 2001; Christensen, 2003), is absent.

However, the morphology of the Meteor Crater gullies is remarkably similar to many Martian gullies. Martian gully alcoves are often at the crest or midway down the talus slopes along crater walls, incise into both bedrock and slope deposits, and contain meter-scale boulders (de Haas et al., 2013, 2015a, 2015b; Núñez et al., 2016). Channels can be V-shaped in cross section (Mangold et al., 2010; Hobbs et al., 2013) and contain fluvial features such as terraces and longitudinal bars (Schon and Head, 2009), or they can contain mass-flow features such as levees (Levy et al., 2010; Johnsson et al., 2014; Sinha et al., 2019). At their distal end, Martian gullies often have lateral levees, lobate or digitate (“finger-like”) deposits, and/or poorly sorted coarse material. De Haas et al. (2015b) observed sedimentological evidence of debris flows on 51 gully fans across the southern midlatitudes of Mars, where exposures in incised channels contained meter-scale boulders randomly distributed in a finer-grained matrix, with occasional lens-shaped and truncated layering, which are

characteristic of the deposits found at Meteor Crater. When Conway and Balme (2016) compared the morphology of Martian gully alcoves (in cumulative/noncumulative area distribution space) to catchments that form from dry granular flows on the Moon and Earth, as well as terrestrial debris flows and fluvial gully catchments, they found that Martian gully source areas are statistically dissimilar to dry granular systems. These observations have led to the general consensus that the majority of gullies on Mars required some fluid or fluidization process to form (Harrison et al., 2015; Conway et al., 2019) and that only very recent flows, often mapped as bright deposits found above angle-of-repose slopes, might be dry in origin (Malin et al., 2006; Kolb et al., 2010).

As current Martian conditions render liquid water highly unstable, constraining gully formation mechanisms and associated fluid volumes and discharges has been a key goal of research into the modern Mars environment for the past two decades. Malin and Edgett (2000) showed that gully heads are often associated with distinct layers within a cliff, which they suggested points to seepage during obliquity-induced freeze-thaw cycles (when ice plugs melt or fracture), but this model (and other variants; e.g., Hartmann et al., 2003; Coleman et al., 2009) requires an unrealistically high-permeability aquifer, a very briny aquifer, or high geothermal fluxes (Goldspiel and Squyres, 2011). Mapping of thousands of gullies across Mars has shown that gullies tend to be located at mid- to high latitudes in both hemispheres and are associated with Martian surface features thought to be related to past and/or present surface ice (e.g., Dickson et al., 2007; Dickson and Head, 2009; Goldspiel and Squyres, 2011; Harrison et al., 2015). As such, some models invoke the melting of near-surface ice, resulting in landsliding or sediment erosion by exfiltration of meltwater (Costard et al., 2002; Gilmore and Phillips, 2002), though the predicted gully morphology is dissimilar to most Martian gullies (Gallagher and Balme, 2011; Conway et al., 2019). A problematic feature for all near-surface ice-melt scenarios is that model results suggest that over the past 5 m.y., the 273 K isotherm has typically been above near-surface ground ice at the mid- to high latitudes (Mellon and Phillips, 2001). Any melting that does occur would primarily occur during the summer (e.g., de Haas et al., 2015a), but active flow events on Mars have been observed mostly during the fall, winter, or early spring (Harrison et al., 2009; Diniega et al., 2010; Dundas et al., 2015). Carbon dioxide is a major constituent of the modern Martian atmosphere and will condense during the midwinter, leading to CO₂ frost cover from the poles to the midlatitudes

($\sim 50^\circ$; Piqueux et al., 2015) and on pole-facing steep slopes between latitudes of 30° and 50° (Vincendon et al., 2010a). There is support for CO_2 -related mechanisms triggering present-day flows, as active gullies are exclusively within the zone of observed or predicted CO_2 ice (Diniega et al., 2010; Dundas et al., 2015; Raack et al., 2015; Pasquon et al., 2016), and it is possible CO_2 frost processes are a significant part of initial gully formation in these regions (Dundas et al., 2022). However, contemporary gully activity has not been observed at latitudes equatorward of where CO_2 frost is observed spectroscopically (i.e., 45°S – 30°S ; Vincendon et al., 2010a, 2010b), so the role of past water cannot be ruled out for all Martian gullies (Dundas et al., 2015). Interestingly, de Haas et al. (2019) showed that for contemporary gully systems, flows triggered by CO_2 result in morphologies similar to those of granular flows on Earth (i.e., like those observed at Meteor Crater). Fluvial models have also been applied to Martian gullies, but they result in large water volumes that are difficult to reconcile with more recent Martian climate (e.g., Parsons and Nimmo, 2010). De Haas et al. (2015a) mapped individual debris-flow events in the young (Johnsson et al., 2014), midlatitude Istok Crater and found that 3–9 mm of liquid water would need to be uniformly spread over individual alcoves ($\sim 4 \times 10^4 \text{ m}^2$ per alcove) to generate modal-size debris-flow events; this is assuming terrestrial debris-flow size-frequency relationships and sediment-water ratios (but as we discussed herein, the water needed to initiate a debris flow event is not accounted for).

Thus, based on their morphology, occurrence of debris-flow and fluvial events, and exposure to a drying climate, the Meteor Crater gullies are relevant to understanding many Martian gullies. Our work highlights several important factors when trying to use geomorphic landforms to assess past hydrologic or climatic conditions on Mars. First, the dominant flow process can change both with climate and surface conditions (e.g., grain-size composition, sediment supply, slope, gully width) such that the modern-day processes we can observe on Mars may only be modifying already existing gullies (e.g., Johnsson et al., 2014). Thus, more work is needed to understand the ranges of conditions and processes that lead to the initial incision of gullies on crater walls, as well as the conditions that “turn off” gully formation. Field sites like Meteor Crater, where we broadly understand its climate history and can use a combination of field mapping, geochronology, and lake core records to obtain better constraints on the linkage between erosion processes and timing and climate, are invaluable to ultimately understanding crater evolu-

tion on Mars. Second, the use of deposits and erosional landforms, at least at Meteor Crater, likely underestimates total water inputs as well as the intensity and duration of those inputs (and the climatic regime those imply). This is an important consideration when doing simple water budget analyses on Mars, like using sediment-to-water ratios that do not account for the additional water required to trigger a debris flow, and inferring paleoclimate conditions (e.g., Mangold et al., 2012; Morgan et al., 2014; Palucis et al., 2014; de Haas et al., 2015a). A better mechanistic understanding of the ways in which sediment is transported on steep slopes, and the associated water volumes and rates, especially under different climate conditions, will aid in our ability to develop more accurate models for Martian systems.

CONCLUSIONS

Since its formation $\sim 50,000$ yr ago, Meteor Crater has undergone significant modification as ~ 71 gully systems incised into its inner walls during the first $\sim 40,000$ yr of pre-Holocene wetter conditions. Fieldwork and cosmogenic dating on an individual gully system combined with a detailed topographic analysis of LiDAR data suggest an alternative explanation to the previously proposed seepage erosion and fluvial incision model for the Meteor Crater gullies. We propose that debris-flow processes dominated 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. When water inputs are not sufficient to generate a debris flow, as was observed during a storm event in September 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 paleoenvironmental factors from the southwestern United States suggesting wetter and cooler conditions. Future study on the timing and evolution of the crater lake, and its influence on gully evolution, will provide further constraints on the evolutionary history of the inner crater walls and floor.

On average, debris-flow events at Meteor Crater consisted of $\sim 200 \text{ m}^3$ of material. To account for the total volume of debris-flow deposits currently deposited on the crater floor, debris-flow frequency was ~ 1 event every 1.5 yr, assuming debris-flow activity terminated $\sim 10,000$ yr ago. Meteor Crater therefore provides us with a general conceptual model

for how gully processes work, highlights the dependence of process on slope and climate, and suggests that analysis of paleodeposits and landforms alone will likely underestimate total water volume inputs, which are important considerations for understanding steep escarpments elsewhere on Earth and Mars.

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