1 Observations of an aeolian landscape: from surface to orbit in Gale Crater

2 **Mackenzie Day^{1*}, Gary Kocurek¹**

3 ¹*2275 Speedway Stop C9000 Department of Geological Sciences, Jackson School of*
4 *Geosciences, University of Texas at Austin, Austin, TX 78712 USA*

5 *Corresponding author: mdday@utexas.edu

7 **Abstract**

8 Landscapes derived solely from aeolian processes are rare on Earth because of the
9 dominance of subaqueous processes. In contrast, aeolian-derived landscapes should typify Mars
10 because of the absence of liquid water, the long exposure times of surfaces, and the presence of
11 wind as the default geomorphic agent. Using the full range of available orbital and Mars Science
12 Laboratory rover *Curiosity* images, wind-formed features in Gale Crater were cataloged and
13 analyzed in order to characterize the aeolian landscape and to derive the evolution of the crater
14 wind regime over time. Inferred wind directions show a dominance of regional northerly winds
15 over geologic time-scales, but a dominance of topography-driven katabatic winds in modern
16 times. Landscapes in Gale Crater show a preponderance of aeolian features at all spatial scales.
17 Interpreted processes forming these features include first-cycle aeolian abrasion of bedrock,
18 pervasive deflation, organization of available sand into bedforms, abundant cratering, and
19 gravity-driven wasting, all of which occur over a background of slow physical weathering. The
20 observed landscapes are proposed to represent a spectrum of progressive surface denudation
21 from fractured bedrock, to retreating bedrock-capped mesas, to remnant hills capped by bedrock
22 rubble, to desert pavement plains. This model of landscape evolution provides the mechanism by
23 which northerly winds acting over ~3 Ga excavated tens of thousands of cubic kilometers of
24 material from the once sediment-filled crater, thus carving the intra-crater moat and exhuming

Mount Sharp (Aeolis Mons). The current crater surface is relatively sand-starved, indicating that potential sediment deflation from the crater is greater than sediment production, and that most exhumation of Mount Sharp occurred in the ancient geologic past.

Keywords: Aeolian processes, Mars, surface, cratering

1. Introduction

On Earth, landscapes evolve through the interplay between tectonics, and water-dominated erosion and deposition. In contrast, Mars has generally been regarded as the most aeolian-dominated planet in the solar system (e.g., Sagan and Bagnold, 1975; Arvidson et al., 1979; Greely et al., 1999). On Earth, landscapes fashioned solely by wind action are rare. For example, although extensive aeolian dune fields blanket the modern Sahara Desert of North Africa, even a cursory examination of this landscape reveals uplands marked by erosional dendritic fluvial networks, and lowlands marked by mid-Holocene lakes and fluvial systems (e.g., Bouchette et al., 2010). Uniquely aeolian erosional and depositional features such as yardangs and dunes are well known, but sand dunes are nearly universally sourced by contemporaneous or earlier wet period fluvial systems (Kocurek and Lancaster, 1999), and yardangs are typically fashioned from lacustrine deposits (Wang et al., 2011). Similarly, although the Dry Valley of Antarctica displays a rich variety of cold-climate aeolian features (Gillies et al., 2013), the fundamental fabric of the landscape is glacial-dominated. While fluvial-dominated and glacial-dominated processes and their landscapes are well described and modeled, with the paucity of aeolian-dominated Earth landscapes, there is no comparable body of work describing landscapes that evolve where wind is the prime or sole geomorphic agent. Resurfacing by aqueous processes and tectonism on Earth precludes the formation of aeolian landscapes over geologic spans of time.

On Mars, landscapes and strata formed in fluvial, deltaic and lacustrine environments are well known from the most ancient Martian surfaces. However, beginning with global climatic change at ~ 3.5 Ga (Bibring et al., 2006), aeolian processes have increasingly dominated. In the absence of liquid water at the surface, wind emerges as the “default” global geomorphic agent, with more localized geomorphic change caused by impacts, ice, and gravity-driven slope retreat, all imprinting over a background of very slow physical weathering. Mars, therefore, may be the best natural laboratory in which to delineate the attributes of an aeolian landscape. In this study we use satellite and surface images to explore the extent to which the landscape of Gale Crater is aeolian-dominated, thus helping to define the attributes of such a system. The collection of aeolian features associated with Gale Crater also serves to reconstruct the history of the wind regime over orders of magnitude of time.

2. Gale Crater study area and data set

The impact that formed Gale Crater is estimated to have occurred ~3.6-3.8 billion years ago (Thomson et al., 2011; Le Deit et al., 2013). At 155 km in diameter, Gale Crater sits just south of the dichotomy boundary near Aeolis Mensae (Fig. 1) (Wray, 2013). The interior of Gale is dominated by the central mound Mount Sharp (Aeolis Mons), which stands roughly 5 km high, or three times the depth of the Grand Canyon. The mound and surrounding features have been extensively mapped (e.g., Anderson and Bell, 2010; Hobbs et al., 2010; Thomson et al., 2011) and are similar to other intra-crater mounds found across Mars (Grotzinger and Milliken, 2012).

Figure 1. Gale Crater context. A. Mars Orbiter Laser Altimeter (MOLA) colorized elevation map showing context of Gale Crater (boxed). Note the change from high elevation in the south to low elevation in the north associated with the global topographic dichotomy. B. Gale Crater shown with HRSC-derived DEM superimposed on HRSC visual basemap. The crater interior is dominated by Mount Sharp (Aeolis Mons). C. Mount Sharp as seen from the surface by *Curiosity*. Foreground wheel tracks are 30 cm wide, and the total relief on the mound is ~5 km (sol 530, Mastcam mosaic starting at frame 0530ML0021050000203319E01_DRCX; all MSL data can be accessed through the JPL PDS archive).

In association with the ongoing Mars Science Laboratory (MSL) mission, an unprecedented number of images have become available for the study of Gale Crater. To date, the rover *Curiosity* has spent over 1,000 Martian sols traversing the floor of Gale, taking images at resolutions down to 15 $\mu\text{m}/\text{px}$ (Edgett et al., 2012). Using the full range of available scales, for this study features were first mapped in orbital images using a High Resolution Stereo Camera (HRSC) basemap, overlain with higher resolution Context Camera (CTX) and High Resolution Imaging Science Experiment (HiRISE) images. Resolution ranged from 25 cm/px in HiRISE coverage, to ~ 10 m/px in HRSC. Surface images were then used to sample aeolian features along the *Curiosity* rover traverse from Bradbury Landing to Pahrump Hills (sols 1-960). Sampling every 200 to 300 m, Navigation Camera (Navcam) mosaics were used to document aeolian features in the landscape. Mast Camera (Mastcam), and Mars Hand Lens Imager (MAHLI) images were used to support identification and characterization of the aeolian features discussed below.

3. Non-aeolian activity in Gale Crater

Gale Crater is fundamentally an impact crater. Smaller-scale cratering blankets the crater interior (Le Deit et al., 2013), disrupting the morphology of aqueous and aeolian features alike. Subaqueous erosional and depositional processes clearly dominated during the early crater history. Although upper portions of Mount Sharp have been speculated to be aeolian in origin (Anderson and Bell, 2010), basal portions of Mount Sharp thus far investigated are interpreted as subaqueous (lacustrine) in origin (Grotzinger et al., 2014). Moreover, the trek of *Curiosity* from its landing site to the base of Mount Sharp revealed deposits of conglomerates, sandstones and mudstones of interpreted fluvial, deltaic and lacustrine origins (Williams et al., 2013; Grotzinger et al., 2014). The crater rim is characterized by dendritic channel systems, which give rise to alluvial fans (Anderson and Bell, 2010; Palucis et al., 2014). Similarly, entrenched channels emanating from Mount Sharp evidence subaqueous erosion on Mount Sharp (Thomson et al.,

2011). Subaqueous erosion and backwasting of the crater rim and central uplift, in addition to a few drainage networks that breach the crater rim, are thought to have sourced the subaqueous strata within Gale Crater through a series of fluvial-deltaic systems and basinal lacustrine body (Grotzinger et al., in press).

The subaqueous history of Gale Crater, however, records an ancient period on Mars with the presence of liquid water and a denser atmosphere. Crater counts suggest that the subaqueous sedimentary strata were emplaced before 3.2-3.3 Ga (Grant et al., 2014; Newsom et al., 2014). For the subsequent ~3 Ga, Gale Crater has arguably been under the domain of aeolian processes, gravity-driven slope processes, and cratering. All subaqueous features are in a current state of erosion. For example, channel networks show a high albedo indicating dust mantling, inverted channels have been exhumed through wind deflation, and fan surfaces are heavily cratered (Palucis et al., 2014).

4. Exhumation of Mount Sharp

Although a depositional origin for Mount Sharp has been suggested (Kite et al., 2013), consistent with observations presented here, most works interpret Mount Sharp as a remnant morphology from the exhumation of a once totally filled crater (e.g., Malin and Edgett, 2000; Grotzinger et al., in press). Gale Crater is one of many craters to host an intra-crater mound separated from the rim by an interpreted erosional moat (Grotzinger and Milliken, 2012). The streamlined morphology of Mount Sharp argues for N-S net transport by wind. The amount of sediment removed by aeolian deflation to yield current Mount Sharp topography can be estimated using the HRSC-derived digital elevation model (DEM). Assuming the crater was once filled to the maximum height of Mount Sharp, the removed volume was calculated as $6.34 \times 10^4 \text{ km}^3$. This does not include the volume of loose sediment in dune fields, sand sheets, transverse aeolian ridges (TARs), and dust.

Loose sediment in Gale Crater is taken as largely derived from erosion of Mount Sharp and the crater rim, consistent with Tirsch et al., (2011) who suggest that intra-crater dune fields are sourced primarily from landscapes within the crater. However, because craters act as temporary depo-centers, the crater has also likely received an unknown quantity of sediment transported into the crater from external sources. Dust would be created with erosion internal to the crater, but the crater must also house dust derived from the global dust budget. In order to estimate the total volume of sediment housed in dune fields, TARs, and sand sheets, major deposits of these were mapped in the crater interior (Fig. 2). The total mapped surface area was multiplied by an assumed average depth (Table 1) and, neglecting porosity, yielded a volume on the order of 10 km^3 . This volume is dominated by dune sediment, so assumptions of equivalent sand depth are not trivial. The largest and smallest barchan dunes and linear dunes in the Bagnold Dune Field were used to bracket estimates of equivalent sand thickness in the dunes. Lee slopes in this field were measured as $\sim 30^\circ$ (Atwood-Stone and McEwen, 2013; Silvestro et al., 2013) and assuming 10° stoss slopes, equivalent sand depths ranged from 3-17 m. Dune spacing is variable and exposed bedrock in interdune areas is estimated to account for $\sim 10\%$ of the mapped area. Taking the average estimated sand depth of 10 m and subtracting the 10% interdune area, the total sand volume is estimated as 6.44 km^3 . Similar considerations were applied to TARs ($5.31 \times 10^{-2} \text{ km}^3$) and sand sheets ($1.85 \times 10^{-2} \text{ km}^3$). Varying estimates of depth and total coverage can alter the total estimate by a scaling factor, but the order of magnitude of the sand volume is estimated to remain the same. In any estimate, the amount of potentially mobile sand housed in Gale Crater is relatively minor, and the crater represents an overall sand-starved system.

Figure 2. Mapped areas of dunes, sand sheets and TARS used to estimate the volume of sediment in Gale Crater. Total volume is of order 10 km^3 . See Table 1 for assumed average depths and surface areas. North-up oriented background HRSC mosaic of Gale Crater is also used in subsequent figures (h1916_0000, h1927_0000, h4235_0001).

Estimating the volume of dust in Gale Crater is less straightforward. A simple estimate can be made assuming a 1 mm thick mantling of dust across the entire surface area approximated by πr^2 . This simple calculation yields a volume of $7.6 \times 10^{-2} \text{ km}^3$, but does not take into account any surface roughness. Here roughness contributions come at a range of scales from the topography of Mount Sharp, to individual grain coatings. The total contribution of dust in Gale could be orders of magnitude higher than this simple calculation, which is given here as a conservative lower bound to the dust mantle volume.

Table 1: Gale Crater interior loose sediment volume estimates and assumptions

Sediment Type	Measured surface area (km²)	Assumed average thickness (m)	Total volume (km³)
Dunes	715	9	6.44
TARs	106	0.5	5.31×10^{-2}
Sand sheets	185	0.1	1.85×10^{-2}
Total volume:			6.51

Assuming a horizontal crater floor (DTM-derived slope is $<1^\circ$), the gross volume of Mount Sharp is estimated as $1.25 \times 10^4 \text{ km}^3$. Volume occupied by a central uplift peak created during impact is unknown and here included in the Mount Sharp volume. This estimate yields a total crater interior volume of $\sim 7.59 \times 10^4 \text{ km}^3$, with Mount Sharp occupying 16% of the interior volume, and the remaining 84% removed. The time over which the postulated exhumation of Mount Sharp occurred is unknown, and erosion rates are expected to have varied, with significantly greater rates early in crater history (Warner et al., 2010; Grotzinger et al., in press) and very low rates currently (Golombek et al., 2010; Farley et al., 2014). An assumption of 3 Ga of excavation yields an average annual flux of $2 \times 10^4 \text{ m}^3/\text{yr}$, whereas assuming 1 Ga yields $6 \times 10^4 \text{ m}^3/\text{yr}$. The Namib Sand Sea on the west coast of southern Africa houses roughly $1.02 \times 10^3 \text{ km}^3$ of sand, with modern sand influx rates of approximately $4 \times 10^5 \text{ m}^3/\text{yr}$ (Lancaster, 1989a). Over the past 3 Ga on Mars, 60 Namib Sand Seas worth of sediment have been excavated from

178 Gale Crater, at average rates of excavation an order of magnitude lower than modern sediment
179 influx rates seen on Earth.

180 **5. Orbital observations**

181 Ripples, dunes, TARs, yardangs, and crater wind streaks can all be observed from orbit in
182 Gale Crater. These features exhibit morphologies characteristic of their formative wind regimes,
183 and cover a wide range of temporal and spatial scales. Considered together, these aeolian
184 features can be used to explore the history of sand-transporting winds spanning a significant
185 period of the geologic past of Gale Crater. Although absolute time scales of formation are
186 unknown, relative time scales can be inferred from superposition of these features (Fig. 3).
187 Ripples orient to a particular wind quickly and can be seen on dune stoss and lee slopes
188 (Howard, 1977). Dunes, in turn, are seen migrating over TARs, which commonly form between
189 yardangs. Finally, yardang morphology is influenced by channel topography, requiring that the
190 channels formed first. Small crater wind streaks also serve to indicate wind direction, but were
191 not observed in superposition with any other features. Each feature is here discussed in turn, and
192 analyzed for a record of formative wind direction.

Figure 3. Superposition of aeolian features. A. Ripples superimposed upon the stoss (1) and lee (2) surfaces of a dune. B. The same dark dune migrating over light toned granule ripples or small TARs. White box indicates position of A. HiRISE image PSP_006143_1745. C. TARs (black arrow) positioned between yardangs (white arrows), transverse to the yardang-formative north-south wind direction. HiRISE image EPS_019698_1750. D. Yardang field formed across a channel valley. Yardang scale and spacing changes dramatically from north to south sides of

valley, suggesting that flow disturbance associated with valley topography affected yardang formation. The valley topography, therefore, necessarily predates the associated yardangs. Note also the intra-valley linear dune (black arrow) covering light toned intra-valley TARs (white arrow). HiRISE images ESP_012195_1750 and PSP_008147_1750. A and B have been contrast-stretched to increase feature definition. Locations are indicated in Figure 5, and north is up in all images.

5.1 Crater wind streaks

Gale Crater exhibits a low albedo wind streak that extends from the southern rim of the crater onto the surrounding plains and into Lasswitz crater (Fig. 4). This wind streak is ~150 km wide and ~370 km long, and is interpreted to have formed from dune sands excavated from the crater interior by northerly winds (Edgett and Malin, 2000; Wray, 2013). Its low albedo contrasts with the bright surfaces seen in the surrounding area (Christensen, 1986), and is interpreted as an absence of dust mantling and an indication of active transport processes (Veverka, 1975).

Bedforms (i.e., ripples or dunes) are not visible on the wind streak, arguing that sediment on wind streak represents a veneer too thin to be organized into bedforms visible at CTX resolution. In addition, wind speeds needed to overcome the topography of the rim and transport sand out of large craters likely occurs only during rare high-speed wind events. Given the size of the wind streak, therefore, wind streak construction must have occurred over significant time.

A separate population of much smaller wind streaks emanate from small craters ($d < 0.5$ km) on the southern rim of Gale Crater (Fig. 5). These bright albedo wind streaks contrast with dark dune sands from the Southern Dunes, and are thought to form by the collection of air-fall dust in the sheltered lee of craters (Fig. 6) (Sagan et al., 1972). Similar wind streaks, observed near the Mars Exploration Rover (MER) *Opportunity* landing site in Meridiani Planum, reversed direction before and after a dust storm, suggesting constructive time scales of weeks to months

(Sullivan et al., 2005). On the north, west and east Gale Crater rim, where no streaks are observed, rim deposits are homogenous in their light tone, thus preventing definition of any dust-mantle wind streaks. On the south rim, measurements of 180 high albedo wind streaks yielded an average orientation of 351° interpreted to form from southerly (171°) winds (Fig. 6B, Table 2). The southerly winds recorded by these features are counter to the dominance of northerly winds observed elsewhere, but match observations in the Southern Dunes discussed below.

Figure 4. Gale Crater wind streak A. Gale Crater with its low albedo wind streak extending into Lasswitz Crater to the south. B. Same image as A, but with extent of the wind streak and craters shown by dashed and solid lines. North-up oriented Mars Orbiter Camera (MOC) wide angle mosaic from Aeolis (MC-23) quadrangle.

Figure 5. Distribution of mapped yardangs, TARs, and crater wind streaks in Gale Crater.

Yardangs are found only on Mount Sharp, and crater wind streaks are seen only on the southern crater rim. Black axial line denotes where yardangs orientations (Fig. 7) were divided between the two halves of the mound. Boxes over HRSC background denote the locations of panels in other figures as indicated. See Table 2 for detailed statistics and sample sizes of each dataset.

Table 2: Wind indicator sample statistics

Feature	Mean orientation (°)	Standard deviation (°)	Min length (m)	Max length (m)	Mean length (m)	Standard deviation (m)	Sample size (n)
Crater wind streaks	351	33	66	962	206	141	180
Yardangs – east	347	24	60	4340	530	508	329
Yardangs – west	359	13	33	1320	349	240	440
TARs total	88	47	9	876	124	103	3222
TARs major	71	24	9	721	125	106	2047
TARS minor	337	22	10	650	93	89	1175
Sand shadows	189	0.5	0.12	3.74	0.59	0.34	69

Figure 6. Southern rim wind streaks. A. Small crater wind streaks on the southern Gale Crater rim. Arrows indicate just a few of the craters with NW trending streaks in this scene. Image is contrast-stretched to make streaks more visible (CTX image P01_001554_1745_XI_05S221W, north-up oriented). B. Histogram of orientations of mapped crater wind streaks. Mean orientation (351°) is interpreted to form from southerly (171°) winds.

254 5.2 Yardangs

255 Yardangs form from persistent aeolian erosion by abrasion, causing elongate flow-
 256 parallel features to be carved out of the bedrock (Hedin, 1903). Typically occurring as fields,
 257 yardangs form slightly asymmetric morphologies, with the upwind end of the feature blunted
 258 into an inverted boat-hull shape. Because some fluvial and mass wasting processes can generate
 259 morphologies similar to yardangs (Ward, 1979), features associated with sinuous channels, or
 260 headward eroding canyons were not included in this study. Positive relief on both sides of the
 261 feature was also required to prevent misidentification of linear textures in layered terrain as
 262 yardangs. A perfect yardang would indicate an absolute wind azimuth with its blunted upwind
 263 hull, but the yardangs observed on Mars commonly do not exhibit this classical asymmetry (Fig.
 264 7). This study, therefore, considers yardangs to be 180° ambiguous indicators of wind direction.

265
 266 Figure 7. Yardangs. A. Yardangs in Gale Crater. These elongated erosional forms have tapered
 267 ends and typically form in groups (Ward, 1979). HiRISE image ESP_018643_1745 is oriented
 268 north up. B and C. Yardang orientations measured on the west (B) and east (C) sides of Mount
 269 Sharp (Fig. 5). Orientations are shown symmetric to account for 180° ambiguity in formative
 270 wind direction. The two sides are statistically distinct and the 12° difference in mean orientation is
 271 interpreted to result from flow divergence around Mount Sharp topography.

Locations and distributions of yardangs mapped in this study (Fig. 5, Table 2) correspond well with independent mapping of Mount Sharp (Anderson and Bell, 2010; Thomson et al., 2011). The presence of yardangs on Mount Sharp and the general absence of well-defined yardangs elsewhere imply that the strata of Mount Sharp are fine-grained and not strongly cemented. Ranging from 33 m to 4.3 km in length, yardangs are elongated N-S with a mean orientation of 354°. Dividing yardangs between the east and west sides of Mount Sharp (Fig. 5) yields two statistically distinct populations of yardangs with means differing by 12° (Fig. 7, Table 2, t-test, $\alpha < 0.01$). This variation could be explained by large-scale flow divergence of northerly winds around Mount Sharp.

5.3 Transverse aeolian ridges

The term “transverse aeolian ridge” (TAR) describes Martian bedforms with size and morphology intermediate between ripples and dunes (Fig. 8) (Wilson and Zimbelman, 2004). Most commonly found in Martian equatorial regions, TARs form symmetric linear ridges with wavelengths between 20 and 60 m (Wilson and Zimbelman, 2004; Zimbelman, 2010). TARs are distinguished from dunes by high albedo, linear crests, lack of activity, and symmetric cross-sectional profiles (Shockey and Zimbelman, 2013). TARs are distinguished from sand shadows by scale and lack of an upwind obstacle, and from ripples by scale, context, and albedo. TARs are commonly found in topographic lows, and are thought to align normal to their formative wind direction (Balme et al., 2008). The formation mechanism of TARs is still debated, with hypothesized terrestrial analogs including granule ripples, reversing transverse dunes (Zimbelman and Scheidt, 2014), anti-dunes (Geissler, 2014), and megaripples (de Silva et al., 2013; Sullivan et al., 2014). To date, lateral migration of TARs has not been observed. Recent work on armored granule ripples in South America has suggested that TARs may be aggrading through the capture and infiltration of dust (de Silva et al., 2013; Bridges et al., 2015). Gravels armoring terrestrial “megaripples” are too large to be transported off the bedform by wind, but

can still trap passing finer material. These fines are kinetically sieved downward as gravels vibrate during wind events, raising the bedform surface. This process is suggested as an analog for TAR aggradation on Mars.

Figure 8. Transverse aeolian ridges. A. TARs confined to a valley on Mount Sharp, and oriented normal to valley walls. TAR orientation is clearly dependent on local topography, making TARs poor indicators of regional-scale winds. These TARs show simple linear and forked morphologies. HiRISE image ESP_031948_1750 is oriented north up. B. Orientations of all TARs mapped in Gale Crater, symmetric to account for 180° formative wind ambiguity. Two overlapping populations of TAR orientations were modeled to have mean NE (71°) and NW (337°) trends. These are interpreted to form from dominant northwesterly (341°) and subordinate northeasterly (67°) winds similar to winds interpreted in the Bagnold Dune Field.

Mapping of TARs in Gale Crater (Fig. 5, Table 2) found that they most commonly form in local topographic lows (e.g., in channels, between yardangs, and in small craters) where flow

is directed by topography. TAR morphologies were dominantly simple (linear) and forked (Balme et al., 2008), but networked morphologies, marked by a plaid pattern of crossing ridges, were seen in HiRISE at the northern base of Mount Sharp. The observed TARs were predominantly confined, enclosed on all sides by topography, or controlled, forming in association with less than circumferential topography (Balme et al., 2008). Some TARs formed on the crater floor and smooth reaches of Mount Sharp, independent of local topography. Ranging from 9 to 876 m long, TARs cover approximately 1% of the Gale Crater interior surface, and are oriented in all directions. To prevent a premature genetic interpretation, TARs are here considered 180° ambiguous indicators of formative wind direction. The mean orientation of all bedforms is 88°, but the set of sampled TARs can be modeled as two overlapping normally-distributed populations, with mean orientations of 71° and 337° for major and minor signals, respectively (Fig. 8, Table 2). These are interpreted to correspond with primary northwesterly and subordinate northeasterly winds.

5.4 Dunes and ripples

5.4.1 Approach

Dune fields in Gale Crater occur within the moat between Mount Sharp and the crater rim. These dune fields have been extensively characterized by Hobbs et al. (2010). Here we more broadly identify four morphologically distinct regions (Fig. 9), and independently map the orientations of dunes and dune brinklines in order to further reconstruct the wind regime in the crater interior. The most detailed analysis is for Bagnold Dune Field because of extensive HiRISE coverage (Silvestro et al., 2013). Analysis for the other regions is necessarily less comprehensive because of the incomplete HiRISE coverage of these fields.

Figure 9. Morphologically distinct dune field regions discussed in section 5.4. White boxes indicate the named dune fields and extent of corresponding context images. Note that the fields are all circumferential to Mount Sharp and associated with steep slopes of either Mount Sharp or the southern crater rim. Background HRSC mosaic.

This work largely concurs with the conclusions of Hobbs et al. (2010) that northerly winds dominated Gale Crater and that these winds are strongly channeled by crater topography. Because dunes and ripples form from mobile sand, their morphology and orientation are thought to reflect the current wind regime in Gale Crater, in contrast to features such as yardangs that require orders of magnitude more time to form. Where sand availability is not restricted and dune reconstitution time is longer than cycles in the wind regime, dune crestlines are oriented to be as perpendicular as possible to all components of the wind regime, the gross bedform-normal orientation of Rubin and Hunter (1987). Where sand availability is restricted, as is commonly the case on Mars, dunes elongate in the mean transport direction from a local sand source (du Pont et al., 2014). In either case, dune crestline trend reflects the total wind regime and net dune migration direction, and not a specific wind direction except where the wind regime is essentially unimodal. Wind ripples, because of their small size and shorter reconstitution times, are more apt to show the last significant wind event, but their orientation is also strongly affected by dune topography (Howard, 1977).

5.4.2 Bagnold Dune Field

The Bagnold Dune Field is situated along the upcoming *Curiosity* traverse, and will be the location of the first surface observations of active Martian dunes. Orbital measurements yielded ripple and dune migration rates in this field of 0.66 m/yr and 0.4 m/yr, respectively (Silvestro et al., 2013). This long and thin field runs 30 km along the northwest basal margin of Mount Sharp, and averages 1 km wide. Following the field from NE to SW, the dune field features a 16 km long main section of well-defined duneforms, followed by a 9 km stretch of bypass where dunes are mostly absent, finally ending in a 5 km long set of re-formed barchan dunes (Fig. 10A).

365 Figure 10. Bagnold Dune Field. A. HiRISE mosaic of the Bagnold Dune Field showing three
366 regions discussed in section 5.4.2. The main area is 16 km long and houses the majority of the
367 dunes. The middle 9 km region is sediment-starved, featuring only sparse bedforms and sand
368 sheets that fill topographic lows. The final 5 km is dominated by asymmetric barchans interpreted
369 to have reformed from sediment sourced from the main field. White boxes show the locations of
370 B and C. The *Curiosity* landing site and traverse through sol 960 are given by the star and white
371 line, respectively. B. Southwest end of the Bagnold Dune Field where SW-facing (Population 2)
372 slipfaces dominate and SE-facing slipfaces (Population 1) are mostly absent. White box shows the
373 location of Figure 13C (HiRISE images ESP_025790_1750 and ESP_012841_1750). C. Sand-
374 starved duneforms in the middle region of sediment bypass. This contrast-stretched image shows
375 the tail end of a linear dune (right) and a tear-drop dune (left) dying out as they migrate to the
376 southwest. Circular dark patches in the image are sand sheets filling small craters (HiRISE image
377 ESP_028823_1755). North is up in all panels.

Figure 11. North-up oriented HiRISE mosaic of the main section of the Bagnold Dune Field with locations of panels in Figures 12 and 13. White star and line indicate the *Curiosity* landing site and traverse. Dashed line corresponds to the dashed line in Figure 14 and marks the dune-dominated surface used in this region to estimate loose sediment volume (Table 1).

Figure 12. Dune morphologies in the Bagnold Dune Field. A. Symmetric barchans (1) migrate and link to become asymmetric barchans (2). B. Symmetric linear dune from the middle of the main section with equally defined slopes to each side of the crestline. C. Progression of dune morphology from west to east in the dune field. Barchans link defect arms downwind to form chains of elongations to the south and southwest (1). Chains become linear dunes, here asymmetric with SE-facing steeper slopes (2). Note how the linear dunes become detached from formative barchans. D. Reversal on linear dune in the SW of the field to have a sharper NW-facing slope. White box shows the location of E. E. Ripples on the linear ridge aligning parallel to ridge direction. Images have been contrast stretched to bring out albedo variation, and north is oriented up in all panels. HiRISE images ESP_024023_1755 (A and C), ESP_033649_1750 (B, D and E).

Dune morphologies in the Bagnold Dune Field range from barchans along the NW margin to linear dunes in the interior (Fig. 11). Small symmetric barchans (100-150 m wide), seen in isolation at the NE end of the field, laterally link with other barchans as they migrate downwind, forming asymmetric dunes as one arm elongates to the SW (Fig. 12A). These elongate arms extend and connect with other defect arms downwind to form chains that coalesce into the SW-NE trending linear dunes (Fig. 12C). Near the beginning of the field these linear dunes (0.5-1.5 km long) are asymmetric, with steep SE-facing slopes (Fig. 12C). To the SW, morphologies grade into symmetric linear dunes (Fig. 12B), and even reverse to have sharper NW-facing slopes (Fig. 12D), arguing for formation by barchan arm elongation in bimodal winds (Bourke, 2010). Rippled sand sheets make up the interdune areas between linear dunes, and ripples observed on linear dunes crests are parallel to the dune ridge (Fig. 12E).

Figure 13. Slipfaces on Bagnold Dunes. A. Tear-drop dune with two steep slipfaces (arrows) bisected by an elongating arm. Note the exposed bedrock and light toned granule ripples in the interdune areas. B. Primary (1) and secondary (2) slipfaces in the NE end of the dune field (Fig. 11). Note the sharp brinkline in the primary face and the subdued brinkline on the secondary face. Crest and ripple orientations are mapped on this image in Figure 14A. A and B taken from HiRISE image PSP_009861_1755. C. SW-migrating barchans in the SW end of the dune field.

415 Note the symmetric (1) and asymmetric (2) morphologies and the SW-elongating arm that
 416 indicates the presence of a northeasterly wind. (See Fig. 10 for location) D. Barchan dune from
 417 the middle of the field (Fig. 11) with no observed primary slipface, and secondary slipface similar
 418 to the morphology observed in B. HiRISE image ESP_033649_1750. All images are contrast-
 419 stretched to bring out the variation in these dark features, and north is up in all panels.

420
 421 Figure 14. Bedform mapping in the Bagnold Dune Field. A. Example mapping superimposed on
 422 the same image seen in Figure 13B. Primary brinkline trend is defined by the sharp south-facing
 423 slipface (red). Secondary trend is defined by steep lee slopes, reworked by ripples (blue). Linear
 424 dune trend is shown stemming from an elongating barchan (green). Ripple trends are shown for

schematic representation of the variation in orientation with position on the dune. B. Bedforms mapped in the main section of the Bagnold Dune Field with the same lateral coverage shown in Figure 11. Asymmetric barchans were also mapped in the extreme southwest, but were relatively few and are not shown here. Populations of orientations are shown in the rose diagram. Note that the primary and secondary orientations are interpreted to form transverse to winds with upwind azimuths of $\sim 344^\circ$ and $\sim 67^\circ$, respectively.

Focusing on the barchan dunes, in all regions the interdune areas between barchans are dominated by exposed bedrock and light-toned granule ripples (Fig. 13A). The isolated symmetric barchans near the beginning of the field exhibit a single sharp brinkline and SE-facing slipface. If the reconstitution time of the barchan dunes is longer than the wind alternation time, multiple slipface orientations could be preserved on a single barchan form. The “tear drop” morphology, where two slipfaces are bisected by an elongating arm, has been modeled to form in bimodal wind regimes with a divergence angle of at least 100° (Parteli et al., 2009), and is seen in abundance on Mars (Hayward et al., 2014). In the Bagnold Dune Field, tear-drop dunes form downwind of symmetric barchans in the main section of the field with SE- and SW-facing slipfaces (Fig. 13A). SE-facing slipfaces (Population 1) have sharp brinklines and smooth (non-rippled) lee surfaces interpreted as grainflow deposits and are hereafter designated the primary slipface population (Fig. 13B). The grainflows seen on these slipfaces indicate formation transverse to flow ($\pm 20^\circ$) (Eastwood et al., 2012). Ripple orientations on dune stoss slopes align roughly parallel near the dune crest with these sharp slipfaces (Fig. 14A). The SW-facing slipfaces (Population 2) are distinctly different. Hereafter referred to as the secondary slipface population, these slopes exhibit superimposed ripples migrating across and reworking the steep deposit (Fig. 13B). Sharp brinklines are absent on secondary lee faces, and no grainflow deposits were observed.

The relationship between the two slipfaces changes along the dune field. The primary slipfaces are the most prominent in the NE of the field, and the secondary slipfaces become dominant as one follows the field to the SW (Figs. 13D, 13C). Rather than a sharpening of the secondary lee faces, this transition occurs as a gradual disappearance of the primary lee faces, which become more subdued to the SW (Fig. 13D). The decay of primary slipfaces roughly corresponds with a reversal of the dominant slope on the linear dunes described above. At the very SW end of the dune field, the final group of barchans is clearly dominated by the secondary slipface population. Here, SE-facing lee slopes are all but absent, and the teardrop morphology is no longer observed. Small symmetric barchans are seen migrating to the SW, and a new morphology of asymmetric barchans dominates. The southern defect arm of SW-migrating crescentic barchans elongates in the transport direction to the SW creating an asymmetric barchan with a short northern and long southern defect arm (Fig. 13C). The morphology of these elongated arms requires that the originally dominant northwesterly wind be still present, but now this wind is clearly subordinate to the northeasterly wind (Tsoar et al., 2004).

The orientations of the two populations of slipfaces and of the linear dunes were mapped across the Bagnold Dune Field (Fig. 14, Table 3). The average orientation of the linear dunes roughly bisects the average upwind azimuth inferred from the two slipface populations. In bimodal winds, barchan arms tend to elongate in the direction of the resultant wind vector (Tsoar, 1989). Assuming the two slipface orientations are approximate indicators of wind direction ($\pm 20^\circ$), an inversion of the theory of gross bedform normal (Fenton et al., 2014; Rubin and Hunter, 1987), yielded a ratio of the two wind directions of ~ 1.3 . The dominant wind direction is therefore interpreted to be from the NW (337°), and the subordinate direction from the NE (74°). This interpreted bimodality is consistent with recent modeling results of barchan morphologies formed in bimodal winds (Parteli et al., 2014), and with the populations of mapped ripples and dunes in the Bagnold Dune Field (Hobbs et al., 2010; Silvestro et al., 2013). This

simplification considers only the mapped brinkline orientations and not the observed morphological shift that suggests a change in wind dominance from one end of the field to the other.

The Bagnold Dune Field is clearly an availability-limited system (Kocurek and Lancaster, 1999). The asymmetric morphologies are consistent with bedform formation in the sand-starved “fingering mode” identified by du Pont et al. (2014). Barchan and linear dunes have low equivalent sand thicknesses relative to other dune types (Lee and Thomas, 1995), and barchan dunes characterize areas of limited sand supply. The 9 km break in the dune field further reflects the limited sand supply in this system (Fig. 10C). In this zone, the dunes lose definition, shrink to small domes, and decay into sand sheets. Although the dunes evidence a clear history of sand transport, on the whole sands are being transported through this region and into the Western Dune Field, making the Bagnold Dune Field a bypass dune field rather than an ultimate sand sink.

Table 3: Summary of brinkline and linear dune measurements

Dune Field	Bagnold Dune Field	Western Dune Field	Northeastern Dune Field	Southern Dunes
Population 1 – Mean brinkline orientation (°)	67 - primary	35	18	339
Standard deviation (°)	15	27	33	28
Number of samples	97	553	38	120
Population 2 – Mean brinkline orientation (°)	344 - secondary	336	324	49
Standard deviation (°)	17	29	23	14
Number of samples	77	246	46	122
Population 3 – Mean brinkline orientation (°)	-	88	50	55
Standard deviation (°)	-	22	14	20
Number of samples	-	105	9	25
Mean linear dune orientation (°)	32	4	327	5
Standard deviation (°)	15	32	16	32
Number of samples	24	76	14	17

489

490 5.4.3 Western Dune Field

491 The Western Dune Field extends along the western margin of Mount Sharp, and is the
492 largest body of sand in Gale Crater (Fig. 15) (Hobbs et al., 2010). This dune field is ~34 km long
493 and is 15 km at its widest. As indicated on Figure 15B, the northern region of the dune field (20
494 km long) is separated from the southern region (10 km long) by a 4 km region of bypass
495 spanning the width of the field. Dunes funnel in and around antecedent topography associated
496 with nearly 700 m of elevation rise across the field (north to south). A mound of exposed
497 layered bedrock can be seen in the middle of the northern section just north of a 1.8 km diameter
498 crater, currently filled with dune sands (Fig. 15B). A dissected ridge marks the boundary
499 between the southern area and the region of bypass, and is associated with a slope break in the
500 local elevation (Hobbs et al., 2010, fig. 5). Clearly deriving sands from the Bagnold Dune Field
501 (Fig. 9), northern reaches of the Western Dune Field align with a retreating scarp (Fig. 16B),
502 possibly laterally equivalent to the “Cratered Surface” terrain unit encountered by *Curiosity*
503 (Anderson and Bell, 2010; Grotzinger et al., 2014). Initiation of the dune field at this scarp may
504 reflect flow deceleration associated with the negative topography, but erosion of this strata may
505 be an additional sediment source for the Western Dune Field.

506 Dune morphologies observed in the Western Dune Field are more diverse than those in
507 the Bagnold Dune Field. Starting in the north, the field is defined by a rampart of peripheral
508 barchans that coalesce to form crescentic and linear dunes with apparent migration toward the
509 center of the dune field (Fig. 16B). Barchans on the western edge of the dune field (Fig. 16A)
510 have well-defined SE-facing slipfaces, whereas barchans on the eastern edge (Fig. 16E) have
511 SW-facing slipfaces. Both sets of slipfaces face the field interior, and elongating barchan arms
512 are always on the north side. Barchans on the western edge coalesce into crescentic dunes (Fig.
513 16A), but lose definition as they migrate inward toward a zone of markedly smooth topography

514 that forms a halo around the dune field interior (Fig. 15B). Barchans on the eastern edge also
515 coalesce and develop extended arms (Fig. 16E), but this pattern yields to subdued linear dunes in
516 the interior on the eastern side of the filled crater (Figs. 15B). The central interior of the northern
517 region shows a complex dune pattern. Large primary linear dunes trend northeast, but
518 superimposed bedforms similar to terrestrial zibars (Nielson and Kocurek, 1986) occur in the
519 interdune areas and intersect the primary linear dunes at a high angle (Figs. 15B, 16C). This
520 complex pattern is reminiscent of star dunes formed in converging flows (e.g., Lancaster, 1989b;
521 Zhang et al., 2012).

522 The middle bypass region (Fig. 15B) is devoid of bedforms, except for scattered barchans
523 that range from 80-260 m in width (Fig. 16D). First observed at the dissected ridge, these
524 symmetric barchans are oriented to migrate north across the exposed plain, and rejoin sands in
525 the northern region of the field.

527 Figure 15. Western Dune Field. A. Bedform mapping in the Western Dune Field. Three
528 populations of slipface orientations were mapped in addition to trends of linear dunes. B. Western
529 Dune Field at the base of just visible light toned Mount Sharp strata to the east. Dashed white line
530 marks the edge of dune-dominated surfaces used in this area to estimate the volume of loose
531 sediment. The three regions discussed in section 5.4.4 and the filled crater are indicated (black
532 arrows). White boxes and annotations indicate the location of panels in Figure 16. CTX image
533 P13_006143_1745_XN_05S223W is oriented north-up. C. Rose diagram of populations of
534 slipfaces and linear dunes corresponding to measurements in A. Populations 1 and 2 are shown by
535 convention in the northern hemisphere, but Population 3 is shown together to prevent splitting of
536 the population.
537

Figure 16. Dune morphologies in the Western Dune Field. A. Peripheral symmetric barchans on the west edge of the Western Dune Field. Barchan dunes link with other barchans as they migrate to the SE, becoming increasingly asymmetric and forming crescentic ridges. Slipfaces on these dunes form Population 1 (Fig. 15A). B. Asymmetric barchan dunes at the northern edge of the

dune field form just downwind of a scarp, hypothesized to source some of the dune sands. C. NE-trending linear dunes with superimposed NW-trending crestlines, forming morphologies similar to star dunes. This morphology argues for convergent flows in this region. D. North-migrating symmetric barchans crossing the sand-starved bypass region. Slipfaces are oriented east-west (Population 3) and interpreted to form from southerly winds. Note the dissected ridge (arrow). E. Asymmetric barchans on the eastern margin of the dune field, opposite barchans in A. SW-facing slipfaces from Population 2 are sharpest in this area. F. Barchans and crescentic dunes in the southern region terminate in bedrock plains to the east. The elongated barchan arm (just left of the scale bar) ends before becoming a linear dune. All images have been contrast stretched to bring out variation in the scene, and north is up (CTX image P13_006143_1745_XN_05S223W).

In the southern region (Fig. 15B), barchan dunes link and coalesce to form a morphological continuum between barchan and crescentic dunes (Fig. 16F). Dunes in this region exhibit north-facing slipfaces (Fig. 16F), matching those of the symmetric barchans in the middle bypass region. Slipfaces on crescentic and barchan dunes become less defined toward the eastern edge of this region where dunes abruptly yield to a sand-free plain (Fig. 16F). Elongating defect arms along this eastern edge terminate before forming symmetric linear dunes.

Slipface orientations of the barchan and crescentic dunes and the trends of linear dunes were mapped in the Western Dune Field (Fig. 15A, Table 3), yielding three populations of barchans/crescentic dunes and a population of linear dunes (Fig. 15C). Population 1 largely consists of the dunes on the western side of the dune field, where their brinklines trend NE (35°) and dune orientation shows migration toward (125°) the dune-field interior. Population 2 largely consists of the dunes on the eastern side of the dune field, where their brinklines trend NNW (336°) and also show migration toward (246°) the dune-field interior. Population 3 characterizes dunes in the southern and bypass regions, where brinklines trend E-W (88°) and dunes are

oriented to migrate north toward (358°) the dune-field interior. The entire population of linear dunes trends N-S (04°), but those in the western interior trend NE, parallel to Population 1, whereas the more subdued linear dunes in the east and superimposed in the west trend NW, parallel to Population 2.

The dune morphologies, their orientations, and the spatial abundance of sand all indicate that wind patterns create a convergence toward the interior dune field in the northern region. Peripheral areas of the dune field show dunes largely resting upon exhumed bedrock (Figs. 16A-E), but in the central interior, dunes are their largest and interdune areas are sand-covered (Fig. 16C). The flow convergence toward the central interior is further manifested by the complex superimposed dune trends.

5.4.4 Northeastern Dune Field

The northeast quadrant of Gale Crater hosts a dark streak indicative of actively transported aeolian sand that stretches 37 km along the northeast basal margin of Mount Sharp and averages ~5 km wide (Fig. 17). High resolution coverage in this dune field is particularly limited, but the area appears to be a mix of dunes, and sand sheets. Following the field from NW to SE, the beginning of the field is marked by symmetric barchans with well-defined east-facing slipfaces. These dunes grade into asymmetric dunes with southern arms elongating to the SE, roughly parallel to the dune field trend (Fig. 18A). Continuing to the south, barchans on the western edge lose definition and grade into apparent sand sheets, such that only about half of the ~5 km width of the dark streak appears to house enough sand to give rise to visible dunes (Fig. 17B). Linear streaks within the sand sheets are oriented parallel to the dune field trend. Further south, bedforms make up increasingly more of the low albedo width. Asymmetric barchans form a rampart margin on the eastern edge of the dune field (Fig. 18B), but their slipfaces are subdued in comparison to the dunes in the north. Where resolvable, these dunes have brinklines trending NW with steep slipfaces toward the interior. Apparent linear dunes also occur within the main

dune field, but their symmetry is difficult to determine with the limited resolution available. About 27 km from its northern margin, the dune field is interrupted by a 4.8 km diameter crater (Fig. 17B). Dune sands fill the crater with a crescentic “bullseye” morphology commonly observed in Martian craters (Hayward et al., 2014). Dark sands are also seen streaming around the exterior of the uplifted crater rim. Finally, at the SE end of the dune field, sands coalesce into several large crescentic dunes with steep SE-facing slopes (Fig. 18C). Ripple orientations on these dunes strongly follow the ridgeline orientation, with interference patterning near the crest attributed to changes in bedform slope (Fig. 18D).

Figure 17. Northeastern Dune Field. A. Bedform mapping in the Northeastern Dune Field. Three populations of slipface orientations were mapped in addition to the linear dune orientations. Rose diagram shows histogram of the measured orientations. Population 3 is shown in the southern hemisphere to limit overlap. B. Northeastern Dune Field at the base of dissected Mount Sharp strata to the SW. Dashed line corresponds with dashed line in A and was used to estimate the

607 volume of dune sand in this region. White boxes show the locations of panels in Figure 18. North-
608 up oriented CTX image G23_027069_1748_XN_05S221W.

609
610 Figure 18. Dune morphologies in the Northeastern Dune Field. A. Barchans becoming
611 increasingly asymmetric and linking to form chains of defect arms elongating to the SE. Note the
612 sharp brinklines defining the slipface Population 1. HiRISE image ESP_026291_1750. B. NW-
613 trending asymmetric linear dunes with steep north-facing slopes. Note the degraded teardrop

barchan (arrow) and its two slipfaces. C. Crescentic dune with steep SE-facing slope forming south of the 4.8 km diameter crater. White box shows the location of D. D. Ripples on crescentic dune surface are parallel to the crest on both the NW stoss and at the base of the steep lee face to the SE. Note the interference patterning near the brinkline. All images have been contrast stretched and are oriented north-up. HiRISE image ESP_026502_1750 covers B, C, and D.

Mapped brinkline orientations north of the 4.8 km crater define three barchans/crescentic dune populations that, as with the Western Dune Field, suggest that the dune field formed by flow convergence (Fig. 17A, Table 3). Population 1 brinklines, characterizing the dunes in the NW (Fig. 18A), trend NNE (18°) with dune migration toward (108°) the interior. Population 2 brinklines, characterizing the eastern periphery dunes, trend NW (324°) with dune migration again toward (234°) the interior. Population 3 brinklines characterize only dunes in the southern reaches of the field, and these trend NE (50°) with dune migration toward (340°) the interior. Linear streaks in the sand sheets and linear dunes in the main dune field trend 327° , roughly parallel to the dune field and potentially define the mean transport direction that gives rise to the migration directions for dunes in Population 1-3, as expected within a sand-limited system (du Pont et al., 2014).

5.4.5 Southern Dunes

The southern floor of Gale Crater hosts a swath of bedforms ~100 km long and 5 km wide (Fig. 19). Unlike the other fields, which form at the base of Mount Sharp, these dunes form far from the mound and instead are associated with steep rim topography. Morphologies in this region are again dominated by a continuum between barchan, crescentic, and linear dunes (Lee and Thomas, 1995). Barchans dunes seen across the field form tear-drop morphologies that elongate into linear dunes (Fig. 20D) or coalesce into crescentic dunes (Fig. 20B). Topography from raised ridges and small hills clearly disrupts bedform morphology in southern Gale Crater

(Figs. 20A, 20C). This region also appears sediment-starved. Bedrock plains can be seen in most interdune areas, and dunes grade into sand sheets to the north and laterally (Fig. 20C).

Figure 19. Southern Dunes A. Southern Dunes distributed across the floor of Gale Crater over north-up oriented CTX mosaic. Dashed line corresponds with the dashed line in B and was used to estimate the volume of dune sand in this region. Dunes are spread along the base of the crater rim and absent close to the base of Mount Sharp. White boxes show the locations of panels in Figure 20. B. Mapping of the Southern Dunes. Three populations of slipface orientation were mapped in addition to the linear dunes. Populations are shown in the rose diagram in both the north and south hemispheres to prevent overlap, and highlight the difference between north-facing Populations 1 and 2 in this region and south-facing Populations 1 and 2 in the other dune fields. Note the variation in linear dune trend from west to east that roughly follows normal to crater rim curvature.

The orientations of resolvable brinklines and linear dunes were mapped across southern Gale Crater (Fig. 19, Table 3). Three populations of brinkline orientations emerged, in addition to the linear dune trends. In the west, the sharpest and most abundant brinklines (Population 1) trend NW (339°) with lee faces oriented toward the NE (69°). In the east, however, the sharpest brinklines (Population 2) trend NE (49°) with lee faces oriented toward the NW (319°). The trend of the linear dunes varies from west to east as well. Trending on average N-S (05°), the linear dunes form perpendicular to the strike of the rim, swinging from a NE-SW trend in the west to a NW-SE trend in the east. The radial pattern in the orientations of both the linear and barchan dunes suggests that the formative winds for these southern dunes is associated with flow down rim topography and toward the crater interior (Hobbs et al., 2010).

A third population of brinklines, trending NE (55°) but with SE-facing (145°) lee faces, was observed only in the western end of the area (Fig. 19). Figure 20A shows where these Population 3 slipfaces interact with Population 1 slipfaces, capturing the transition from dominantly north winds in the Western Dune Field to south winds in the Southern Dunes region.

Figure 20. Dune morphologies in southern Gale Crater. A. Teardrop barchan dunes with slipfaces from Populations 1 (solid arrows) and 3 (dotted arrows) depicting the transition from dominantly north to southerly winds. Note divergence around hill topography and gradation into linear dunes streaming toward the crater center (NNE). CTX image P15_006855_1746_XN_05S222W. B. Crescentic ridges with steep NW-facing lee slopes form along from the opposite end of this region from A. Note that the orientation has swung around such that inferred wind direction is still normal to the crater rim. CTX image P01_001554_1754_XI_05S221W. C. Asymmetric barchans elongating to the north and evidencing two populations of slipfaces (Populations 1 and 2). Note the bare interdune areas and gradation to sand sheets to the north. CTX image

P01_001620_1749_XI_05S222W. D. Sinuous crescentic ridge splitting into tear-drop dunes as defects elongate to the north. Note the two populations of slipfaces also observed in C. CTX image P14_006644_1747_XI_05S222W. All images have been contrast stretched and are oriented north-up.

6. Surface observations

The traverse of *Curiosity* from its landing site at Bradbury Landing to the base of Mount Sharp at Pahrump Hills allowed for field observations of landscapes that can otherwise only be viewed from orbit. Toward the goal of characterizing this interpreted aeolian landscape, terrain along the first 960 sols of the traverse was surveyed using Navcam mosaics to determine landscape morphologies typical of this area of Gale Crater (Fig. 21). The observed landscapes can be grouped into four broad categories: rocky plains, undercut bedrock, capped mesas, and remnant hills (Fig. 22).

As of sol 960, *Curiosity* had traveled 8.3 km across the floor of Gale Crater, moving generally up-section for a total elevation gain of ~70 m. The traverse aimed to capture the diversity of landforms and chemistry observed from orbit, targeting several named waypoints to characterize distinct units in the stratigraphy (Fig. 21). The four landscapes discussed in this section are not unique to any one terrain or stratigraphic unit (Anderson and Bell, 2010; Grotzinger et al., 2014), but rather are broadly representative of processes repeatedly seen shaping the crater interior. In addition, some of the features observed in these landscapes indicate a formative wind direction, thus adding to the body of wind directions inferred from orbital observations.

Figure 21. *Curiosity* rover traverse from sols 1-960 covering Bradbury Landing to Pahrump Hills over HiRISE mosaic. Waypoints discussed in section 6, and approximate sol when they were first reached, are indicated by the black dots. This image also covers barchan and linear dunes in the main region of the Bagnold Dune Field. North is up.

Figure 22. Orbital view of the four landscapes classified in section 6 interacting near the Kimberley waypoint. This view shows all four stages of the erosional process. Capped mesas (solid line) once overlaid the Kimberly (star), but have retreated to the northwest, leaving behind remnant hills (dashed lines) and rocky plains (dotted area). At the eastern base of the remnant hills we see where plains have eroded down to a new bedrock level, already forming low scarps (dotted line) that will eventually become the capped mesas. White line shows the approximate rover traverse, and the star indicates the rover position when Mastcam mosaic in Figure 29 was taken. North-up oriented HiRISE image ESP_036128_1755.

713 6.1 Rocky plains

714 The most common landscape observed along the traverse can be characterized as a flat
715 plain with a varying amounts of scattered rock sitting atop a hard, dusty surface (Fig. 23A).
716 Rocky plains are captured in nearly all scenes from the first 500 sols, after which they are less
717 dominant, but still ubiquitously observed. The “Rugged” and “Hummocky Plains” terrain units,
718 mapped in Grotzinger et al. (2014), are overwhelmingly dominated by this landscape. The
719 distribution of rock across the surface appears pseudo-random with no clustering or preferential
720 alignment between rocks. Unassociated with outcrop, these rocks come from no obvious near-
721 field source and range in size from pebbles to decimeter-sized cobbles. Transport by wind of a
722 ~3 cm clast was observed by the Viking 1 lander after a substantial storm (Arvidson et al., 1983),
723 but many cobbles on these plains are too large to have been transported by winds. Rather, rocks
724 on these plains are most similar to terrestrial desert pavements. Formed when fine material is
725 deflated away from a surface faster than rock is weathered and eroded, desert pavements armor
726 planar surfaces with tightly packed rock, slowing and ultimately halting aeolian deflation of the
727 surface (Cooke, 1970). In some cases, desert pavement can even be aggrading, raising surface
728 stones with kinetic sieving as trapped fines are vibrated to below the armored surface (McFadden
729 et al., 1987), similar to the hypothesized TAR formation mechanism (Bridges et al., 2015). The
730 Gale Crater surface rock armor is typically less closely-packed than some interlocking
731 pavements observed on Earth (e.g., Death Valley, CA; Haff, 2001), but the spatial density of
732 clasts on the surface (Fig. 23C) is sufficient to raise the aerodynamic surface roughness to
733 preclude most deflation, making these surfaces stable.

Figure 23. Rocky plains landscape. A. Mastcam mosaic of flat cobbled plains taken on sol 541 (mcam02134). Plains show dust-filled fractures from underlying bedrock (1) and exposures of light toned bedrock (2). Foreground wheel track (3) is 30 cm wide (Mastcam mosaic starting at 0541ML0021340010203723E01_DRCX). B. Ventifact observed on sol 694. Note the elongated pits and fluting on the right hand face as compared with the blunted (upwind) left face. Image is ~40 cm wide (Mastcam image 0694MR0029330020402220E01_DRCX). C. Light toned dust

mantling rocky cobble surface is disturbed by wheel tracks that cut into the surface and reveal dark-toned regolith (Mastcam mosaic from sol 456 starting at 0456ML0018370000201931E01_DRCX).

Pebbles and cobbles littering the rocky plains commonly exhibit fluting, lineations and the distinctive polish characteristic of ventifacts formed from persistent aeolian abrasion (Fig. 23B), reflecting the long exposure of these stones on the surface (Laity and Bridges, 2009). Indicators of wind direction, measured in ventifacts along the first 100 sols of *Curiosity*'s traverse, recorded a relatively wide NW-SW range of winds (Bridges et al., 2014). Comparison between ChemCam observations of the rock called Jake Matijevic, a well-defined ventifact near the Rocknest outcrop in Yellowknife Bay, and less abraded bedrock indicated that the high lithium rind seen on the bedrock outcrop is absent on highly ventifacted surfaces (Bridges et al., 2014). This argues that abrasion has removed the coating, or is resurfacing the rock faster than the rind can form, confirming the dominance of physical weathering over chemical. Heavily ventifacted rocks have also been observed in similar landscapes of rocky plains seen at the Pathfinder (Greeley et al., 1999), Viking 1 (Arvidson et al., 1983), and Mars Exploration Rover (MER) *Spirit* (Sullivan et al., 2008, fig. 50) landing sites.

In Gale Crater, rocky plains landscapes are typically devoid of bedforms, consistent with the interpretation that these surfaces are highly deflated, and show a paucity of transport-available sand at the surface. Unlike the abundant ripples observed on rocky plains by *Spirit* (Sullivan et al., 2008, fig. 28), only occasional sand shadows were encountered along *Curiosity*'s traverse across the rocky plains (Fig. 24A). Sand shadows are linear accumulations of sand that are deposited in the lee of obstacles even on sand-starved surfaces (Hesp, 1981), thus making them reliable indicators of the last significant wind direction. Measurement of the orientation of 69 sand shadows, from Yellowknife Bay to Pahrump Hills, generated a mean orientation of 09°

with standard deviation of 0.5° (Fig. 24B, Table 2), thereby indicating a narrowly constrained north formative wind. Ranging from 12 cm to 3.74 m long, these accumulations are similar to those observed at the Pathfinder landing site (Greeley et al., 1999, fig. 2). Despite an abundance of available obstacles, sand shadows and other bedforms were absent from the plains from sols 361-369, 429-439, and 634-644.

The Rocknest sand shadow, observed in Yellowknife Bay from sol 57 to 100, is the most studied sand shadow to date in conjunction with the first scooping activities by *Curiosity* (e.g., Blake et al., 2013). Also interpreted to have formed by northerly winds, this sand shadow exhibits a 2-3 mm thick armored surface formed by 1-2 mm rounded, dust-rimmed grains, which formed cohesive rafts during the excavations (Minitti et al., 2013). In contrast to the armored surface, the sand shadow interior consists of fine sediment largely below the resolution of MAHLI ($\sim 150\ \mu\text{m}$). This contrast between the coarse armored surface and the fine dusty interior suggests that the sand shadow may have originated as a coarse-grained feature that subsequently grew by the infiltration of dust and fine sand into the interior, similar to the dynamics of TARs described above.

Figure 24. Sand shadow surface observations. A. Sand shadow (50 cm long) accumulating behind a float rock obstacle in rocky plains (sol 333 Mastcam image 0333ML0013490070107979E01_DRCX). B. HiRISE map of the *Curiosity* traverse with measured sand shadow orientation superimposed on location of observation. Where multiple sand shadows were observed in a single mosaic, the average orientation is shown. Rose diagram shows the orientation and inferred upwind azimuth of all 69 measured sand shadows.

788 6.2 Undercut bedrock

789 Bedrock outcrops are seen consistently every few hundred meters for the length of the
790 traverse. Exposures are typically heavily fractured, with fractures forming local sediment traps
791 filled with dust or sand (Fig. 25). Similarly fractured outcrops were observed near Home Plate by
792 *Spirit* (Sullivan et al., 2008, fig. 43), and at the Burns formation by *Opportunity* (Grotzinger et
793 al., 2005, fig. 4). Exposures of bedrock are commonly undercut, forming small scarps, and
794 decimeter-scale relief shelters N-S trending sand shadows (Fig. 25). Float rocks can be seen on
795 top of some exposed bedrock, and in clusters at the base of small scarps, tending to decrease in
796 size away from the outcrop. Outcropping bedrock is typically surrounded by rocky plains, and
797 scalloping and polish of bedrock surfaces evidence long exposure to aeolian abrasion.

798
799 Figure 25. Undercut bedrock outcrop at Cooperstown (sol 440 Navcam mosaic centered on image
800 NLB_436560260EDR_F0220000NCAM00259M). Float rock is visible atop bedrock exposure
801 (1). Exposed surface is heavily fractured with sand and dust filling most fractures (2). Arrows
802 indicate small sand shadows forming in the lee of scarp float boulders, including the 40 cm
803 boulder indicate by (3).

804

805 Close examination of exposed bedrock in Yellowknife Bay revealed streamlined nodules
806 in the bedrock surface. Similar to concretions observed by *Opportunity* in the Burns formation

(Grotzinger et al., 2005), these diagenetic features range from 1-3 mm in diameter. Nodules in the Sheepbed mudstone (Grotzinger et al., 2014) appear partially exhumed with one blunted edge and opposing tail of bedrock all aligned locally in the same direction (345°) (Fig. 26). These features are interpreted as caused by direct wind abrasion of bedrock from NW winds (Stack et al., 2014). A surface exposure age of 78 ± 30 Ma was measured on the Sheepbed mudstone at the Cumberland drill target (Farley et al., 2014). Cumberland is located ~1 m from an overlying scarp of Gillespie Lake unit, implying a relatively slow modern erosion rate of ~1 m of scarp retreat per 100 Ma.

Figure 26. Streamlined nodules in the Sheepbed mudstone found at Yellowknife Bay. A. In-place nodules being exhumed from bedrock (sol 298 Mastcam mosaic starting at 0298MR0012480590203737E01_DRCX). B. Same image as A but with orientations of streamlines superimposed on leeward tails. Average orientation is 345° with inferred wind from the NE.

824 6.3 Capped mesas

825 Capped mesas become increasingly common along the traverse (Fig. 27). First observed
826 on either side of Dingo Gap, mesas commonly form along the edges of the “Cratered Surface”
827 terrain unit (Grotzinger et al., 2014), and on opposing sides of valleys (e.g., Hidden Valley,
828 Dingo Gap). These flat-topped landforms are defined by 2-4 m high hills capped by scarps of
829 retreating competent rock typically ~1 m thick. Below the scarps, decimeter- to meter-sized
830 boulders, apparently sourced from the caprock, armor dusty slopes. Similar observations of
831 eroded float rock are seen coming from capping layers in the Burns Formation along the
832 *Opportunity* traverse (Grotzinger et al., 2005). In Gale Crater, ventifacting was observed on float
833 rock near the base of slopes, but fluting and polish textures could not be resolved nearer to the
834 hilltops. Sand shadows were commonly observed in association with basal slope boulders, and in
835 some cases the topography sheltered fields of ripples (Fig. 27).

836
837 Figure 27. Retreating scarps cap the northwest wall of Hidden Valley, forming a flat-topped mesa
838 (1). Boulders on the lower slopes (2) are partially covered by ripples (3) and appear to be sourced

839 from the resistant caprock. Vertical relief in this image is ~ 4 m (sol 707 Mastcam images
840 0707ML0030050030304604E01_DRCX and 0707ML0030050020304603E01_DRCX).

841
842 Figure 28. Dingo Gap TAR. A. Valley-spanning TAR at Dingo Gap as observed from orbit
843 (HiRISE image ESP_036128_1755). B. East-facing Mastcam mosaic of *Curiosity* tracks across
844 Dingo Gap TAR. Tracks are ~30 cm wide (sol 538, starting image
845 0538ML0021240000203532E01_DRCX). C. MAHLI image of Dingo Gap surface. Rounded
846 grains cover ~20% of the TAR surface in contrast to more tightly packed armoring seen on the
847 Rocknest sand shadow. Note the thin coating of dust on all grains. Crescent impression on the

surface is from *Curiosity's* Alpha Particle X-ray Spectrometer (APXS) on target Barker (sol 531, 0531MH0002990010201151C00). D. MAHLI image wheel track wall target Dampier. Wall shows subsurface to be dominated by dust-sized fines, rafting at the dusty surface implies some cohesion of dust (sol 531, 0531MH0003470010201183C00).

On sol 535 of its traverse, *Curiosity* crossed the TAR named Dingo Gap, spanning ~12 meters between two capped mesas. MAHLI images revealed the TAR to be topped with rounded 1-2 mm grains covering ~20% of the surface, similar to the Rocknest sand shadow but not as tightly packed (Fig. 28). These armoring grains are similarly coated in dust that fell away after contact with rover hardware. Imaging of the wheel tracks revealed the interior of the TAR to be dominated by the fine dust particles, with some rounded grains supported in a dust matrix. Surface armoring was also observed by *Spirit* on the El Dorado ripple field, but surface grains were only ~300 microns, and much more closely packed (Sullivan et al., 2008). Despite their differences, bedforms such as the Rocknest sand shadow, the Dingo Gap TAR, and the El Dorado ripples appear to represent a common class of Martian bedforms sharing coarse-grained armored surfaces and fine interiors. The well-rounded grain surfaces indicate that these sands are not freshly eroded from bedrock, but rather have been transported sufficiently to generate near-spherical rounding. As described above, these bedforms may also share a dynamic of surface stabilization with continued vertical growth by dust infiltration.

6.4 Remnant hills

Small hills, not capped by erosionally resistant rock layers, increase in frequency along the traverse in conjunction with the increasing frequency of the capped mesas. Characterized by rounded tops, slopes on remnant hills in Gale Crater are typically shallower than slopes below capped mesas, but total relief is similar (1-3 m). These hills exhibit the same armoring of decimeter- to meter-scale boulders seen below capped mesas (Fig. 29). Float rock was measured in a ~50 m transect from the top of a hill just beyond Dingo Gap, down the slope, and across the

873 plains until float rock size became uniform with the surrounding area (Fig. 30). Choosing the
874 largest float rocks in the scene, rocks identified in the high resolution Mastcam mosaic were
875 measured using stereo Navcam mosaic. Distance from the rover was used as an approximation
876 for distance along the transect, and is differenced to be relative to the hilltop in Figure 30.
877 Measured diameters decreased in size away from the hilltop, following roughly an exponential
878 decay. Though resolution in this area was not high enough to resolve ventifact textures,
879 ventifacts were observed at the base of remnant hills later in the traverse (e.g., at the Kimberley).
880 No observations of larger hill-boulders were captured in enough resolution to see fluting or pits.
881 Small accumulations of sand in the form of ripples or sand shadows were again found against
882 these hills, but remnant hill TARs have not yet been observed. *Spirit* observations of McMurdo
883 capture rock-covered hillslopes with similar morphology (Greeley et al., 2008).

884
885 Figure 29. Remnant hill observed at the Kimberley. Boulders cover the rounded hillslopes (1),
886 and outcrops of retreating bedrock scarps can be seen in the foreground (2). Relief in exposed
887 bedrock shelters 10-20 cm sand shadows (3). East-facing image was taken at the location

888 indicated in Figure 22 (sol 603Mastcam mosaic starting at image
889 0603ML0025440010301349E01_DRCX).

890
891 Figure 30. Float rock measurements on a remnant hill A. Graph shows decreasing rock size with
892 distance from the hilltop shown in B. Rover-relative distances were calculated and used as a
893 proxy for distance along the transect. The size decrease is interpreted to reflect less first-cycle
894 aeolian erosion on young hilltop rocks and more extensive erosion on older rock further from the
895 degraded scarp. B. Transect and measured float on a remnant hill just beyond Dingo Gap (sol
896 538 Mastcam image 0538ML0021250000203544E01_DRCX). White lines indicate the
897 diameters measured on the largest boulders at various elevation levels in the scene.

898
899 *6.5 Landscape evolution by wind and gravity*

900 The four landscapes observed along the traverse are proposed to represent a spectrum of
901 progressive stages in a cycle of landscape evolution where aeolian erosion and gravity-driven
902 mass wasting, coupled with physical weathering, compete to level the surface. These processes,

operating over the last 3 Ga, are taken as necessary and sufficient for the removal of the $\sim 6 \times 10^4$ km³ of sediment from Gale Crater needed for the exhumation of Mount Sharp.

Physical weathering by impact bombardment and thermal stressing appears to be the main cause of ubiquitous fractures observed along the *Curiosity* traverse and elsewhere on Mars (e.g., Sullivan et al., 2008; Grotzinger et al., 2005). Gale Crater has been subjected to impact bombardment since its formation at ~ 3.6 Ga (Le Deit et al., 2013; Hartmann and Neukum, 2001). At the surface, small craters in Gale are seen ringed by blocks of fractured rock (Newsom et al., 2015, fig. 8). Millennia of thermal cycling further weathers competent Martian rock as repeated expansion and contraction stresses and fractures the material (Viles et al., 2010). Mars temperature fluctuations occur on a variety of scales, from diurnal to Milankovitch, around a cold mean annual equatorial temperatures of ~ 220 K (Schorghofer, 2008). Bedding planes and polygonal fractures provide additional zones of weakness along which bedrock can fracture.

Unlike on Earth, first-cycle aeolian erosion of bedrock is significant on Mars, and occurs because of long exposure of surfaces and the increased energy of grain impacts (Kok, 2010) even with infrequent significant wind events. Aeolian abrasion is evident in Gale Crater from yardangs, ventifacts, fluting and polished surfaces of both bedrock and float, but first-cycle abrasion of rock and continued physical weathering must also account for the diminishing of clast size on the surface with distance from bedrock. In all the traversed landscapes, sand was absent except where it was trapped by local topography, and the volume of sand held in dune fields is a tiny fraction of the amount of sediment created during the exhumation of Mount Sharp. This relative paucity of loose sand on the surface of Gale Crater indicates that potential sand transport from the surface greatly exceeds the production of sand on the surface (“availability-limited” system of Kocurek and Lancaster, 1999). Although rates of erosion have varied greatly in Gale Crater, over geologic time wind transport has been sufficient to exhume nearly all of weathered material from the crater.

The progressive landscape evolution evident along the *Curiosity* traverse is idealized in Figure 31. Exposed fractured bedrock is subjected to aeolian abrasion and continued physical weathering. Fractures in the outcrop form nucleation points for increased weathering and erosion, funneling local flows that widen and deepen fractures by sand abrasion. Continued breakdown of bedrock and aeolian undercutting cause gravity failure of overhanging scarps. Weathering and abrasion then reduce the size of these fallen clasts. When no longer covered by resistant caprock, underlying less-resistant layers deflate at a much faster rate. Areas initially flooded by resistant caprock deflate down as caprock retreats to mesas. When the caprock is completely eroded, remnant hills with a residual cover of disjointed rocks remain. Hill topography is smoothed to flat, and the final surface resulting from this denudation is a desert pavement. On this armored surface, deflation is severely limited, but breaks in the pavement pattern provide holes where erosion is enhanced and the pavement can be eroded from both the surface and sides. Disruptions in the pavement can stem from impacts, fluvial dissection, or the natural uneven distribution of surface clasts. Over geologic time, first cycle erosion by abrasion can also break through the pavement, shrinking cobbles and exposing the surface below. Continued surface deflation by physical weathering and abrasion of surface clasts must ultimately lead to underlying bedrock exposure and a renewal of the cycle.

Despite the absence of fluvial or tectonic processes, the proposed landscape evolution model reflects the same patterns of erosion in classical terrestrial landscape evolution models (e.g., Davis, 1899; Gilbert, 1909). Both historical and the proposed aeolian model show the back-stepping that occurs in differentially resistant layers, and the progressive rounding of hillslopes to flat surfaces. The observation of King (1953) that all fluvially-sculpted landscapes share the same morphologies seems to hold true even when the sculpting fluid is air. The major difference is instead interpreted to be the time scale over which these landscapes form.

Figure 31. Landscape evolution model. A. Bedrock is exposed at the surface and fractures form nucleation points for enhanced erosion. B. Fractures in bedrock widen and cut downward through the competent layer, eventually undercutting the outcrop and pieces of outcrop spall off. Some sand shadows accumulate in the lee of local topography. C. Underlying layers erode faster than bedrock, creating mesas capped by fractured bedrock. Float rock spalled off of the capping unit

coats the hillslope. In general, rocks further from the hilltop are older, more heavily ventifacted, and smaller. Sand shadows and larger bedforms are sheltered behind meter-scale topography. D. Eventually the capping unit is eroded away, leaving a remnant hill covered by ventifacted fragments of the capping unit. Float rock far from the hilltop has been subjected to more aeolian abrasion than more recently formed float near the hilltop, creating a decrease in rock size down the hill. Sand still accumulates in this meter-scale relief. E. Hill regolith is winnowed away, leaving a heavily ventifacted desert pavement that slowly abrades and deflates until new fractured bedrock is exposed.

7. Discussion and conclusions

7.1 Reconstruction of Gale Crater wind regime over time

The range of aeolian features in Gale Crater must have formed over orders of magnitude of time. Although their exact age is unknown, the oldest wind indicators are the exhumation of Mount Sharp, its streamlined morphology, and the position of the Gale Crater wind streak, all of which are consistent with northerly formative winds. These long-term wind indicators require first-cycle erosion of bedrock by the wind, as also directly reflected by the yardangs and ventifacts. Yardangs are overwhelmingly oriented N-S, consistent with northerly winds. At much finer spatial and temporal scales, streamlined nodules carved from bedrock at Yellowknife Bay also indicate a northerly wind. Ventifact measurements, some of which are on bedrock, are less consistent and the interpreted westerly winds are at odds with the other wind indicators. Bridges et al. (2014) suggested that the ventifacts may reflect rare high-speed winds, but given the consistency of the other wind indicators, the ventifacts more likely reflect flow affected by local topography or subsequent clast movement. With the exception of ventifacts, therefore, the long time-scale wind indicators overwhelmingly argue that winds from the north have persisted over the geologic-scale time spans over which Mount Sharp formed (Fig. 32).

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Figure 32. Inferred wind directions from aeolian features forming over long time-scales in Gale Crater. Yardang orientation varies slightly (12°) from the east to west of Mount Sharp, and is interpreted to result from flow divergence around the 5 km high mound. TAR orientations are shown here in representation of the two dominant modes at locations where TARs are most dense. Both TARs and ventifacts cover a wide range of inferred winds and are thought to be poor indicators of regional circulation because of the influence of topography. The Gale Crater wind

streak extends south of this image (Fig. 3). North-up oriented HRSC background mosaic is also used in all following figures.

Inferred wind directions from features forming over shorter time-scales reveal a regional north wind that is profoundly modified by crater-scale secondary flow. Crater rim and mound topography delineate the crater moat, and features in the moat are divided between northern and southern halves (Fig. 33). Primary wind from the north is argued by the: (1) deflationary pavements and absence of sand dunes at the sand-starved northern apex of Mount Sharp, (2) orientation of sand shadows, and (3) net southward transport of dunes in the Bagnold and Northeastern dune fields. In the northern half of the crater, secondary flow deflection around the topography of Mount Sharp is reflected by the: (1) channelization of dune fields along the flanks of Mount Sharp, (2) SW-facing slipfaces in the Bagnold Dune Field, and (3) SE-facing slipfaces in the Northeastern Dune Field. A second component of secondary flow is evident by: (1) crater-rim downward flow evident in all the northern half dune fields, and (2) mound downward flow evident in the Western Dune Field. The interplay of the primary and secondary winds, plus the intrusion of southerly winds characteristic of the southern half of the crater moat, appear responsible for the flow convergence evident in the Western and Northeastern dune fields. In the southern half of the crater floor, small wind streaks on the crater rim and the dominant orientation of the Southern Dunes must arise from crater-rim downward flow. The convergence of northern and southern moat winds regimes is again evident in the northwestern area of the Southern Dunes. The TARs are not addressed here because their orientations appear strongly controlled by local topography, thus making them poor general wind indicators. It is nevertheless notable that TAR orientations, measured largely in the northern half of the crater, are dominated by two directions (71° and 337°) interpreted to result from the NW and NE winds evident from the other indicators.

Figure 33. Inferred recent wind directions from aeolian features forming on short time-scales in Gale Crater. Wind indicators are divided, with interpreted northerly winds dominating the northern half and southerly winds dominating the southern half. Note the relative association of wind direction and crater rim orientation in the south, and the apparent divergence of wind directions around Mount Sharp in the north.

1022 Integrating all of the wind indicators argues for a crater wind regime with components of
1023 regional winds, topography-deflected secondary winds, and topography-created katabatic winds
1024 (Fig. 34). The relative impact of each component has evolved with exhumation of the mound
1025 and, potentially, global climate. Regional primary winds from the north must have dominated
1026 during the span of time over which Mount Sharp was exhumed, and emergence of Mount Sharp
1027 by excavation would have greatly altered the wind pattern through time. As the moat was
1028 formed, topography-deflected secondary flows included funneling of the wind along the mound
1029 flanks, and development of a wake in the southern lee of Mount Sharp. Downward flows on the
1030 crater-rim and mound are interpreted as topography-created katabatic winds, which have been
1031 emphasized in circulation models of Gale Crater (Tyler and Barnes, 2013; Hobbs et al., 2010).
1032 Katabatic winds form from atmospheric density gradients caused by fluctuations in temperature,
1033 and can set up in opposition to large-scale regional flows (Rafkin et al., 2001). Such winds
1034 would flow radially into the crater moat from the top of the crater rim, as is evident from the
1035 recently-formed wind indicators, or from the mound, as is evident in the Western Dune Field
1036 (Fig. 33). Katabatic winds would reinforce regional winds in the north, and lee return flows
1037 behind Mount Sharp in the south. Current wind indicators argue for a modern dominance of
1038 crater-rim katabatic winds (Fig. 33), in contrast to the long-time wind indicators that argue for a
1039 dominance of regional northerly winds (Fig. 32).

Figure 34. Circulation model for Gale Crater interior. Regional northerly winds (solid lines) diverge as they encounter Mount Sharp, funneling between mound and rim to the east and west. Regional flows shed eddies and turbulent vortices in the wake of Mount Sharp. Katabatic slope-flow winds (dashed lines) flow inward down rim topography, dominating short time-scale morphologies. Eddies and katabatic flows oppose regional winds and account for convergence in

the Western and Northeastern Dune Fields. Line thickness and arrow size are schematic representations of relative wind strength.

The above analysis is based upon the interpretation of geomorphic features that span a range of spatial and temporal scales. Modeling of wind circulation in Gale Crater is typically conducted for a particular set of conditions, rather than integrated over geologic spans of time. Hobbs et al. (2010) modeled circulation in Gale at ~4 km scale resolution under dust-storm atmospheric conditions. The model revealed a dominance of regional NNW winds and topography-driven deflection in agreement with the above results. During the two modeled Martian sols, both katabatic (downslope) and anabatic (upslope) winds occurred, but these were weak compared to the regional NNW winds. Tyler and Barnes (2013) modeled the mean diurnal cycle in Gale Crater over 20 sols and at the same ~4 km resolution. Their model highlights the diurnal changes in wind regime, characterized by regional north winds deflected by topography in the daytime, with strong katabatic winds flowing down Mount Sharp and the crater rim at night. These models both agree with the conclusions of this study, but further indicate that variations in the crater wind regime are ongoing, rather than changing gradually with the exhumation of Mount Sharp. The suggested daily variations in wind regime are perhaps poorly represented in even the shortest time-scale aeolian features, and as such, our interpretation represents an integrated view of winds in Gale Crater over geologic time.

7.2 Aeolian-dominated landscape evolution

At all scales of observation, from orbital images to surface views by *Curiosity*, the landscape in Gale Crater is dominated by aeolian features. Evidence for subaqueous processes occurs only in the stratigraphic record and in relict or exhumed features, whereas no contemporary features were observed characteristic of landscapes eroded by water (e.g., dendritic drainage networks). This supports the representation of the landscape of Gale Crater as

one fashioned by physical weathering, gravity, and aeolian processes. Most of the landscape represents a sand-starved, erosional system, but the available sand is organized into aeolian bedforms comprising a relatively small volume of sediment. This observation is consistent with the interpretation that the inferred processes (i.e., physical weathering, downslope wasting, and aeolian abrasion and deflation) are sufficient for moat excavation and Mount Sharp exhumation from an initially filled crater.

The landscape patterns observed at the surface argue that denudation occurs in progressive stages from continuous bedrock-capped layers, to separated bedrock-capped mesas, to remnant hills capped by bedrock clasts, to desert pavements on rocky plains. Processes that drive the system include bedrock fracturing by impacts and thermal stress, aeolian abrasion that diminishes bedrock and widens fractures, and aeolian deflation that removes sediment. Preferential aeolian erosion of less resistant beds creates overhanging scarps that fail due to gravity. One implication of these processes is that first-cycle erosion of bedrock, directly evident by yardangs and ventifacts, is otherwise widely present on the Martian surface. A second implication is that desert pavements, generally considered as stable features on Earth, must undergo continued aeolian abrasion and deflation, and are ultimately removed on Mars. Given the documented extremely slow weathering rates on Mars and the apparent infrequency of high-energy wind events, landscape evolution in Gale Crater must be accredited to either the geologically long periods over which a surface is exposed, or much higher rates of surface denudation under a very different climatic regime at the time of initial excavation.

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