RESEARCH ARTICLE

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Key Points:

- Dark topographic slumps are observed on the slopes of Juventae Chasma on Mars
- Ten active slumps have been observed in JC, all of which formed in or near the same season
- Intriguing spatial and temporal correlations with recurring slope lineae and atmospheric obscurations are observed

Supporting Information:

- Supporting Information S1
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Seasonal Slumps in Juventae Chasma, Mars

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Abstract Dark topographic slumps several meters wide, tens of meters in length, and up to a meter in depth are observed on the slopes of Juventae Chasma (JC), Valles Marineris (VM), Mars. These slumps usually originate near the terminal points of recurring slope lineae (RSL). Near their initiation points, the slumps have topographic depressions due to the removal of materials; near their lowermost reaches, new materials are deposited in lobes. Over the course of three Mars years, 10 active slumps have been observed in JC, all of which formed in or near the same season (areocentric longitude: L_s 0°–120°). Mars Color Imager (MARCI) observations show low-altitude atmospheric obscurations confined within the topography of the VM and JC in the seasons when the slumps form. In one instance, data from the Compact Reconnaissance Imaging Spectrometer for Mars and MARCI show evidence of H₂O ice in the atmospheric obscuration, likely due to the formation of a low-level afternoon cloud above a dust storm, or mixing of condensate clouds with a diffuse dust cloud. The presence of atmospheric obscurations with H₂O ice near times when the slumps form is intriguing, but no direct evidence currently exists to support that they aid in slump formation. Further monitoring of this site will help establish if RSL and/or atmospheric events play a role in the creation of contemporary slumps.

1. Introduction

The contemporary surface of Mars experiences modifications by a variety of active processes. Contemporary surface processes such as dust devils (Cantor, Kanak, & Edgett, 2006; Grant & Schultz, 1987; Malin & Edgett, 2001; Verba et al., 2010) and slope streaks (Aharonson, 2003; Chuang et al., 2007; Schorghofer & King, 2011; Sullivan et al., 2001) modify the Martian surface by removing the surficial dust and exposing a substrate with contrasting albedo and/or grain size. Over 500 new impact craters detected within the last 11 years on Mars have modified the surface by excavation, fragmentation, and redistribution of the regolith (Byrne et al., 2009; Daubar et al., 2013, 2016; Dundas & Byrne, 2014; Malin et al., 2006). Other stochastic features such as boulder tracks that form due to boulders falling from steep slopes and extending to shallow slopes also actively modify the Martian surface, although they are much rarer (Diniaga et al., 2014). Gully formations modify the surface of Mars by carving new channels and/or extension of existing channels (Diniaga et al., 2013; Dundas et al., 2012; Hansen et al., 2011; Heldmann & Mellon, 2004; Mangold & Costard, 2003). There are also polar processes such as frost-dust avalanches (Russell et al., 2008) and spiders (Piqueux, Byrne, & Richardson, 2003; Portyankina, Hansen, & Aye, 2017) that actively modify the present-day surface of Mars.

The slopes of Valles Marineris (VM) and other equatorial sites, southern midlatitudes, and Chryse and Acidalia Planitia host hundreds of thousands of recurring slope lineae (RSL), which are narrow, elongated, and low-albedo surface markings that incrementally grow downslope on steep terrains of Mars (Chojnacki et al., 2016; Grimm, Harrison, & Stillman, 2014; McEwen et al., 2011, 2014; Ojha et al., 2014, 2015; Stillman et al., 2014, 2016; Stillman, Michaels, & Grimm, 2017). RSL form and incrementally grow on warm Sun-facing slopes, and gradually fade and completely disappear, usually when the temperature decreases (Chojnacki et al., 2016; McEwen et al., 2011, 2014; Ojha et al., 2014; Stillman et al., 2014, 2016, 2017). The temperature-dependent formation of RSL has led many to hypothesize that volatiles such as brines may play a role in their formation (Chevrier & Rivera-Valentin, 2012; Chojnacki et al., 2016; Grimm et al., 2014; Levy, 2012; Massé et al., 2016; McEwen et al., 2011, 2014; Ojha et al., 2014, 2015; Stillman et al., 2014; Stillman et al., 2016, 2017; Wang et al., 2015; Watkins et al., 2014), although dry processes have also been recently proposed (Dundas, McEwen, & Milazzo, 2016; Schmidt et al., 2017).

Figure 1. (a) Thermal Emission Imaging System (THEMIS) day IR (Edwards et al., 2011) with Mars Orbiter Laser Altimeter (MOLA) (Smith et al., 2001) color mosaic of Juventae Chasma. (b) HiRISE observation of two hills inside Juventae Chasma. The small black box in Figure 1a show the location of the HiRISE image in Figure 11b. Repeat imaging of the region in Juventae Chasma by HiRISE. (c–h) HiRISE view of ESP_042597_1755 (longitude: 298.6°E, latitude: −4.7°N) (L_s : 34°, sol: 69 of Mars Year 33), ESP_043309_1755 (L_s : 59°, sol: 123), ESP_043876_1755 (L_s : 78°, sol: 166), ESP_044153_1755 (L_s : 87°, sol: 187), ESP_044364_1755 (L_s : 95°, sol: 203), and ESP_044997_1755 (L_s : 117°, sol: 251). The small black box shows the location of images shown in Figures 1c–1h. The atmospheric obscuration is observed in Figure 1c at sol 123, and a new topographic slump is observed in the next image acquired by HiRISE (Figure 1e), after 43 sols (red arrow). The black arrow shows a faded slump with similar morphology.

Figure 2. HiRISE IRB color observation of the slump 85 sols after the initial observation. The red arrow shows the direction of incoming sunlight (i.e., incidence angle). The white arrows point to the shadow cast by the slump wall scarp, indicating topographic change associated with slumping. The black arrow points to the area where slumping initiated.

The magnitude of the surface modification by RSL is not entirely understood. RSL may contribute negligibly toward surface modification, and their dark appearance could be simply due to the transient presence of minor water, which reduces the surface albedo of regolith in the visible wavelengths (Massé et al., 2014; Pommerol et al., 2013), analogous to the dark appearance of wet soil on Earth (Balsam, Deaton, & Damuth, 1998; Head et al., 2007; Levy et al., 2011). Surface modification by RSL has been rarely observed at High Resolution Imaging Science Experiment (HiRISE) resolution (Chojnacki et al., 2016). Small topographic slumps have been detected to form on the slopes below active RSL in Garni crater within Melas Chasma, on the walls of Coprates Chasma, and on hills of Juventae Chasma (JC) (Chojnacki et al., 2016); however, it is unknown if the topographic slumps are caused by RSL activity or if these are simply steep equatorial slopes with the most repeat high-resolution images.

Here we describe the observations of 10 new topographic slumps in JC by the High Resolution Imaging Science Experiment (HiRISE) (McEwen et al., 2007). We describe the geographical distribution, seasonality, and temperature dependence of the topographic slumps and discuss their possible links to RSL. We also describe the observations of atmospheric obscuration with H₂O-ice detections within the formation period of the topographic slumps. We perform calculations to constrain the amount of water vapor required for ice nucleation on top of a dust cloud and discuss potential roles that atmospheric obscurations may have in the slumps' formation.

2. Data Sets and Methodology

2.1. High Resolution Imaging Science Experiment (HiRISE)

The HiRISE camera aboard Mars Reconnaissance Orbiter (MRO) acquires high-resolution (~30 cm/pixel) images over a swath ~5 km wide, with a center swath (~1 km) with coverage in three colors (McEwen et al., 2007). HiRISE has acquired

Figure 3. The relative albedo of the slump as a function of time, along with various theoretical fits. (a) Locations of the slump (red), and a control area (in blue) normalized by a “background” area (yellow). (b) The relative albedo of the slump (in red) and the control area (in blue) as a function of time. While the relative albedo of the slump increases over time, the relative albedo of the control area (blue) remains close to unity in all images, suggesting minor influence from different viewing geometries. (c–e) Relative albedo of the slump as a function of time and various theoretical fits, labeled with the corresponding R^2 values.

Table 1
HiRISE Observation Constraint on the Active Slumps in Juventae Chasma

	Before image	After image	GIF	Before L _s	After L _s	MY	Confidence	Resolved topography?
Slump 1	ESP_032997_1755	ESP_034342_1755	S.SF1 (Movie S1)	5	53	32	10	Yes
Slump 2	ESP_043309_1755	ESP_043876_1755	S.SF1 (Movie S1)	59	78	33	10	Yes
Slump 3	ESP_042597_1755	ESP_043309_1755	S.SF1 (Movie S1)	34	59	33	8	No
Slump 4	ESP_041318_1755	ESP_042386_1755	S.SF1 (Movie S1)	345	26	32	5	No
Slump 5	ESP_034342_1755	ESP_036399_1755	S.SF2 (Movie S9)	53	125	32	10	Yes
Slump 6	ESP_042597_1755	ESP_043309_1755	S.F1 (Movie S2)	34	59	33	10	No
Slump 7	ESP_032997_1775	ESP_034342_1755	SE. F1 (Movie S4)	5	53	32	10	Yes
Slump 8	ESP_034342_1755	ESP_036399_1755	S.SF2 (Movie S9)	53	125	32	10	Yes
Slump 9	ESP_049401_1755	ESP_050390_1755	NA	314	356	33	10	Yes
Slump 10	ESP_049401_1755	ESP_050390_1755	NA	314	356	33	10	Yes

Note. NA, not available.

32 images of JC (as of 30 May 2017) to monitor RSL. The elevation measurements were done using a HiRISE digital terrain model (DTM) constructed from HiRISE stereo pairs (ESP_030795_1755 and ESP_030373_1755) using BAE System photogrammetry software (BAE Systems, 2008) similar to earlier change detection campaigns (Chojnacki et al., 2016, 2017; McEwen et al., 2011, 2011). These DTMs were built using BAE System photogrammetry software yielding a post spacing of 1 m and vertical precision of ~17 cm (Kirk et al., 2008).

Figure 4. (left) Before and (right) after images show the formation of slumps 1 to slump 4 in Juventae Chasma. The red arrows point to the area where failure initiated. Slump 3 and slump 4 are classified as candidate slumps because they lack topography and are barely discernable at HiRISE resolution.

The expected vertical precision is a function of image resolution (25 cm/pixel), subpixel matching quality (1/5 pixel correlation), and stereo viewing geometry (e.g., convergence angle of $\sim 16^\circ$) and provides a quantitative assessment of post-to-post confidence (Kirk et al., 2003). The DEM was also registered to Mars Orbiter Laser Altimeter (MOLA) (Smith et al., 2001) shot points for absolute elevation, which yields a mean difference from MOLA of -4.17 ± 16.63 m. Following a large slumping event, a second DTM was constructed using ESP_044997_1755 and ESP_044364_1755 with the original DTM as a seed and a vertical precision of ~ 28 cm.

All monitoring images were also orthorectified to the parent DTM to facilitate change detection. New topographic slumps were detected by applying a change detection algorithm on HiRISE orthorectified images. The algorithm first matches the histograms of a pair of images and quantifies differences between the images by subtracting the pixel values of a first-date image from a second-date image. A difference map is produced, in which regions with temporal changes have high pixel values, and regions without any temporal change will have small pixel values. Depth estimates of one topographic slump were based on shadow geometry (Figure 2). Additionally, this was confirmed to within ~ 28 cm on a BAE System photogrammetry software workstation (BAE Systems, 2008), which allows elevations to be extracted while a user confirms using the stereo viewer. Fading rate calculations were made following the methodology described in *Daubar et al.* (2016). HiRISE images were crudely atmospherically corrected by subtracting the darkest pixel value (i.e., dark shadow) from the entire HiRISE scene, and changes were measured by normalizing the HiRISE digital-number value of a region-of-interest by an adjacent unmodified background area (Daubar et al., 2016).

2.2. Compact Reconnaissance Imaging Spectrometer for Mars

The composition of the atmospheric obscurations (when present) was investigated using hyperspectral data acquired by the Compact Reconnaissance Imaging Spectrometer for Mars (CRISM) instrument on MRO. CRISM acquires hyperspectral cubes with 544 spectral channels between $0.4 \mu\text{m}$ and $3.92 \mu\text{m}$ (Murchie et al., 2007). CRISM images were preprocessed using ENVI's (Exelis Visual Information Solutions) CRISM Analysis Tool to reduce atmospheric effects and map-project the images (McGuire et al., 2009). Dust and ice aerosols lead to increased atmospheric radiance in the CRISM wavelengths at higher emission angles, corresponding to longer path lengths (Clancy, 2003; Murchie et al., 2007). Particularly for ice aerosols, this effect leads to asymmetric emission angle dependence on brightening, whereby positive emission angle results in higher radiance due to increasing forward-scattering (Clancy, 2003). Therefore, to investigate the composition of the atmospheric obscuration and to minimize the effect of varying emission angle, the image was normalized by dividing the value at each wavelength in the line dimension by the median value at that wavelength over the entire line dimension. A 3 by 3 kernel spline function was applied to the normalized spectra to smooth out the residual noise. CRISM consists of two detectors, one is the visible and near-infrared (VNIR) channel spanning $0.4\text{--}1.0 \mu\text{m}$, with the other being the infrared (IR) channel covering $1.0\text{--}3.92 \mu\text{m}$ (Murchie et al., 2007).

2.3. Context Camera

MRO's Context Camera (CTX) acquires images of Mars with a $5\text{--}6$ m/pixel scale (Malin et al., 2007). The areal extent of atmospheric obscuration was interpreted from the CTX images.

2.4. Mars Color Imager

MRO's Mars Color Imager (MARCI) is a wide-angle, multispectral camera that acquires a daily global view of Mars in seven channels (five in visible and two in the ultraviolet (UV)) (Bell et al., 2009). The Martian surface has a very low contrast in the UV, which is further reduced in the presence of any atmospheric obscurations. Water ice appears bright in the MARCI UV and blue channels (Bell et al., 2009; Wolff et al., 2010). We therefore manually checked the MARCI UV and blue images to seek evidence of atmospheric water ice in VM and Juventae Chasma. The details of MARCI calibration, data products, and methodology can be found in Bell et al. (2009) and Wolff et al. (2010).

2.5. Constraint on the Atmospheric Water Abundance

The calculation of the expected water vapor content (in precipitable microns) during an ice-cloud event can be made by the simple assumption of a saturated ice-cloud layer and the assumption that there is no water vapor outside of the ice-cloud layer (making this a minimum estimate). Starting with the topographically

Figure 5. (left) Before and (right) after images show the formation of slumps 5–10 in Juventae Chasma. The red arrows point to the area where failure initiated. Slump 6 lacks topography and is barely discernable at HiRISE resolution making it a candidate slump.

inferred 600 m (altitude above the mean elevation of JC floor) upper bound of the ice-cloud layer, two different thicknesses are considered: a 100 m thick layer extending down to 500 m altitude and a 600 m thick layer extending down to the surface. The full diurnal cycle of pressures and temperatures at the altitudes of interest are calculated using the Martian Climate Database (MCD) model (Forget et al., 1999; Millour et al., 2015). These temperatures and pressures are utilized to calculate saturation vapor pressures at these altitudes with the Clausius-Clapeyron relation. In our calculation, the vapor pressure in the ice-cloud layer is assumed to be 120% or 250% of the saturation pressure, two values that have been reported

Figure 6. HiRISE ESP_032997_1775_RED showing the region in Juventae Chasma where the slumps occur.

for the threshold saturation for ice nucleation onto dust particles (Field et al., 2006; Iraci et al., 2010). These vapor pressures are then used to calculate water vapor mixing ratios at the top and bottom of the ice-cloud layer. The mixing ratios within the ice-cloud layer are assumed to vary linearly between the top and bottom values, an assumption that is justified given the thinness of the layer. These mixing ratios are then integrated over the height of the layer and divided by the density of liquid water and Mars's surface gravity to obtain the water content.

3. Results

3.1. Observations of Active Slumps and RSL

In an effort to constrain the seasonality of active RSL in JC, a coordinated MRO campaign was carried out providing simultaneous HiRISE, CRISM, and CTX observations of a hill in Juventae Chasma during late southern fall of Mars Year (MY) 33 (solar longitude, $L_s = 58^\circ$, Earth date: 23 October 2015) (Figure 1). A dark topographic slump with a length of ~ 45 m and an area of ~ 400 m² was observed by HiRISE (Figure 1; also see Movie S1 in the supporting information). Near its point farthest upslope, the slump is characterized by a topographic depression due to the removal of materials; near its lowermost point, new materials are deposited in a fan-like morphology (Figure 2), similar to adjacent features and slumps seen elsewhere in VM (Chojnacki et al., 2016). At its deepest point, the slump has an estimated depth of 1 m, implying a volume of removed materials of approximately 400 m³. The depression is clearly resolvable in terrain shaded relief

and orthoimages and was confirmed using the stereo images viewed in the photogrammetry software. While this depth value approaches the post spacing of the DTM, the finer-scale vertical precision, slump shading geometry, and relatively large area of the slump with posts lower than the surrounding terrain yield confidence for this measurement. The fading timescale of the low albedo of the slump is rapid; a spectral contrast of less than 10% compared to the background is observed within 85 Martian solar days (sols) (Figures 1 and 3). This fading timescale is comparable to the fading of RSL once they cease growth (Stillman et al., 2014, 2017).

Our HiRISE imaging campaign of JC also captured the formation of nine other active slumps on this hill (Table 1, Figures 4–6, and Movies S2–S9). Slump-3 and slump 6 are marginally but discernibly resolved by HiRISE, and seems to lack topography. Slump-9 is clearly discernable at HiRISE resolution but lacks resolvable topography. We therefore refer to slump 3, and slump 6 as candidate slumps. The other seven slumps have widths greater than several meters, and their topography are clearly resolved in HiRISE images. There are two prominent hills in Juventae Chasma, which we refer to as the northern and southern hills (Figure 1). It should be noted that all the active slumps and almost all paleo-slumps are located on the northern hill (Figure 6) of JC. The northern hill is slightly steeper than the southern hill (Figure 7), but the slumps do not begin on the steepest slopes, and therefore that cannot be the limiting factor. Both of these hills feature abundant RSL.

The intensive observation campaign of JC by HiRISE has enabled us to put a tight constraint on the seasonality of the active slumps and RSL (Figure 8 and Table 1). RSL in JC are observed in all slope aspects of the north and the south hill, but here we only present seasonality of the RSL in the south facing slopes of the north hill where the active slumps are present (Figure 6). In MY 31, RSL growth is observed between L_s 247° and 345° in the south facing side of the north hill. No discernable slumps formed during this time (Figure 8). RSL are mostly faded by L_s 4° of MY 32 (see Movies S1 and S2), and slump 1 and slump 7 formed sometime between L_s 5° and 53°, when RSL are mostly faded and disappearing. Slump-5 and slump 8 formed sometime between L_s 53° and 125°; neither of the HiRISE observations from L_s 53° and L_s 125° shows any RSL on the south facing slopes. In MY 32, RSL growth starts around L_s 213° and continues to L_s 34° of MY 33. By L_s 34°, most RSL in the south facing slopes have started to fade, and by L_s 58°, the south facing slopes are devoid of any RSL. Slump-4 formed between L_s 345° and 26°. Slump-3 and slump 6 are observed to form between L_s

Figure 7. Slope maps derived from the HiRISE DTM (DTEEC_030373_1755_030795_1755_A01) of Juventae Chasma overlaid on an orthorectified image. The color bar shows the slopes of the terrain in degrees. (top) Northern hill. (bottom) Southern hill. The red outlines show the location of the paleoslumps.

Figure 8. Seasonality of RSL and the 10 slumps in Juventae Chasma. The slumps formed between the L_s marked by the colored dots. The first row shows the seasonality of RSL in the south facing slope of the north hill in JC.

34° and 59°. Slump-2 formed between L_s 59° and 78°, when RSL are absent on the south facing slope. In MY 33, RSL are actively growing between L_s 194° and 314°. By L_s 314°, most RSL in the south facing slopes have faded, and two new slumps are observed between L_s 314° and 356°. In summary, all 10 slumps are observed to form after RSL have mostly faded or disappeared and eight of the 10 slumps formed between L_s 345° and 120°, less than 40% of a Martian year. No active slumps are observed to form in HiRISE observations acquired between L_s 120° and 314° in MY 31–33. Based on the surface temperatures in MCD, L_s 0°–120° corresponds to the coldest time of the year in JC (Figure 9). The MCD model does not take high-resolution topographic differences into account and the Sun is in the northern hemisphere, so the surface temperature of the south facing slopes where the 10 active slumps formed (Figure 6) should be even lower than predicted by the model.

There are ~40 additional “paleo” slumps that formed prior to the HiRISE imaging campaign of JC (Figure 10). Unlike the 10 slumps whose formation was constrained by repeat HiRISE observations, no constraints can be placed on the formation timescale of the paleo-slumps. The morphologies of some of these paleo-slumps are similar to the active slumps (Figure 11), suggesting formation through a similar mechanism.

3.2. Observation and Composition of the Atmospheric Obscuration

A coordinated MRO campaign carried out by HiRISE, CRISM, CTX, and MARCI spanning the formation period of slump 2 shows obscuration of the surface by an atmospheric phenomenon. A CTX sequence (Figure 12) shows that the area was obscured by clouds at L_s 58°. A concurrent MARCI observation from 23 October 2015 shows the atmospheric obscuration to be confined within the topography of Valles Marineris and Juventae Chasma (Figure 13), ruling out high-altitude clouds as the source of the obscuration. Dust storms are observed outside the canyon a day before (22 October 2015) and are likely the source of the atmospheric obscuration in VM (Bruce Cantor, personal communication). To further constrain the elevation of the atmospheric obscuration, we analyzed the concurrent CRISM observation. In the CRISM VNIR observations, clouds and dense atmospheric obscurations have higher apparent reflectivity. Unlike in CTX images of the area taken before or after 23 October, on that date in multiple CRISM VNIR bands, the low-lying area to the south-east appears more reflective than the summit of the central hill (~600 m in elevation) (Figure 14). Collectively, these lines of observation are interpreted as evidence for a low-elevation atmospheric obscuration, extending up to approximately 600 m in height.

In the CRISM IR observation (image ID FRS00038BDE), a major absorption band at ~3.1 μm is observed along with the possible presence of

Figure 9. Surface temperature range as a function of the season at Juventae Chasma from the MCD model. The top and bottom lines are the daily maximum and minimum temperature profiles as a function of the solar longitude (L_s).

Figure 10. Paleo slumps observed on a hill in Juventae Chasma and their spatial distribution. A slope aspect map is shown in color, overlaid by mapped paleo slumps. Note the terrain artifacts (orange and yellow polygons on north slopes), which were avoided for our analysis.

weak absorption band at $1.5\ \mu\text{m}$ (Figure 15). We infer that the absorption band at $3.1\ \mu\text{m}$ results from the high attenuation coefficient (~ 1) of water ice at this wavelength (Vincendon et al., 2011; Warren, 1984). The absorption bands at $1.2\ \mu\text{m}$ and $1.5\ \mu\text{m}$ band (which are the expected positions of vibrational overtone/combination bands in H_2O) are extremely weak or absent, which is to be expected given that detections at those wavelengths require a high signal-to-noise ratio for fine-grained clouds (Vincendon et al., 2011). The $3.1\ \mu\text{m}$ absorption band in the CRISM observation is similar in strength to the $3.1\ \mu\text{m}$ band of Martian mesospheric water ice-clouds observed elsewhere on Mars (Figure 16), implicating the presence of water ice in the atmospheric obscuration. An earlier CRISM observation (MY: 31, L_s : 345°) of the same region does not show any atmospheric obscuration (Figure 17 and Table 2), nor were clouds observed in concurrent MARCI observations. This earlier CRISM observation, normalized in similar fashion as 38BDE, does not show any $3.1\ \mu\text{m}$ absorption band, proving that the source of the strong $3.1\ \mu\text{m}$ absorption band in FRT00038BDE is the atmospheric obscuration, and not the ground underneath (Figure 17). There are three additional clear-sky CRISM IR observations of this site, and none of them show any evidence for a strong $3.1\ \mu\text{m}$ absorption band in their spectra (Table 2).

The band-7 (centered at $320\ \text{nm}$) UV channel of MARCI is particularly sensitive to atmospheric H_2O -ice, and the MARCI image from 23 and 24 October 2015 shows bright areas inside VM and Juventae Chasma, with ice optical depth of 0.10–0.15, providing

additional evidence for the presence of H_2O in the atmospheric obscuration (Figure 18; also see Movie S10). There are many other examples of atmospheric obscuration events with H_2O ice in MARCI observations of VM and JC. In these atmospheric obscuration events, the floors of eastern VM and JC are covered by an optically thick layer that appears bright in the MARCI VIS and UV observations. We present a few examples from MY 30 (L_s : 60°) and MY 32 (L_s : 42°) (see Movies 11–16). We examined MARCI observations to look for atmospheric obscuration events between the formation periods of each slumping event and find multiple examples in eastern VM and JC (Figure 19).

Figure 11. Examples of paleo slumps in the south facing slopes of Juventae Chasma (ESP_030940_1755_RED). The red arrows point to the area where the slumping initiated. The black arrows point to nearby recurring slope lineae (RSL).

Figure 12. CTX view of the low-altitude atmospheric obscuration in Juventae Chasma. (a–c) Parts of F17_042597_1758 (L_s/sol : $34^\circ/69$), F19_043309_1753 (L_s/sol : $58^\circ/123$), and F21_043876_1744 (L_s/sol : $78^\circ/166$), respectively, showing the hills and low-lying plains in Juventae Chasma. Figures 12a and 12c show the sand-encircled hills under haze-free conditions. The white-box in Figure 12a shows the areal extent of CRISM observation in Figure 14.

4. Discussion

4.1. Topographic Slumps and Their Potential Relationship to RSL

The slumps in JC are distinct due to their topographic depression and elongated, triangular morphology (Figures 2, 4, and 5). The morphology of the slumps is similar to scars generated by terrestrial loose snow

Figure 13. MARCI view of Valles Marineris (red outline) and Juventae Chasma (blue outline) from 20 October ($L_s:57^\circ$) to 25 October ($L_s:59^\circ$) of 2015. The atmospheric obscuration in the HiRISE and CTX image of Juventae Chasma was observed in 23 October 2015. The geographical extents of the VM and JC are outlined in red and blue, respectively.

Figure 14. CRISM VNIR and CTX observation of the fog in Juventae Chasma. The four panels show reflectance value at 440 nm, 530 nm, 600 nm, and 770 nm, respectively, from CRISM observation FRS00038BDE.

avalanches, which form in thin layers of dry snow with low particle cohesion, and often start at or near the surface from a small area and spread out laterally as they move downslope. This results in a shallow triangular scar with a sharply pointed head scarp as more snow is pushed downslope and entrained into the slide (McClung & Schaerer, 1993; Perla, 1980). These scars bear similarity to avalanche scars formed in laboratory experiments with thin layers of cohesionless glass beads near the angle of repose. These experiments have revealed that the initiation of a failure at a point in a thin layer of beads results in an avalanche that propagates only downhill, in which collisions between neighboring grains promote lateral entrainment. This produces an avalanche scar with a sharply pointed triangular head scarp (Daerr & Douady, 1999). By contrast, the initiation of an avalanche in a thicker layer of beads causes a loss of support for grains uphill of the avalanche initiation point, which promotes the upslope propagation of the head scarp at the same time the avalanche flows downhill. This produces avalanche scars with rounded head scarps, rather than the triangular head scarps that form in thin layers (Daerr & Douady, 1999; McClung & Schaerer, 1993).

The triangular morphology and the shallow topographic depression of the Martian slumps are thus consistent with slump formation in a thin layer of granular material with weak particle cohesion. On Mars, other slope features such as slope streaks and avalanche scars also exhibit triangular morphologies and have been hypothesized to form by similar mechanisms (Chuang et al., 2007; Gerstell, Aharonson, & Schorghofer, 2004; Sullivan et al., 2001). Based on the triangular morphology, Sullivan et al. (2001) propose a formation mechanism in which slope streaks form as dust avalanches due to oversteepening by dusty air fall deposits. While avalanche scars have depths up to 10 m and are much thicker than slope streaks, which typically have depths of a few meters (Gerstell et al., 2004), avalanche scars have been proposed to form under a similar mechanism to that proposed here for the observed slumps, in which the flow occurs within a loose granular layer bounded by a distinct interface below.

Figure 15. CRISM observation of the atmospheric obscuration in Juventae Chasma and spectra. (top) Map-projected view of CRISM FRS00038BDE (R: Band 233 (2.6 μm), G: Band 78 (1.51 μm), R: Band 13 (1.1 μm)), stretched with a Gaussian mode to show the contrast between areas highly obscured by atmosphere and hills with less obscuration. The red color boxes show the location from which the red spectrum was extracted. (bottom) Spectrum from red color box shown in Figure 15 (top) divided by the median spectrum from the line dimension. The observed data are plotted in dashed lines, and smoothed spectra in solid lines. The dotted vertical lines are at 1.2 μm , 1.5 μm , and 2.0 μm .

Although there are some morphological similarities between slumps, slope streaks, and avalanche scars, there are several major differences between them. The slumps described here have a distinct seasonality in the timing of their formation, whereas both slope streaks and avalanche scars are caused by sporadic events of small spatial extent, and lack seasonal preference for their formation (Chuang et al., 2007; Schorghofer & King, 2011). Further, the albedos of the slumps in VM fade extremely rapidly (tens to hundreds of sols), whereas slope streaks fade over years to decades (Schorghofer et al., 2007). These differences suggest that these features differ in the characteristics that control their fading timescales despite the potential similarities in their formation mechanisms.

The southern slopes of the northern hill in JC, where slumps occur, also host hundreds of RSL. The RSL on south facing slopes of JC are most active (i.e., forming and incrementally growing) during $L_s \sim 180^\circ$ – 360° , which corresponds to the warmest time of the year, and are absent or faded when the slumps appear to be forming, which falls during the coldest time of the year ($L_s \sim 0^\circ$ – 120°) (Figures 8 and 9). All 10 slumps described here either formed while RSL were fading or when they were completely absent. While RSL tend to initiate from near the top of the hills, topographic slumps usually initiate in the mid-slope of the hill, which is shallower ($\sim 25^\circ$ – 30°) and where RSL terminate (see Movie S4). The effective friction coefficient at these slopes can be estimated using Heim's ratio which is defined as ratio of the drop height (H) to the runout distance (L) (Heim, 1932). For the seven slumps with resolvable topography, the Heim's ratio is between 0.49 and 0.54 (Table 3). This is consistent with Heim's ratio expected for dry landslides with volume less than 10^6 m^3 (Lucas, Mangeney, & Ampuero, 2014).

Initially, each slump appears dark, but its color then fades rapidly (Figure 3). The fading rate calculation in JC could only be done for slump 2; most other slumps were already faded at the time of HiRISE observation although their topographic signature is still resolvable. A rapid fading rate was also observed for a slump in Garni crater (Chojnacki et al., 2016). RSL and large associated fans are also observed to change their albedo at similar timescales to the slumps (Chojnacki et al., 2016; McEwen et al., 2011; Ojha et al., 2013).

Figure 16. CRISM observation of mesospheric water ice clouds. (a) Map-projected view of CRISM HRS0000B3DA (R: Band 233 (2.6 μm), G: Band 78 (1.51 μm), R: Band 13 (1.1 μm)). (b) Map-projected view of CRISM FRT000173B5 (R: Band 233 (2.6 μm), G: Band 78 (1.51 μm), R: Band 13 (1.1 μm)). (c) Spectra from the blue looking region in Figures 16a and 16b. (d) Same as Figure 16c with the spectral range extended out to 3.9 μm . The symbols and smoothing schemes are identical to those in Figure 3. The numerator and denominator pixels used to produce the spectra are annotated in the figure.

In both cases, however, it is unknown if the surface darkening is due to chemical effects (e.g., surface wetness by brines) or textural effects (e.g., grain size sorting), although rapid fading is most easily explained by drying. In most cases, slumps form once RSL have started fading or disappeared completely, although some counterexamples were found in neighboring chasmata (e.g., Garni crater in Melas Chasma) (Chojnacki et al., 2016).

The slopes where RSL and the slumps form (McEwen et al., 2011, 2011) are close to the angle of repose for cohesionless sediment (e.g., sand to pebbles) (Atwood-Stone & McEwen, 2013). It is conceivable that the RSL activity somehow triggers the slumping, perhaps by oversteeping the slopes from cumulative deposition. The exact cause and the mechanism (dry versus wet) of such triggering are not obvious. If subsurface volatiles are involved in RSL formation, as has been proposed based on the spectral detection of hydrated salts in some RSL locations (Ojha et al., 2015), then evaporation of brines may lead to a net decrease in cohesion and cause the observed slumping. On the other hand, if slumps are related to the atmospheric obscuration events, then slope instability arising from the volumetric expansion of deliquescent salts and/or increased pore pressure in response to enhanced near-surface humidity could be another viable formation mechanism. Dark patches appear inside slumps when RSL form and grow, and they could represent areas of concentrated hygroscopic salts that deliquesce (Figure 20). However, the relatively large volume of excavation makes this formation scenario unlikely, requiring volumetric expansion of salts at a depth of ~ 1 m. If significant volumes of water are flowing through the surface layer to create RSL (Stillman et al., 2017), then some

Figure 17. CRISM observation of Juventae Chasma and spectra. (top) Map-projected view of CRISM FRS0002A0D8 (R: Band 233 (2.6 μm), G: Band 78 (1.51 μm), R: Band 13 (1.1 μm)), stretched with a Gaussian mode. The colored red box show the location from which the spectrum was extracted. This box is approximately at the same location as the red box in Figure 15. (bottom) Spectra from colored boxes shown in Figure 17 (top). The symbols and smoothing schemes are identical to those in Figures 15 and 16.

of that water could seep to a depth of ~ 1 m, increasing pore pressure and reducing friction, so slumping is possible. However, it is difficult to explain how such volumes of water could be produced on Mars today. Further, the estimation of Heim's ratio (0.4–0.7) for the slumps are within the bounds expected for dry landslides and do not require any special mechanism to reduce friction (Lucas et al., 2014).

Slumps could also form due to ground shaking by regional seismicity or bolide impacts (Gerstell et al., 2004) and may have no genetic connection with atmospheric events or RSL activity. This hypothesis, however, is not supported by observations—no fresh craters have been observed in the area—and it does not explain the seasonality of the slumps.

The RSL and their small gullies, typically above the slumping locations (Figure 1), may be related to the slumps as follows. If RSL form by eroding the slopes, then they would promote gully formation and fan deposition below. This buildup of deposited sediment on steep slopes eventually leads to failure. However, it is not clear why this would occur only in the coldest season. If RSL and slumps are related, then it is also unclear why slumps are only present at a few RSL sites and not in others.

An alternative hypothesis is that RSL and slumps are genetically unrelated to each other, and their distinct seasonality may be mostly due to the surface and atmospheric temperature conditions. In other words, warm temperatures trigger RSL activity on steep slopes and cold temperatures trigger slumping, without any direct connection between RSL and slumps. In VM alone, there are over 200 candidates and confirmed RSL sites (Stillman et al., 2017), yet topographic slumps have only been observed at a handful of sites. A focused study to examine all known RSL sites to look for evidence of active and paleo slumps is underway to examine if RSL and slumps are genetically related. The HiRISE camera has uniquely covered $\sim 2\%$ of Mars in the last 11 years, so a detailed study like this is not

Table 2
Coordinated CRISM-HiRISE Observations of Juventae Chasma

CRISM_ID	Coordinated_HiRISE	Acquisition date	MY	L_s	Sol
0002A0D8	ESP_032496_1755	7/2/2013	31	345	640
0002 AC25	ESP_032997_1755	8/10/2013	32	5	9
00031402	ESP_037388_1755	7/18/2014	32	164	342
00036FD5	ESP_041318_1755	5/21/2015	32	345	641
00038BDE	ESP_043309_1755	10/23/2015	33	59	123

Figure 18. Band-7 UV observation by MARCI on 23 October 2015. The bright areas in the image show the location of atmospheric water ice. A zoomed-in view of the area in the red-box is shown in the bottom panel. The red arrow points to Valles Marineris and Juventae Chasma. The legend shows the I/F values.

possible at all sites. Although observation of active slumps requires frequent monitoring, paleoslumps are easily discernable and future studies aimed at cataloging paleoslumps will help narrow down the sites for more frequent observations. Future observation of numerous other buttes and hills inside JC will also help assess if slumps are more pervasive in this region of Mars than others.

Figure 19. Number of atmospheric obscuration events observed in MARCI images between the formation period of each slump. Note that slump 1/slump 7, slump 2/slump 4, slump 3/slump 6, slump 5/slump 8, and slump 9/slump 10 formed during the same L_s interval.

4.2. Atmospheric Obscurations

MRO's collective observations of the atmospheric obscuration in Juventae Chasma were acquired at L_s 58°, 3:00 p.m. local time—at this time the MCD model predicts the surface and the air temperature at a 600 m level to be ~250 K and ~230 K, respectively, which should prevent formation of atmospheric ice particles or sublimate ice particles that formed at higher altitudes (Figure 21). The observation of H_2O ice in CRISM and MARCI could be due to the formation of a low-level afternoon cloud above a dust storm as has been previously observed (Cantor et al., 2001; Malin et al., 2008), or mixing of condensate clouds (that formed elsewhere) with a diffuse dust cloud. The lack of the 3.1 μm spectral absorption bands in the earlier observations of JC (e.g., Figure 17) precludes the surface as the source of the 3.1 μm absorption band. The surface temperature and the vertical temperature profile up to a few kilometers are too warm for CO_2 ice (condensation temperatures of around 145 K) to condense during the observation period ($L_s = 58^\circ$) (Figure 21). An alternative source of the 3.1 μm absorption could be suspended hydrous-dust particles in the atmosphere; however, dust particles would appear less blue and more diffuse in MARCI UV images (Bell et al., 2009; Wolff et al., 2010).

Ice nucleation in the atmosphere results from the deposition of water vapor onto suspended dust particles. The threshold saturation (S_{crit}) of water vapor required for ice nucleation can vary with the dust contact geometry, dust composition, and atmospheric temperature. For the MCD modeled temperatures and pressures in Juventae Chasma, and assuming S_{crit} of 120% (Field et al., 2006; Whiteway et al., 2009), we find that a 13 precipitable micrometers (pr μm) water vapor column will initiate nucleation. This is of the same order as the seasonal average of 10 pr μm for this time period (Smith, 2002), although we note that the 13 pr μm is required just in the confined ice-cloud layer of 100 m thickness and that the total atmospheric column would likely be much higher considering the ambient water vapor in the rest of the column outside of the

Table 3
Measurement of H/L (Height of Drop/Length of Runout) for Slumps in Juventae Chasma

Slump	Height of the top of the slump	Height of the toe of the runout	Height	Length	H/L
1	−3431	−3479	48	94	0.51
2	−3314	−3391	77	150	0.51
3	NA	NA	NA	NA	NA
4	NA	NA	NA	NA	NA
5	−3297	−3378	81	148	0.54
6	NA	NA	NA	NA	NA
7	−3451	−3583	132	241	0.54
8	−3592	−3682	90	182	0.49
9	−3254	−3334	80	148	0.54
10	−3458	−3843	385	737	0.52

Note. NA, not available.

Figure 20. Repeat imaging of a slump in JC by HiRISE. Top left to bottom right: ESP_030940_1755 (L_s : 275°), ESP_031863_1755 (L_s : 318°), ESP_032074_1755 (L_s : 327°), ESP_032219_1755 (L_s : 334°), ESP_032496_1755 (L_s : 345°), and ESP_032997_1755 (L_s : 5°). The arrow in the second figure is pointing to dark patches inside the slump.

ice-cloud layer. If S_{crit} of 250% is assumed (Iraci et al., 2010), an atmospheric water vapor content of 27 pr μm is required for ice nucleation near 3 PM (Figure 22). Both estimates assume that the ice cloud is 100 m thick. The same calculation repeated assuming a ground-hugging fog, as has been suggested by the spectral data from the Planetary Fourier Spectrometer (Möhlmann et al., 2009) on board the Mars Express spacecraft requires water vapor columns of >600 pr μm (Figure 22). These estimates are similar to those of Leung et al. (2016), who found that the air inside the canyon would require a 4–7 times enrichment in water vapor at saturation compared to outside the canyon. They noted that this suggests that a local source of water vapor is required to explain water ice fog appearing within the confines of Valles Marineris on Mars (Leung et al., 2016).

These results suggest that a relatively thin ice cloud layer can be produced with water vapor columns that are of the same order as the observed seasonal averages, while ground-hugging fog would require orders of

Figure 21. Modeled diurnal temperature profile at Juventae Chasma. (a) Surface temperature as a function of the local time. (b) Vertical temperature profile as a function of altitude from the surface at 3:00 PM local Martian time. (c) The temperature at 600 m above the surface as a function of the local time at $L_s = 58$.

magnitude higher water vapor columns than have been observed on Mars. If the ice cloud layer were at even slightly higher altitudes than the Juventae Chasma event, the cooler atmospheric temperatures could make it conceivable that a smaller water vapor column comparable to the Juventae Chasma event could be sufficient for ice nucleation.

Although ice clouds are observed in VM and JC during the formation time of slumps, it is currently unknown if the two events are related to each other. Both slumps and atmospheric obscuration events with H_2O ice are temperature-related phenomena, so their seasonal correlation could be independently driven by seasonal temperature. On the other hand, an increase in near-surface humidity by the presence of an ice cloud could provide viable pathways for deliquescence, perhaps triggering the formation of slumps. Deliquescence may be more likely in the shallow subsurface than at the surface (Gough, Chevrier, & Tolbert, 2016; Wang et al., 2017). Alternatively, the slumps may expose a damp subsurface layer (to explain darkening and fading),

Figure 22. The water vapor column required to nucleate ice-clouds as a function of local time. The pressure and temperature conditions are for $L_s = 58$ and are calculated from the MCD model. Two different ice-cloud thicknesses are considered, a layer extending from 500 m to 600 m, or from the surface to 600 m. Two S_{crit} (threshold saturations) are also considered.

without this triggering the slumps. Daily observations of Mars by MARCI are currently being examined to catalogue the frequency of atmospheric obscuration events in JC and VM. Future monitoring of this site is crucial to understand relations (if any) between these phenomena.

5. Conclusions

HiRISE observations of a hill in JC show multiple topographic slumps that are several meters wide, tens of meters in length, and up to a meter in depth. Ten active slumps have been observed in JC and are observed to form in roughly the same season. Low-altitude atmospheric obscurations (likely dust) confined within the topography of the VM and JC have been observed at times when slumps form and in some instances, evidence of H₂O ice in or on top of the atmospheric obscuration is present. The slumps also display intriguing spatial correlation to RSL in JC. The formation mechanisms of both the slumps and RSL are unknown, and no causal relation to the atmospheric obscuration or RSL can yet be established. We suggest that additional monitoring through coordinated HiRISE/CRISM/MARCI observations, as well as modeling, may constrain the formation mechanism for the slumps, as well as help to further explore the possibility of any genetic connection with the atmospheric obscuration events.

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