Sinton crater, Mars: Evidence for impact into a plateau icefield and melting to produce valley networks at the Hesperian–Amazonian boundary

Gareth A. Morgan, James W. Head III *

Department of Geological Sciences, Brown University, 324 Brook Street, Providence, RI 02912, USA

A R T I C L E   I N F O

Article history:
Received 19 December 2007
Revised 3 November 2008
Accepted 9 February 2009
Available online 6 March 2009

Keywords:
Mars, surface

A B S T R A C T

The majority of martian valley networks are found on Noachian-aged terrain and are attributed to be the result of a ‘warm and wet’ climate that prevailed early in Mars’ history. Younger valleys have been identified, though these are largely interpreted to be the result of localized conditions associated with the melting of ice from endogenic heat sources. Sinton crater, a 60 km diameter impact basin in the Deuteronilus Mensae region of the dichotomy boundary, is characterized by small anastomosing valley networks that are located radial to the crater rim. Large scale deposits, interpreted to be the remains of debris covered glaciers, have been identified in the area surrounding Sinton, and our observations have revealed the occurrence of an ice rich fill deposit within the crater itself. We have conducted a detailed investigation into the Sinton valley networks with all the available remote data sets and have dated their formation to the Amazonian/Hesperian boundary. The spatial and temporal association between Sinton crater and the valley networks suggest that the impact was responsible for their formation. We find that the energy provided by an asteroid impact into surficial deposits of snow/ice is sufficient to generate the required volumes of melt water needed for the valley formation. We therefore interpret these valleys to represent a distinct class of martian valley networks. This example demonstrates the potential for impacts to cause the onset of fluvial erosion on Mars. Our results also suggest that periods of glacial activity occurred throughout the Amazonian and into the Hesperian in association with variations in spin orbital parameters.

© 2009 Elsevier Inc. All rights reserved.

1. Introduction

Valley networks commonly occur as extensive systems in the Noachian highlands (Carr, 1996) and are often connected by large open-basin lakes (e.g., Fassett and Head, 2008), suggesting widespread fluvial erosion and drainage toward the northern lowlands. Most valley networks ceased forming at about the Noachian-Hesperian boundary (e.g., Fassett and Head, 2007). The few that are observed to have formed during the Hesperian appear in different settings (e.g., volcanic edifices and plateaus near outflow channel sources) and appear to be due to local environmental conditions (e.g., Fassett and Head, 2005, 2006; Mangold et al., 2008) rather than a major reversion to Noachian climate conditions.

Sinton is located within an isolated plateau along the northern dichotomy boundary at Deuteronilus Mensae (Fig. 1). The majority of the valleys display broad, steep walls and anastomosing patterns around streamlined landforms. Despite extensive surveys of the surfaces of the surrounding plateaux and highlands, no other fluvial erosional features were identified in the vicinity. This suggests that the valley formation was related to the impact that created Sinton crater. Crater counts along the ejecta blanket of Sinton on which the valleys are situated provides a maximum age for the impact event close to the Hesperian/Amazonian boundary.

The study region is also host to extensive deposits of lineated valley fill (LVF) and lobate debris aprons (LDA) which have been interpreted to represent the remains of debris covered glacial systems (Lucchitta, 1984; Head et al., 2006a, 2006b; Morgan et al., 2009). Evidence exists to suggest that multiple episodes of LVF emplacement have occurred in the study area (Morgan et al., 2009). We therefore explore the possibility that the impact occurred at a time when the plateau surface was covered in ice. Such a scenario represents a special case of the impact induced valley formation model first proposed by Brackenridge et al. (1985), and thus the Sinton valleys should appear to represent a new class of valley network associated with local environmental conditions. If our interpretation is correct it implies that snow and ice deposition along the dichotomy boundary was extensive during the Late Hesperian–Early Amazonian. This suggests that Mars has undergone significant ‘glacial’ periods most likely associated with obliquity variations through large portions of its history. The associated climatic variations may have been responsible for significant modification of the dichotomy boundary. In the following sections we discuss...
the morphology of the valleys, review the evidence for their formation and develop criteria to recognize similar features elsewhere on Mars.

2. Geology of the Sinton region

The study area is located within the Deuteronilus Mensae region of the northern dichotomy boundary (Fig. 1). This portion of the boundary forms an escarpment separating the highlands from the northern plains. The region is characterized by fretted valleys, which divide the most northern reaches of the highlands into plateaus and mesas that become progressively smaller to the north. These remnants fragments of the highlands are considered to be typical of the material that comprise Noachis Terra to the south and are interpreted to be composed of both sediments and volcanic material that has been mixed and reworked by impacts (Tanaka et al., 2005). The fretted valleys within the study region and throughout the northern dichotomy boundary are filled with Lineated Valley Fill (LVF) and the walls of the plateau and mesas are flanked by Lobate Debris Aprons (LDA). Much debate has surrounded the origin of the LVF/LDA deposits studies, though most authors agree that ice was involved in their formation (Carr and Schaber, 1977; Squyres, 1978; Lucchitta, 1984; Head et al., 2006a, 2006b). Comprehensive examinations of the morphological properties of the deposits have argued that they are the remnants of debris-covered glaciers (e.g., Lucchitta, 1984; Head et al., 2006a, 2006b). Morgan et al. (2009) report on detailed investigations into the LVF/LDA deposits in this region and found evidence for integrated flow patterns interpreted to represent extensive valley glacial deposits, supporting previous work concluding that Amazonian glaciation played a role in the degradation of the fretted valleys (e.g., Head et al., 2006a, 2006b).

2.1. Plateau surface

The main plateau on which Sinton crater is situated extends for over 300 km in a NW by SE direction and is ~100 km at its widest extent (Fig. 2). It is surrounded on all sides by steep slopes (Fig. 2b) that rise over 1 km from the surrounding LDA and LVF. The plateau has been mapped by Tanaka et al. (2005) to consist of highland material and represents the same unit as the elevated terrain to the south. From the outer flanks of the plateau, there is a more gradual rise to the center, which culminates in a peak elevation of ~700 m above martian datum. This topographic rise represents the western rim of an ancient highly degraded 350 km Ismeniae Fossae impact basin. The largest crater on the plateau is the well-preserved Sinton crater which is flanked to the east by a ~30 km depression (presumably a relict impact crater) that is degraded to the point that it is only clearly identifiable in MOLA topographic data (Fig. 2). The resulting topography between these two impact features is a large escarpment tens of kilometers in length that runs in a SE direction from Sinton crater. Alcoves, several kilometers across, have been eroded into the flanks of the plateau, and extend for up to tens of kilometer into the plateau center, forming elongated valleys. The majority of alcoves are located along the south outer flanks, though some are found along cliffs facing all orientations.

2.2. LVF/LDA in the study region

Large-scale deposits of LVF/LDA are abundant throughout the region surrounding the main plateau. The large ~20 km wide valley to the south of the plateau is completely filled with LVF, as are the valleys to the east (Fig. 2). LDA deposits predominately emanate from the northern flanks of the plateau and are also located around all of the surrounding isolated mesas (Fig. 2). Detailed studies of the region by Morgan et al. (2009) have demonstrated that both LDA and LVF systems are fully integrated and likely represent different manifestations of the same material. Small lobes of LVF (~3 km wide) which emanate from within the alcoves merge with, and feed the main deposits of LVF/LDA (Morgan et al., 2009). Lineations of flow can be traced across the surface of the LVF from the alcoves to where the deposits open up as LDA onto the surrounding plains. This demonstrates that the deposits of LVF/LDA represent large scale integrated systems that extend over areas of 30,000 km² within the study area (Head et al., 2006b; Morgan et al., 2009). Integrated systems of this scale are consistent with terrestrial glacial landsystems, which originate within sheltered cirques and open up into large-scale valley glaciers. Head et al. (2006a, 2006b) have argued that surface lag deposits have preserved underlying ice from sublimation in a manner analogous to the sublimation tills found on the surface of debris-covered glaciers in the McMurdo Dry Valleys of Antarctica (Marchant and Head, 2007).

Martian general circulation models indicate that the formation of glaciers could have been initiated during previous climatic regimes associated with the cyclic transition from higher (~35°) to lower (present) obliquity values (Madeleine et al., 2007). Under these conditions there is a redistribution of volatiles from large-scale glacial systems on the flanks of the Tharsis Mons (e.g., Head.

Fig. 1. Context MOLA topographic map of the northern dichotomy boundary at Deuteronilus–Protonilus Mensae; the location of the study region is highlighted by the red box. Note the gradual decrease in topography from the southern highlands into the northern lowlands. The boundary area is characterized by the occurrence of numerous mesas which become increasingly smaller to the north. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
Valley network formation from ice/impact interactions

Fig. 2. (a) Sinton crater and vicinity. The map highlights the extent of the LVF and LDA systems within the region (in valleys with the outer boundary indicated by the dashed white line), which completely surround the plateau on which Sinton is located (portion of HRSC image 1589, overlain by MOLA topographic data). (b) Slope map. The fretted terrain within the study area consists of relatively smooth-surfaced plateaus and mesas surrounded by steep flanks exhibiting slopes of over 23° (derived from the DTM). Smooth crater-fill deposits line the floor of Sinton crater and the terraces along its interior flanks. The LVF/LDA deposits surrounding the main plateau exhibit almost entirely horizontal surfaces, which consist of slopes less than 0.5° (dark blue). Note the presence of a steep topographic ridge on the surface of the main plateau that extends to the southeast from Sinton crater. Slope map overlain on a portion of HRSC image 1589. (c) Map of the distribution of: (i) Valley networks. Note that the valleys are located around Sinton crater, and the majority originate close to crater rim. The regions depicted in white represent topographic hollows within the plateau. Note that valleys originate close to the NE topographic ridge along the plateau surface that can be seen in (b). (ii) Secondary craters and ejecta material (highlighted in yellow) interpreted to have been formed by the impact that created Sinton crater. HRSC image 1589. (d) 300 m contour lines overlain on a Viking image mosaic. Note that the LVF/LDA surrounding the main plateau can be seen clearly in the image (compare with (a)). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

and Marchant, 2003; Shean et al., 2007) to the dichotomy boundary where it is deposited as snow.

Some source lobes of LVF within the study area (Morgan et al., 2009) emerge from alcoves and are superimposed on, rather than integrated with, the main trunk LVF/LDA deposits. These occurrences argue for a subsequent period of ice emplacement to the one that deposited that main trunk LVF/LDA deposits. This supports previous observations of similar features to the east (Levy et al., 2007; Dickson et al., 2008a), suggesting that multiple episodes of glacial activity have occurred along the study region and the northern dichotomy boundary as a whole.

2.3. Sinton crater

The classification of craters on Mars is made difficult by the diversity of primary impact morphologies relative to the other terrestrial bodies (Strom et al., 1992); nevertheless, Sinton is clearly a typical complex crater with a diameter of 63 km, a central peak and internal wall terraces. The complex topography of the main plateau results in the crest of the crater rim having an uneven elevation. The highest sections are to the southeast and exhibit an elevation of ~660 m, which is ~1990 m above the lowest to the north (Fig. 2). Such a setting has undoubtedly had an effect on the modification stages of the crater development, causing the wall terraces to be most developed along the southern flanks of the crater.

The interior of the crater (Fig. 3) as well as the surface of some of the terraces are covered by a material that appears largely smooth at HRSC resolution, implying that some form of deposition has occurred in the crater since the formation of Sinton (Fig. 4). Although largely smooth, there are some interesting features within the crater fill material. MOC images reveal a complex dissected surface texture, comprised of a range of morphologies from small pits and buttes, tens of meters across, to more gradual undulating surfaces (Fig. 4a). To the south of the central uplift there is a region
of pitted terrain (Fig. 3) visible on coarser-resolution images that consists of closely spaced depressions of the order of hundreds of meters across (Fig. 4b). Such crater fill is common within craters of this size (30–100 km) within the mid-latitudes (Crown and Bleamaster, 2007). The morphology is similar to the nature of surfaces of the LVF and LDA within the study area and across the dichotomy boundary as a whole. Such surface textures are interpreted to represent the sublimation of ice from within the LVF/LDA (Carr, 2001; Malin and Edgett, 2001; Mangold, 2003; Levy et al., 2009). Gradual small-scale mass wasting along fractures within the surface of LVF deposits has been envisioned as the process of formation; at temperatures exceeding the frost point, interstitial ice in contact with the atmosphere sublimes, destabilizing solid grains of debris and causing the lateral regression of the fracture to produce a pit (Mangold, 2003). Although the interpretation of LVF/LDA hypothesized by these authors is still debated, the presence of potential sublimation features points to the important role that ice appears to play in the development of the landforms. The ice-related interpretation of such features is also supported by the occurrence of degraded craters with a unique morphology that are found exclusively on the surface of LDA/LVF (Mangold, 2003). Such craters, called ‘oyster shell’ and ‘ghost’ craters, are present within the surface of the crater fill (Figs. 4c and 4d) and do not appear to have experienced conventional modes of erosion, but are interpreted to be the result of the loss of volatiles from within the material into which the projectiles impacted (Mangold, 2003), or primary crater forms (ring-mold cratering) subsequently degraded (Kress et al., 2008).

Patterns indicative of flow within the crater fill material are evident along the interior crater flanks. This also supports the interpretation that Sinton crater fill is composed of the same material as LVF/LDA. Lobes bounded by concentric raised ridges are present within the fill material occupying crater rim terraces close to the top of crater rim (Fig. 4e). These are morphologically similar to the lobes of LVF identified within the alcoves that surround the plateau (Morgan et al., 2009) and thus further suggest that the crater fill consists of LVF/LDA material. The lobes are up to 3 km in length, though their size seems to be controlled by the scale of the terrace on which they occur. The presence of LVF/LDA-like material within Sinton crater suggests that the emplacement of ice within the study region is not just restricted to the fretted valleys, but also occurred on the main plateau itself. This is further supported by the occurrence of fill material within other relatively large craters on the plateau.

2.4. Ejecta deposit and distribution

The distribution of ejecta associated with Sinton crater is made complex by the fact that the projectile that formed it impacted an isolated plateau that is of a size comparable to the resulting crater. Nevertheless the crater displays an uneven and largely poorly defined ejecta blanket around the remainder of the plateau (Fig. 2). To the southeast, there is an apparent radial pattern produced by chains of secondary impacts that form striation-like gouges oriented perpendicular to the crater rim. Such ejecta-related disturbances across the plateau surface are particularly evident at MOC resolution. To the north of the crater on the adjacent lowland plains there is an accumulation of debris which is likely to have been formed from material ejected during the impact, representing the northern extent of the ejecta deposit (Fig. 2).

The influence of the impact is also present on the highland surfaces to the south (Figs. 2 and 5). This is predominantly observed in the form of secondary craters. These are identified by their characteristic morphology, being both shallow and elliptical with their longest axis oriented perpendicular to the crater rim. Despite the extensive distribution of the ejecta to the south, there is no disruption present on the LVF surface within the valley to the south of the main plateau, or in other areas of LVF in the immediate vicinity of the plateau flanks (Fig. 5d). The lack of ejecta and secondaries on the LVF to the south is interpreted to be due to resurfacing subsequent to the impact event.

The most outstanding aspect of the plateau surface surrounding the crater is the occurrence of dense valley networks (Fig. 2). This characteristic makes it truly unique relative to other craters within the region. Fig. 5 compares Sinton with three different craters displaying the range of morphologies present in craters of a similar
Valley network formation from ice/impact interactions

Fig. 4. Surface morphologies of Sinton crater fill deposits. (a) Comparison between the surface texture of the crater fill material along the floor of Sinton crater (i) and the LVF deposits along the floor of the southern valley system at MOC resolution. The similarity between the two surface morphologies (consisting of knobs and pits) suggests that the deposits share a similar origin. The pitted texture is interpreted to be caused by the action of sublimation (e.g., Mangold, 2003; Levy et al., 2009) and thus suggests that the deposits once contained a significant amount of ice. MOC images: M0300514 (a), R0700879 (b). (b) Image of the 100 m scale pits in the surface of the crater fill south of the central peak. HRSC image 1589. (c) Image of a ‘ghost’ crater in the crater fill material. MOC image: M0300514. (d) Image of ‘oyster shell’ craters in the crater fill material (see also “ring-mold craters” of Kress et al., 2008). MOC image: M0300514. (e) Examples of flow features in crater fill within one of the terraces in the crater wall. The white arrows represent the implied flow direction, which is consistent with the local downslope direction. THEMIS image V02146006.
size and freshness across the northern plains, where uneven topography does not have a major affect on the distribution of ejecta. Sinton ejecta shares some similarities with Arandas crater (Fig. 5c); which has a strong radial fabric and the unnamed crater in Fig. 5d; which has a ballistic ejecta containing strings of secondaries reminiscent of the secondary chains observed on the surface of the plateau to the south of the study area. The major difference though between Sinton and other craters of a similar size is that the ejecta deposit around Sinton is very poorly defined and far less apparent than the other craters despite similar levels of freshness. This suggests that subsequent to the impact, unusual erosional processes have operated to modify the Sinton crater ejecta deposit.

3. Valley networks

3.1. Valley networks on the plateau surface

The majority of valley networks are located directly to the south and southeast of Sinton, though some are found as far as 75 km from the crater rim crest (Figs. 2 and 5). Many of the valleys have an orientation in a direction perpendicular to the crater rim. The valleys exhibit a range of widths from 30–500 m (the smaller ones being only evident in MOC/HiRISE images), and the longest networks extend >40 km (Figs. 6–10). The valleys are generally steep-sided and have flat floors (Figs. 7, 9, 10), although it is not clear if the flat floors are primary or are a result of the deposition of material within them. No channels are observed within any of the valleys at any of the available resolutions (Figs. 6–10). Tributary systems are a common occurrence for valleys of all scales and in some instances valleys diverge into several branches that remerge downslope, producing anastomosing patterns (Fig. 10). The source of the valleys is poorly defined; the majority of the valleys either emerge from broad topographic depressions close to the rim of Sinton, or simply originate from plains adjacent to the crater rim crest (Figs. 2 and 6). Valleys are observed to disappear into smooth hollows that are located across the plateau and reemerge from the down-slope side (Figs. 2, 7, 11). These depressions occur at a range of scales (of the order of several kilometers) and shapes, and their origin is uncertain, although they may predate or have been formed during the impact of Sinton. The valleys are found in three broad regions around the outer rim of Sinton: to the southeast, directly to the south and to the west (Fig. 2). We investigated each of these areas individually with all available data sets and describe our findings in detail below.
3.1.1. South east of Sinton crater

The southeastern portion of the plateau displays the longest valleys, extending for ~50 km from the rim of Sinton to the plateau outer flanks (Fig. 6). The plateau in this region is characterized by modest gradients of ~1.5° and there are no significant changes in slope associated with the erosional features. The widest valleys are ~500 m across, but fluvial erosion features of all scales are present. Despite the meandering displayed by most of the valleys, all of the erosional features are strongly orientated along a northeast–southwest direction, which is perpendicular to the rim of Sinton (Figs. 6–9). This is especially evident in the smaller networks, but is also observed in the large valleys. These are aligned in a northeast direction for almost their entire length apart from short, sharp deflections consisting of 90° meanders (Figs. 2, 6–9). This suggests that there may be some form of structural control associated with Sinton ejecta that has affected the valleys orientation.

The source of the southeastern valleys is not readily apparent, a situation similar to other valleys across the plateau. Some of the valleys to the east are found to originate close to the drainage divide formed by the summit of the escarpment that runs in a northeastern direction from the main crater rim (Fig. 10). Small
valley networks that feed some of the long valley networks are present close to the southern edge of the drainage divide. The northern and eastern side of the escarpment is relatively steeper (>6°) than the southern side (<3°) and is host to small-scale valleys, ~5 km in length. These small-scale networks originate close to the edge of the divide and drain into depressions aligned with the foot of the escarpment (Fig. 10).

The most elevated portions of the valleys are spaced ~5 km apart, and are joined by smaller-scale tributaries (Fig. 7). Teardrop-shaped landforms, ~200 m long, are present within the wider channels and close to tributary junctions (Fig. 7). Many of the tributaries of the upper portions of the networks join the main trunk valleys as hanging valleys (Fig. 7). Further downslope the valleys coalesce to form complex anastomosing patterns, which open up into broader confluences that are filled with numerous streamlined ‘islands’ (Figs. 8 and 9). Despite the multiple interactions with the smaller scale valleys, the wider and most deeply incised valleys exhibit near constant widths for the majority of their lengths.

Smaller-scale erosional features are resolvable within the valley networks within CTX images. Terraces hundreds of meters wide are present along the outer edge of some of the wider confluences (Fig. 8). Longitudinal groves and etched troughs can also be seen on the floors of channels, although deposition of material within the valleys since their formation may be obscuring other such features elsewhere on the plateau.

3.1.2. South of Sinton crater

Large numbers of valley networks and associated fluvial features are present along the southern portion of the plateau (Fig. 11). The valleys originate along the outer edge of Sinton and extend to the southern flanks of the plateau. The gradient of this portion of the plateau is slightly steeper, with a value ~2.5°, than to the northeast, although higher values (~5°) are encountered close to the outer rim crest of Sinton where the longer valley networks originate. As was the case to the east, the valleys display anastomosing patterns, and streamlined landforms are prevalent.

In many instances the channels are found to drain into the ~5 km alcoves cut into the southern cliffs of the plateau (Figs. 2 and 11); this is not only the case for the valleys directly south of Sinton but is also seen to the southeast (Fig. 6) thus raising the possibility that the erosion that generated the channels also played a role in creating or modifying the alcoves (Fig. 11). The occurrence of alcoves without channels elsewhere in the region (Fig. 2), argues that there is no direct relationship between channels and alcoves except that they provide local topographic lows into which the channels drain. There is also no evidence for any form of disturbance on the surface of the LVF within the alcoves associated with the mouths of the valley systems, indicating that the LVF postdates the valleys. High-resolution MOC/HiRISE images of the alcoves reveal that a large number of the channels do not actually cut down into the edge of the alcoves themselves, but stop abruptly 10–100 s of meters away from the alcove edge within a region marked by a distinctive transition in surface texture (Fig. 12). The area immediately surrounding the alcove appears smoother at MOC resolution than that of the surface the valley networks incise, which consist of more hummocky terrain, that likely represents ejecta material from Sinton crater (Fig. 12). This suggests that some erosional process has operated to remove the downstream portion of the valleys and the surrounding surfaces as it seems inconceivable that the valleys would abruptly terminate 100s of meters short of the edge of the alcoves. The removal of material around the alcove argues...
for post-valley network formation erosion. This is consistent with an episode of alcove enlargement that occurred during a phase of LVF (glacial) activity (Morgan et al., 2009). Indeed the transition between the two units may represent the maximum extent of LVF within the alcove during a period of ice accumulation that occurred subsequent to the formation of the valley networks.

3.1.3. West of Sinton crater

There are significantly fewer erosional features present directly west of Sinton crater compared to the south and southeastern portions of the plateau. One wide valley network (650 m) does exist and extends due west for 15 km from Sinton, where its terminus is obscured by the ejecta of a fresh 1 km diameter impact crater (Fig. 2). The network exhibits many of the features present in the other valleys, including multiple conduits, steep walls and broad valley floor. A 2 km ‘island’ has been formed by the divergence and subsequent convergence of the valley networks. The northern most valley is less defined than the one to the south and is comprised of several smaller anastomosing channels (∼100 m wide).

3.2. Valley networks within Sinton crater

Small-scale discontinuous valley networks are present on the exposed sections of the internal flanks of the crater (Fig. 13). These valleys are ∼300 m at their widest extent and the longest uninterupted sections are ∼10 km in length. The level of incision and density of the valley networks is significantly lower than those on the plateau surface outside of Sinton. The spatial resolution of MOLA data is too low to permit the precise measurement of the depths of the valleys, but estimates from THEMIS images indicates the depths to be of the order of tens of meters. Their distribution is limited to the crater interior walls which are located below the highest elevated portions of the crater rim to the southeast and northeast (Fig. 3). The valleys are found along the length of these slopes. The widest valleys are located close to the rim crest (Fig. 14). It is difficult to identify the source of these higher-elevation portions of the valleys as the majority emerge from depressions within the upper terraces. No valley networks have been observed at any of the available resolutions on the central peak of Sinton, although they do occur at the same elevations on the surrounding crater wall slopes. The complexity of the valley morphology varies throughout the crater, and appears to be related to slope angle, with the most dendritic drainage configurations consisting of multiple tributaries occurring along the crater terraces, which exhibit slopes close to one degree (Figs. 2 and 14). Where the slopes become steeper than this, the valleys show simpler configurations, consisting of increasingly straight channels that are orientated parallel to each other in a downslope direction.

At a number of locations close to the base of the internal slopes of the crater, valley networks open up into elongated alcoves, of the order of a kilometer wide and 5–10 km long. Fan-shaped features several kilometers wide protrude out from the mouths of the alcoves and extend out for ∼3 km over the plains of the crater floor (Fig. 14). A similar-sized and shaped landform is also present further up the crater interior flanks above the crater floor where there are breaks in slope (#3 in Fig. 14). Valley networks terminate at the upslope portion of the apex of the feature and other networks originate at the downslope margins where there is a prominent increase in slope of ∼5°. At THEMIS resolution (∼18 m/pixel), the surface of all of these features appears to consist of a smooth texture that is similar in nature to the material which occupies the surfaces of the terraces within the internal crater flanks (e.g., #2 in Fig. 13). Other valley networks also terminate into the similarly smooth textured material comprising the fans, and there are several more examples of accompanying valley networks that originate on the downslope side. In one case the downslope portion of the valley emanates from a theater-shaped depression (Fig. 14).

4. Interpretations of the valley networks and comparisons with other martian erosional features

4.1. Valley networks on the plateau surface

The availability of high-resolution data sets of the martian surface over the last decade have enabled multiple valley networks and associated erosional features of comparative size to those that surround Sinton to be identified and analyzed in detail. These have included examples present in both Noachian (e.g., Fassett and Head, 2005) and Hesperian (e.g., Ansan and Mangold, 2006; Fassett and Head, 2006) aged terrain, and have been interpreted to have formed over differing periods of time as a result of varying degrees of activity, from prolonged episodes of pluvial activity (e.g., Mangold et al., 2004) to the catastrophic releases of water (Mangold et al., 2008).
A variety of valleys with diverse morphologies have been identified elsewhere on Mars on relatively gentle slopes (similar to those of the surface of Sinton plateau, <3°). Typical Noachian valley networks with sizes comparable to the Sinton external valleys display well-developed meanders and dendritic patterns that consist of multiple tributaries (Carr, 1996). Morphological and morphometric comparisons with terrestrial networks suggest that the martian networks were formed gradually during warmer and wetter climate conditions (Carr, 1996). The identification of Noachian interconnected paleo-crater lakes, requiring sufficient influx of water from feeder valleys to fill the crater so that breaches can occur, further favors clement conditions (Fassett and Head, 2005, 2008). Localized episodes of fluvial activity may have occurred during the Hesperian in the form of similar dendritic valley systems on 2.8–3.4 Gyr old terrain to the west of Echus Chasma (Mangold et al., 2004). The morphology of these valleys contrasts somewhat to the Sinton valleys, which exhibit only limited meandering and are never characterized by more than three clear Strahler's stream orders, four orders less than the valleys west of Echus Chasma.

Valley networks tens of kilometer in length can also form within small-scale outflow channel systems as a result of high discharge, low frequency events. A 60 km long, 50 m deep valley...
network is present along the outer reaches of an outflow channel south of Nili Fossae in Syrtis Major, interpreted by Mangold et al. (2008) to be the result of subsurface discharge of water due to volcanic activity. This valley shows some similarities with the Sinton valleys (in that limited tributaries feed it), but it differs morphologically in that they begin in theater shaped heads. The scale of incision is also more substantial as the valley is $\sim 30 \text{ m}$ deeper that the largest Sinton valley and up to 300 m wider. This is especially significant since the outflow valleys are cut into volcanic plains that are substantially more resistant to erosion than the poorly consolidated ejecta material of the Sinton crater valleys. In addition, the Nili Fossae outflow channel is accompanied by other erosional landforms (e.g., groves and tear drop shaped islands) that are external to the valley itself. Similar features are absent on the Sinton plateau at this scale, perhaps due to a much higher flux of water in Syrtis Major relative to the Sinton valley networks.

The flanks of some of the relatively small volcanoes (180–270 km edifice diameter) in Tharsis and Elysium have valleys that extend radially from the outer edge of the summit calderas (Fassett and Head, 2006, 2007). These valleys range in width from 200 m to 2.5 km and have been interpreted to be fluvial in origin. The steeper slopes on the volcanoes ($\sim 8^\circ$) relative to the Sinton plateau ($1.5–3^\circ$) are interpreted to have caused the valleys on the volcanoes to consist of sub-parallel, immature drainage patterns (Fassett and Head, 2007). Despite this difference in slope, there are some similarities between the two networks. Both systems display valleys that begin abruptly and have near-constant widths and depths over their entire lengths. This is a characteristic that has also been noted in fluvial features in the McMurdo Dry Valleys of Antarctica (Fassett and Head, 2006), a region regarded as an applicable terrestrial analog (e.g., Anderson et al., 1972; Gibson et al., 1983; Mahaney et al., 2001; Wentworth et al., 2005).

The general morphology of the Sinton crater valleys is unique relative to other similar-sized valleys previously identified on Mars: they consist of broad flat floors and steep walls; anastomosing drainage patterns; hanging valleys; streamlined islands; and a strong spatial association with an impact crater. The broad expansion of the fluvial features of all scales along the plateau surface (Fig. 2), and especially to the southeast (Fig. 6) indicate that fluvial erosion must have been rather intense, though concentrated on the plateau surface surrounding Sinton crater.

The occurrence of main trunk valleys (more deeply incised into the plateau surface than the smaller-scale fluvial features that surround them) suggests that these larger valleys formed over a longer time period as a result of the initiation of stable drainage configurations concentrating fluid flow along the most efficient drainage pathways (i.e. those with a larger hydraulic radius) (Fig. 15). This would have been accompanied by the abandonment of the smaller (presumably sediment clogged) fluvial pathways, leaving them as relict features and hanging valleys perched on the plateau surface surrounding the main valleys. However, because the smaller valleys have been preserved and thus were not removed by the progressive down cutting and widening of the larger valleys, the fluvial erosion could not have been prolonged sufficiently to enable mature drainage configurations to develop on the plateau surface (Fig. 15). This is further supported by the constant width of the large valleys, which demonstrates their lack of maturity. We therefore interpret the Sinton external valleys to represent a distinct type of martian valley network that was formed and modified by processes potentially differing from those responsible for the erosion of the other types of valleys on Mars.
Fig. 13. 500 m wide valley network present on the surface to the west of the main plateau. The valley system is located directly to the west of Sinton crater and constitutes the widest valley network on the plateau. The source of the valley is unclear, as the upstream portions of the valley widen and become indistinguishable from the plateau surface ~1 km from the outer rim of Sinton. Despite the scale of the valley, it exhibits many of the features present in the other networks (Figs. 6 and 11), including anastomosing channels, steep walls and a broad valley floor. North is at the top of the figure. THEMIS image V12506004.

4.2. Estimation of discharge

Estimating the discharge responsible for the erosion of fluvial features from remotely sensed data sets requires several steps, and contains a large degree of uncertainty. Due to the small scale of the valleys, it is difficult to establish their cross-sectional dimensions accurately. Nevertheless, establishing estimates for discharge is important as it allows us to place some constraints on the geologic processes responsible. Komar's (1979) modification of the Manning equation which corrects for martian gravity, has commonly been applied to the calculation of martian fluvial networks:

\[
Q = A \left( \frac{g_m R^{4/3}}{g_e n^2} \right)^{1/2},
\]

where \( A \) is the cross sectional area, \( R \) is the hydraulic radius (the ratio of flow cross sectional area to wetted perimeter), \( g_m \) and \( g_e \) are the values for gravity on Mars and Earth respectfully and \( n \) is the Manning coefficient. As there is no means to establish the depth of fluid in the valleys, we assume bankfull discharge, which will provide an estimate of maximum discharge. Valley depth was estimated from individual MOLA points and the width was measured directly from the highest resolution images. The valleys were assumed to have had a rectangular cross section. Slope measurements were derived from high resolution HRSC DEMs (~200 m). The Manning coefficient involves additional uncertainty as it is determined empirically for terrestrial channels and cannot be measured directly from the available data. We used the approximation of Wilson et al. (2004) for the coefficient (0.0545) that is more appropriate for martian conditions. It is unknown whether the fluvial erosion was continuous or if there were several episodes of activity. As discussed in the previous section, there appears to have been a concentration of flow in the larger valley networks at some point after the initiation of the fluvial features (Fig. 15). Therefore, for the purposes of calculating the discharge we have only considered the flow within the largest valleys from the southeastern, southern and western portions of the plateau. These valleys are also transected by suitable individual MOLA orbital tracks which permitted cross section dimensions to be estimated. The measurements we made and our estimates of discharge are summarized in Table 1 below.

The largest valleys surrounding Sinton crater have large estimated peak discharge values of \( 2.5 \times 10^5 - 1.4 \times 10^5 \) m\(^3\) s\(^{-1}\) (Ta-
there is uncertainty in the valley cross-sectional profiles and the widest network, situated directly west of Sinton (Fig. 13). Although ble 1), with the upper limit of this range corresponding to the widest network, situated directly west of Sinton (Fig. 13). Although there is uncertainty in the valley cross-sectional profiles and the nature of fluvial erosion under martian conditions (e.g., Fassett and Head, 2005, 2007), the discharge estimates for the Sinton valleys are over two orders of magnitude greater than the 80 km long valley network that enters Jezero crater. The Jezero crater valley was interpreted to have formed over relatively prolonged time periods (minimum time of ∼10–18 years) under Late Noachian climatic conditions (e.g., Fassett and Head, 2005). The Sinton discharge estimates are more consistent with valley networks formed as the result of volcanic thermal anomalies releasing significant amounts of water from snow/ice deposits (e.g., Ceraunius Tholus, Fassett and Head, 2007) and groundwater sources (e.g., Syrtis Major Planum outflow channel, Mangold et al., 2008). In a later section, we assess the most probable source of water that would have been capable of supplying such high peak discharges.

4.3. Valley networks within Sinton crater

The drainage configurations of the internal valleys appear to be correlated with slope, with the more dendritic patterns present on low slopes (<6°) and the simpler, straighter valleys found on the steeper sections (10–20°). Such drainage patterns are consistent with other networks on Mars and on the Earth.

The morphology and smooth texture of the fan shaped features which occur where the valleys terminate onto flat surfaces (Fig. 14) is consistent with the material being depositional in nature and is interpreted to represent sediment that was deposited when the valley networks were active. We interpret the fan shaped features to be alluvial fans deposited by stream flow during this period of activity. The presence of such broad fans suggests that they were built up gradually, perhaps as a result of a discrete low frequency events.

The intriguing aspect of these valley networks compared to other martian examples is their: (1) small scale, (2) localized nature, (3) low drainage densities, and (4) discontinuous nature in the downslope direction. Valleys exhibiting discontinuities within their networks are not uncommon on Mars, largely due to the effects of post-formation erosion and deposition over several billion years (e.g., Carr, 1996). In the case of the Sinton interior valleys, their intermittent nature suggests that surface runoff infiltrated into the ground and traveled downslope as throughflow before reemerging at the surface to continue eroding the valley. For example, a valley network can be seen emerging from directly below the outer margins of the fan deposit located along the crater terrace (Fig. 14), supporting this interpretation. Similar relationships between channels and fans are also observed within the small martian gullies first observed by Malin and Edgett (2000). These landforms are slightly smaller (∼1.5 km difference in length) than those within Sinton, but exhibit similar morphological characteristics including sinuous channels and depositional fans. The Malin and Edgett (2000) gullies also consist of isolated drainage configurations and are commonly located within crater interiors. Progradational and abandoned channels are prominent on some of the gully fan systems, suggesting that they have experienced multiple episodes of activity. This activity has been attributed to the melting of volatiles during periods of high obliquity (e.g. Chirstensen, 2003; Dickson et al., 2007), and the gullies are morphologically similar to small-scale fluvial features in the McMurdo Dry Valleys of Antarctica that have been produced by summer meltwater from snowpacks (e.g. Morgan et al., 2007, 2008). Observations of the Antarctic gullies found that channel runoff would frequently infiltrate into fans and resurface downslope forming theater shaped channel systems (Morgan et al., 2007, 2008; Marchant and Head, 2007) that are analogous to those within Sinton crater. Hence, we interpret these features to have been sourced from water infiltrating into the smooth sediments deposited along the terrace, and trav-

<table>
<thead>
<tr>
<th>Valley system</th>
<th>Width (m)</th>
<th>Depth (m)</th>
<th>Slope (°)</th>
<th>Q (m³ s⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sinton Southeast</td>
<td>320</td>
<td>20</td>
<td>1.5</td>
<td>8 × 10⁴</td>
</tr>
<tr>
<td>Sinton South</td>
<td>234</td>
<td>12</td>
<td>2.8</td>
<td>2.5 × 10⁴</td>
</tr>
<tr>
<td>Sinton West</td>
<td>480</td>
<td>50</td>
<td>2</td>
<td>1.4 × 10⁵</td>
</tr>
<tr>
<td>Valley entering Jezero Crater, Nili Fossae</td>
<td>170–400</td>
<td>–</td>
<td>–</td>
<td>500–900</td>
</tr>
<tr>
<td>Ceraunius Tholus Valleys</td>
<td>250–700</td>
<td>20–60</td>
<td>5–8</td>
<td>4 × 10⁴–</td>
</tr>
<tr>
<td>(Fassett and Head, 2005)</td>
<td></td>
<td></td>
<td></td>
<td>6 × 10⁵</td>
</tr>
<tr>
<td>Syrtis major planum outflow channel</td>
<td>500</td>
<td>50</td>
<td>0.7</td>
<td>5 × 10⁵</td>
</tr>
<tr>
<td>(Mangold et al., 2008)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Fig. 15. Conceptual model of the evolution of the valley networks on the surface of the plateau. 1. Anastomosing small-scale valleys form in response to the introduction of surface runoff on the plateau surface. 2. Over time, the runoff becomes concentrated into two of the valleys as the system attempts to maximize efficiency. This causes the valleys to become increasingly incised and widened as a result of the increase in erosion. The other channels are left clogged with sediment and hanging above the main valleys as flow within them ceases. 3. If fluvial erosion continued over longer periods (years–centuries) under relatively constant discharge, the valleys would have become progressively downcut into the plateau and widened in response to mass wasting along the valley flanks. This would cause erosion and eventual removal of the smaller abandoned valley networks. The preservation of smaller valleys on the plateau surface argues that the valleys never reached stage 3, and thus represent relatively short-lived activity.

Table 1

Estimates of channel-forming discharge for the largest valley within each region of the plateau (using Komar, 1979 modification of the Manning equation, and Wilson et al., 2004 estimate of the Manning coefficient for martian channels) compared to estimates for similar-sized martian valleys.
The differences between the internal and external valleys may be due to the different environments in which they are located (i.e. crater interior compared to plateau surface respectively). Alternatively it is possible the valleys did not form at the same time and by the same processes that formed the external networks. Internal valley networks have been identified within other impact craters, such as Lyot (Dickson et al., 2008b), but these craters lack the extensive valley networks on their outer flanks, further suggesting that they formed independently of the external Sinton valleys.

5. Age relations

Dating the landforms in the study area is essential to the understanding of the geological history of the region. The strong spatial correlation between Sinton crater and the valley networks suggest that the impact was related to their formation. The lack of superposition by the crater or any of its ejecta on the valley networks indicates that the valleys postdate its formation, an interpretation that is supported by the occurrence of valley networks within the crater itself. The event that formed Sinton was instantaneous, and thus if the impact can be dated, it would place a time constraint on the development of the region and provide a maximum age for the formation of the valley networks. Successful impact crater size-frequency distribution studies require suitably-sized areas in which to carry out the counts. Due to the absence of a well-defined ejecta deposit surrounding the main crater it is not possible to use this as a unit to constrain the date of the impact by crater-counting techniques. The actual crater interior itself, despite being of sufficient surface area (∼3100 km²), is also unsuitable due to the presence of widespread post-impact deposits lining its walls and floor (Figs. 2 and 3).

Therefore, an alternative method is to directly derive age estimates for the valley networks themselves, an approach which is difficult due to their small surface area relative to the catchments in which they occur. New techniques for directly dating valley networks have been presented by Fassett and Head (2008), utilizing the areas within buffer zones. This method, however, requires larger-scale valley networks than are present in our study area, in order to provide sufficiently large areas in which to count. Nevertheless, due to the high density of networks to the south and southeast of the main crater which completely cover portions of this area (especially to the southwest, Fig. 9) and because the valleys display a uniform level of degradation, it is reasonable to undertake crater counting over this area to provide a minimum age for these sets of networks.

The longitudinal extent of the crater counting area was restricted to include only the area with the highest drainage density. The edge of the summit of the plateau escarpment served as the boundary to the south as did the rim of Sinton crater to the north. This resulted in a crater counting area to cover a region of the plateau of ∼3201.5 km² (Fig. 16). During the counting process any craters found to be intercepted by valley networks or associated erosional landforms were not counted. The results of the crater counting (Fig. 17) revealed that for craters >0.75 km, the size frequency distribution curve fits the isochrons most closely along the Hesperian/Amazonian boundary (∼3 Gyr using isochrons defined by Hartmann, 2005), thus indicating that the minimum age for the formation of the valleys is within the Late Hesperian.

The crater size-frequency plot also revealed that the distribution curve rolls over at lower crater diameters, causing the data points corresponding to the smallest craters (<0.375 km) to cross the isochrons (Fig. 17a). This is interpreted to be due to the occurrence and continual operation of degradation and resurfacing processes that occurred subsequent to the initial emplacement, causing the preferential obliteration of smaller craters. This could be due to the result of aeolian activity and other such small-scale erosional/depositional activity expected to have affected the martian surface over the last 3 billion years.

To further constrain the age of the valley networks, a second area to the west of Sinton crater was also counted (Fig. 15). Although the drainage density is significantly lower than that of the southern area, ejecta from Sinton and valley networks are present on this portion of the plateau (Fig. 13). Therefore, this part of the plateau should be of a similar age or younger than the region to the south. The crater counting area was defined to within one Sinton crater diameter to the west, corresponding to the extent to which the impact would have deposited the greatest volume of ejecta, and provided a crater counting surface area of ∼2030 km² (Fig. 16). The size-frequency distribution of the counts within this area are consistent with that of the counts to the south, and
Vibrated areas of the fill deposits are located along the upper terraces deposits of LVF/LDA material in the study region (the most ele-
tablish the minimum age of emplacement for the most elevated
(Fig. 17). This is consistent with the age estimates provided by crater counts on the LVF deposit in the valley to the south of the main plateau by Morgan et al. (2009) (Fig. 18b) as well as other counts conducted on other LVF/LDA deposits across the northern dichotomy boundary (e.g., Mangold, 2003; Levy et al., 2007). This suggests that crater fill emplacement within Sinton occurred at the same time that LVF/LDA emplacement was occurring across dichotomy boundary during the Late Amazonian. Indicating that if the climate models are correct, the redistribu-
tion of volatiles from the poles to the lower latitudes during high obliquity resulted in the deposition of snow along the dichotomy boundary, causing widespread ice rich accumulation, which af-
fected all elevations across the boundary.

6. Water sources and mechanisms of fluvial erosion

Valley networks and associated erosional features have been identified on Mars since the initial Mariner missions of the late 1960s early 1970s (Masursky, 1973) and their origin has been a source of debate ever since. These valley networks have been dated as having formed primarily during the Late Noachian (Fassett and Head, 2007). Due to the implied involvement of surface runoff they have been cited as important morphological indices of the martian paleoclimate, suggesting that early warm and wet condi-
tions once dominated the climatic regime of the planet (Masursky, 1973). At times later than the Noachian/Hesperian boundary, the general consensus is that Mars became cold and dry (e.g., Carr, 1996). Large-scale outflow channels attributed to the catastrophic release of ground water have been the most dominant form of flu-
val erosion since this early phase of wetter conditions. However, early studies by Gulick and Baker (1989, 1990) and Gulick (2001) showed the presence and environments of some younger valley networks. Recent research using newer high resolution spacecraft datasets have also discovered younger valley networks across the martian surface; a prominent example being the Hesperian-aged networks documented by Mangold et al. (2004) around the Valles Marineris area. As previously discussed, these are of a comparative scale to those in the Sinton crater, but are eroded into near flat surfaces and are found to exhibit high drainage dens-
ties and highly ordered dendritic tributary systems. Mangold et al. (2004) attributed such morphological characteristics to the work of precipitation-fed runoff, suggesting either a transitional climate from the Noachian or episodic periods of wetter conditions. How-
ever, there are other means by which valley networks can form that do not rely on the prevalence of conditions warmer that the current climate (Mangold et al., 2004). These include: groundwater sapping, water-lubricated debris flows and hydrothermal activity. Therefore, detailed analysis was carried on the Sinton crater val-
ley networks and associated erosional features in order to identify the most suitable source of water and derive a model of formation that is consistent with the geological evidence.

6.1. The case for groundwater

A groundwater source that excludes the necessity for an atmospherically-derived water source is inconsistent with both the topographic setting of the valleys and their morphology. The iso-
lated plateau on which the valleys are situated is in excess of

![Fig. 17. Crater count plots for (a) the valley networks south of Sinton crater, and (b) the western portion of the plateau within one crater diameter of Sinton crater. The isochrons are plotted according to the Hartmann (2005) convention. (a) Craters were counted on the portion of the plateau that was most densely covered in val-
ley networks, thus the age derived from the crater distributions is considered to be a minimum for the valley forming period. The plots best fit the isochrons close to Hesperian/Amazonian boundary. (b) The age correlates with that of (a) suggesting that valley networks to the west of Sinton crater also have a minimum age close to the Hesperian/Amazonian boundary. Both plots show that the crater size distribu-
tions cross the isochrons at smaller crater diameters (<500 m) indicating that some process(es) operated to remove the smaller craters.

also provided a best fit along the Hesperian/Amazonian isochron (Fig. 17).

The internal deposits within Sinton crater were dated to est-
ablish the minimum age of emplacement for the most elevated deposits of LVF/LDA material in the study region (the most ele-
vated areas of the fill deposits are located along the upper terraces

of the crater and are ~545 m above the highest sections of the valley LVF that surround the plateau). The geological map (Fig. 3) shows the extent of the count area as the crater fill was treated as a single unit (including the ‘pitted’ fill) and formed a count-
ing area of ~1113 km. Only craters smaller than 250 m were
found to be present on the surface of the crater fill. Neverthe-
less the crater size-frequency distribution plot shows a best fit to the right of the 100 Myr isochron consistent with an age of >100 Ma–500 Ma (Fig. 18a). This is consistent with the age esti-
mates provided by crater counts on the LVF deposit in the valley to the south of the main plateau by Morgan et al. (2009) (Fig. 18b)
as well as other counts conducted on other LVF/LDA deposits across the northern dichotomy boundary (e.g., Mangold, 2003; Levy et al., 2007). This suggests that crater fill emplacement within Sinton occurred at the same time that LVF/LDA emplacement was occurring across dichotomy boundary during the Late Amazonian. Indicating that if the climate models are correct, the redistribu-
tion of volatiles from the poles to the lower latitudes during high obliquity resulted in the deposition of snow along the dichotomy boundary, causing widespread ice rich accumulation, which af-
fected all elevations across the boundary.

6. Water sources and mechanisms of fluvial erosion

Valley networks and associated erosional features have been identified on Mars since the initial Mariner missions of the late 1960s early 1970s (Masursky, 1973) and their origin has been a source of debate ever since. These valley networks have been dated as having formed primarily during the Late Noachian (Fassett and Head, 2007). Due to the implied involvement of surface runoff they have been cited as important morphological indices of the martian paleoclimate, suggesting that early warm and wet condi-
tions once dominated the climatic regime of the planet (Masursky, 1973). At times later than the Noachian/Hesperian boundary, the general consensus is that Mars became cold and dry (e.g., Carr, 1996). Large-scale outflow channels attributed to the catastrophic release of ground water have been the most dominant form of flu-
val erosion since this early phase of wetter conditions. However, early studies by Gulick and Baker (1989, 1990) and Gulick (2001) showed the presence and environments of some younger valley networks. Recent research using newer high resolution spacecraft datasets have also discovered younger valley networks across the martian surface; a prominent example being the Hesperian-aged networks documented by Mangold et al. (2004) around the Valles Marineris area. As previously discussed, these are of a comparative scale to those in the Sinton crater, but are eroded into near flat surfaces and are found to exhibit high drainage dens-
ties and highly ordered dendritic tributary systems. Mangold et al. (2004) attributed such morphological characteristics to the work of precipitation-fed runoff, suggesting either a transitional climate from the Noachian or episodic periods of wetter conditions. How-
ever, there are other means by which valley networks can form that do not rely on the prevalence of conditions warmer that the current climate (Mangold et al., 2004). These include: groundwater sapping, water-lubricated debris flows and hydrothermal activity. Therefore, detailed analysis was carried on the Sinton crater val-
ley networks and associated erosional features in order to identify the most suitable source of water and derive a model of formation that is consistent with the geological evidence.

6.1. The case for groundwater

A groundwater source that excludes the necessity for an atmospherically-derived water source is inconsistent with both the topographic setting of the valleys and their morphology. The iso-
lated plateau on which the valleys are situated is in excess of
one kilometer above the surrounding terrain, making it unlikely that its surface could be internally connected to any large-scale regional groundwater sources that have been proposed to exist beneath the cryosphere (Clifford, 1993; Clifford and Parker, 2001). Perched aquifers have been suggested to exist on Mars to explain the occurrence of ∼1 km long gullies (Malin and Edgett, 2000), although in the absence of a significant means of recharge, such localized aquifers within the plateau would be insufficient to supply the >50 km long Sinton valleys. Moreover, the occurrence of valley heads at the crest of steep ridges running along the plateau (Fig. 10) is also inconsistent with an internal water source eroding the valleys. The lack of theater-shaped heads, which would be expected if sapping had occurred (Laity and Malin, 1985) or the absence of chaotic terrain indicative of the rapid release of aquifers (e.g., Baker and Milton, 1974) further argues against an internal water source.

6.2. The case for rainfall

The concentration of valley networks on the main plateau, and their spatial association with Sinton crater itself (Fig. 2), is not typical of widespread pluvial activity. Estimates of discharge are two orders of magnitude higher for the Sinton valley networks (Table 1) compared to similarly scaled Noachian valleys considered to be formed by prolonged fluvial activity (Fassett and Head, 2005). The study of precipitation-fed terrestrial valley networks is based on the work of Horton (1954) who derived a system of quantification based on morphometric parameters, which has since been applied to martian examples (e.g., Mangold et al., 2004). Fluvial geometry has been found empirically to be dependent on a range of factors but is critically related to slope. On Earth, dendritic patterns are only observed on relatively flat surfaces, below angles of ∼1°. Above this value, valleys become increasingly more subparallel. As a result of the lower surface gravity on Mars such a value would be different than its terrestrial counterpart. However, in the absence of experimental data or observations of active valley networks, it is not possible to derive a satisfactory value with confidence. Therefore, at values close to critical slope angle one has to be cautious when attempting to make direct morphometric comparisons (i.e., such as bifurcation ratios) with terrestrial precipitation-fed valley networks in terms of the origin of the valley geometry. However, there are other aspects of the morphology, which can be utilized to assess a precipitation origin. Regardless of the angle of slope, precipitation-fed valleys will tend to converge to larger, higher order valleys in the downslope direction. The occurrence of anastomosing drainage patterns and broad, flat valleys in our study region is not consistent with this prediction.

6.3. The case for the melting of snow/ice deposits

Atmospherically derived snow and ice deposits are an alternative source of water, providing that the deposits could be heated sufficiently to generate the meltwater required for erosion. Such a scenario is also consistent with the occurrence of ice-rich remnant deposits that have been found at a range of elevations both below and above the average plateau surface (Head et al., 2006b; Morgan et al., 2009; Figs. 2 and 3) and would also explain the immature drainage configurations of the valleys (Fig. 15). Volcanically heated snow and ice deposits have been interpreted to release sufficient water volumes to generate discharges comparable to 10^2–10^3 m^3 s^−1 estimated for Sinton (Fassett and Head, 2006, 2007). The thermal anomaly associated with the impact event that formed Sinton crater would have provided an alternative but substantial energy source, capable of melting surface ice deposits, if they had existed on the plateau prior to the impact. This would also explain the close spatial association between the valleys and Sinton crater ejecta. Hence, interaction between the energy of impact and residual ice deposits presents a mechanism for inducing surface runoff across the plateau surface. Fluidized ejecta around craters on Mars (Fig. 5b) have also been interpreted to result from impact/ice interactions, through the remobilization of ice within the regolith (e.g., Squyres, 1989). These features, however, do not show evidence for fluvial erosion. Hence, we argue that surficial ice deposits were required to generate the Sinton valleys.
A period of Late Amazonian glaciation is well established through the extensive research that has been carried out on LVF/LDA deposits across the northern dichotomy (e.g., Head et al., 2006a, 2006b). At least two periods of glaciation appear to have occurred through the redistribution of volatiles from the poles to lower latitudes during high obliquity (Madeleine et al., 2007). Mars is likely to have experienced large obliquity variations throughout its history (Laskar et al., 2004). Therefore, it is conceivable that ice deposition also occurred during earlier periods, including the Hesperian.

Hydrothermal systems can be established whenever conductive heat transport is initiated by the coexistence of a fluid phase (usually water) with a heat source (Farmer, 1996). Impact cratering processes provide significant heat sources that can support hydrothermal systems on the Earth and Mars (Newsom, 1980). Detailed studies of the 23 km Haughton impact structure in Devon Island, Arctic Canada (a well established and utilized Mars analog) have documented evidence for post-impact hydrothermal alteration in concentric fault systems that surround the crater rim (Osiniski et al., 2005a, 2005b). Mineralogical analysis of the alteration indicates that temperatures reached at least 200 °C, and that the fault systems provided pathways for hot fluids, potentially for several kilometers away from the heat source. Brackenridge et al. (1985) argued that analogous hydrothermal systems may have been widespread early in Mars history, during the period of heavy bombardment, and may have been responsible for Noachian valley networks, instead of a warmer, wetter climate period. Brackenridge et al. (1985) further hypothesized that ground ice was exploited by impact hydrothermal systems, producing valley networks through headward sapping and down valley fluid flow. This model has been challenged recently by the dating of Noachian valley networks; the age of most of the valley networks is Late Noachian/Early Hesperian, ~0.3 Gyr after the late heavy bombardment (Fassett and Head, 2008). This separation in time is too great to maintain active hydrothermal systems, even if one takes into consideration the several million-year estimates for the lifetime of hydrothermal systems within large scale terrestrial impact basins (e.g., the Chicxulub crater, Mexico; Abramov and Kring, 2006).

In light of the evidence for: (1) high discharge rates, (2) immature anastomosing drainage configurations, and (3) extensive fluvial features in close proximity to a large crater in a region interpreted to have experienced significant glacial activity, we propose that the valley networks originated from the release of water due to the deposition of hot ejecta over snow/ice deposits present on the plateau during the impact event. Although the melting of near-surface ground ice (which presently is estimated to begin at a depth of >1 m deep at Sinton’s latitude; Jakosky and Haberle, 1992) may also have contributed water to the fluvial activity, the large volumes required to produce the $10^{4}$−$10^{5}$ m$^{3}$ s$^{-1}$ estimated discharges and the lack of similar fluvial features around other craters of comparative size, argues that ground ice alone is insufficient. The release of water from ground ice due to persistent hydrothermal systems could potentially have supplied the valley networks for an extended period after the initial melt of ice. However, if hydrothermal activity similar to that outlined in Brackenridge et al. (1985) did persist after the melting of the initial external ice deposits, it was not sufficient to enable the valleys to develop into mature networks (Fig. 15).

7. Model

On the basis of the analysis of the geology, geomorphology, age relations and topography of the Sinton crater area, we interpret the valley networks to have formed by the transfer of heat associated with the ejecta of the Sinton impact to pre-existing surficial snow and ice deposits on the plateau. We hypothesize that enough impact kinetic energy was transferred to the ejecta in the form of heat that the burial of snow and ice in hot ejecta was sufficient to produce melting and drainage, creating the valleys. Here we present a conceptual model that describes the main stage of valley formation developed from our observations and models of the cratering process (Fig. 19).

7.1. Pre-impact

Prior to the impact, we hypothesize that the dichotomy boundary was experiencing an active ‘glacial period’ in which ice-rich deposits were present across the plateau (Fig. 19a). Such a scenario is supported by the current occurrence of ice-rich and ice-related deposits throughout the study region in the form of: (1) LVF/LDA deposits within the fretted valleys and plains (Fig. 2), (2) ice-related crater fill within Sinton (Fig. 3), and (3) a mantling deposit on the plateau surface. The occurrence of mantling deposits have been interpreted to be partly responsible for the paucity of small craters within the crater count plots (Fig. 17). These relatively recent ice-rich deposits formed during the late Amazonian in conjunction with periods of relatively higher obliquity (Head et al., 2006a, 2006b; Mustard et al., 2001). The accumulation of significant quantities of snow and ice at northern mid-latitudes during certain portions of the obliquity history of Mars has been modeled by Madeleine et al. (2007). They find that at obliquities of 35° and moderate atmospheric dust opacities, widespread precipitation of snow occurs over the dichotomy boundary and persists throughout the year, ultimately producing plateau and valley glacial deposits (Madeleine et al., 2007). Modeling of the obliquity history of Mars shows that these obliquity conditions are not unusual and that this condition is likely to have recurred frequently in the past history of Mars (Laskar et al., 2004). Furthermore, recent evidence suggests that the current LDA/LVF deposits represent only a remnant of a much larger and more extensive plateau glaciation that occurred in the late Amazonian (e.g., Dickson et al., 2008a). Thus, there is a high likelihood that extensive ice and snow deposits occurred on the Sinton plateau at different times in the past history of Mars.

7.2. Impact

As the projectile made contact with the surface, energy was transferred into the plateau target material as a combination of shock waves and associated rarefaction waves, causing the formation of a transient cavity and emplacement of ballistic ejecta (Fig. 18b). Estimates of the energy released by the impact can be made by calculating the kinetic energy of a typical impacting projectile by estimating its mass (m) and velocity (v):

$$KE = \frac{1}{2}mv^2.$$  

Assuming a mass based on chondritic densities, a diameter approximately one tenth of the crater diameter, and an average velocity of Mars-crossing-asteroids ($\sim$10 m s$^{-1}$) (Ivanov, 2001), the impact would have provided substantial energy, of the order of $10^{18}$ J (Table 2). This value would be an order of magnitude higher if Sinton was the result of a higher velocity, long-period comet impact (40 km s$^{-1}$; Steel, 1998).

During the excavation stage (Fig. 19c) ice deposits would have been blanketed in hot ejecta, which would have contributed to melting the ice, causing water to accumulate at the ejecta-ice interface. The proportion of the initial kinetic energy that is converted to thermal energy is poorly constrained for large impacts. One approach for testing whether the energy from Sinton would have been sufficient to generate the valley networks is to calculate...
Fig. 19. Model of impact-induced valley formation at Sinton crater depicting a series of idealized west-east cross sections through the main plateau throughout the formation process. Note that the diagrams are vertically exaggerated. (a) Ice is deposited during a period of high obliquity, forming ice-rich deposits across the plateau. (b) A projectile comes into contact with the surface, causing the excavation of a transient cavity. The ice in contact with the impact along with the projectile and a proportion of the plateau surface are vaporized. (c) During the excavation stage ejecta and impact melt leave the crater. The ice deposits are blanketed in hot ejecta which melts the ice, causing the release of water out onto the recently emplaced ejecta. This modifies and erodes the ejecta deposit compared to other relatively similar size craters (Fig. 5). Linear gouges cut into the surface of the plateau by the emplacement of material during the excavation of the crater (secondary crater chains) would have provided natural conduits through which meltwater could flow. This is consistent with channels to the southeast of the crater exhibiting an orientation perpendicular to the crater rim for almost their entire length, apart from short sharp deflections (Fig. 6). (d) In the period after the impact the meltwater would have drained off the plateau carving the valley networks and leaving behind the remnants of the ejected material. Some late stage water may have reached the surface through impact-related hydrothermal systems (e.g., Osinski et al., 2005a, 2005b). Such a process could have maintained activity within some of the valley networks. Modified after Melosh (1989) and Osinski et al. (2005b).
the proportion of the original kinetic energy required to generate the volume of meltwater necessary to carve the channels. The mass of ice melted by a given amount of energy is:

\[
M_{\text{Ice}} = \frac{E}{L}
\]  

\(M_{\text{Ice}}\) is the energy available to the melting of ice, and \(L\) is its latent heat of fusion (for which we use 335 kJ kg\(^{-1}\)). Equation (3) can be used to calculate the mass of ice (and from which the water volume can be calculated) that can be produced for a range of thermal energies supplied to the plateau as a proportion of the original projectile kinetic energy (Table 2). Table 2 demonstrates that only a small proportion of the initial projectile energy is required to generate substantial amounts of melt. Only 5% of the initial energy \((\sim 10^{17}\text{kJ})\) would need to be utilized by the melting of ice to provide a value comparable with the estimates of the volcanic energy available for snowmelt production and Hesperian valley formation attained by Fassett and Head (2006) for the Hecates Tholus volcano. If we only consider the area of the plateau with valley networks (assuming that any ice in the region where the transient crater formed would have been vaporized; this gives an area of \(5 \times 10^3\text{m}^2\)), 5–10% of the initial energy would be sufficient to melt an ice deposit 500 m to 1 km thick (Table 2). This is comparable to the current thickness of dichotomy boundary LDA/LVF deposits (Li et al., 2005).

The sporadic release of water associated with the melting would account for the formation of multiple small anastomosing channels that have no apparent source (Figs. 6–9, 15A). As the flow developed through the continual release of meltwater, runoff would have been concentrated in the most efficient distribution systems as the other channels become clogged with sediment derived from the erosion of the ejecta. This could concentrate erosion and lead to the development of the observed broad-flat channels (Fig. 15B). The broad depressions that connect the valleys (Fig. 10) may represent where the melting of underling ice was concentrated and caused the collapse and further erosion of overlying ejecta material.

During the outflow of water, portions of the ejected material would be transported downslope by the meltwater and preferentially deposited across the plateau surface and into the surrounding valleys. This could help to explain the highly modified ejecta deposit surrounding Sinton crater compared to the more distinctive ejecta deposits surrounding other similar-sized craters (Fig. 5). During the valley formation, water flow may have been controlled by the micro-topography of the original ejecta patterns (such as the radial fabric patterns produced by ballistic secondary impacts). This is consistent with the larger channels to the southeast of the crater exhibiting an orientation perpendicular to the crater rim for almost their entire length, apart from short, sharp local deflections (Fig. 6).

The lack of more mature drainage configurations and broader valley networks suggests that the erosional process was not prolonged and that runoff ceased after the ejecta had cooled or the ice/snow source of meltwater had been depleted. Nevertheless some hot springs may have developed at the location of the extensonal faults. These can act as pathways for hot fluids and steam generated at hydrothermal systems associated with the center of the thermal anomaly under the floor of the crater (e.g., Osinski et al., 2005b).

Based on the discharge estimates calculated in Section 4.2, the volume of ice that could be melted by the impact in which 5–10% of the initial energy was partitioned for this purpose, would have supplied the 15 largest valley networks for \(\sim 80–150\) days. The values of discharge are for peak bankfull conditions and so the duration of flow would likely have been significantly longer, by as much as ten times the value predicted above if we consider the uncertainties surrounding the cross sectional dimensions and surface roughness. Estimates from large scale terrestrial impacts (such as the one that formed the Chicxulub crater) indicate that as much as 50% of the projectile original kinetic energy was converted into heat (e.g., Ryder et al., 1996). Therefore, although there are uncertainties in both the discharge and energy estimates, the calculation does at least demonstrate the physical plausibility of the model and suggests that the fluvial activity was likely to have been short-lived relative to the mature Noachian valleys.

The uneven distribution of the valley networks could represent a combination of the location of ice on the plateau surface prior to the impact, the dispersal of ejecta (related to the azimuth and zenith angles of the projectile trajectory) and the underlying plateau topography. In our model we argue that the internal valley networks formed after the formation of external valley networks and thus are related to different times and processes. Their morphologic similarities with smaller scale gully systems may suggest that they have formed due by similar processes (e.g., Christensen, 2003; Dickson et al., 2007).

8. Summary and conclusions

Previous analysis of the dichotomy boundary region have shown an array of evidence for significant glacial activity occurring in the form of LDA and LVF throughout the region, and specifically in the area surrounding Sinton impact crater (Morgan et al., 2009). Large scale integrated LVF and LDA systems surrounding the main plateau have been identified and mapped and found to originate from within multiple alcoves that are cut into the plateau flanks (Head et al., 2006b; Morgan et al., 2009). Ice related deposits are also prevalent across the surface of the plateau within topographic lows, such as impact craters. Sinton crater itself is host to a large-scale unit of crater fill, exhibiting a range of ice and water related flow features (including kilometre-scale lobes) that have been identified within other similarly sized craters in the mid-latitudes. Our data therefore suggests that significant amounts of ice have been deposited within the study area on the surface of the plateau in the past (see Fig. 3).

The spatial relationship between the valleys and the main crater, and the apparent control its ejecta has imposed on the orientation of valley formation, suggest that the two are related. The thermal anomaly associated with the impact provides a means of melting ice deposits that were present on the plateau at the time of impact. Hence an ice-ejecta interaction forms the basis of our valley network formation model; the release of water is initiated by the melting of ice from the deposition of hot ejecta deposits over its surface. Such a mechanism provides a means of generating intermediate scale fluvial features (i.e. hundreds of meters wide by tens of kilometers long) in the absence of a climatic regime favorable for fluvial activity.

Previous work has suggested that there was a major period of climate change in the Hesperian, transitioning from a potentially “warm and wet” Noachian, to a “cold and dry” Amazonian (e.g., summarized in Carr, 1996). Our observations of Sinton crater pro-

Table 2
Energy balance calculations that give the volume and duration of flow for a given percentage of initial projectile kinetic energy supplied to the melting of ice.

<table>
<thead>
<tr>
<th>Percentage of initial projectile kinetic energy</th>
<th>Ice thickness that could be melted (m)</th>
<th>Volume of water (km³)</th>
<th>Duration of flow within the valleys (days)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>109</td>
<td>500</td>
<td>16</td>
</tr>
<tr>
<td>3</td>
<td>328</td>
<td>1500</td>
<td>47</td>
</tr>
<tr>
<td>5</td>
<td>547</td>
<td>2500</td>
<td>78</td>
</tr>
<tr>
<td>10</td>
<td>1090</td>
<td>5000</td>
<td>157</td>
</tr>
<tr>
<td>20</td>
<td>2188</td>
<td>10,000</td>
<td>314</td>
</tr>
</tbody>
</table>

\[M_{\text{Ice}} = \frac{E}{L}\]
viding evidence that a significant cover of snow and ice existed on the dichotomy boundary plateau during at least part of the Hesperian (see Fig. 20). Our results also demonstrated how the interaction of ice with localized heat sources can generate intermediate-scaled fluvial landforms and thus is consistent with the interpretation that other such thermal anomalies (e.g., such as those associated with volcanic edifices; Fassett and Head, 2007) are capable of generating runoff through the melting of atmospherically deposited snow/ice. Glacial activity could therefore have accounted for one of the dominant erosional processes operating along the dichotomy boundary since its formation.

Future work will concentrate on answering two outstanding questions: (1) What was the volume and spatial extent of ice during a ‘glacial’ period? (2) How many glacial cycles were there and how long did the activity last? One approach that could be envisaged was: (1) Is there evidence for a glacial period? (2) How many glacial cycles were there and for how long did the activity last? One approach that could be envisaged for one of the dominant erosional processes operating along the dichotomy boundary plateau during at least part of the Hesperian (as presented by Irwin et al., 2004): 1. Dichotomy boundary formation (Frey et al., 2002). 2. Erosion and burial of the dichotomy boundary (Neukum and Hiller, 1981; McGill, 2000). 3. Formation of large-scale valley networks on Mars (dated by Fassett and Head, 2007). 4 and 5. Possible multiple episodes of ice deposition. 6. Pre-impact glacial period. 7. Impact event, forming Sinton crater. 8. Rapid melting of ice deposits and subsequent valley formation. 9. Potential hydrothermal activity around Sinton crater. 10. Possible multiple episodes of ice deposition. 11. Most recent phase of glaciation associated with main trunk of LVF (Morgan et al., 2009). 12. Latest phase of glaciation associated with superimposed lobes (Morgan et al., 2009).

Acknowledgments

We gratefully acknowledge financial support from the NASA Mars Data Analysis program (NNX07AN95G), the NASA US participation in the Mars Express High-Resolution Stereo Camera (HRSC) (JPL1237163), and the NASA Applied Information Systems Research Program (NN05CA61G). Thanks are also extended to Caleb Fassett, James Dickson and Wes Patterson for their technical assistance in data analysis and contribution to scientific discussions during preparation of the manuscript.

References


Valley network formation from ice/impact interactions

59