Glaciovolcanism in the Tharsis volcanic province of Mars: Implications for regional geology and hydrology

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ABSTRACT

The Tharsis region of Mars is a vast volcanic plateau which hosts the immense Tharsis Montes shield volcanoes. The Tharsis region is suggested by multiple lines of morphologic and modeling evidence to have been a site of coincident volcanic and glacial activity throughout the majority of the geologic history of Mars. The prolonged and overlapping histories of volcanism and glaciation within the Tharsis region raise the likelihood of widespread surficial glaciovolcanism, a possibility which is supported by the recognition of glaciovolcanic landforms in the region by past investigations. Given this likelihood, we perform an exploratory study to assess the potential role that surficial glaciovolcanism may have played in the geologic and hydrologic history of the Tharsis region. We first review the history and characteristics of volcanism and glaciation in the Tharsis region, as well as previously documented evidence for past glaciovolcanic activity, in order to outline relevant conditions and parameters. The outlined volcanic and glacial conditions are then used in conjunction with results and predictions from past modeling of surficial glaciovolcanic processes to assess the potential role of glaciovolcanism in the formation of the Tharsis region’s major tectonic and hydrologic features. We conclude that surficial glaciovolcanism may plausibly have contributed to the formation of many of the tectonic and fluvial features in the Tharsis region, offering advantages over prior formation models, particularly for the large basin/chaos-sourced outflow channels concentrated in the area. The formation of a range of investigated features in the Tharsis region by surficial glaciovolcanism does not require ambient warm and wet climate conditions therefore suggesting potential consistency between the observed features and a predominantly cold and icy climate. Consequently, this analysis represents an incremental contribution to better understanding the climatic and hydrologic evolution of Mars. However, analyses performed in this work indicate that the glaciovolcanic origin models we considered are unable to viably account for the complete range of explored features, indicating that future work is required to better resolve the processes involved in their formation.

1. Introduction

The Tharsis rise of Mars (Fig. 1a) is the most prominent volcanic province on the planet and arguably within the solar system. The Tharsis region also represents the most active center of volcanism on Mars, with a history of volcanic activity spanning from the Noachian to the very recent Late Amazonian period (discussed in detail in the following section). Rising to a mean elevation of ~2.3 km (based on our de

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Following the loss of the thicker atmosphere (Carr and Head, 2010; Lammer et al., 2013; Jakosky et al., 2017) and thus adiabatic cooling conditions, abundant evidence (later outlined in detail) suggests regional accumulation of snow and ice continued throughout the Amazonian period within the Tharsis rise region. Therefore, the Tharsis region is a site where coincident volcanic and glacial activity are predicted to have occurred throughout the majority of the history of Mars, raising the possibility for widespread and sustained glaciovolcanism (a comprehensive term referring generally to any processes of glacial and volcanic interaction; Smellie and Edwards, 2016).

Here we perform an exploratory study to assess the potential role of glaciovolcanism (Allen, 1979; Squyres et al., 1987; Smellie and Chapman, 2002; Head and Wilson, 2007; Wilson and Head, 2007a,b, 2009; Edwards et al., 2013, 2014; Wilson et al., 2013; Scanlon et al., 2014; Smellie and Edwards, 2016) in the geologic and hydrologic history of the Tharsis region. To perform this assessment, we first review the history and characteristics of volcanism and glaciation in the Tharsis region, as well as previously documented evidence for past glaciovolcanic activity, in order to constrain relevant conditions and parameters. We then evaluate the potential role of glaciovolcanism in the formation of major tectonic and hydrologic features within the region including: volcanic aureole apron units, the Solis-Planum mega-thrust, valley networks, and outflow channels (Fig. 1). Analysis of the role of glaciovolcanism in the formation of these features is carried out by first outlining formation criteria through reviews of the geomorphology of each feature. The outlined formation criteria of each feature are then evaluated for consistency against predictions from possible genetic surficial glaciovolcanic models informed by previous modeling work (Cassanelli and Head, 2016, 2018a). Ultimately, the objective of this analysis is to test the consistency of a cold and icy climate with an array of geologic and hydrologic features contained within the Tharsis region as an incremental step toward understanding the climatic and hydrologic evolution of Mars. Given this particular objective, many of the analyses in this manuscript are performed by assuming cold and icy conditions for the adoption of relevant model predictions and data (e.g., cold and icy climate model scenarios; Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016; Palumbo et al., 2018). It is important to note that these model-dependent predictions and data are, in many cases, subject to inherent assumptions and limitations which hold consequent implications for the results presented in this work. While a complete discussion of these assumptions and limitations is outside the scope of this study, important implications are noted in the manuscript where relevant and references are provided to sources containing more detailed descriptions for each instance of utilized model-dependent predictions and data. All data utilized in this study are derived from published sources or from publically available data repositories and are referenced accordingly throughout the manuscript.

2. Tharsis region volcanism

The nature of surficial glaciovolcanic activity is controlled in large part by the rate and volume of lava emplacement (e.g., Squyres et al., 1987; Wilson and Head, 2002a; Wilson and Head, 2007a,b; Smellie and Chapman, 2002; Head and Wilson, 2007; Wilson et al., 2013; Edwards et al., 2014; Cassanelli and Head, 2016, 2018a; Smellie and Edwards, 2016), and thus by the prevailing style of volcanism. Therefore, in order to evaluate the nature of possible surficial glaciovolcanic processes in the Tharsis region, we first outline the styles and characteristics of volcanism which have occurred throughout the history of the area.

The Tharsis rise is a vast volcanic plateau covering a total area of \( \sim 3 \times 10^7 \text{ km}^2 \) (nearly 25% of the surface of Mars) (Phillips et al., 2001; Tanaka et al., 2014), comprised of an estimated \( \sim 3 \times 10^8 \text{ km}^3 \) of igneous material (Phillips et al., 2001), and containing 13 large, distinct volcanic edifices (Tanaka et al., 2014) (Fig. 1a). Here we define the extent of the Tharsis region through the aggregation of mapped units (Tanaka et al., 2014) on the basis of volcanic origin, unit elevation, and unit surface morphology. As a result, we therefore include Olympus Mons, its surrounding apron unit, and Lunae Planum in our definition of the Tharsis region, yielding a total defined area of \( \sim 2.8 \times 10^7 \text{ km}^2 \) (which lies within the range of extents previously defined for the Tharsis region). While there are no firm constraints on the inception of the accumulation of the Tharsis rise, the feature has often been interpreted to have been largely in-place by the end of the Noachian period \( \sim 3.7 \text{ Ga} \) (Phillips et al., 2001; Carr and Head, 2010) (although alternative interpretations have suggested that the bulk of Tharsis formation may have occurred later; Bouley et al., 2016; Citron et al., 2018; Bouley et al., 2018). Following the Noachian, volcanic activity in the Tharsis region continued at a reduced rate throughout the Hesperian and into the Amazonian
periods (Tanaka et al., 1988, 2014; Werner, 2009), to as recently as the last several 1–10s of Ma (Hartmann et al., 1999; Neukum et al., 2004; Hauber et al., 2011; Mangold et al., 2010; Richardson et al., 2017). Thus, volcanic activity within the Tharsis region has been sustained throughout most of the geologic history of Mars.

The Tharsis rise is interpreted to have formed through hot-spot-like volcanism above a mantle upwelling, resulting from either predominantly crustal uplift (Carr, 1974; Greeley and Spudis, 1981; Phillips et al., 1990) or volcanic construction by the accumulation of lavas (Solomon and Head, 1982). The uplift origin of the Tharsis rise is supported (Carr, 1974; Plescia and Saunders, 1980, 1982; Phillips et al., 1990) by the broad domical shape of the feature (Fig. 1a), the presence of ancient cratered terrains at high elevations within the region, and extensional fractures in the region exhibiting generally radial trends. In an uplift origin scenario (Carr, 1974; Plescia and Saunders, 1980), the majority of the Tharsis topography is attributed to lithospheric uplift, with the surface later veneered by deposition of relatively thin lavas through plains volcanism, accompanied by the construction of the massive shield volcanism (Fig. 2a). However, the uplift origin is challenged by disagreement between the orientation of tectonic features in the region and stresses predicted for lithospheric uplift of Tharsis, and by unreasonable thermal and chemical anomaly requirements for mantle support (Solomon and Head, 1982). In a constructional origin scenario (Solomon and Head, 1982), the Tharsis topography is attributed chiefly to the accumulation of volcanic deposits, with a substantial additional component of isostatically-compensated deposits residing at depth.

Here, we are interested in assessing the operation of glaciovolcanic processes in the Tharsis region, and the contributions they may have had toward the regional geology. Therefore, we adopt the constructional origin interpretation of Tharsis rise (Solomon and Head, 1982), a scenario which would involve a greater extent of extrusive volcanic activity and thus surficial glaciovolcanic activity. This particular interpretation has been adopted because it provides a useful end-member scenario to test the potential involvement of surficial glaciovolcanism in the formation of geologic features in the Tharsis region (in accordance with our study objectives). We note that, if instead Tharsis originated primarily by uplift, the amount of glaciovolcanic activity that would have taken place in the region would be consequently reduced (with the possibility of increased subsurface glaciovolcanism generated by increased intrusive magmatism; Phillips et al., 1990).

The Tharsis region as we have defined (Fig. 1a) encompasses a range of geologic units (Tanaka et al., 2014) emplaced in a range of volcanic styles throughout the history of the Tharsis region including shield, plains, and flood basalt volcanism (Solomon and Head, 1982) (Fig. 2a). The volcanic edifices within the region and the main constituents of the Tharsis bulge itself (the Tharsis Montes and the adjacent plateau plains including Daedalia, Sinai, Solis, Syria, and Thaumasia Planum) are interpreted to have been formed through a combination of shield and plains volcanism (Fig. 2a). Volcanic units emplaced in a flood basalt mode, as inferred by their association with the Hesperian ridged plains volcanic units (Head et al., 2002) evidenced by their surface ages and shared wrinkle-ridge morphology (Solomon and Head, 1982), are contained predominantly in the eastern portions of the Tharsis region (Fig. 2a) (exemplified by Lunae Planum). Here we define the extent of surface units in the Tharsis region emplaced by flood basalt style volcanism through the aggregation of mapped (Tanaka et al., 2014) volcanic units in the region which exhibit wrinkle-ridge surface morphology (Fig. 2b). Explosive volcanism is an additional style of volcanic activity expected to have taken place within the Tharsis region during its course of development due to the prevalence of low atmospheric pressures (e.g. Mougins-Mark et al., 1982; Edgett, 1997; Head and Wilson, 1998; Wilson et al., 1998; Hynek et al., 2003; Wilson and Head, 2007a, b; Wilson and Head, 2009; Kerber et al., 2012). Past analyses (Wilson and Head, 2009) indicate that the superplutonic deposition of both heated and cooled volcanic tephra generated by explosive eruptions acts almost exclusively to protect the ice against ablation. Therefore, we have not directly considered this process in our analysis of surficial glaciovolcanic interactions, though we note, where relevant, the potential important implications of ice preservation by tephra burial.

The history of the emplacement of these various volcanic units within the Tharsis region (Solomon and Head, 1982) likely began in the Noachian period with intense, extensive shield and plains volcanism constructing the bulk of the rise. Following the early phase of Tharsis construction in the Noachian period, shield and plains volcanism in the region continued at a diminished rate throughout the Hesperian with the additional emplacement of the extensive flood basalt volcanic plains in the eastern portions of the region (Fig. 2a). Volcanic activity in the Tharsis region then continued throughout the Amazonian period of Mars history (spanning the last >3 Gyr) through further plains volcanism and development of the large shield constructs of the region. This emplacement history, and the distribution and timing of volcanism within the Tharsis region, is illustrated by the mapped (Tanaka et al., 2014) current surface ages of the region’s volcanic units (Fig. 2b) as determined through stratigraphic and crater age dating. The mapped surface ages (Fig. 2b) indicate that the majority of the Tharsis region is covered by plains volcanic units dating from the Hesperian to Amazonian (Fig. 2b) period. The oldest volcanic units exposed at the present surface reside...
predominantly in the eastern and southeastern portions of the Tharsis rise and are associated with volcanic and flood basalts plains. The most recent volcanic activity in the Tharsis region is associated with the large shield volcanic constructs of the area which have been resurfaced by volcanic deposits into the Late Amazonian (Fig. 2b) and which have remained centers of continued volcanic activity to as recently as the last 1–10s of Ma (Hartmann et al., 1999; Neukum et al., 2004; Hauber et al., 2011; Mangold et al., 2010; Richardson et al., 2017).

Critical parameters required for our analysis of surficial glacio-volcanic processes include the total thickness and area of lava accumula-
tion as well as the thicknesses and areas of individual lava flows. Under the construc
tional Tharsis origin interpretation we have adopted, the total ac
cumulated thicknesses of lava resulting from the operations of
shield and plains volcanism in the region are on the order of ~1–10 km (i.e. the approximate range of total Tharsis topography) (Solomon and Head, 1982). The total thickness of the flood basalt volcanic plains located on the eastern side of the Tharsis region (Fig. 2a) are estimated at ~1–3.5 km (Plescia and Saunders, 1980; De Hon, 1982; Watters, 1991) and cover a total area of ~4.5 × 10⁶ km² (Fig. 2a). Measurements and estimates of the thicknesses and areas of individual lava flows produced by shield volcanism in the Tharsis region yield typical values on the order of ~50 m and ~500 km², ranging from ~1 to 220 m and ~2–25,000 km² (Baloga et al., 2003; Warner and Gregg, 2003; Basilevsky et al., 2005; Hiesinger et al., 2007; Mouginis-Mark and Rowland, 2008; Richardson et al., 2017). Typical thicknesses and areas of individual lava flows generated by plains volcanism in the Tharsis region, and in other similar deposits on Mars (chiefly Elysium Planitia which contains some of the youngest and best preserved volcanic plains deposits on Mars; Plescia, 2003; Keszthelyi et al., 2004; Vaucher et al., 2009a), average to ~20 m and ~10,000 km², and range from ~1 to 200 m (Scott and Tanaka, 1980; Simon et al., 2014) and ~1000–2.5 × 10⁶ km² (Scott and Tanaka, 1980; Plescia, 1990, 2003; Sakimoto and Gregg, 2001; Berman and Hartmann, 2002; Fuller and Head, 2002; Keszthelyi et al., 2000, 2004; Werner et al., 2003; Lanagan, 2004; Vaucher et al., 2009b, 2009a; Jaeger et al., 2010; Simon et al., 2014).

Summary: In summary, the volcanic history of the Tharsis region is characterized by widespread and intense shield and plains volcanism in the Noachian period during which time the bulk of the rise was constructed. The intensity of volcanic activity in the area generally declined into the Hesperian with the exception of the extensive flood basalt volcanic plains which were emplaced in the eastern portions of the region during this period. Volcanic activity within the Tharsis region continued throughout the Amazonian period at a diminished rate and became increasingly focused toward the development of the major shield constructs of the region. The characteristics of each of the different volcanic styles which have operated in the Tharsis region, including the timeframe of activity, the areas and thicknesses of total lava accumulation, and the areas and thicknesses of individual lava flows, are summarized in Table 1. Having outlined the nature and characteristics of the volcanic history of the Tharsis region we now turn to a review of the glacial history of the region.

3. Tharsis region glaciation

Due to the significant elevation of the Tharsis region (Fig. 1a), the majority of which is generally agreed to have formed by the end of the Noachian period (Phillips et al., 2001; Carr and Head, 2010), models of the early Mars climate (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016) predict Tharsis to have been a location of regional snow and ice accumulation during the Noachian and early Hesperian periods. This is because, in the presence of a thicker early Mars CO₂ atmosphere, adiabatic cooling results in substantially decreased mean annual surface temperatures at the elevations of the Tharsis region causing the area to act as a cold trap (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016). This is a robust prediction found to occur not only in models predicting a cold and icy climate (where the global mean annual surface temperature is ~225 K; Wordsworth et al., 2015), but also in forced warm and wet end-member climate models where the global mean annual surface temperature is raised above 273 K through the addition of a uniformly absorbing grey gas (Wordsworth et al., 2015). Under these predicted early Mars climate conditions, snow and ice become sequestered within the Tharsis region due to a lack of ablation, thus leading to the accumulation of snow and ice deposits to thicknesses which are limited by the available global supply of surface/near-surface water. The current available surface/near-surface water inventory on Mars is equivalent to a ~34 m global equivalent layer (GEL) of ice (Carr and Head, 2015). Plausible ranges for the early Mars, Late Noachian/Early Hesperian available surface/near-surface water inventory range from ~1–5X the current available global surface water inventory, yielding a range of ~34–170 m GEL (Fastook and Head, 2015). Recent evaluation (Carr and Head, 2015) of the available global surface/near-surface water supply during these early periods of Mars history, however, suggests a total global available inventory not signifi-
cantly different from the present (~34 m GEL). Within these global available surface/near-surface water inventory constraints, climate (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016) and glacial (Fastook and Head, 2015) modeling predict the accumulation of regional ice sheets above a ~1 km equilibrium line altitude (ELA) (Fig. 1b) to average thicknesses of ~278–725 m, with maximum ice sheet thicknesses reaching ~0.7–2.8 km within deep craters. Detailed discus-
sion on the processes involved in the predicted transfer and sequestration of water-ice within the highland terrains can be found in the referenced studies (e.g. Forget et al., 2013; Wordsworth et al., 2013, 2015; Fastook and Head, 2015; Wordsworth, 2016).

During the later periods of Mars history, beyond the Late Noachian to Early Hesperian, the mass of the martian atmosphere declined by escape and loss processes resulting in a decrease of the surface pressure from ~1 bar toward that of the present day, ~6 mbar (Carr and Head, 2010; Lammer et al., 2013; Jakosky et al., 2017). This loss in atmospheric mass and surface pressure would have been accompanied by a decay in the strength of the adiabatic cooling effect, leading to a mean annual surface temperature distribution governed by radiative equilibrium (Wordsworth et al., 2013). Consequently, the distribution of mean annual surface temperatures would be controlled predominantly by latitudinal position, with minimal influence from elevation. Additionally, the loss of the insulating greenhouse effect produced by the thicker CO₂ atmosphere would lead to globally reduced mean surface temperatures. Climate models of the present day atmospheric conditions of Mars (Madeleine et al., 2014), with a 6 mbar CO₂ atmosphere, indicate a mean annual global surface temperature of ~210 K, a substantial reduction from the ~225 K which is sustained under the early thicker 1 bar CO₂ atmosphere (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016). Under these later atmospheric conditions, the highly elevated Tharsis region would no longer act as a cold trap, ceasing the preferential

Table 1

<table>
<thead>
<tr>
<th>Volcanic Style</th>
<th>Activity Timescale</th>
<th>Total Area (km²)</th>
<th>Total Thickness (km)</th>
<th>Individual Flow Area (km²)</th>
<th>Individual Flow Thickness (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Shield</td>
<td>Noachian-Amazonian</td>
<td>1 × 10⁶</td>
<td>1–10</td>
<td>500</td>
<td>50(1–220)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>(2.5–5 × 10⁶)</td>
<td></td>
</tr>
<tr>
<td>Plains</td>
<td>Noachian-Amazonian</td>
<td>2.8 × 10⁶</td>
<td>1–10</td>
<td>1 × 10⁴</td>
<td>20(1–200)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>(10¹–10²)</td>
<td></td>
</tr>
<tr>
<td>Flood Basalt</td>
<td>Noachian-Hesperian</td>
<td>4.5 × 10⁶</td>
<td>1–3.5</td>
<td>1 × 10⁴</td>
<td>20(1–200)</td>
</tr>
<tr>
<td></td>
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<td>(10¹–10²)</td>
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regional accumulation and sequestration of snow and ice. However, there are several lines of evidence which suggest that more localized accumulation of snow and ice continued throughout later Mars history within the Tharsis rise region. Evidence supporting this continued accumulation of snow and ice includes: (1) Climate modeling studies predicting the accumulation of ice within the Tharsis rise region (Forget et al., 2006; Madeleine et al., 2009, 2014). (2) The presence of extensive fan-shaped deposits superposed upon Amazonian-aged lavas on the western flanks of the Tharsis Montes which are inferred to have been produced through cold-based tropical mountain glaciers (Head and Marchant, 2003; Shean et al., 2005; Forget et al., 2006; Fastook et al., 2008; Kadish et al., 2008; Head et al., 2010). (3) The presence of ring-mold craters on the Tharsis Montes interpreted to form due to impacts into an ice-rich substrate (Head and Weiss, 2014). (4) Indications of extensive glaciation within Valles Marineris during the Late Noachian to Early Hesperian timeframe (Gournon et al., 2014). (5) The distribution of ice-bearing, latitude-dependent mantling deposits to mid-to-low latitudes (Kreslavsky and Head, 2002; Head et al., 2003a, 2016; Madeleine et al., 2009; Schon et al., 2009; Fastook and Head, 2014; Dunders et al., 2018). (6) The preservation of ice-cored pedestal craters within the Tharsis region and at comparable latitudinal bands which are interpreted to form through ejecta armorng of surface snow and ice (Kadish et al., 2010; Kadish and Head, 2011; Schon and Head, 2012; Fastook and Head, 2014). (7) The widespread presence of layered ejecta craters interpreted to have required surface snow and ice to form (Weiss and Head, 2013, 2014, 2017). (8) Deposits of ice-bearing linedatey valley fl ows, lobate debris aprons, and concentric crater fl ow (Head et al., 2006; Madeleine et al., 2009; Levy et al., 2010, 2014; Fastook et al., 2011; Fastook and Head, 2014). (9) The detection of abundant excess ground-ice contained within the nearby Arcadia Planitia through Mars Reconnaissance Orbiter Shallow Radar (SHARAD; Seu et al., 2004) instrument measurements (Bramson et al., 2015).

These lines of evidence can be divided into two broad categories. (1) Evidence for the formation of cold-based tropical mountain glaciers on the Tharsis Montes, indicative of more concentrated/localized snow and ice accumulation (items 1–3). The formation of cold-based tropical mountain glaciers on the western flanks of the Tharsis Montes is predicted to result from preferential snow and ice deposition due to westerly atmospheric circulation and orographic upwelling (Forget et al., 2006; Fastook et al., 2008). The distribution of the fan-shaped deposits on the western flanks of the Tharsis Montes suggest that cold-based glaciers covered an area of at least ~1.6 × 10^5 km^2, with predictions from glacial modeling (Fastook et al., 2008) indicating total areas as large as ~5 × 10^5 km^2 with ice thicknesses ranging from ~100 to 4000 m, averaging ~2000 m. (2) Evidence provided by widespread regional features, such as the pedestal craters, indicating a more distributed/regional pattern of snow and ice accumulation (items 4–9). This class of features is inferred to result from the redistribution of volatiles from the polar regions toward mid-to-low latitudes during periods of increased planetary obliquity (Kreslavsky and Head, 2002; Head and Marchant, 2003; Head et al., 2003a, 2005; Forget et al., 2006; Madeleine et al., 2009, 2014). While these features indicate the occurrence of distributed snow and ice accumulation, they do not provide precise constraints on the total areal distribution. The average thicknesses of surface ice indicated by these features is ~50 m, ranging from ~1 to 100 m with the lower bound corresponding to typical thicknesses of latitude-dependent mantle deposits (Kreslavsky and Head, 2002; Schon et al., 2009) and the upper bound to maximum estimates of surface ice thickness derived from pedestal craters (Kadish et al., 2010).

Summary: In summary, model-dependent evidence suggests that the glacial history of the Tharsis region is characterized by the accumulation of widespread, regional ice sheets during the Noachian and Early Hesperian periods, transitioning to more focused/localized glaciation during the Hesperian and throughout the Amazonian periods. The timeframe, general distribution, and ice thicknesses associated with these various forms of glacial activity occurring throughout the history of the Tharsis region are summarized in Table 2. With the history and nature of both volcanic and glacial activity in the Tharsis region characterized, we now review documented evidence for the past operation of glaciovolcanic processes in the region.

4. Tharsis region glaciovolcanism

In this study, we explore the operation of large-scale surficial lava-ice interactions within the Tharsis region. This assessment is complimented by previous studies which have proposed and recognized the past activity of other forms of glaciovolcanic activity in the Tharsis region, namely interactions between intrusive volcanism and ground-ice as well as subglacial eruptions. Many past studies have proposed a role for subglacial eruptions and the interaction of intrusive volcanism and ground-ice in the formation of a wide suite of landforms in the Tharsis region including: (1) The circum-Chryse (and other) outflow channels (McKenzie and Nimmo, 1999; Chapman and Tanaka, 2002; Wilson and Head, 2004; Leask et al., 2007b). (2) Chasmata and chaos suggested to be the products of large sub-ice eruptions (Chapman and Tanaka, 2002). (3) Steep-sided plateau-like features with smooth capping layers interpreted to be possible tuyas (Chapman and Tanaka, 2001; Komatsu et al., 2004; Head and Wilson, 2007; Kadish et al., 2008; Martínez-Alonso et al., 2011; Scanlon et al., 2014, 2015). (4) Steep-sided mounds and ridges in the fan-shaped deposits on the flanks of the Tharsis Montes interpreted to be moberg or tindar (Head and Wilson, 2007; Kadish et al., 2008; Scanlon et al., 2014). (5) Steep-sided, leveed fl ows interpreted to be subglacial lava fl ows with chilled margins (Head and Wilson, 2007; Scanlon et al., 2014, 2015). (6) Digitate lava fl ows interpreted to have interacted with an upslope glacial margin (Head and Wilson, 2007; Scanlon et al., 2014, 2015). (7) Low, widespread mounds suggested to be preserved remnants of effusions of pillow lavas formed during subglacial eruptions (Scanlon et al., 2014, 2015). (8) Sinuous valleys and ridges interpreted to be subglacial eskers and meltwater channels (Head and Wilson, 2007; Kadish et al., 2008; Scanlon et al., 2014, 2015). (9) Various surface materials interpreted to be deposits of tephra and volcanic material derived from subglacial eruptions and the interactions of intrusive volcanic features with ground-ice (Chapman and Tanaka, 2002; Wilson and Head, 2004; Martínez-Alonso et al., 2011).

While this range of features is inferred to have involved a different subset of glaciovolcanic processes from those investigated in this assessment, they support the past occurrence of contemporaneous volcanic and glacial activity and the operation of glaciovolcanic processes in the Tharsis region (though it is worth noting that alternative interpretations have suggested the formation of many of these features exclusively through volcanic activity; Leone, 2014, 2017; Leverington, 2004, 2006, 2009, 2011, 2014, 2018). Therefore, these features lend additional justification for the glaciovolcanic scenario we explore. With the characteristics of regional volcanism and glaciation within the Tharsis region outlined, and the history of glaciovolcanism now established, the details of the specific glaciovolcanic processes we assess are next outlined.

<table>
<thead>
<tr>
<th>Style</th>
<th>Activity Timeframe</th>
<th>Average Thickness (m)</th>
<th>Thickness Range (m)</th>
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<tbody>
<tr>
<td>Regional</td>
<td>Noachian-Early Hesperian</td>
<td>275–725</td>
<td>0–2800</td>
</tr>
<tr>
<td>Localized</td>
<td>Mid-Hesperian-Amazonian</td>
<td>2000</td>
<td>100–4000</td>
</tr>
<tr>
<td>Regional</td>
<td>Mid-Hesperian-Amazonian</td>
<td>50</td>
<td>1–100</td>
</tr>
</tbody>
</table>
5. Glaciovolcanic model description

In this section, we provide a brief, high-level introduction to the general glaciovolcanic model and concepts developed through prior work and applied in this evaluation. More detailed descriptions of the fundamental glaciovolcanic model and concepts can be found in the original referenced sources (cited below where appropriate) as well as in following sections wherein more specific models are described. In our evaluation of glaciovolcanism in the Tharsis region, we specifically assess a lava-ice interaction geometry involving the emplacement of lava flows atop surficial snow and ice deposits, as proposed in the ice sheet lava heating and loading mechanism (Cassanelli and Head, 2016, 2018a; Fastook and Head, 2018). This particular lava-ice interaction geometry is assessed because, while other lava-ice interaction geometries are possible, supra-ice lava emplacement optimizes heat transfer to the surface ice (Head and Wilson, 2002, 2007; Wilson and Head, 2002a; Wilson and Head, 2007a,b; Cassanelli and Head, 2016, 2018a) thereby providing testable maximum end-member estimates for melting rates. Furthermore, recent terrestrial studies which have documented the interaction between extrusive lava flows and surface snow and ice on Earth (Edwards et al., 2012, 2013, 2014, 2015) have found the supra-ice geometry to be a dominant mode of lava flow emplacement. Supra-ice emplacement of extrusive lava flows can occur by (1) direct advancement of lava flows onto the ice from topographically-elevated terrain, as has been observed in glaciated terrestrial volcanic settings (Edwards et al., 2012, 2013, 2014, 2015), (2) supraglacial erosions fed by dikes emplaced through surface ice due to high strain rates (Head and Wilson, 2002, 2007; Wilson and Head, 2002a; Wilson and Head, 2007a,b) or, (3) eruptions from tuyas that have pierced through surface ice (Smellie, 2007; Kadish et al., 2008; Scanlon et al., 2014).

This particular glaciovolcanic process, supra-ice lava flow emplacement, is predicted to result in significant meltwater production through top-down contact and bottom-up geothermal heating (which could serve as a source for the formation of fluvial features) as well as surface deformation through the subsidence and collapse of lava flows (due to melting and removal of underlying ice) (Cassanelli and Head, 2016, 2018a). We now explore the possible involvement of large-scale, surficial glaciovolcanism through supra-ice emplacement (and the predicted resultant processes) in the formation of tectonic and fluvial features in the Tharsis region. The primary objectives of these exploratory analyses are to: (1) consider a novel suite of glaciovolcanic models and processes that may plausibly account for the formation of the range of expected features, and (2) evaluate surficial glaciovolcanism as a potentially advantageous explanation for their formation as a basis to refine current understanding and to help guide further investigations.

6. Tectonic features

In our exploratory evaluation of the role of glaciovolcanic processes in the formation of tectonically-related features in the Tharsis region, we focus on the aureole apron formation surrounding the Olympus Mons edifice and the enigmatic Syrria-Thaumasia block/Solis Planum mega-thrust, beginning with the former.

6.1. Volcanic edifice aureoles

The Olympus Mons shield edifice (Fig. 1) is almost entirely encircled by a scarpl 6–8 km in height (Basilevsky et al., 2005) and the extensive aureole apron units which extend up to ~700 km from the edifice bounding scarpl, predominantly off the northwestern margin (Fig. 3a and b). The aureole units of Olympus Mons date to ~2.54–3.5 Ga (Morris and Tanaka, 1994; Isherwood et al., 2013) and are comprised of at least 9 individual rocky deposits up to 2 km thick characterized by lobate, sub-circular plandform and hummocky texture (Tanaka, 1985; De Blasio, 2018) (Fig. 3a and b). The formation of the aureole units of Olympus Mons has been attributed to a multitude of mechanisms including lava flows (McCusker et al., 1972), subglacial deposits (Hodges and Moore, 1979), ash flows (Morris, 1981), deep-seated deformation (Francis and Wadge, 1983), subaerial landslides (Harrison and Grimm, 2003), subaqueous landslides (De Blasio, 2018), and ice lubricated landslides (Tanaka, 1985). However, consensus among recent analyses of the morphology of the Olympus Mons aureole units suggests an origin by a complex catastrophic landslide (De Blasio, 2018). Here we do not aim to falsify or support any of the aforementioned proposed formation models. These have been presented in the interest of acknowledging all plausible models, with our objective being to assess the viability of an additional, and previously unconsidered aureole formation process by surficial glaciovolcanism.

There are two plausible ways in which the formation of the aureole units could have resulted from, or been facilitated by, large-scale surficial lava-ice interactions. (1) Subsidence and collapse of an accumulated sequence of Olympus Mons lava flows emplaced atop surrounding ice sheets due to melting (Cassanelli and Head, 2016). (2) Rapid emplacement of lava flows due to basal lubrication from melting of underlying ice, similar in nature to the previously proposed ice-lubricated landslide model (Tanaka, 1985). In both cases, active snow and ice accumulation would be required around the base of the Olympus Mons scarp during the Hesperian to Early Amazonian periods (as indicated by the estimated age of the aureole units) and at elevations below the ~1 km equilibrium line altitude predicted to characterize snow accumulation under Late Noachian climate conditions (Fastook and Head, 2015). Within these constraints, ice accumulation could have occurred by the redistribution of polar volatiles equatorward during high obliquity excursions, a process thought responsible for the formation of the ice-cored latitude-dependent mantle deposit (Head et al., 2003a). The potential for the accumulation of ice near the base of the Olympus Mons edifice during the Hesperian to Early Amazonian is broadly supported by recent detections of excess ground-ice buried within the nearby Arcadia Planitia (and other mid-latitude locations) (Bramson et al., 2015; Dundas et al., 2018), proximal, ice-related viscous flow features (Milliken et al., 2003), and the extent of the latitude-dependent mantle (Kreslavsky and Head, 2000).

Lava flow subsidence/collapse: Formation of the Olympus Mons aureole deposits could plausibly be accounted for by the subsidence and collapse of lava flows emplaced atop ice surrounding the Olympus Mons basal scarp due to the substantial lava flow deformation predicted (Cassanelli and Head, 2016) to result from this process. The deformation of supra-ice emplaced lava flows, resulting from subsidence and collapse due to melting and removal of underlying ice, could produce a deposit with a heavily disrupted surface morphology potentially similar to that exhibited by the Olympus Mons aureole units (Fig. 3a and b). In order to account for the formation of the Olympus Mons aureole units by the subsidence-collapse of a supraglacial lava sequence, a surrounding ice sheet ~6–8 km thick along the edifice basal scarp is required to produce the elevation offset between the scarp and the surface of the aureoles by melting and removal (Fig. 3). The accumulation of an ice sheet to ~6–8 km in thickness would require highly asymmetric, localized ice accumulation and would also likely require an unreasonably large available surface/near-surface water inventory, as this ice thickness exceeds all maximum estimates of martian ice sheet thickness under a wide range of conditions (see Section 3 and Table 2). Moreover, an ice sheet 6–8 km thick, even at the low mean annual surface temperatures at the surface elevation around the base of Olympus Mons (under typical Late Noachian conditions around the 3.5 Ga age of the apron units; Wordsworth et al., 2013), would be thermally unstable. This can be simply demonstrated by the steady-state one-dimensional heat conduction equation:

\[ G = K_\text{f} \frac{dT}{dz} \]  

(1)

where \( G \) is the geothermal heat flux (~55 mW/m²) during the Late Noachian period; Solomon et al., 2005), \( K_\text{f} \) is the thermal conductivity of...
ice (~2 W/m K under Late Noachian martian conditions; Cassanelli and Head, 2015), T is temperature (K), and z is depth (m). With a mean annual surface temperature of ~230 K at the base of the Olympus Mons Scarp under typical Late Noachian conditions (Wordsworth et al., 2013), the solution of (1) indicates a thermal-equilibrium ice sheet thickness of ~1.5 km. The accumulation of an ice sheet beyond this thermal equilibrium thickness would result in the onset of basal melting. Therefore, unless the rate of ice accumulation greatly exceeds the rate of basal melting (unlikely, as under typical Late Noachian Mars conditions, bottom-up melting rates are comparable to predicted accumulation rates; Wordsworth et al., 2013; Cassanelli and Head, 2015, 2016), ice sheet thickness could not grow beyond this equilibrium value. If rapid accumulation allowed an ice sheet to grow beyond the equilibrium thermal thickness, the rate of basal melting would consequently increase and act to drive ice sheet thickness down toward the equilibrium value. If the rate of ice accumulation subsequently waned (possibly due to exhaustion of the surface/near-surface ice inventory), the ice sheet thickness would decline back toward the equilibrium value. Therefore, due to the difficulties in the creation and maintenance of an ice sheet of the required 6–8 km thickness, the formation of the Olympus Mons aureole units by the subidence and collapse of lava flows emplaced atop surrounding surface ice is not favored.

Frictional reduction and lava flow sliding: A second possible surficial glaciovolcanic origin of the Olympus Mons aureole units is by rapidly emplaced supraglacial lava flows (broadly similar to the previously proposed ice-lubricated landslide model; Tanaka, 1985) in the following manner (Fig. 3c): (1) lava flows advance rapidly over surface ice deposits surrounding the Olympus Mons edifice due to formation of a lubricating basal melt lens, (2) once the flows advance beyond the surface ice margin a rampart is produced due to the higher frictional basal surface and emplacement deceleration, (3) ice underlying the emplaced lava flow is then melted leading to subsidence and collapse, (4) surface ice deposits are reformed on the collapsed lava flow, (5) subsequent lava flows are emplaced in a similar manner, back-stepping toward the edifice generating lobate scarps (from the ramparts) and a disrupted surface morphology. This mechanism may also help account for the extensive distances over which the aureole units are deposited away from the Olympus Mons basal scarp (Fig. 3a and b), as the basal lubrication of supraglacial lava flows by melting has been observed to enhance lava flow emplacement rates under terrestrial conditions (Edwards et al., 2013). We note, however, that these observations were collected under laboratory conditions for much thinner lava flows on steeper slopes (Edwards et al., 2013). Given this issue of scale, we do not base any results or conclusions on these observations, and instead later apply simple scaling relationships to measurements from natural terrestrial supraglacial lava flows to generate approximate emplacement velocity estimates. Nevertheless, these experimental observations (Edwards et al., 2013) generally indicate increased lava flow emplacement velocities over an ice substrate.

To perform a first-order evaluation of the viability of this mechanism to produce the aureole units, we assess whether a typical Tharsis shield lava flow could reach the 700 km distal end of the aureole units (Fig. 3) prior to cooling and solidification. This criteria can be addressed in a very simple fashion by taking a ratio of the timescale required for thermal diffusion across the lava flow thickness (assuming lava flow cooling predominantly by conduction and neglecting advective effects which
would enhance cooling due to material movement during emplacement) and viscous flow over the required flow distance (assuming Newtonian rheology for simplicity) which yields:

$$\frac{2L \kappa \mu}{H^2 \rho g \sin \alpha}$$  \hspace{1cm} (2)

where \( L \) is a flow distance of interest (~700 km in this case), \( \kappa \) is the thermal diffusivity (for which a typical value of \( 10^{-6} \text{m}^2/\text{s} \) is adopted), \( \mu \) is the lava viscosity (for which a value of \( 10^5 \text{Pa}\text{s} \) is used, consistent with measurements from terrestrial basalts; Harris and Allen, 2008). \( H \) is the lava flow thickness (for which the typical 50 m value is utilized for Tharsis shield volcanism; Table 2), \( \rho \) is the lava density (assuming a typical value of 3000 kg/m\(^3\)), \( g \) is gravitational acceleration (3.71 m/s\(^2\) on Mars), and \( \alpha \) is the slope (for which a value of 0.3° is used, the approximate average slope across the longest section of the aureole units). Substituting the relevant parameters into (2) indicates a ratio value on the order of \( 10^{-5} \), suggesting that, for the conditions of interest, viscous flow occurs over a more rapid timescale than diffusive lava flow cooling.

As a second test of the ability of supraglacial lava flows to reach the 700 km distal end of the aureole units, emplacement velocities of terrestrial supraglacial can be utilized. Observations of terrestrial supraglacial flows documents by Edwards et al., (2012) exhibited an approximate average emplacement velocity of 35 m/hr. To produce an estimate for supraglacial lava flow emplacement velocities on Mars, we scale this emplacement velocity by a factor of \( \left( \frac{\sin \alpha}{\sin \alpha} \right)^{\alpha} \left( \frac{g}{g} \right)^{\alpha} \left( \frac{H_m/H_2}{1} \right)^{\alpha} \) (from the Bingham plastic lava rheology) to account for the differences in slope, gravity, and lava flow thickness. Substituting appropriate parameters (\( a_m \sim 0.3^\circ \); \( a_e \sim 5^\circ \); \( g_2 \sim 9.81 \text{m/s}^2 \); \( H_m \sim 50 \text{m} \); \( H_e \sim 5 \text{m} \); Edwards et al., 2012) yields a scaling factor of ~2 which translates to an average emplacement velocity of ~70 m/hr (~0.02 m/s). Thus, assuming this as the constant emplacement velocity, ~1 yr would be required for the lava flow to reach a distance of 700 km. This emplacement timescale is well below the characteristic diffusive cooling timescale for a 50 m thick lava flow (the typical lava flow thickness produced by Tharsis shield volcanism; Table 1), given by \( H^2/\kappa \) where \( \kappa \) is the lava thermal diffusivity (with a typical value of \( 10^{-6} \text{m}^2/\text{s} \), which is ~80 yr. It is important to note, however, that these simple evaluations of lava flow emplacement neglect the influence of advective heat transfer which would result in more rapid lava flow cooling and a consequential reduction in the possible emplacement length. Additionally, the increases in lava viscosity which would occur due to cooling during lava flow emplacement over this timescale and space are not accounted for, a factor which would result in decreases of the emplacement velocity and length. While both of these factors would lead to reductions in the maximum distances over which supraglacial lava flows could be emplaced, it is not uncommon for martian lava flows to exceed 700 km in length, with some individual flows stretching over distances in excess of 1–2x10\(^6\) km (Keszthelyi et al., 2000, 2004; Warner and Gregg, 2003; Chapman et al., 2010b). Therefore, we conclude that the supraglacial emplacement of lava flows over the 700 km length of the aureole units is plausible.

We next examine the morphometry of the aureole units as a further test of the glaciovolcanic supra-ice lava flow emplacement origin. Variations in the topography across the surface of the aureole units, produced by the ridges and scarps (Fig. 3c), is characteristically on the order of ~100 m with some scarps showing elevation offsets of up to ~1 km (Fig. 3c). Given the lava flow emplacement origin we assess (Fig. 3c), and typical lava flow thickness from Olympus Mons of ~50 m, these topographic variations require some form of distal lava flow thickening by a factor of ~2–20. This could have been accomplished by a rapid change in the resistance to the lava flow from either a change in basal friction as the flow advanced beyond the glacial margins or by encountering a thickened distal rampart from a previously emplaced lava flow (Fig. 3c). In plan view, the aureole deposits exhibit a wide, fan-shaped lobe morphology radial to the Olympus Mons edifice extending over widths of several hundred kilometers. In a supraglacial lava flow origin scenario, this morphology could most plausibly have been produced by lava flow emplacement geometry initially controlled by radial spreading from the Olympus Mons edifice with later central flow focusing and extension due to marginal chilling. A final morphometric consideration of note is the apparent lack of preserved phreatomagmatic features within the aureole deposits. Generally, the advancement of lava flows over a volatile substrate is anticipated to result in molten fuel-coolant interactions generating rootless cones through explosive steam production (Dundas and Keszthelyi, 2013). However, significant phreatomagmatic activity and rootless cone formation were not documented in recent observations of the advancement of terrestrial lava flows over surface ice (Edwards et al., 2012, 2014, 2015). While observations are currently limited, the lack of phreatomagmatic activity may be attributed to steep topographic gradients and permeable snow allowing for rapid drainage of meltwater preventing major steam production. These same factors could also have been responsible for inhibiting the production of rootless cones during the formation of the aureole units, particularly given the porous state likely to characterize martian surface ice deposits during the time period of formation (Cassanelli and Head, 2015).
2009). (6) Incipient plate tectonism on Mars (Baker, 2006; Baker et al., 2007; Yin, 2012a, b; Dohm et al., 2013, 2018). As in an earlier section, the proposed explanations listed above have been presented in the interest of acknowledging all plausible models. Here our objective is again to assess the viability of an additional, and previously unconsidered, formation process by surficial glaciovolcanism.

A possible surficial glaciovolcanic origin scenario for the Syria-Thaumasia block is thin-skinned crustal deformation along a weak detachment layer of volcanically buried ice (Fig. 4d), akin to the mechanisms proposed by (Webb and Head, 2002) and (Montgomery et al., 2009). This glaciovolcanic origin scenario could have resulted from the supraglacial emplacement of the plateau plains-forming lava flows (and potentially explosively generated volcanic tephra; Wilson and Head, 2009) with incomplete top-down melting of the underlying ice, thereby producing a buried ice layer (Cassanelli and Head, 2016) (Fig. 4d). In order for the surface ice to have been preserved through supraglacial lava flow emplacement, the thickness of the plains forming lava flows could not have greatly exceeded ~10 m (Cassanelli and Head, 2016). This is because at these thicknesses the individual lava flows do not contain enough heat energy to melt relatively significant thicknesses of the Noachian/Hesperian regional ice sheets (which are predicted to be ~275 thick on average; Table 2), and as lava flows continue to accumulate, top-down heat transfer is rapidly attenuated by the presence of intervening layers of previously emplaced chilled lava. The attenuated top-down heat transfer leads to reduced top-down melting during the emplacement of subsequent lava flows, and thus the preservation of the buried ice (Cassanelli and Head, 2016). Typical plains volcanism lava flow thicknesses in the Tharsis region are ~20 m (Table 2) and therefore, are relatively consistent with this constraint, allowing the possibility for preservation of surface ice buried beneath supra-glacially emplaced volcanic plains. Additionally, the preservation of surface ice during volcanic burial could have been facilitated by the supraglacial deposition of explosively generated volcanic tephra (as was noted in the preceding section) protecting the ice from ablation (Wilson and Head, 2009) and allowing the accumulation of thicker lava flows.

As a result of the limited top-down melting during supra-ice emplacement of ~10 m thick lava flows (Cassanelli and Head, 2016), the overall thickness of the Noachian/Hesperian regional ice sheets will not be significantly reduced, and the primary effect of the accumulating lava flows will be the establishment of a thermally insulating layer atop the ice sheets. The additional insulation provided by the accumulating lava flow sequence will act to raise the ice-melting isotherm from depth toward the base of the buried ice, which upon encountering the ice will result in the initiation of basal melting (Cassanelli et al., 2015; Cassanelli and Head, 2016). Depending upon the geothermal heat flux and the rate and total thickness of lava flow accumulation, the onset of basal melting can be deferred beyond the lava flow emplacement period by timescales up to the order of 1 Myr (Cassanelli and Head, 2016). After the onset of basal melting, a meltwater lens could reduce the basal friction and lead to slip along the basal surface, causing movement and deformation of the overlying volcanic plains. It is worth noting that deformation occurring in this particular manner would require a relatively impermeable substrate material beneath the ice to prevent meltwater infiltration and allow the formation of a basal melt lens. However, the typical infiltration capacity of the martian crustal substrate is predicted to considerably exceed bottom-up ice sheet melting rates over a wide range of reasonable...
geothermal heat flux estimates and thermophysical ice sheet properties (Cassanelli et al., 2015; Cassanelli and Head, 2015, 2016). Alternatively, deformation could have taken place within the weak buried ice layer (Montgomery et al., 2009) without the requirement of ice sheet basal melting. In either case, however, the depth of deformation would be constrained by the thermal equilibrium depth of buried ice in the Tharsis region. This is because after the base of the buried ice surpasses the thermal equilibrium depth due to accumulation of superposed lavas, basal melting of the buried ice would initiate and reduce the ice thickness back toward the equilibrium depth. If superposed lavas accumulated to a thickness sufficient to exceed the ice equilibrium thermal depth, the entire buried ice layer would be subject to bottom-up basal melting by geothermal heat input. It is possible for a thickness of superposed lavas exceeding the ice thermal equilibrium depth to have been accumulated if the rate of construction was more rapid than the rate at which the ice-melting isotherm could ascend toward the base of the buried ice. This “deferred” bottom-up basal melting scenario results from the limitations in total geothermal heat input to the base of the buried ice layer (Cassanelli and Head, 2016). However, given that the maximum timeframe for deferred melting is on the order of 1 Myr, it is unlikely that a thickness of superposed lava greatly exceeding the ice equilibrium thermal depth could have been accumulated.

The ice thermal equilibrium depth can be determined with the use of equation (1) following the procedure outlined in Section 6.1. With a nominal Late Noachian-Early Hesperian mean annual surface temperature in the region of ∼230 K (Forget et al., 2013; Wordsworth et al., 2013, 2015), heat flux of ∼55 mW/m² (Solomon et al., 2005), and an upper-end estimate of the overburden volcanic plains thermal conductivity of 3 W/m K, we find a thermal equilibrium depth of ∼2.4 km. It is possible for this nominal thermal equilibrium depth estimate to have been influenced by elevated regional heat fluxes in the Tharsis region due to volcanic activity (Cassanelli et al., 2015) as well as variations in the thermal conductivity of the plains materials. At a regionally elevated geothermal heat flux of 100 mW/m² (Cassanelli et al., 2015) and a reduced volcanic plains thermal conductivity value of 2 W/m K, the thermal equilibrium depth is reduced to ∼0.9 km. While supraglacial volcanic tephra deposition could have aided in initially preserving surface ice through volcanic burial, the reduced thermal conductivity of the porous tephra material relative to consolidated lava flows would further reduce the thermal equilibrium depths. Therefore, for the reasons earlier discussed in this section, the depth of deformation in the glaciovolcanic origin scenario for the Thaumasia-Syria block could not greatly exceed these predicted ice thermal equilibrium depths, thus requiring relatively thin-skinned detachment/deformation. These ice thermal equilibrium depths lie at the lower end of the detachment/deformation depth range estimated for the Syria-Thaumasia block deformation which spans from ∼1 to 15 km (Anguita et al., 2006; Okubo and Schultz, 2004; Montgomery et al., 2009). The lower end of the estimated ∼1–15 km detachment/deformation depth range is inferred primarily for the regions wrinkle ridge features (Mangold et al., 1998; Montgomery et al., 2009) while the deeper values at the upper-end correspond to the large-scale tectonic features such as Coprates Rise and the Thaumasia Highlands, both interpreted as surface exposures of thrust faults (Plescia and Saunders, 1982; Schultz and Tanaka, 1994; Webb and Head, 2002). Thus, in order for the proposed surficial glacioclastic origin scenario (Fig. 4d) to account for the complete range of tectonic deformation associated with the Syria-Thaumasia block, and the inferred depths of detachment/deformation, very rapid supraglacial lava accumulation to a total thickness of ∼10 km is required. Given the required accumulation of such a large thickness of volcanic material over a geologically short timescale of ∼1 Myr, formation of the Syria-Thaumasia block/Solis Planum mega-thrust by detachment along a volcanically-buried layer of ice (Fig. 4d) is not favored.

Summary: We tested a surficial glacioclastic origin scenario for the Syria-Thaumasia block by thin-skinned crustal deformation along a weak detachment layer of volcanically-buried ice (Fig. 4d). Formation by this mechanism requires accumulation of the volcanic plains which constitute the Syria-Thaumasia block plateau by individual lava flows with thicknesses that do not greatly exceed ∼10 m in order to preserve the buried ice (or possibly thicker lava flows if explosive tephra was first deposited onto the ice). This constraint is generally consistent with the estimated lava flow thicknesses for the plains volcanic style responsible for the construction of the plateau (Table 1). Alternatively, in order for crustal deformation to have occurred along a weak buried ice layer, the depth of detachment could not greatly exceed the thermal equilibrium depth of ice within the subsurface as beyond this depth ice would be unstable to melting and removal. A simple analysis of the thermal equilibrium depth for ice within the Syria-Thaumasia block plateau indicates a maximum depth of ∼2.4 km which lies at the lower end of the ∼1–15 km depth range estimated for the detachment/deformation of the Syria-Thaumasia block/Solis Planum mega-thrust. However, the lower end of this estimated detachment/deformation depth range is associated with the detachment depths of the regions wrinkle ridge features, while the larger tectonic features such as the Thaumasia highlands and Coprates Rise, both interpreted as surface exposures of thrust faults (Plescia and Saunders, 1982; Schultz and Tanaka, 1994; Webb and Head, 2002), require deeper detachment toward the higher end of the ∼1–15 km range. Therefore, in order for the proposed surficial glacioclastic origin scenario (Fig. 4d) to account for the complete range of tectonic deformation associated with the Syria-Thaumasia block, and the inferred depths of detachment/deformation, very rapid supraglacial lava accumulation to a total thickness of ∼10 km is required. Given the required accumulation of such a large thickness of lava over a geologically short timescale of ∼1 Myr, formation of the Syria-Thaumasia block/Solis Planum mega-thrust by detachment along a volcanically-buried layer of ice (Fig. 4d) is not favored.

7. Valley networks

The valley networks (Carr and Clow, 1981; Gulick and Baker, 1989; Baker et al., 1992; Carr, 1996, 2012; Hynek et al., 2010) (Fig. 1) are branching systems of linear-to-sinuous depressions which have persisted as one of the strongest lines of evidence for past fluvial activity on the surface of Mars. Valley networks are widely distributed (Hynek et al., 2010) across the surface of Mars, and have been dated by crater-counting analyses (Fassett and Head, 2008) to predominantly the Late Noachian-Early Hesperian period of Mars history. The valley networks can be broadly divided into two populations, (1) a generally older population of systems widely distributed across the cratered highland terrains of Mars, and (2) a younger population of more isolated systems incised into volcanic edifices across the planet (though there exist some exceptions to this division including localized, relatively young valley network systems within the highland terrains; Wilson et al., 2016). The valley networks associated with the volcanic edifices of Mars typically exhibit younger ages (Fassett and Head, 2008), slightly elevated drainage densities (Hynek et al., 2010), and distinct morphologies (Fassett and Head, 2007) when compared against the older, more ubiquitous valley network systems that occur throughout the cratered Noachian highland terrains. The Tharsis regions contains valley networks systems from both of these broad populations, with those associated with the broader Noachian highland population located principally within the Thaumasia highlands, and the distinctly younger systems associated with volcanic edifices.

While it is nearly universally agreed that the origin of the valley networks is a result of fluvial activity, whether they formed under relatively warm and wet or cold and icy climate conditions remains unresolved. Accounting for the formation of the valley networks by warm and wet conditions has been problematic due to predictions of climate models suggesting the climate of early Mars was generally cold and icy (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016; Palumbo et al., 2018) (though it is possible for the valley networks to have formed in response to short-term warming excursions, a subject of much ongoing
research beyond the scope of consideration in this assessment; e.g. Baker, 2001, 2009; Toon et al., 2010; Halevy and Head, 2014; Palumbo et al., 2018). These difficulties are amplified for the valley network systems of the Tharsis region because: (1) the Tharsis valley networks systems reside at significant elevations (Fig. 1b) in areas predicted (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016; Palumbo et al., 2018) to exhibit highly depressed mean annual surface temperatures, and (2) the valley networks incised upon the Tharsis volcanic edifices exhibit relatively young Hesperian/Amazonian ages and were therefore formed in a predominantly cold and dry period of Mars history (Carr and Head, 2010), beyond the timeframe of the potentially warm and wet Noachian. Given these challenges, and the occurrence of the Tharsis valley network systems in a region of intense, prolonged volcanic activity, we examine the possible role of glaciovolcanism in accounting for their formation, an origin which does not require problematic warm and wet climate conditions. We first assess the regional valley network systems of the Tharsis region.

7.1. Regional valley network systems

The regional valley networks of the Tharsis region (Fig. 5a and b), concentrated within the Thaumasia highlands, are among the most mature fluvial systems on Mars, with systems exhibiting both high drainage densities and stream orders (Hynek et al., 2010). The formation of the regional valley networks in the Thaumasia highlands of the Tharsis region have previously been attributed to: (1) precipitation in a relatively warm and wet climate (Ansan and Mangold, 2006), (2) melting of snow in response to transient heating events in an ambient cold and icy climate (Gulick et al., 1998; Gulick, 2001; Scanlon et al., 2013), (3) groundwater sapping (Mangold and Ansan, 2006), (4) groundwater discharge due to hydrothermal activity (Dohm et al., 1998; Gulick, 1998, 2001; Gulick et al., 1998; Tanaka et al., 1998; Dohm and Tanaka, 1999), and (5) groundwater discharge produced by seismic activity from impact events (Gulick et al., 1998; Tanaka et al., 1998; Dohm and Tanaka, 1999). In this assessment we focus specifically on Warrego Valles, a particularly well-developed valley network system (Gulick, 2001) which is incised into the southern face of the Thaumasia highlands (Fig. 5a–d).

A possible surficial glaciovolcanic mechanism to account for the formation of the Warrego Valles valley network (and by extension, the other systems of regional valley networks in the Thaumasia highlands), is the dewatering of crust and volcanically-buried ice uplifted to the surface by a Solis Planum mega-thrust (as discussed in the preceding section and originally proposed by Montgomery et al., 2009) (Fig. 5e). Formation of valley networks by this mechanism (Fig. 5e) predicts: (1) Valley network heads to begin below the peak elevation of the highland thrust material by a distance equivalent to the depth of the ice melting isotherm below the original surface prior to uplift, as thawed material would need to be exposed at the surface to cause discharge (this prediction assumes that the uplifted material has not been subject to significant erosion and downwasting). (2) Valley network heads over a narrow range of elevations corresponding to the thickness of thawed subsurface material (which was originally at a depth sufficient to heat the material above the }

![Image](image_url)
melting point) exposed at the surface by thrusting. (3) Valley networks that exhibit characteristic groundwater sapping channel morphology (as opposed to that characteristic of runoff/overland flow sourced channels) with amphitheater-shaped head walls and flat floors. (4) Contemporaneous ages of faulting and valley network formation within the Warrego Valles region. Crater-count age dating estimates (Fassett and Head, 2008) suggest an approximate age for the Warrego Valles system of ~3.7 Ga, near the Noachian-Hesperian boundary. This is broadly consistent with the estimated Late Noachian-Early Hesperian age of faulting in the Warrego region derived from stratigraphic constraints (Anderson et al., 2001; Dohn et al., 2001).

In order to evaluate the consistency of Warrego Valles with the first of these outlined predictions (the initiation of valley heads below the peak elevations of the uplifted host terrain by a distance equal to the depth of the ice-melting isotherm below the original surface) measurements of the distance of the uppermost valley tributary heads from the peak elevations of the host terrain (as indicated by gridded MOLA topography) are taken and compared to estimates of the depth of the ice-melting isotherm below the original surface. The depth of the ice-melting isotherm below the original surface is estimated through the use of the steady-state heat conduction equation following the procedure described in Section 6.1. Assuming a plausible, inclusive range of substrate thermal conductivities from 1 to 3 W/m K, geothermal heat fluxes from 55 to 100 mW/m² (Solomon et al., 2005; Cassanelli et al., 2015) and a predicted (Forget et al., 2013; Wordsworth et al., 2013, 2015) mean annual surface temperature in the Thaumasia highlands region of ~230 K yields equilibrium ice-melting isotherm depths of ~0.5–2.4 km below the original surface. To determine if the Warrego Valles valleys head below the local highland peaks by an amount equal to or greater than these ice-melting isotherm depths measurements of the elevation at the head of each tributary within the Warrego Valles system have been collected using gridded MOLA topographic data (Fig. 5f). Given the 0.5–2.4 km range of estimated ice melting isotherm depths, the valley network head elevation measurements indicate that the Warrego Valles system generally satisfies this first morphologic requirement as the majority of the valleys head at elevations (Fig. 5d and f) > ~1 km below the peak elevations of the local highlands (~8–9 km).

We next utilize the Warrego Valles tributary head elevation measurements (Fig. 5f) to evaluate if the system exhibits a sufficiently consistent band of valley network head elevations, in accordance with the second genetic model prediction. The histogram of the elevations of the Warrego Valles tributary heads indicates a range of head elevation from ~3 to 8.5 km, with a majority of the system tributaries (~95%) heading over a 3.5 km elevation range from ~4 to 7.5 km (Fig. 5f) (this compares to the total elevation range of the Warrego Valles system, the elevation difference between the mouth of the main branch to the head of the most elevated tributary (Fig. 5c and d), of ~2.5–8.5 km). Given the 3.5 km range in valley network tributary head elevations, and the prior criteria requiring a minimum total uplift of ~0.5–2.4 km (in order for material below the ice-melting isotherm depth to have been exposed at the surface allowing valley networks to form by the proposed dewatering mechanism), the crustal materials of the Thaumasia highlands must have been uplifted by a minimum total of ~4–5.9 km. In the area of Warrego Valles, the height of the Thaumasia highlands above the adjacent volcanic plains is ~5 km, but varies from ~1 to 6 km along portions of the highlands containing valley network systems. Therefore, the total uplift of the Thaumasia highlands is generally within the minimum range required for the valley networks to have formed by the dewatering mechanism, but some localized valley network systems reside in portions of the highlands below the required uplift threshold. Thus, these valley network systems require the operation of a different formation mechanism or substantial erosion, downwasting, or subsidence of the local host highlands terrains.

The final prediction of the dewatering mechanism (Fig. 5e) is the production of valley networks exhibiting groundwater sapping morphology, as the source of the water for the valleys in this scenario is the subsurface. The morphology of the Warrego Valles system, however, is more consistent with that of valleys formed by surface runoff and overland flow (Ansan and Mangold, 2006), suggestive of a precipitation source, either rainfall or snowfall with subsequent melting. Despite the overland flow morphologic characteristics, a rainfall precipitation origin of Warrego Valles has been challenged by the lack of similarly well-formed valley systems in immediately adjacent terrain of the same geologic age (Gulick et al., 1998; Gulick, 2001), indicating a more localized water source (alternatively, the lack of drainage development on adjacent terrains could be the result of variations in surface infiltration capacity, where areas of reduced infiltration favor enhanced drainage development; Baker and Partridge, 1986). Therefore, given the lack of sapping channel morphology, and the evidence against a distributed rainfall precipitation source (assuming relatively uniform regional surface infiltration capacity), a favorable alternative mechanism to account for the formation of the Warrego Valles valley network system is the localized melting of surface snow deposits in response to elevated geothermal heat fluxes due to volcanic activity and tectonic uplift (Gulick et al., 1998; Gulick, 2001).

Summary: Using the Warrego Valles valley network system as a case example, we assessed a possible surficial glaciovolcanic origin of the regional highland valley networks concentrated within the Thaumasia highlands of the Tharsis region (Fig. 5). Specifically, we evaluated a genetic mechanism for Warrego Valles (and by extension, the other systems of regional valley networks in the Thaumasia highlands) by dewatering of crust and ice uplifted to the surface by a Solis Planum mega-thrust. The formation of the highland valley networks by this mechanism predicts: (1) valley network heads to begin below the peak elevation of the uplifted Thaumasia highlands by a distance equivalent to the depth of the ice melting isotherm below the original surface (as thawed material would need to be exposed at the surface to cause discharge), (2) a narrow range in valley network head elevations corresponding to the thickness of dewatering subsurface material exposed at the surface by thrusting, and (3) characteristic groundwater sapping channel/valley morphology. The first two of these predicted criteria are generally met by Warrego Valles, however, the valley morphology of the system is more consistent with an origin by runoff/overland flows associated with an origin by precipitation. Therefore, given the lack of sapping channel morphology, and evidence against a source by rainfall precipitation (assuming relatively uniform regional surface infiltration capacity), a favored origin of the Warrego Valles valley network system is the melting of surface snow deposits in response to elevated geothermal heat fluxes produced by the uplift of heated material from depth during thrusting to form the Thaumasia highlands.

7.2. Volcanic valley network systems

Many of the volcanic edifices within the Tharsis region are incised by systems of valley networks (Fig. 6a and b). Given that these valley network systems head at or near the summits of the highly elevated Tharsis volcanic edifices (Fig. 6a and b), it is predicted (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016; Palumbo et al., 2018) that, even under even under relatively warm and wet global climate conditions, the formation of these particular valley network systems would have occurred in the presence of local mean annual surface temperatures well below the freezing point of water. Additionally, the volcanic valley network systems of the Tharsis region have been age dated by crater counting techniques (Fassett and Head, 2008; Hynek et al., 2010) to ~0.6–3.7 Ga. These relatively young inferred ages indicate that the formation of these volcanic valley network systems took place beyond the presence of a thicker early atmosphere and the potentially warmer and wetter conditions of the Noachian period, but rather under the cold and dry climate conditions characteristic of the Hesperian and Amazonian periods of Mars history (Carr and Head, 2010) (though, as noted earlier, these valley network systems may have formed as a result of short-term warming excursions, a possibility not directly considered in our analysis). Therefore, accounting for the formation of
the volcanic valley network systems of the Tharsis region by rainfall precipitation is challenging. This difficulty has led to the alternative suggestion for the formation of these volcanic valley network systems by basal melting of snow and ice deposits in response to elevated geothermal heat fluxes near the edifice summits (Fassett et al., 2006, 2007). The specific objective of this section is to argue for a new, possibly advantageous, surficial glaciovolcanic origin model. We begin by first outlining a number of difficulties faced by the ice sheet basal melting origin model for the volcanic valley networks which motivate the proposal of this new mechanism.

In this assessment, we perform a case study on the Ceraunius Tholus valley network system (Fig. 6c–f) as it poses the most rigorous test of any formation mechanism. This is because the Ceraunius Tholus valley network system exhibits the highest drainage density (~0.15 km/km²) and lowest drainage area (~1200 km², the upslope area between the heads of each valley) of the volcanic valley networks in the Tharsis region (Hynek et al., 2010) and also exhibits an anomalously young Late Hesperian to Early Amazonian age of ~2.96 Ga (Fassett and Head, 2008).

Prior investigations have suggested the formation of the Ceraunius Tholus valley network systems by basal melting at or near the summit of Fig. 6. (a. & b.) The volcanic valley network systems of the Tharsis regions concentrated on the flanks of shield edifices. The valley networks are mapped as the (a.) blue and (b.) white lines. (c. & d.) The Ceraunius Tholus valley network system mapped as (c.) blue and (d. & e.) black lines with context provided in panel (a.). (f.) MOLA Topographic profile, from (d.), across the long axis of the Ceraunius Tholus edifice. (g.) A possible glaciovolcanic scenario for the origin of the valley networks which are carved into the flanks of the Ceraunius Tholus edifice involving the eruption of extrusive lavas atop surface snow and ice near the summit of the edifice with subsequent melting and discharge. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)
the edifice in response to elevated peak geothermal heat fluxes of ~220 mW/m² produced by an active magma chamber (Fassett et al., 2006, 2007). At this estimated peak geothermal heat flux, snow and ice deposits of ~400–600 m thick would be required to initiate basal melting for a reasonable surface snow/ice thermal conductivity value of 2 W/m K (Cassanelli and Head, 2015) and predicted (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016; Palumbo et al., 2018) local mean annual surface temperatures of ~230 K during the Noachian–Early Hesperian to ~205 K in the Late-Hesperian–Amazonian periods (Table 2), problems remain regarding the formation of the valley networks by basal melting of surface snow and ice. (1) There is no characteristic morphologic evidence of wet-based/polythermal glacialization on the flanks of Ceraunius Tholus that would have likely been produced by basal melting of the requisite ~400–600 m thick ice sheets. Evidence of wet-based glacialization might not have been produced by basal melting of thinner snow packs, but these would likely be too thin to have undergone basal melting. (2) The meltwater production rates produced by bottom-up ice sheet basal melting are generally low (Cassanelli and Head, 2016), making the formation of valley networks by this mechanism alone difficult. The rate of meltwater production per unit area of ice undergoing basal melting can be estimated in a simple manner with the following relationship (Russell and Head, 2007):

\[
R = \frac{G}{L_f} - \frac{K_A T}{L_f z}
\]

(3)

where \( R \) is the melting rate (kg/m² s), \( L_f \) is the latent heat of fusion of ice (3.35 × 10⁶ J/kg), \( z \) is the thickness of surface snow/ice (m), and the remaining parameters are as previously defined. Ice sheets accumulating on the Ceraunius Tholus edifice in the presence of an ambient 55 mW/m² geothermal heat flux (an average Late Noachian value; Solomon et al., 2005) and mean annual surface temperatures of 205 and 230 K (representing the Amazonian and Noachian climate conditions; Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016; Palumbo et al., 2018), would reach thermal equilibrium at thicknesses of ~2 and 1.5 km, respectively (assuming a constant ice thermal conductivity of 2 W/m K; Cassanelli and Head, 2015). If these ice sheets were then exposed to the peak, magmatically-enhanced geothermal heat flux of 220 mW/m² at the summit of Ceraunius Tholus (Fassett and Head, 2007), basal melting rates would be ~4.5–4.9 × 10⁶ kg/m² s (assuming a constant ice thermal conductivity of 2 W/m K; Cassanelli and Head, 2015). Applied over the ~1200 km² summit drainage area of Ceraunius Tholus (the total area above the heads of the individual system valleys), these melting rates yield total melt production rates of ~0.6 m³/s. These low total meltwater production rates, generated over an area of 1200 km², are not likely to exceed the infiltration capacity of the substrate or produce substantial erosion of the surface (Cassanelli and Head, 2016, 2018a, 2019) and are insufficient to supply the discharge needed to form the valley networks of Ceraunius Tholus, the single largest of which is estimated to have involved peak discharges of 4 × 10⁶ to 6 × 10⁶ m³/s (Fassett and Head, 2007). Therefore, the formation of the Ceraunius Tholus valley networks by meltwater production from basal melting of surface snow and ice requires the collection and release of accumulated meltwater reservoirs to produce elevated discharge. This difficulty was previously addressed by invoking the formation and catastrophic drainage of a Ceraunius Tholus caldera lake at elevated discharge rates (Fassett and Head, 2007).

A possible alternative surficial glaciovolcanic formation mechanism for the Ceraunius Tholus valley networks is the direct top-down heating and melting of surface snow and ice by supra-ice emplacement of extrusive volcanic lava flows (Fig. 6g) (Cassanelli and Head, 2016, 2018a, 2018b). In this origin scenario, lava flows emplaced atop surface snow and ice can lead to volumetrically substantial meltwater production at rates substantially greater than by bottom-up basal melting (Cassanelli and Head, 2016). The emplacement of lava flows atop ice sheets accumulated on the flanks of the Ceraunius Tholus edifice could have occurred by either: (1) erupted volcanic material and tephra falling and accumulating atop the ice with potential for subsequent congealing and flow (if insufficiently cooled during emplacement), and (2) direct advancement of erupted lava flows onto the surrounding ice (at lower elevations) as has been observed in some glaciated volcanic terrains on Earth (Edwards et al., 2012, 2014, 2015). Observations of supra-ice lava flow emplacement on Earth (Edwards et al., 2012, 2015) have documented little recognizable post-emplacement evidence to specifically indicate supraglacial lava emplacement. Therefore there is no characteristic morphologic signature of this surficial glaciovolcanic process on the flanks of Ceraunius Tholus does not preclude the operation of this mechanism. Assuming a typical Tharsis region individual shield lava flow thickness of ~90 m (Table 2), earlier models of supra-ice lava emplacement (Cassanelli and Head, 2016, 2018a) suggest peak melting rates on the order of 10⁻⁷ kg/m² s. Applied over the 1200 km² drainage area, this melting rate yields a total meltwater production rate on the order of 10⁶ m³/s, several orders of magnitude greater than that basal melting. While the discharge provided by top-down melting greatly exceeds that of basal melting, reaching the peak discharges required to form the channels incised into the flanks of Ceraunius Tholus (maximally, ~4 × 10⁶ to 6 × 10⁶ m³/s; Fassett and Head, 2007) still requires meltwater accumulation and release. A number of possible meltwater confinement and release mechanisms can occur as a result of glaciovolcanic processes leading to sufficiently elevated discharges. These mechanisms are reviewed and assessed in greater detail by (Cassanelli and Head, 2018a, b), include the formation and drainage of surficial supraenglacial meltwater reservoirs. The formation and drainage of a Ceraunius Tholus caldera lake falls under this broad category of mechanisms, and has previously been demonstrated (Fassett and Head, 2007) to supply sufficient discharges and volumes to account for the formation of the Ceraunius Tholus valley networks.

**Summary:** The possible surficial glaciovolcanic formation of the volcanic valley networks carved into the flanks of the shield volcanoes within the Tharsis region was evaluated through a case study of the Ceraunius Tholus valley network system (Fig. 6). Crater count age dating indicates formation of this system took place during the Late Hesperian to Early Amazonian, ~2.96 Ga, a time dominated by cold and dry surface conditions. Due to this relatively young age and the elevations of the host terrain, accounting for the formation of the volcanic valley network systems of the Tharsis region by rainfall precipitation is challenging. As a result of this difficulty, the formation of the Ceraunius Tholus valley network system has previously been attributed to basal melting of surface snow and ice in response to elevated geothermal heat fluxes produced by magmatic activity in the edifice. For a predicted range of elevated geothermal heat fluxes, basal melting of surface snow and ice requires deposits accumulated on thicknesses of ~400–600 m. However, no characteristic evidence of wet-based glaciation, expected for deposits of this thickness undergoing basal melting, is preserved on Ceraunius Tholus. Moreover, meltwater produced by bottom-up basal melting is generated at a very limited rate and tends to infiltrate into the subsurface. As a result, sustained timescales and special conditions are required to allow meltwater accumulation and release to provide sufficient discharges to erode the Ceraunius Tholus valley network system. As an alternative, we explored an origin of the Ceraunius Tholus valley network system of surficial glaciovolcanic mechanism of surficial melting of surface snow and ice by supra-ice lava flow emplacement (Fig. 6g). This mechanism is better able to account for the formation of the valley network system by: (1) providing greatly elevated meltwater production rates, and (2) producing no morphologic signature of wet-based glaciation because the surface snow and ice remains cold-based, thereby also preventing major meltwater infiltration. While the surficial glaciovolcanic mechanism is able to provide increased meltwater production rates, accumulation and release is likely still required to...
supply adequate discharge to erode some of the Ceraunius Tholus system valleys, most plausibly by the formation and drainage of a caldera lake.

8. Outflow channels

The outflow channels of Mars are enormous erosional channels interpreted to have formed through catastrophic fluvial/diluvial activity (Baker and Milton, 1974; Carr, 1979, 1996; Baker, 1982; Baker et al., 1992; Clifford, 1993; Komatsu and Baker, 1997; Head et al., 2003b; Hanna and Phillips, 2005; Andrews-Hanna and Phillips, 2007; Pacifici et al., 2009). The outflow channels are distinct from the other main group of fluvial features on Mars, the valley networks, in their size, age, and geomorphology. The outflow channels are significantly larger than the valley networks, formed generally later in martian history during the Hesperian and Amazonian periods, originate from discrete point sources as opposed to branching tributary networks, and contain abundant bedform morphology largely absent in the valley network systems (Baker and Milton, 1974; Sharp and Malin, 1975; Masursky et al., 1977; Carr, 1979, 1996, 2012; Carr and Clow, 1981; Baker, 1982; Baker et al., 1992; Clifford, 1993; Carr and Malin, 2000; Wilson et al., 2004; Coleman and Baker, 2009). The formation of the martian outflow channels has often been attributed to catastrophic outbursts of groundwater released from confined aquifers (Baker and Milton, 1974; Carr, 1979, 1996, 2002; Baker, 1982; Baker et al., 1992; Clifford, 1993; Head et al., 2003b; Clifford, 1995; Carr, 1979, 1996, 2002; Head et al., 2003b; Carr, 1979, 1996, 2002; Hanna and Phillips, 2005; Wang et al., 2006) (along with a range of other mechanisms not considered here, but more thoroughly outlined in other contributions; e.g. Baker, 1985; Baker et al., 2015). Mechanisms proposed to have been responsible for the release of pressurized groundwater to the surface include: intrusive volcanism and dike emplacement (Head et al., 2003b; Leask et al., 2007b), tectonic pressurization and fracturing (Hanna and Phillips, 2006), over-pressurization (beyond the confining lithostatic pressure) by downward cryosphere advancement (Hanna and Phillips, 2005; Wang et al., 2006) or excess hydraulic head (Marra et al., 2014), and the collapse of confining overburden materials (Rodriguez et al., 2005, 2015, 2016).

The Tharsis region is the host of 6 such outflow channels (Fig. 1a) (Tanaka et al., 2014) which can be broadly classified into those sourced by fractures/fossae and those sourced by basins/chaotic terrain. Here we specifically examine Mangala Valles (Figs. 1a, 7a and 7b) (Zimbelman et al., 1992; Wilson and Head, 2004; Hanna and Phillips, 2006; Leask et al., 2007a, 2007b; Bargery and Wilson, 2011; Keske et al., 2015), the only clear example of a fracture/fossae-sourced outflow channel in the Tharsis region, and Kasei Valles (Figs. 1a, 8a and 8b) (Baker and Kochel, 1993).
Mangala Fossa (part of the larger Memnonia Fossae graben system) near comprised of a series of channels and valleys which emanate from the Valles out to a depth of ~150 m within the surrounding terrain (Mangala Valles system at the Mangala Fossa is ~10 km wide and incised contained within have been interpreted (water at a peak discharge rate of ~10$^3$ m$^3$/s), hydraulic estimates (North where it drains into Amazonis Planitia (out)). Keszthelyi, 2014). 8.1. Fracture/fossae-sourced outflow channels: Mangala Valles

Mangala Valles (Fig. 7a and b) is a ~900 km outflow channel system comprised of a series of channels and valleys which emanate from the Mangala Fossa (part of the larger Memnonia Fossae graben system) near the southwest margin of the Tharsis region (Fig. 1a). The head of the Mangala Valles system at the Mangala Fossa is ~10 km wide and incised to a depth of ~150 m within the surrounding terrain (Fig. 7a and b). The outflow channel gradually widens and shallows downstream toward the North where it drains into Amazonis Planitia (Fig. 7a and b). Paleohydraulic estimates (Leask et al., 2007a, 2007b) indicate that the Mangala Valles outflow channel was formed by a minimum of ~15,000 km$^3$ of water at a peak discharge rate of ~10$^3$ m$^3$/s. The geomorphology of the Mangala Valles outflow channel system and the numerous subchannels contained within have been interpreted (Leask et al., 2007a, 2007b) to suggest that, throughout the majority of the duration of the formative flood, the discharge was sustained near the estimated peak values (~10$^3$ m$^3$/s), implying a total cumulative formation timescale of ~17 days. Detailed mapping and crater age dating efforts (Keske et al., 2015) suggest that the formation history of Mangala Valles involved at least three major volcanic events at ~0.7–1, 0.4, and 0.35 Ga, and two major fluvial events at ~0.7–0.8, and 0.35–0.4 Ga, indicating a relatively young Late Amazonian system age.

Due to the characteristics of the Mangala Valles outflow channel system outlined above, accounting for the formation of the system specifically by the groundwater outburst model faces several difficulties as indicated by various lines of model-dependent evidence: (1) A liquid groundwater system must have been preserved at great depth beneath a very thick cryosphere into the Late Amazonian against efficient vapor diffusive loss to the cryosphere (Clifford, 1991). (2) The groundwater system must have been supplied by a source of active recharge to resupply the aquifer to prevent depletion and to maintain sufficient hydraulic head to fuel discharge (the most plausible source being the Tharsis Rise, still unlikely late in the Amazonian period near the time of formation of the Mangala Valles outflow channel; Russell and Head, 2007; Cassanelli et al., 2015). (3) The groundwater system must have been saturated to the base of the cryosphere to allow confinement and pressurization, potentially requiring significant volumes of liquid water (Clifford, 1993). (4) Dike emplacement (Wilson and Head, 2002b) or

![Image](image-url)
tectonic extension must have produced and maintained an open conduit across the entire ~5–10 km thick cryosphere (Clifford et al., 2010) to allow water outflow to the surface. (5) In order to supply the discharges estimated for the Mangala Valles system the permeability of the aquifer from which the floods originated must have been unreasonably large, far exceeding those common of terrestrial aquifers at similar depths and scales (Head et al., 2003b). (6) Groundwater models predict cumulative erosion of the channel by a series of many smaller individual flooding events (Hanna and Phillips, 2006), inconsistent with the channel geomorphology indicating that the peak flux was sustained throughout most of the duration of channel formation (Leask et al., 2007a, 2007b), implying the formation of the channel by a single, or small number of large floods.

Given these challenges to the traditional groundwater origin model for Mangala Valles, we explore an alternative possible surficial glaciovolcanic formation model (Cassanelli and Head, 2016, 2018a, 2018b) involving supra-ice lava flow emplacement from Mangala Fossa and large-scale lava-ice interactions (Fig. 7c and d) (similar to that previously applied to the Reull Valles channel in Hesperia Planum and the Athabasca Valles channel in Elysium Planitia; Cassanelli and Head, 2018a, 2018b). The formation of outflow channels by this glaciovolcanic mechanism (Fig. 7c and d) is predicted to proceed by meltwater generation and release from both direct top-down contact heating and melting, and by insulation-induced bottom-up heating and melting (Cassanelli and Head, 2016, 2018a, 2018b). Here we disregard bottom-up heating and melting due to the very significant thicknesses of lava and ice required to initiate bottom-up melting, the low rates of meltwater production, and the tendency of bottom-up generated meltwater to infiltrate into the subsurface (Cassanelli et al., 2015; Cassanelli and Head, 2015, 2016).

In order to determine the viability of this glaciovolcanic formation model (Fig. 7c and d), we estimate the volumes and discharges of water that can be produced by the proposed supra-ice lava flow emplacement mechanism. The volumes and discharges of meltwater that can be produced by this glaciovolcanic mechanism, involving supra-ice lava flow emplacement and top-down contact heating and melting (as proposed in the ice sheet lava heating and loading model; Cassanelli and Head, 2016, 2018a, 2018b), depends upon a several factors including: (1) the thickness of surface snow/ice, (2) the thickness of the individual accumulating lava flows, (3) the total thickness of lava accumulated, and (4) the area over which top-down heating and melting occurs (constrained by the extent of surface ice and lava flows). During the timeframe of outflow formation (Late Amazonian; Keske et al., 2015), surficial snow and ice deposits in the vicinity of the Mangala Valles channel head are likely to have been related to either the latitude-dependent mantle, or regional ice sheets accumulated at increased planetary obliquity (see Section 3). As a result, the likely range of possible surface ice thicknesses is ~1–100 m (Section 3, Table 2). Volcanism sourced by the Mangala Fossa fracture is interpreted to have involved voluminous, low-viscosity lava flows (Keske et al., 2015), characteristic of flood basalt or plains volcanism. Therefore, the likely thicknesses and extents of individual lava flows emplaced within the Mangala Valles head region are ~20 m and 10^5 km^2, respectively (Section 2, Table 1). With a typical individual lava flow thickness of ~20 m, complete melting of the thickest surface ice likely to have been present at the Mangala Valles head, ~100 m, is predicted to require the supra-ice emplacement of only two lava flows (Cassanelli and Head, 2016, 2018b). Consequently, the total volume of meltwater which can be supplied by glaciovolcanic interactions within the Mangala Valles head region is likely to be limited by the amount of surface ice present. Therefore, we assume any surface ice present within the region to have undergone complete melting in response to the emplacement of the volcanic plains, and evaluate the melting area required to supply the volume of water needed to form Mangala Valles over the plausible range of surface ice thicknesses from 1 to 100 m (Section 3, Table 2). In order to assess the discharge rates which could be produced by supra-ice lava flow emplacement in the Mangala Valles head region, we employ results from previous lava-ice interaction thermal models (Cassanelli and Head, 2016, 2018a, 2018b) for the melting rates produced during the emplacement of a sequence of two 20 m thick lava flows. Results of prior modeling indicate peak top-down melting rates of ~10^-3 kg/m^2s and average rates of ~10^-4 kg/m^2s over the course of lava flow emplacement and heat transfer to underlying ice.

We find that, for a range of surface ice thickness from 1 to 100 m, a total area of melting of ~10^5–10^6 km^2 is required to supply the minimum 15,000 km^3 of water to carve the Mangala Valles outflow channel (Fig. 7e) while an area of ~10^5–10^6 km^2 is required to supply the necessary discharge (10^4 m^3/s) (Fig. 7f). These required areas of melting are prohibitively large given the ~10^5 km^2 extent of the volcanic plains surrounding the head of Mangala Valles. Therefore, in order for the Mangala Valles outflow channel to have been produced by surficial glaciovolcanic interactions and supra-ice lava flow emplacement (Fig. 7c and d), multiple events of meltwater generation, accumulation, and release are needed to generate the necessary volumes and discharges (as has been found in the assessment of the formation of other outflow channels by this mechanism; Cassanelli and Head, 2018a, 2018b). A multitude of processes can occur in conjunction with surficial glaciovolcanic interactions and supra-ice lava flow emplacement to result in the confinement and release of meltwater in bursts at greatly elevated discharge rates. These mechanisms and conditions are reviewed and assessed in further detail by Cassanelli and Head (2018a, b). The most likely of these processes to have occurred under the conditions relevant to the Mangala Valles system and the surficial lava-ice interaction scenario we investigate, given the regional extent of the surface ice and the supraglacial lava flow emplacement geometry being assessed, are: (1) supra/englacial meltwater accumulation confined by an ice dam, with release triggered by flotation of the confining ice allowing drainage beneath the dam (Fig. 7c and d) (much like the processes responsible for the generation of glacial lake outburst floods on Earth, such as the glacial Lake Agassiz events; Clarke et al., 2004), and (2) supra/englacial meltwater accumulation retained by the ice surface topography, released by subsequent top-down supraglacial overflow (Smellie, 2006) leading to the erosion (by both mechanical and thermal processes; Marston, 1983; Russell et al., 2001) of a channel in the confining ice (similar to processes observed in some Icelandic jökulhaip events; Russell et al., 2001, Smellie, 2006) (Fig. 7c and d).

Utilizing simple analyses of these meltwater accumulation and release processes described in prior work (Cassanelli and Head, 2018b), we find that the fluxes needed to supply the Mangala Valles could be provided by outflow beneath a confining ice dam through a conduit ~30 km in width, and by a supraglacial sheet flow ~100 km in width, narrowing to ~2 km if a supraglacial channel 50 m in depth were eroded into the ice sheet surface. Therefore, given the ~10 km length of the Mangala Valles channel head, these processes are broadly consistent with the geomorphology of the system (if drainage in the supraglacial scenario was rapidly concentrated into a narrower channel). In order to supply the requisite minimum volume of 15,000 km^3 by surface ice melting over the ~10^5 km^2 area of volcanic plains around the Mangala Valles head, a total of 15–150 melting, accumulation, and release events would need to occur for surface ice thicknesses of 10–100 m. This estimated range of total flooding events required to form Mangala Valles falls within the recurrence range of floods produced by similar processes on Earth (Baker and Bunker, 1985; Waitt, 1985; Benito and O’Connor, 2002). However, the multitude of flooding events required by this mechanism are not entirely consistent with prior geomorphologic assessments suggesting formation of the channel through a single, or small number of large floods (Leask et al., 2007a, b), a difficulty requiring further investigation.

Summary In order to assess the fossae/fracture-sourced outflow channels of the Tharsis region we performed a case study of the Mangala Valles system, the only clear example of this class of outflow channels in the Tharsis region (Figs. 1a and 7). The formation of the Mangala Valles outflow channel has historically been attributed to the catastrophic effusion of groundwater from aquifers confined by an ice-cemented
cryosphere. However, due to the relatively young Late Amazonian age of the channel system (~0.35–0.8 Ga), this canonical groundwater outflow channel formation mechanism is problematic, requiring: (1) Preservation of a liquid groundwater system at great depth beneath a very thick cryosphere into the Late Amazonian against efficient vapor diffusive loss to the cryosphere (Clifford, 1991). (2) A source of active recharge to resupply the aquifer to prevent depletion and maintain sufficient hydraulic head for outflow (the most plausible source being the Tharsis Rise, still unlikely late in the Amazonian period near the time of formation of the Mangala Valles outflow channel). (3) Significant volumes of liquid water to saturate the groundwater system to the base of the cryosphere to allow pressurization. (4) Creation and maintenance of an open conduit across the entire ~5–10 km thick cryosphere to allow water outflow to the surface. (5) Unreasonably high aquifer permeability, far exceeding those common of terrestrial aquifers at similar depths and scales. (6) In some models, cumulative channel erosion by a series of many smaller individual flooding events, inconsistent with the channel geomorphology implying the formation of the channel by a single or small number of large floods. Due to these difficulties with the traditional groundwater origin model for Mangala Valles, we explored an alternative possible surficial glaciovolcanic formation model involving supra-ice emplacement of Mangala Fossa lava flows, with subsequent meltwater accumulation and release by outflow beneath a confining ice dam or supraglacial overflow (Fig. 7). We find that, to a first order, this alternative formation model is able to supply the water volumes and discharges needed to erode the channel while conforming to the geomorphic constraints of the system. Therefore, this glaciovolcanic mechanism provides a viable alternative explanation for the formation of the outflow channel system without the difficult conditions required of the classical groundwater model.

8.2. Basin/chaos-sourced outflow channels: Kasei Valles

Kasei Valles (Fig. 8a and b) is the largest outflow channel on Mars, situated along the eastern boundary of the Tharsis region (Fig. 1a). The channel heads at the Echus Chasma trough of Valles Marineris, extending ~1600 km to the north where it turns east and continues a further 900 km to drain into Chryse Planitia (Fig. 1a). The Kasei Valles outflow channel is estimated to have been eroded by a minimum of ~1.36 × 10^6 km^3 of water. Prior estimates suggest the channel system was formed by peak discharges as high as 10^6–10^7 m^3/s (Carr, 1996), and also involved the occurrence of smaller floods with discharges ranging from ~8 × 10^6 to 2 × 10^6 m^3/s (Williams et al., 2000) as indicated by the morphometry of interior nested channels.

Kasei Valles is historically interpreted (Robinson and Tanaka, 1990; Carr, 1996) to have been sourced from Echus Chasma through the catastrophic effusion of groundwater or release of an impounded lake. More recent morphologic and mapping investigations (Williams et al., 2000; Chapman et al., 2010a, 2010b), however, indicate that the history of the Kasei Valles outflow channel was far more complex. These studies suggest that Kasei Valles (Williams et al., 2000; Chapman et al., 2010a, 2010b) developed gradually through multiple flooding events, from multiple sources, over an extended period of time with interspersed periods of volcanic and fluvo-glacial activity occurring from the Noachian into the Amazonian as recently as ~70 Ma. Specifically, unit geomorphology, stratigraphy, and crater age dating suggest a detailed history of the Kasei Valles system as follows (Chapman et al., 2010a, 2010b): (1) Hesperian flood basalt lavas from the Tharsis rise are emplaced at ~3.7 Ga forming ridged plains in the area. (2) At ~3.6 Ga the first event of fluvo-glacial activity occurred with the erosion of the northern, east-trending section of the channel by Early Hesperian-aged floods sourced from the Tharsis Montes area to the west. (3) Several episodes of volcanic and fluvo-glacial activity occurred at ~3.4 Ga, including a separate event of flooding within the northern, east trending-section of Kasei Valles again sourced from Tharsis region to the west, as well as a period of volcanic activity to the south near Echus Chasma and Syria Planum. Contemporaneously, water being released by numerous channels from the Syria Planum plateau toward Echus Chasma. During this time, the presence of surface ice and ice-rich materials within the Kasei Valles system are suggested by the identification of possible eskers, U-shaped glacial gullies, and glaciovolcanic features such as tindar ridges. (4) Also at ~3.4 Ga, an additional volcanic event resulted in the emplacement of flood lavas within the channel system. (5) From ~1.8 to 1.0 Ga, glacio-fluvial activity continued within the Kasei Valles system with lavas sourced from the Tharsis Rise and Echus Chasma areas developing the north trending portion of the Kasei Valles channel. (6) A prolonged period of volcanic activity in the Late Amazonian lasting until ~89 Ma which involved the emplacement of voluminous lavas sourced from Echus Chasma and the Tharsis Montes. Some of the Amazonian-aged lava flows produced during this period of activity extend up to 2100 km in length and exhibit a platy surface morphology. The platy surface morphology bears resemblance to the play-ridged terrain of lava flows in the Central Elysium Planitia region of Mars which have been suggested to be a possible product of surficial glaciovolcanic interactions and therefore an indication of the presence of surface snow and ice (Cassanelli and Head, 2018b). (7) A final major phase of fluvo-glacial occurred at ~70 Ma involving small floods from the Echus Chasma trough.

This more detailed understanding of the history of the Kasei Valles outflow channel indicates that the formation of the system involved numerous episodic of volcanic, fluvial, and glacial activity sustained over a large portion of Mars history into the very recent Late Amazonian period. As a result of this history of channel development, accounting for the formation of the Kasei Valles system by the canonical groundwater outburst model faces many of the same issues outlined for the formation of the Mangala Valles system (discussed in Section 8 and 8.1). Given these same difficulties, and the history of Kasei Valles formation (Chapman et al., 2010a, 2010b) indicating sustained coincident volcanic, fluvial, and glacial activity, as well as the possible evidence for glaciovolcanic processes, we explore a possible surficial glaciovolcanic origin of the system. In particular, we evaluate the potential for volcanic activity associated with the construction of the Tharsis region to have interacted with surface ice deposits to generate numerous events of meltwater production and runoff which was collected and released from the north-trending portion of Kasei Valles and Echus Chasma to form the outflow channel (Fig. 8c and d).

To determine if the Kasei Valles outflow channel system could have been formed by meltwater derived from surficial glaciovolcanic interactions in the Tharsis region (Fig. 8c and d) we first delineate the total drainage area contributing to the system. The Kasei Valles drainage area was determined using gridded MOLA data at a resolution of ~470 m/pixel, yielding a total contributing drainage area of ~6 × 10^6 km^2 (Fig. 8e and f) (it is worth noting that this drainage area has been derived from modern topography and is likely to have been modified by volcanism, tectonism, and erosion throughout the course of Kasei Valles formation). The drainage basin of the Kasei Valles system delineated through this analysis occupies much of the northeastern Tharsis region (Fig. 8e and f), areas comprised predominantly of plains and flood basalt volcanic deposits (Fig. 2a). Therefore, we assume surficial glaciovolcanic interactions that may have taken place in the Kasei Valles drainage basin to have involved the supra-ice emplacement of lava flows with an average thickness of ~20 m (Table 1). Current understanding suggests prolonged development of the Kasei Valles from the Late Noachian to the very recent Amazonian period. Consequently, the Kasei Valles drainage basin may have hosted regional surface ice deposits over a plausible range of average thicknesses from ~1 to 275 m (Table 2). Given the significant total accumulated thicknesses of lava in this region, on the order of kilometers (Table 1), we assume that any present surface ice within the Kasei Valles drainage basin would have undergone complete melting by surficial glaciovolcanic interactions. Thus, over the total drainage area (~6 × 10^6 km^2; Fig. 8e and f), glaciovolcanic melting of surface ice deposits 1–275 m thick could generate ~5.5 × 10^3–1.5 × 10^6 km^3 of
meltwater. This range of meltwater volumes encompasses the estimated minimum volume of water required to erode the Kasei Valles system, \( \sim 1.36 \times 10^7 \, \text{km}^3 \), which equates to the cumulative melting of \( \sim 250 \, \text{m} \) of ice across the Kasei Valles drainage basin. This total cumulative melt thickness of ice is within the range of thicknesses predicted for surface ice deposits formed within the Tharsis region (Table 2), and could plausibly have occurred through multiple phases of glaciovolcanic melting of relatively thinner surface ice deposits.

Prior analysis (Cassanelli and Head, 2018a, 2018b) of the supra-ice emplacement of \( \sim 20 \, \text{m} \) thick lava flows under relevant martian conditions predicts peak melting rates on the order of \( 10^{-2} \, \text{kg/m}^2\cdot\text{s} \) and average melting rates over the course of two lava flow emplacement events of \( 10^{-3} \, \text{kg/m}^2\cdot\text{s} \). The emplacement of two 20 m thick lava flows is sufficient to melt the thickest surface ice deposits likely to have been present in the Tharsis region throughout the Amazonian and much of the Hesperian period (\( \sim 100 \, \text{m} \); Table 2). However, in the Late Noachian, regional surface ice sheets in the Tharsis regions are predicted (Fastook and Head, 2015) to exhibit average thicknesses of \( \sim 275 \, \text{m} \) (assuming a total available surface/near-surface water inventory on Mars equivalent to that of the present day; Table 2), a greater thickness than models (Cassanelli and Head, 2018a, 2018b) predict could be melted by sequential emplacement of two 20 m lava flows. As a result, complete melting of surface ice of this thickness would require a greater number of lava flow emplacement events and would decrease the predicted average melting rate. For the level of this analysis, however, we utilize the previously determined average melting rate predicted for the emplacement of two 20 m lava flows noting that the average meltwater production rates produced will represent over-estimates in a Late Noachian scenario of complete top-down surface ice melting by glaciovolcanism.

To derive a maximum estimate of the rate of meltwater production that can be produced by glaciovolcanic melting in the Kasei Valles drainage basin, we assume simultaneous glaciovolcanic melting across the basin. Under this maximum assumption, application of the predicted peak and average glaciovolcanic melting rates yields total meltwater production rates of \( \sim 6 \times 10^7 \, \text{m}^3/\text{s} \). It is, however, improbable that glaciovolcanic melting occurred across the entire Kasei Valles drainage basin during any given time period, with melting likely limited by the extent of individual lava flows emplaced within the catchment area. Given a typical individual lava flow area of \( \sim 10^7 \, \text{km}^2 \) in the Kasei Valles drainage basin region (which contains chiefly plains and flood basalt lava flows; Table 1), rates of meltwater production within the catchment are reduced to \( \sim 10^3 \, \text{m}^3/\text{s} \). These values are below the minimum discharge estimated to have been involved in the formation of Kasei Valles, \( \sim 8 \times 10^3 \) to \( 2 \times 10^4 \, \text{m}^3/\text{s} \), therefore requiring meltwater collection into reservoirs with subsequent release at elevated discharges. In the Kasei Valles system, meltwater produced within the drainage basin could plausibly have been collected into reservoirs within the southern, smooth portions of the channel (where substantial channel incision is not apparent) by ice-impoundment and within Echus Chasma (Fig. 8c and d). We define the extent of the possible meltwater reservoir within the north-trending portion of Kasei Valles (Fig. 8c and d) along the \( \sim 500 \, \text{m} \) contour (which very closely follows the edge of the channel) truncated at the juncture between the north-trending and east-trending sections of Kasei Valles where the smooth channel floor morphology transitions to a more highly dissected/incised state. This defined meltwater reservoir extent assumes closure by ice-impoundment along the northern margin of the reservoir at the location indicated in Fig. 8d. Given the width of the defined reservoir, an ice dam of this extent would require an ice dam width a few hundred kilometers in width oriented transversely to the channel system. However, the initial channel width is likely to have been less than that presently observed, thereby reducing the required ice dam width. While complete failure of this large ice dam is unlikely, expansion of an initially localized failure could have supplied water across the broader channel extent. The extent of a meltwater reservoir which could be contained within Echus Chasma is defined by the 3.5 km contour which directly traces the boundary of the trough (Fig. 8c and d). Importantly, drainage and collection of meltwater within these two reservoirs assumes their existence prior to the development of the Kasei Valles system. While the pre-existence of the channel head reservoir is unlikely, it is possible for the Echus Chasma basin to have been formed by tectonic extension with subsequent melt accumulation and overflow leading to initial channel (and head reservoir) development.

The potential volume of meltwater contained within both of these reservoirs is \( \sim 10^7 \, \text{km}^3 \) each, thereby requiring a total of \( \sim 13 \, \text{fill} \) and release events to supply the minimum volume of \( \sim 1.36 \times 10^7 \, \text{km}^3 \) needed to erode the Kasei Valles channel. This estimated number of required flooding events is consistent with the recurrence of floods produced by similar processes observed on Earth (Baker and Bunker, 1985; Waft, 1985; Benito and O’Connor, 2003). Under the possible reservoir geometries examined, the minimum \( \sim 8 \times 10^7 \) to \( 2 \times 10^6 \, \text{m}^3/\text{s} \) discharge inferred (Williams et al., 2000) to have been involved in the formation of Kasei Valles could readily be met given the significant \( 400 \, \text{km} \) width of the possible Kasei Valles channel reservoir (Fig. 8c and d) (refer to discussion on the effect of reservoir outlet geometry on flood discharge contained in Section 8.1), while the adequacy of floods from Echus Chasma to form Kasei Valles has been established in prior studies (Robinson and Tanaka, 1990; Tanaka and Chapman, 1992).

Summary and implications for other basin-sourced channels: Here we examined the formation of the Kasei Valles system (Fig. 8) as an example of the basin-sourced outflow channels in the Tharsis region. We find that the formation of Kasei Valles can be plausibly accounted for by a surficial glaciovolcanic origin involving the generation of meltwater within the Kasei Valles watershed by supra-ice lava flow emplacement with subsequent collection and release from reservoirs formed in Echus Chasma and the southern, north-trending portion of the channel itself (Fig. 8c and d). This formation mechanism is a plausible consequence of the extensive and sustained volcanic and glacial activity in the region and is able to: (1) supply the volumes and discharges of liquid water required for channel formation, (2) account for the distributed sources inferred (Chapman et al., 2010a, 2010b) to have contributed to the system, (3) account for the numerous flooding events suggested by the channel morphology without requiring active groundwater recharge and a recurring mechanism for cryosphere breaching and groundwater release, and (4) provide a source of abundant liquid water over the prolonged Noachian-Amazonian history of channel development (Chapman et al., 2010a, 2010b) under continuously cold climate conditions (Madeleine et al., 2009, 2014; Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016).

In addition to Kasei Valles, a number of other similar basin-sourced outflow channels exist within the Tharsis region, most notably Simud and Tiu Valles (Fig. 1a) which appear to emanate from Valles Marineris. The formation of these outflow channels has previously been attributed to the catastrophic drainage of lakes formed within Valles Marineris (Coleman et al., 2007; Harrison and Chapman, 2008; Warner et al., 2013). The surficial glaciovolcanic origin we explored for Kasei Valles offers a viable mechanism to account for the formation of these liquid water reservoirs within Valles Marineris under the cold climate conditions predicted to characterize the majority of martian history (Madeleine et al., 2009, 2014; Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016). Therefore, the surficial glaciovolcanic outflow channel origin we explore, involving the catastrophic drainage of reservoirs fed by runoff from glaciovolcanic melting of surface ice, offers a plausible explanation for the formation of the basin-sourced Tharsis region outflow channels.

9. Conclusions

The Tharsis region is the largest and most long-lived volcanic province on Mars, with a history of volcanic activity spanning \( \sim 4 \, \text{Gyr} \). Intensive and widespread volcanic activity began in the region during the Noachian period with the construction of the vast Tharsis rise plateau and continued throughout martian history into the recent Late
Amazonian through further plains volcanism and shield development. Throughout this timeframe of volcanic activity, the Tharsis rise is also predicted to have been a location of regional and local ice accumulation under a wide range of climatic conditions, from the relatively warm and wet conditions often suggested for the Noachian period, to the cold and dry conditions which characterize the current climate. As a result, the Tharsis region of Mars is predicted to have been a site of coeval volcanic and glacial activity throughout a majority of the geologic history of the planet, raising the possibility for widespread, sustained glaciovolcanic interactions. Here we assessed the possible role of large-scale surficial glaciovolcanic interactions in the formation of several classes of tectonic and fluvial/diluvial features throughout the Tharsis region.

9.1. Tectonic features

The tectonic features of the Tharsis region assessed in this analysis include the aureole apron units of Olympus Mons (Fig. 3) and the Syria-Thaumasia block/Solis Planum mega-thrust (Fig. 4). Edifice aureoles: We find that a surficial glaciovolcanic origin of the Olympus Mons aureole apron units (Fig. 3) by the sequential emplacement of lava flows (and possible volcanic tephras) emanating from the edifice atop surrounding regional surface ice (Fig. 3c) could plausibly account for the morphologic characteristics of the units. Supra-ice emplacement allows rapid advancement of lava flows over the great distances the aureole units are deposited (~700 km), and could result in the construction of the apron ridges through the formation of distal ramparts due to emplacement deceleration by flow beyond surface ice margins or encountering rampart obstacles from prior lava flows (Fig. 3c). Syria-Thaumasia block/Solis Planum mega-thrust (Fig. 4) involving detachment and deformation along a weak, volcanically-buried ice layer produced by the accumulation of superposed lava flows (similar to previous models invoking detachment and deformation along a weak buried salt layer) (Fig. 4d). Simple thermal analyses indicate that buried ice in the Syria-Thaumasia region would be thermally unstable below ~2.4 km, well below the estimated average ~10 km mega-thrust detachment depth. Consequently, formation of the Syria-Thaumasia block/Solis Planum mega-thrust in this manner is not favored as it requires geologically rapid accumulation of volcanic materials to thicknesses of ~10 km over ~1 Myr to outpace ice removal by bottom-up melting.

9.2. Fluvial/diluvial features

The fluvial/diluvial features of the Tharsis region assessed in this analysis include regional highland valley networks, exposed primarily in the Thaumasia highlands, the younger, more isolated volcanic valley networks incised on the regions shield edifices, and the large outflow channels sourced by both fractures/fossae systems and by basins/chaos terrain. Regional/highland valley networks: Using the Warrego Valles system as a case example, we assessed a possible surficial glaciovolcanic origin of the regional highland valley networks (which, in the Tharsis region, are concentrated along the Thaumasia highlands) (Fig. 5) by dewatering of crust and ice uplifted to the surface by the Solis Planum mega-thrust (Fig. 5e). This mechanism predicts: (1) valley network heads initiating at a distance below the Thaumasia highlands peaks equal to the depth of the ice-melting isotherm below the original surface, (2) a narrow range in valley network head elevations, (3) a contemporaneous age of faulting and fluvial erosion, and (4) characteristic groundwater sapping channel/valley morphology. While the first three of these predicted criteria are met by Warrego Valles, the system does not exhibit characteristic sapping channel morphology and thus a favored origin of the Warrego Valles system is the geothermal melting of surface snow deposits (as proposed by previous investigators).

Volcanic valley networks: Through a case study of the Ceraunius Tholus valley network system (Fig. 6), we explored the possible origin of the volcanic valley networks by surficial glaciovolcanic interactions involving the top-down melting of surface snow and ice by supra-ice lava flow emplacement (Fig. 6g). We find that this mechanism is better able to account for the formation of the volcanic valley network systems than prior ice sheet basal melting models by: (1) providing greatly elevated meltwater production rates, and (2) producing no morphologic signature of wet-based glaciation. While the surficial glaciovolcanic mechanism is able to provide increased meltwater production rates, melt accumulation and release is likely still required to supply adequate discharge to erode some of the Ceraunius Tholus system valleys, most plausibly by the formation and drainage of a caldera lake.

Fossae/fracture-sourced outflow channels: The possible surficial glaciovolcanic origin of the fossae/fracture-sourced outflow channels was assessed through a case study of the Mangala Valles system (Figs. 1a and 7). The formation of the Mangala Valles outflow channel has previously been attributed to the catastrophic effusion of groundwater from aquifers confined by an ice-cemented cryosphere. However, due to the relatively young Late Amazonian age of the channel system (~0.35–0.8 Ga), formation by groundwater outflow is problematic (refer to section 8.1 for further detail). Due to the difficulties associated with the traditional groundwater origin model for Mangala Valles, we explored an alternative possible surficial glaciovolcanic formation model (Fig. 7c and d) involving supra-ice emplacement of Mangala Fossa lava flows, with subsequent meltwater accumulation and release by outflow beneath a confining ice dam or supraglacial overflow. We find that, to a first order, this alternative formation model is able to supply the water volumes and discharges needed to erode the channel while conforming to the geomorphic constraints of the system. Therefore, this glaciovolcanic mechanism provides a viable alternative to explain the formation of the outflow channel system without the difficult conditions required of the groundwater outflow model.

Basin/chaos-sourced outflow channels: Our final analysis focused on the basin/chaos-sourced outflow channels of the Tharsis region, utilizing Kasei Valles (Fig. 8) as a case example. Kasei Valles is the largest outflow channel on Mars and is interpreted to have developed through numerous episodes of volcanic, fluvial, and glacial activity from multiple sources, over an extended period of time from the Late Noachian to the very recent Late Amazonian. As a result of this history of channel development, accounting for the formation of the Kasei Valles system by the classical groundwater outburst model faces many of the same issues as noted for the Mangala Valles system. Given these same difficulties, and the history of Kasei Valles formation indicating sustained coincident volcanic, fluvial, and glacial activity, as well as evidence for glaciovolcanic processes, we explored a possible surficial glaciovolcanic origin of the system (Fig. 8c and d). In particular, we evaluated the potential for volcanic activity associated with the construction of the Tharsis region to have interacted with surface ice deposits to generate numerous events of meltwater production and runoff which was collected and released from the north-trending portion of Kasei Valles and from Echus Chasma to form the outflow channel (Fig. 8c and d). This formation mechanism (Fig. 8c and d) is a plausible consequence of the extensive and sustained volcanic and glacial activity in the region and is able to: (1) supply the volumes and discharges of liquid water required for channel formation, (2) account for the distributed sources inferred to have contributed to the system, (3) account for the numerous flooding events suggested by the channel morphology, and (4) provide a source of abundant liquid water over the prolonged Noachian-Amazonian history of channel development. Under continuous cold climatic conditions, therefore, the surficial glaciovolcanic outflow channel formation model we explored (Fig. 8c and d) offers a potentially advantageous explanation for the origin of the basin-sourced outflow channels of the Tharsis region.

To conclude, we find that glaciovolcanic processes resulting from the extended and overlapping histories of volcanic and glacial activity in the Tharsis region may have contributed to the formation of many of the tectonic and fluvial features in the region offering advantages over previously applied formation models.


