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VOLCANO-ICE INTERACTION AS A MICROBIAL HABITAT ON EARTH AND MARS

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Abstract

Volcano-ice interaction has been a widespread geologic process on Earth that continues to occur to the present day. The interaction between volcanic activity and ice can generate substantial quantities of liquid water, together with steep thermal and geochemical gradients typical of hydrothermal systems. Environments available for microbial colonization within glaciovolcanic systems are wide-ranging and include the basaltic lava edifice, subglacial caldera meltwater lakes, glacier caves, and subsurface hydrothermal systems. There is widespread evidence of putative volcano – ice interaction on Mars throughout its history and at a range of latitudes. Therefore, it is possible that life on Mars may have exploited these habitats, much in the same way as has been observed on

Earth. The sedimentary and mineralogical deposits resulting from volcano-ice interaction have the potential to preserve evidence of any indigenous microbial populations. These include jökulhlaup (subglacial outflow) sedimentary deposits, hydrothermal mineral deposits, basaltic lava flows, and subglacial lacustrine deposits. Here, we briefly review the evidence for volcano-ice interactions on Mars and discuss the geomicrobiology of volcano-ice habitats on Earth. In addition, we explore the potential for the detection of these environments on Mars and any biosignatures these deposits may contain.

1. Introduction

The detection of extraterrestrial life has become a major goal in modern space exploration, with Mars in particular being recognized as an appropriate target. The search for life on Mars during the past few decades has been significantly aided by research into life within martian analogue environments on Earth (e.g., Cavicchioli 2002). Environments that have received considerable attention as proxies for past or present martian habitats include the Antarctic Dry Valleys (Wierzchos *et al*., 2005; Walker & Pace 2007), the Atacama Desert (Navarro-Gonzalez *et al*., 2003), evaporite environments (Rothschild 1990; Edwards *et al*., 2006), and permafrost (Gilichinsky *et al*., 2007). These environments have shown an array of resilient microbial communities that thrive under harsh environmental conditions and provide a framework from which to develop life-detection strategies for Mars.

The martian crust is predominantly igneous in nature and ranges from basaltic to andesitic in composition (McSween *et al.,* 2009). Therefore, it is imperative to understand martian volcanic environments in terms of their habitability and potential for microbial colonisation. In particular, where volcanism interacts with liquid water, there is the potential to support life, as seen on Earth (e.g., Boston *et al*., 1992). Liquid water is unstable at the martian surface today and has been for a considerable part of its history. Water currently exists as a largely continuous global cryosphere within, or below, the regolith (Clifford 1993; Kuzmin 2005; Clifford 2010), with the largest known reservoirs of water today frozen at the poles (Carr 1987; Jakosky & Phillips 2001; Hvidberg 2005; Clifford 2010) and within a latitude dependent mantle (Levy *et al*. 2010). Differences in localized lithospheric heat flow and crustal thermal properties are likely to result in spatial variation in the cryosphere thickness (Clifford 2010). This cryosphere, coupled with volcanic activity, has the potential to produce several kinds of environments for life on Mars with a wide range of thermal and chemical conditions, particularly through the generation of hydrothermal systems (Chapman *et al.,* 2000; Head & Wilson 2002; Schulze-Makuch *et al*., 2007). It has previously been suggested that where regions of volcano – ice interactions are found, suitable sites may exist to search for evidence of martian life (Boston *et al*., 1992; Farmer 1996; Hovius *et al*., 2008; Gulick 1998; Payne & Farmer 2001). Here, we review glaciovolcanism on both Earth and Mars within the context of assessing the range of microbial habitats that exist through volcano – ice interaction, as well as the potential for biosignature preservation within these environments.

2. Glaciovolcanism on Earth

The interaction between volcanism and ice on Earth is on-going and widespread. *Glaciovolcanism* specifically describes any interaction between volcanism and ice, including glaciers, snow, firn (recrystallised snow), and ground ice (Smellie 2006; 2007). Chapman *et al.,* (2000) described three types of volcano – ice interaction: "Type 1" is an alpine interaction with volcano summit snow and valley glaciers; "Type 2" a continental ice sheet/glacier interaction; and "Type 3" involves the interaction with lava and surface ground ice. Type 2 includes subglacial volcanism in its true definition, which is specific to volcanic eruptions beneath thick glaciers and ice sheets (Smellie 2006), and on Earth subglacial volcanism is a common feature of volcanically active, high latitude terrains. Examples of widespread subglacial volcanism today include those found in Iceland (Gudmundsson *et al.* 1997; Bourgeois *et al*., 1998), British Columbia (Edwards *et al*., 2002), and Antarctica (Smellie & Skilling 1994; Smellie *et al.,* 2008). In Iceland in particular, many volcanoes are situated beneath the

Vatnajokull ice cap (Chapman *et al,*. 2000), some of which maintain subglacial meltwater lakes (see Section 4.1). Geomorphological products indicative of basaltic subglacial volcanism include tuyas (Figure 1a) and moberg/hyaloclastite ridges. Tuyas form as a result of central vent eruptions into an overlying thick ice sheet (Bourgeois *et al*., 1998), while hyaloclastite ridges result from a series of fissure eruptions beneath ice, which form long ridges that follow the strike of the rift. These eruptive features display a distinctive elevated topography in contrast to the surrounding terrain due to the restrictive role of the ice into which the lava was erupted, preventing the lateral flow of lava away from the eruptive center. Subsequent retreat of ice reveals these distinctive volcanic landforms (Figures 1a and b).

During an eruption, conductive heat flow melts the surrounding ice, while the low temperatures of the ice begin to solidify the magma under high water pressure, typically forming effusive pillow lava formations (Jakobsson & Gudmundsson 2008). Convection also plays a large role in the transfer of heat from the magma body to the overlying ice (Höskuldsson & Sparks 1997), which produces a growing zone of meltwater. Over time, a subglacial edifice can grow within this meltwater "lens'"(Figure 2a), broadly consisting initially of pillow basalts (Figure 1b), and then hyaloclastite beds and palagonite tuffs as the confining pressure reduces and the eruption becomes more explosive (Smellie & Skilling 1994;

Jakobsson & Gudmundsson 2008). If the edifice becomes large enough to break through the ice, a cap rock of horizontal subaerial lava may be deposited (Figure 2b).

When the eruption is smaller, perhaps the result of a fissure, entirely subglacial hyaloclastite ridges or pillow mounds (Figure 1b) will form. These edifices will remain beneath the glacier until exposed and eroded. As magma flow diminishes, the growth of the lava edifice ceases, but the overlying ice continues to melt due to the convective transfer of heat through the liquid water interface between the magma and the ice (Head & Wilson 2002). Figure 2 summarizes these processes and associated environments. Additionally, subglacial hydrothermal systems may continually melt the base of the glacier, which would sustain a subglacial caldera lake between eruptions (Björnsson 2002). Such caldera lakes, and meltwater generated during an eruption, are typically catastrophically released as jökulhaups (Roberts 2005; Figure 2c). In the case of Eyjafjallajökull - the Icelandic volcano that erupted in April 2010 - the eruption was initially subglacial beneath the small ice cap, but after a few hours this changed to phreatomagmatic activity coupled with meltwater discharge, with the lava eventually emerging from the eruption site ~ 1 week after the initial eruption, having melted through the ice (Gudmundsson *et al.*, 2010).

3. Glaciovolcanism on Mars

3.1 Volcanism and the cryosphere

Volcanism on Mars has occurred throughout its history (Fassett & Head 2010); evidence of volcanic activity (e.g., lava flows) spans from the Noachian right up to the very recent Amazonian (Hartmann, 2005; Werner 2009). Indeed, at specific localities such as Olympus Mons and Hectes Tholes, the ages of lava flows span ~80% of martian history (Neukum *et al.* 2004). Evidence of past glaciation is also widespread, both spatially and temporally, with evidence of large polar ice caps in the Hesperian and low-latitude Amazonian glaciations (Kargel & Strom 1992; Carr & Head 2010). Likewise, the subsurface cryosphere has been a long-lived and widely distributed source of ice (Clifford *et al.,* 2010). Therefore, it is highly probable that these major processes have interacted in the past (Chapman *et al.,* 2000; Head & Wilson 2002) and may even continue to do so today deep within the subsurface (Schulze-Makuch *et al.,* 2007). As a result, volcano – ice interaction may represent an environment that has persisted over a significant part of martian history.

3.2 Glaciovolcanism through Martian history

The processes and occurrences of volcano – ice interactions on Mars have been reviewed and discussed in depth by Chapman *et al.,* (2000), Head & Wilson (2002; 2007), and Chapman (2003), and involve the emplacement of sills, dykes, lava flows, and large magma bodies into cryospheric permafrost or into an existing ice cap. It has been suggested that glaciovolcanic activity has occurred throughout the history of Mars (Head & Wilson 2007; Chapman *et al*., 2000), and there are many topographic features on Mars that have been interpreted as products of volcano-ice interaction (Table 1, Figure 3). Allen (1979) identified many putative subglacial volcanoes in both the northern plains and near the south polar cap of Mars. Since then, more candidate subglacial volcanoes and regions of volcano – ice interaction have been identified (examples summarized in Table 1 and their locations shown in Figure 3). These include flat topped tuyas/edifices (Figure 5a; Ghatan & Head 2002; Head & Wilson 2007), lava ridges/dykes (Figure 5b; Ghatan *et al*., 2003; Head & Wilson 2007), pseudocraters (Figure 5f; Lanagan *et al*. 2001; Fagents & Thordarson 2007), major outflow channels typical of glacial outburst floods caused by geothermally melted ice (jökulhlaups) (Figures 5c, d, Figure 5b)**;** Head & Wilson 2002), and marginal drainage channels (Head & Wilson 2007). Jökulhlaups in particular have been proposed as an explanation for some of the numerous outflow channels and valleys (Figures 5d & 6b) apparently carved by liquid water (Carr & Head 2003; Fassett & Head 2007; Baker 2001; Rice and Edgett 1997), with the large flood deposits and catastrophic outwash plains identified on Mars as comparable to those generated by Icelandic jökulhlaups (Hovius *et al.,* 2008; Fishbaugh & Head 2002). As illustrated by Gulick (1998), much of the fluvial erosion on Mars is spatially and temporally related to volcanic activity. Baker (2001) and Burr *et al.* (2002) also observed that catastrophic flood channels and volcanic lava flows are closely associated in the Cerberus Rupes and Marte Vallis region. This further demonstrates the potential importance of volcanism in the generation of liquid water available to life on Mars.

It has been widely suggested that Noachian Mars (Figure 4) represented a warmer, and perhaps more clement, period of martian history (e.g., Craddock & Howard 2002; Chevrier *et al.,* 2007; McKeown *et al.,* 2009) that was followed by a change to acidic, cold, and desiccating surface conditions at the beginning of the Hesperian (Bibring *et al.* 2006). If true, and if life did indeed evolve in the very early history of Mars, glaciovolcanic environments during the Hesperian and Amazonian may have provided a subsurface refuge as an alternative to the increasingly hostile surface conditions. Here, both Hesperian and Amazonian examples of glaciovolcanism are described (see Table 1, and Figures 3, 5, and 6 for localitions and images).

At the Hesperian volcano Ceraunius Tholus (see map in Figure 3), there is clear evidence for drainage valleys and a depositional fan originating from the caldera rim (Figure 6d). The geometry of this rim is such that it would favor the accumulation of meltwater sourced from the geothermal melting of snowpack at the summit of the volcano (Fassett & Head 2007). Likewise, the south polar Dorsa Argentea Formation (Figure 6a) has been interpreted several times to be an area of multiple subglacial volcanic eruptions with associated meltwater accumulation and drainage (Ghatan & Head 2002; 2003; Milkovich *et al.* 2002; Dickson & Head 2006). This Hesperian-aged, volatile-rich deposit displays evidence for significant melting (e.g., channels, eskers, Figure 6c), with valleys interpreted to have been outflow regions that drained significant quantities of meltwater from a thinning southern circumpolar ice sheet, induced by volcanic activity (Ghatan & Head 2004; Milkovich *et al.* 2002; Head & Pratt 2001). Finally, interior layered deposits (ILDs; Figure 5e) within the late-Hesperian Juventae Chasma have been interpreted by some workers to be the result of sub-ice volcanism (Chapman & Tanaka 2001; Chapman 2003). These, and other nearby "light-toned layered deposits" (LLDs) have been found to contain a number of hydrated minerals, including monohydrated sulphates, opal, and ferric sulphates, along with mafic minerals that include pyroxenes and olivine (Bishop *et al.* 2009). It has been hypothesized that hydrothermal processes may have been involved in the deposition of opal, with subice volcanism providing the necessary heat source (Bishop *et al.* 2009).

More recently, possible subglacially emplaced dyke swarms (Figure 5b), and potentially also moberg ridges, have been identified between the Elysium Rise and Utopia basin (Pedersen *et al.,* 2010), while Levy et al. (2010) identified features at Galaxias Fossae that bear a striking similarity to volcanogenic glacial cauldrons on Earth (Figure 5g). The martian cryosphere has of course changed significantly over time, due to a combination of local and global climate change (Baker *et al.,* 1991; Clifford *et al.,* 2010), and effects of obliquity variations (Forget *et al.*, 2006). Evidence for Amazonian glaciation at mid – low latitudes due to high martian obliquity is now well recognized (Head *et al.,* 2003; Neukum *et al.,* 2004; Schorghofer 2007; Fassett *et al.,* 2010). As such, glaciovolcanic products have been identified in the equatorial regions of these terrains (Chapman 2003; Leask *et al.,* 2006; Kadish *et al.*, 2008) as well as in more polar latitudes. In particular, both glaciation and volcanism are thought to have occurred as recently as the late Amazonian (Dickson *et al.* 2011), and Head *et al.* (2003) identified deposits consistent with a possible martian ice-age 2.1 - 0.4 million years ago. At Olympus Mons, Neukum *et al.* (2004) found the youngest lava flows to be <30Ma in age and identified multiple episodes of volcanic and glacial activity, with associated hydrothermal water release caused by the melting of ground ice by magma intrusion.

4. Glaciovolcanic microbial habitats

The importance of subglacial volcanism for martian exobiology lies in the observation that basaltic subglacial eruptions on Earth generate large volumes of liquid water that can be stored and transported beneath the overlying glacier (Wilson & Head 2002), and that many of the environments that result from such volcanism exist within the subsurface. In particular, the interaction between geothermal heat flow and an overlying cryosphere or ice cap is highly conducive to the generation of hydrothermal systems (Schulze-Makuch *et al*., 2007), both during and between eruptions (Björnsson 2002; Wilson & Head 2007). Subglacial volcanic habitats range from the overlying cryosphere to deep within the lava edifice, and these are discussed here individually. Examples of the microbiota and physicochemical characteristics of selected environments are also provided in Table 2.

4.1. Subglacial caldera lakes

During and between subglacial eruptions, meltwater can be confined as a subglacial caldera lake (Gudmundsson *et al.,* 1997). Such caldera lakes exist in Iceland (Björnsson 2002; Johannesson *et al*., 2007) and have been inferred to have existed on Mars (Fassett & Head 2007). The lakes in Iceland are inhabited by a specialized population of psycrotolerant and chemotrophic bacteria (Table 2) in the lake water and volcanic sediments that lie at the bottom of the lake (Gaidos *et al*., 2004; 2008). One of these caldera lakes is characterized by a largely anoxic mixture of glacial meltwater and sulphidic geothermal fluid (Gaidos *et al*. 2008). A bacterial community based on acetogenesis, sulphate reduction, sulphide oxidation, and potentially methanogenesis is tentatively inferred, with acetogenesis in particular hypothesized to be an important input of carbon into this ecosystem (Gaidos *et al*. 2008). These caldera lakes can exist as a habitable environment until catastrophically drained as a jökulhlaup and can be highly dependant upon the underlying geometry of the volcanic edifice and overlying ice (Gudmundsson *et al.,* 1997). At Grimsvotn, the topography is such that meltwater can accumulate and form a relatively stable lake until either there is an eruption event or the ice damming the lake is breached (Björnsson 2002). Conversely at Gjálp, continual drainage of ~20°C temperature meltwater away from the eruption site has been observed, with no subsequent ponding of water (Gudmundsson *et al.*, 2004; Jakobsson & Gudmundsson 2008). These subglacial caldera lakes represent one of the most potentially exciting environments within the volcano - ice system.

4.2. Subglacial lava edifices

Basalt, combined with localized areas of hydrothermal activity, has the potential to be colonized by a chemosynthetic-based ecosystem on Mars (Boston *et al.,* 1992). Mild hydrothermal activity within the volcanic edifice is thought to occur in the several years following an eruption, based on observations of modern subglacial eruptions in Iceland, such as Gjálp (Jakobsson & Gudmundsson 2008). Basalt is the most abundant geological substrate on Earth and Mars, and as such a significant amount of work has focused on exploring life that inhabits this environment on Earth. Terrestrial basaltic habitats exist predominantly at, and below, the sea floor, within the continental subsurface environments (e.g., aquifers), and as subaerial substrates (e.g., lava flows). Oceanic basaltic lava flows in particular have been the subject of much investigation regarding their microbiota over the past few decades. Fresh basalt erupted from mid-ocean ridge systems is widely found to be colonized and altered by a range of bacterial and archaeal chemosynthetic microbial communities. These can exploit the redox gradients between reduced species and oxygenated sea-water, such as for Fe oxidation (Edwards *et al.* 2003), as well as employ anaerobic pathways such as methanogenesis, S^0 reduction, sulphate reduction, and Fe reduction (Martin *et al.* 2008). Additionally, basaltic habitats within the terrestrial deep subsurface have been of interest in terms of understanding subsurface ecosystems on Earth and potentially on other planets, such as Mars (Stevens & McKinley 1995; McKinley & Stevens 2000).

Volcanic edifices that currently exist beneath glaciers on Earth are directly analogous to those that may have existed on Mars, but these environments are yet to be explored regarding their microbiota. Those edifices that have been exposed by glacial retreat have been found to host surprisingly diverse bacterial communities. Recent work by Cockell *et al.*, (2009a; 2009b), and Herrera *et al.*,

(2009) demonstrated the exploitation of subglacially erupted basaltic hyaloclastites as a favorable volcanic habitat for crypto- and chasmoendolithic life (see Table 2). This widespread utilization of basaltic environments on Earth suggests that any potential biological colonization of subglacial volcanic systems on Mars is likely to exploit the basaltic volcanic edifice as both a physical substrate on which to attach and as a source of energy.

4.3. Cryospheric hydrothermal environments

Glaciers and permafrost on Earth are known to contain a diverse array of psycrophillic and mesophillic life, particularly in basal ice (Table 2) or at the ice-rock boundary (Priscu & Christner 2004). Such communities could be incorporated into temporary hydrothermal systems within the cryosphere. Martian permafrost also has the potential to provide a habitable environment through the interaction with elevated geothermal heat and the subsequent production of meltwater. This is especially true where magma intrusions have a large surface area/volume ratio, such as dykes and sills (Head & Wilson 2002). Although no present-day geothermal anomalies have been detected (Christensen *et al.* 2003), the widespread evidence of significant volcanism and endogenic hydrothermal activity suggests higher heat flow in the past (Schulze-Makuch *et al.* 2007). McKenzie & Nimmo (1999) calculated that a 16 km wide dyke intrusion into a 5 km thick permafrost layer (ice fraction 0.2 by volume) would produce a subsurface lens of meltwater with a volume of 6.5 km^3 for each kilometer length of the dyke, and that such a meltwater zone would not start to refreeze until ~8 Ma after the dyke intrusion. Similarly, Travis *et al*. (2003) showed that hydrothermal circulation can occur on Mars with sufficient geothermal heat interacting with the overlying permafrost and also suggested that these upwelling hydrothermal plumes could provide a suitable environment for chemosynthetic life (Travis *et al.,* 2004). Such permafrost hydrothermal systems would remain within the subsurface, except for directly above magma intrusions or where springs breach the surface (e.g., along fractures) (Chapman *et al.,* 2000).

In addition, where a volcanic eruption has taken place beneath a glacier, there is the potential for glacier caves to form within the ice itself, carved by the drainage of hydrothermal fluids and meltwater. Little is known about the processes that occur at the glacier base in volcano - ice settings, including the formation of these glacial caves (Tuffen *et al*., 2002). Some of the best described caves are those at Mount Rainier, where fumarole interaction with overlying firn and snow produced caves over 1.5km in length (Kiver & Mumma 1971; Zimbelman *et al*., 2000). Some of these caves were observed to be steam-filled through fumarolic activity (Zimbelman *et al.,* 2000), and meltwater was seen to drip continuously from cave walls and ceilings (Kiver & Mumma 1971). A small crater lake was also observed within part of the cave system (Kiver & Steele 1972). Glacier caves associated

336 with subglacial volcanism also exist in Iceland (Figure 1d), and similar "ice
337 towers" have been identified at Mt. Erebus, in Antarctica (Hoffman & Kyle
338 2003). These caves provide an ice, and water-rich subsurface environment,
339 potentially coupled with fumarolic input. Such environments would be highly
340 favorable for microbial colonization on Mars, and exploration into the
341 microbiology of those on Earth would shed significant light on this issue.
342
343 Finally, high localized geothermal heat flow can also melt the overlying glacial
344 ice or permafrost in isolated areas at the surface and form glacial springs (Figure
345 1f) and intraglacial meltwater lakes (Figure 1e) that interact with surface
346 fumaroles. Such volcanically driven environments exist in the Atacama (Costello
347 *et al*., 2009), Antarctica (Soo *et al*., 2009), and Iceland (Olafsson *et al.,* 2000), and
348 often produce "islands" of biodiversity within an otherwise highly hostile
349 environment (Costello *et al.,* 2009). Martian hydrothermal systems have been
350 suggested many times as an environment suitable for microbial life (e.g., Rathbun
351 & Squyres 2002; Varnes *et al.,* 2003; Pope *et al*., 2006), and those generated
352 through volcano – ice interaction are no exception.
353

354 5. Biosignature preservation

355 The generation of widely varying environments through volcano – ice interaction
356 results in a diverse range of deposits within the geological record. Evidence for

putative glaciovolcanism appears to span almost the entirety of martian geological history (Table 1, Figure 4), although the majority of examples are from Hesperian and Amazonian terrains where geomorphological features are best preserved. Those features that are consistent with the generation and ponding of meltwater are perhaps the most optimum targets, regardless of their age. In particular, deposits representative of environments analogous to the subglacial caldera lakes seen in Iceland (such as jökulhlaup deposits) could be primary targets. These and other products of volcano - ice interaction, including basaltic lavas and hydrothermal mineral deposits, are discussed below regarding their biosignature preservation potential.

5.1 Jökulhlaup deposits

Evidence for life in volcano – ice systems could be recorded via the presence of biomolecules within subglacially erupted basalt and jökulhlaup deposits. Data from the orbiting hyperspectrometers CRISM and OMEGA show the presence of phyllosilicate minerals at the martian surface, with smectite clay minerals such as montmorillonite and nontronite having been identified (e.g., Poulet *et al.,* 2009). It has been proposed that clay-rich deposits may be suitable sites of organic preservation on Mars (Ehlmann *et al.,* 2008). Such minerals are ubiquitous among subglacially erupted basaltic lavas, due to the widespread breakdown of volcanic glass to palagonite and smectite clays through contact with liquid water (Stroncik

& Schmincke 2002). Phyllosilicate detection on Mars to date has been restricted to a few Noachian terrains such as Nili Fossae (Mustard *et al.* 2009) and Mawrth Valles (Michalski & Noe Dobrea 2007). It has yet been found to coincide with putative volcano – ice geomorphological features, although a recent study by Martinez-Alonso *et al.* (2011) tentatively indicates Mg-smectite clays to be associated with mesas interpreted to be subglacial tuyas. Volcano - ice landforms therefore could be considered suitable spectroscopic targets for future investigation. If such deposits coincide with volcano – ice interaction terrains on Mars, these could be prime geological formations to search for evidence of life. Indeed, Warner & Farmer (2010) used visible–near infrared and shortwave infrared remote sensing to spectrally identify low-temperature hydrothermal mineralogical assemblages within Jökulhlaup deposits in south Iceland. As suggested by these authors, such “mineralogical fingerprints” can be used to identify potentially past habitable conditions within a subglacial volcanic system and are therefore ideal astrobiological targets. An example of such a target includes the drainage valleys and deposits at the edge of the Dorsa Argentea Formation (DAF; Figure 6a). Here, sinuous channels lead away from the bases and margins of candidate subglacial volcanoes (Ghatan & Head 2002; Head & Wilson 2007). This terrain is thought to be formed much in the same way as Icelandic jökulhlaup deposits, where drainage channels leading away from the

DAF are interpreted to represent volcanism-induced subglacial meltwater release (Ghatan & Head 2004).

5.2 Hydrothermal deposits

Hydrothermal systems on Earth are noted for their ability to preserve detailed microbial fossils, particularly within silica (Preston *et al.,* 2008) and carbonate (Allen *et al*., 2000) systems. Indeed, silica deposits of possible fumarolic or hydrothermal origin have been identified by the MER Spirit landing site (Squyres et al. 2008). However, such preservation is dependant upon the deposition of mineralized or solute-rich fluids and the subsequent precipitation of the mineral phases and preservation through fossilization of the in situ microbial community. There are examples where concentrated mineral deposits form within, or as a direct result of, volcano - ice interaction. At the Bockfjord volcanic complex in north-west Spitsbergen (Norway), the subglacially erupted volcanoes Sigurdfjell and Sverrefjell contain basaltic lavas with hydrothermal carbonate cement deposits (Blake *et al*., 2010). These carbonates demonstrate a potential mechanism for the preservation of microfossils and organic biosignatures within a volcano - ice system on Mars. Additionally, subglacially erupted pillow lavas in central Iceland (Figure 1b) have been found to contain gypsum deposits within the lava vesicles, most likely precipitated during hydrothermal circulation within the subglacial edifice following eruption (Storrie-Lombardi *et al.* 2009). Such deposits, if found on Mars, would suggest a once-habitable subsurface hydrothermal environment that may have preserved signatures of life.

Alternatively, where subsurface silica-charged hot spring fluids are frozen through eruption into a sub-zero environment, cryogenic opal-a is precipitated between ice crystals, which produces distinctive cryogenic particle morphologies (Channing & Butler 2007). As suggested by Channing & Butler (2007), this precipitation may fossilize any microorganisms present within the hot-spring fluid, which are partitioned out of the growing ice crystals and into the surrounding liquid vein network along with the silica (Mader *et al.,* 2006; Channing & Butler 2007).

Finally, the subglacial volcano Kverkfjöll in Iceland is associated with several hydrothermal systems (Cousins *et al*., 2010). One of these - the hot spring Hveragil - has thick deposits of calcite along the floor of the gully that the hot spring flows along (Ólafsson *et al*., 2000) and, as with many hot spring mineral deposits, is likely to contain biosignatures such as microfossils, organics, or both. Little is known regarding the preservation of biosignatures within such systems generated by volcano – ice interaction, and this represents a significant area of research with direct implications for the search for life on Mars. One significant drawback, however, is the often small-scale and highly localized nature of such mineral deposits (e.g., to an isolated spring), which could potentially hinder their discovery.

5.3 Subglacially erupted lavas

Glassy basaltic lavas on Earth often contain intricate tubular and pitted structures, which have been widely interpreted to be formed by the activities of euendolithic microbes at the glass – palagonite interface (Furnes *et al*., 2007 and references therein; McLoughlin *et al*., 2009). Typically, 80 - 90% of hyaloclastite is glass (Jakobsson & Gudmundsson 2008), which leads to the possibility for the significant production of microbial bioalteration textures so commonly seen in submarine glassy lavas (McLoughlin *et al*., 2009). It has been previously suggested that these bioalteration textures would make suitable biosignatures when looking for life on Mars (Banerjee *et al*. 2006; McLoughlin *et al*., 2007), particularly when they are preserved by minerals such as zeolites and titanite infilling tubular textures (Furnes et al., 2004; Izawa *et al*., 2010). However, while these putative biosignatures appear to be ubiquitous in lavas within an oceanic setting, an abundance of such textures is yet to be found in basalt of a subglacial origin, despite sharing the same glassy lithologies (pillow lavas and hyaloclastites). A recent study by Cousins *et al*. (2009) demonstrated a possible environmental control on the formation of bioalteration textures and in particular showed that the subglacial environment was not as conducive to their formation as those that are oceanic. Likewise, while Cockell *et al.,* (2009a) described biogenic pitting in subglacially erupted hyaloclastites in Iceland, they also note an absence of the characteristic tubular textures seen in oceanic lavas. Bioalteration textures therefore may not necessarily be the most suitable biosignature for identifying past life within subglacial basaltic lavas, and other alternative options, such as geochemical biosignatures, should also be explored. For example, distinctive trace element (Zr, Sc, and Mn) signatures have been found to result from the utilization of organic acids to dissolve basaltic substrates (Hausrath *et al.,* 2007; Hausrath *et al.,* 2009). Likewise, sulphur isotope (^{32}S, ^{33}S, ^{34}S) compositions can provide evidence of microbial sulphate reduction within altered oceanic basalts (Rouxel *et al.,* 2008).

6. Discussion & Conclusions

An active volcano – ice system can potentially provide all the necessary ingredients for life. The continual release of geothermal heat into an overlying glacier can sustain a subsurface meltwater environment, while the release of volcanic gases such as H_2S, CO, CO_2, and H_2 could support a variety of chemosynthetic metabolisms. The presence of this heat flow will also mean that a continual convective system will create a cycling of material through the different environments, which will remove waste products from some niches and deliver nutrients to others.

It is clear that the presence of liquid meltwater is key to glaciovolcanic systems being suitable for life, but there are significant differences between terrestrial and martian systems that need consideration. On Mars, the melting efficiency of water ice is much reduced due to the low initial temperature of the ice (Hovius *et al*., 2008), which perhaps suggests volcano – ice systems on Mars were not as viable as those on Earth. Indeed, there are locations on Mars interpreted to be the result of subglacial volcanism that display a distinct lack of evidence for meltwater. Such places include the proposed subglacial lava flows at Ascraeus Mons, where rapid re-freezing of a cold-based glacier would prevent any significant basal melting (Kadish *et al*., 2008). However, it is thought that the temperature of the meltwater is highly influential on the formation of jökulhlaups, whereby higher heat flow enables the enlargement of subglacial drainage tunnels (Gudmundsson *et al*., 1997). The occurrence of jökulhlaup-like flows and deposits on Mars therefore suggests that subglacial eruptions can lead to significant subglacial melting, even with cold-based glaciers (Head & Wilson 2007).

While habitable environments potentially may exist in this subglacial volcanic setting on Mars, they are most likely to be transient and isolated. On Earth, any new body of liquid water will be rapidly colonized due to the widespread and globally connected biosphere (Cockell & Lim 2005). While it remains possible that pockets of martian life could exist, as yet there is no evidence for a martian biosphere. As a result, the delivering of martian life, should it exist, to newly formed habitable environments remains a problem. It can be seen that, in Iceland, features indicative of subglacial volcanic activity often occur in clusters (e.g., Alfaro *et al*., 2007), which suggests that localized habitable regions may exist within a close enough proximity to allow transport of microorganisms between individual niches. On Earth, regions of high heat flow are rarely isolated to just one volcano. Indeed, in the case of Iceland, Vatnajökull (glacier) overlies seven individual volcanic centers. Additionally, it has been observed that rapid vertical transport of hydrothermal fluid occurs beneath Mýrdalsjökull via faults within the ice (Björnsson 2002). Therefore, it is possible to envisage such subglacial systems to be connected via fractures and channels within the ice, where meltwater (and any microbial life it may carry) may circulate, distributing microorganisms from one system to another.

The vast majority of the terrestrial biosphere is dependant upon photosynthesis, either directly or indirectly (Varnes *et al*., 2003). Photosynthesis on Mars, however, would be hindered by the exposure to UV radiation and by the increased distance to the Sun, which reduces the flux of photosynthetically active radiation (PAR) to ~55% of that typically experienced on Earth (Cockell & Raven 2004). If photosynthetic communities were to exist within a subglacial volcanic system, they would be limited to the near-surface ice and specifically use blue and green wavelengths due to the high absorbance of red light within ice (Hawes & Schwarz 2000). Cockell & Raven (2004) showed experimentally that the maximum depth within snow-pack at which the minimum level of PAR can penetrate is ~24cm. Additionally, work on ice-covered lakes in Antarctica has shown there to be benthic photosynthetic communities residing at ~16m water depth beneath 3.5 – 5m of ice cover (Vopel & Hawes 2006), which is much shallower than the depths of many subglacial volcanic systems, which are typically beneath several hundred meters of overlying ice (Wilson & Head 2002). At depths of 100m within glacial ice, PAR is entirely absent (Warren *et al.*, 2002). Subglacial volcanic environments, therefore, are not suited to a photosynthesis-based community. This limits the primary producers of this environmental setting to chemosynthetic pathways.

On Earth, chemoautotrophs are major contributors for communities residing within dark, extreme environments, such as deep sea vents (McCollom & Shock 1997; Van Dover 2000). Specifically to Mars, anaerobic chemolithoautotrophs can potentially inhabit subglacial volcanic environments through the oxidation of inorganic compounds and fixation of carbon dioxide as the carbon source (Boston *et al.*, 1992). Numerous chemosynthetic pathways could be exploited due to the chemical disequilibrium that results from the mixing of high- and low-temperature fluids (Gaidos & Marion 2003). On Earth, the majority of the chemosynthetic microbial communities residing in present-day hydrothermal systems are indirectly dependent on photosynthetically produced O_2 (Varnes *et al*., 2003). However, an estimated 1 – 2% of these communities obtain chemical energy from redox reactions that are completely independent of photosynthesis (Varnes *et al*., 2003), and it is these microorganisms and their metabolic pathways that are potentially suitable for survivability on Mars, particularly within subglacial hydrothermal systems.

In conclusion, the conditions that exist as a result of volcano – ice interaction provide a wide range of habitats for life on Earth and may have provided a possible subsurface haven for life on Mars during past epochs. While there is still much work to be done with regard to understanding the thermal and geochemical conditions of such environments on Mars, the combination of basaltic lava, liquid water, and hydrothermal activity provides a possible subsurface haven for life. The wide range of geological deposits – be it jökulhlaup sediments, hydrothermal minerals, or subglacial basalt – provides several mechanisms for the preservation of any biosignatures for future discovery.

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FIGURES & TABLES

FIG. 1. Icelandic examples of subglacial volcanic products and environments. **a)** Tuya (Herðubreið) with car for scale; **b)** Pillow mound, in central Iceland (people on left for scale); **c)** Sandur subglacial outwash plain in south Iceland (road bridge for scale); **d)** Glacier cave at Kverkfjöll; **e)** Subaerial glacial meltwater lake (~500m across) above the subglacial volcano Kverkfjöll; **f)** fumaroles and hot springs interacting with the glacier surface at Kverkfjöll volcano.

FIG. 2. Simplified diagram showing the processes of volcano – ice interaction: **a)** initial subglacial eruption into overlying ice, creating a meltwater lens and ice cauldron at the glacier surface; **b)** emergent eruption, whereby sustained volcanic activity eventually melts through the ice, resulting in subaerial capping lavas when the edifice becomes higher than the waterline, forming a tuya morphology;

c) subaerial hot springs and fumaroles at the glacier surface, sourced by surface glacial meltwater and underlying geothermal activity.

FIG. 3. Map of localities given in Table 1 (Image credit: National Geographic Society, MOLA Science Team, MSS, JPL, NASA).

FIG. 4. Geological time scales for Mars (left) and the corresponding divisions for Earth (right). Boundary variations for both the Neukum ('N') or Hartmann ('H') chronology models (Hartmann & Neukum 2001) are shown.

FIG. 5. Examples of putative glaciovolcanic features on Mars as described in the text and given in Table 1. **a)** Mesa at Acidalia Planitia interpreted as a possible tuya (Martinez-Alonso *et al.,* 2011; HiRISE image PSP_009497_2210_RED) ; **b)** putative subglacially erupted dykes near the Elysium Volcanic Province (Pedersen *et al*., 2010; HiRISE image PSP_006591_2165); **c)** Chasma Boreale (Fishbaugh & Head 2002; MOLA shaded relief map overlain on a THEMIS IR day 100 m mosaic); **d)** large drainage channels in the Athabasca Valles (Burr *et al.*, 2002; MOC image M2101914); **e)** Interior Layered Deposit in Juventae Chasma (Chapman 2003; MRO CTX image B18_016712_1762_XN_03S061W); **f)** pseudocraters or "rootless cones" north of the Cerberus plains (Fagents *et al.,*

2002; MOC image M08/01962); **g)** ice-cauldron morphology at Galaxias Fossae (Levy *et al.,* 2010; HiRISE image PSP_005813_2150).

FIG. 6. More examples of putative glaciovolcanic features on Mars as described in the text and given in Table 1. **a)** broad view of the Dorsa Argentea Formation with numerous inverted linear features, shown in more detail in Figure 5c (Ghatan & Head 2004; (MOLA shaded relief map overlain on a THEMIS IR day 100 m mosaic); **b)** Chaos terrain and drainage channels at Xanthe Terra (Chapman & Tanaka 2002; MOLA shaded relief map overlain on a THEMIS IR day 100 m mosaic); **c)** Inverted linear features interpreted as eskers (Ghatan & Head 2004, MRO CTX image P13_006282_1046_XN_75S043W); **d)** volcano Ceraunius Tholes showing evidence of floodwater originating at the caldera rim (Fassett & Head 2007; MRO CTX image B04_011399_2045_XN_24N097W).

Table 1. Examples of candidate volcano – ice interaction features identified on Mars, and their associated terrestrial analogue (where given). Alternative interpretations of these sites are also shown for comparison.

Table 2. Physicochemical characteristics and resident biota present within terrestrial volcano - ice related environments.

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Table 1. Examples of candidate volcano – ice interaction features identified on Mars, and their associated terrestrial analogue (where given). Alternative interpretations of these sites are also shown for comparison.

Region on Mars	**Approximate age**	**Description**	**Terrestrial analogue cited (in Iceland, unless otherwise stated)**	**Reference**	**Alternative hypotheses**
Ascraeus Mons (Tharsis)	Mid- to Late Amazonian	Flat topped ridges typical of subglacial volcanism, fan-shaped glacial deposit.	Herðubreið (Figure 2a)	Kadish *et al.,* (2008)	Gravity sliding and pyroclastic activity (Edgett *et al*., 1997).
Chasma Boreale and Abalos Colles	Amazonian < 20 000 years	Flat topped ridges, subglacial volcanism with possible caldera lake formation, and catastrophic outflow of subglacial meltwater (jökulhlaup) resulting in a large water incised chasm.	Gjalp 1996 subglacial eruption	Fishbaugh & Head (2002); Greve (2008); Hovius *et al.,* (2008).	Long term wind erosion and solar ablation of ice-rich units (Warner & Farmer 2008a,b); non-uniform accumulation of the north polar layered deposit (Holt *et al.,* 2010).
Cerberus Fossae and Athabasca Valles	Late Amazonian <100Ma	Extensive aqueous flooding in close association with large fissures and lava flows, instigated by dike emplacement into the cryosphere.	Channelled Scabland in the northwestern United States	Head *et al.* (2003)	Release of a subsurface liquid water aquifer via volcanotectonic fissures (Burr *et al.,* 2002a; Burr *et al.,* 2002b)
Elysium Volcanic Province	Early Amazonian	Subglacially emplaced dikes and moberg ridges, possible evidence for an ice cauldron.	Gjalp and Grimsvotn	Pedersen *et al*., (2010)	-
Chryse and Acidalia Planitiae	Late Hesperian to Early Amazonian	Mesa-like features, drumlins, eskers, kettle holes, and inverted valleys. Orbital spectral data consistent with hydrous alteration of volcanic materials.	Icelandic subglacially-erupted tuyas including Herðubreið and Hlodufell	Martinez-Alonso *et al.* (2011)	Sedimentary deposits resulting from mass flow and mass-wasting (Tanaka (1997)
Juventae Chasma and Interior Layered Deposit mounds	Late Hesperian to Amazonian	Lava capped ridge and kettle holes, interpreted as jökulhlaups following a sub-ice eruption	Gjalp 1996 subglacial eruption, Skaftafjell.	Chapman *et al.* (2003); Chapman & Tanaka (2001)	Evaporite deposit prior to the development of the chasma, or aeolian deposition of volcanic sulphate aerosols (Catling *et al*., 2006).
Southern polar region (Dorsa Argentea Formation)	Hesperian	Morphological features displaying a linear trend, many of which are steep sided and flat topped. Interpreted to be a subglacial eruption into a previously much larger polar ice cap, with channels forming possibly from basal meltwater drainage.	Drainage channels from the Vatnajokull and Mydralsjokull ice caps, Gjalp 1996 subglacial eruption	Dickson & Head (2006); Ghatan & Head (2002; 2003; 2004)	Outlet channels and floodplains, inverted channel deposits, and debris flows (Tanaka & Kolb 2001)
Ceraunius Tholus & Hecates Tholus (Tharsis)	Hesperian	Volcano surrounded by valleys and depositional fans. Potential for a caldera lake. Interpreted to be a magmatic intrusion resulting in basal melting of snowpack	Katla - a subglacial volcano beneath the Mydralsjokull ice cap, and the Aniakchak caldera in Alaska (Ceraunius); meltwater streams in the	Fassett & Head (2006; 2007)	-

			Antarctic Dry Valleys (Hecates).		
Xanthe Terra (Aromatum Chaos)	Late Noachian – Early Amazonian	Chaos terrain and outflow channels, interpreted to be the result of cryospheric disruption via the intrusion of a volcanic sill and/or dikes.	Icelandic tuyas, jökulhlaups, and sandur plains	Chapman & Tanaka (2002); Leask *et al.*, (2006)	Burial of ice sheets in a confined basin resulting in basal melting, eventually being released catastrophically (Zegers *et al.*, 2010) ; dehydration of sulphates triggered by geothermal heating (Montgomery & Gillespie 2005); catastrophic release of groundwater from an over-pressured aquifer (Carr 1979).

Table 2. Physicochemical characteristics and resident biota present within terrestrial volcano - ice related environments

Terrestrial Environment	**Location**	**Environment characteristics**	**Biota**	**Reference**
Skaftá subglacial caldera lake	Iceland	Anoxic lake bottom, 3.5 - 6°C (whole lake), pH 5.22, 6:1 mixture of glacial meltwater and sulphidic geothermal fluid.	Bacteria dominated community with no archaea so far detected. Community dominated by *Acetobacterium* species. Putative metabolic pathways include homoacetogenesis ($H_2 + CO_2 \rightarrow$ acetate, hydrogen oxidation, and sulphate reduction.	Gaidos *et al.* (2009); Johannesson *et al.* (2007)
Grimsvotn subglacial caldera lake and tephra	Iceland	Fresh, oxygenated, pH 4.87-5.13, dominated by glacial melt with acidification by volcanic CO_2. Little hydrothermal input. Coarse volcanic tephra at lake bottom.	Microorganisms adapted to growth at low temperature. Distinct communities within the tephra and lake water, with higher biomass in the lake tephra. Presence of microbial carbon fixation (autotrophs).	Gaidos *et al.* (2004)
Mt. Erebus fumaroles, Tramway Ridge	Antarctica	Geothermally heated ice-free ground on the flank of the volcano with CO_2-rich steam fumaroles. Near-neutral to acidic soil pH, temperatures between 2.5 - 65°C (over <0.6m distance). Low total C and N. High in Fe and Mn.	Low sequence identity to environmental bacteria and cultured isolates, suggesting the site is dominated by yet-to-be described bacterial groups. Both bacterial and archaea present, exhibiting high and low biodiversity respectively.	Soo *et al.* (2009)
Subglacially-erupted basalt (now subaerially exposed)	Iceland	High porosity (25.8%) basaltic-composition volcaniclastic rock substrate, rich in palagonite	Diverse community dominated by Actinobacteria, with many bacteria genetically similar to those from a variety of soil environments.	Kelly *et al.* (2010)
John Evans Glacier, Ellesmere Island	Canada	Base of a polythermal glacier with basal melting. External mean temperature of -14.5°C.	Psycrophillic organisms including aerobic chemoheterotrophs, anaerobic nitrate reducers, sulphate reducers, and methanogens.	Skidmore *et al.* (2000)

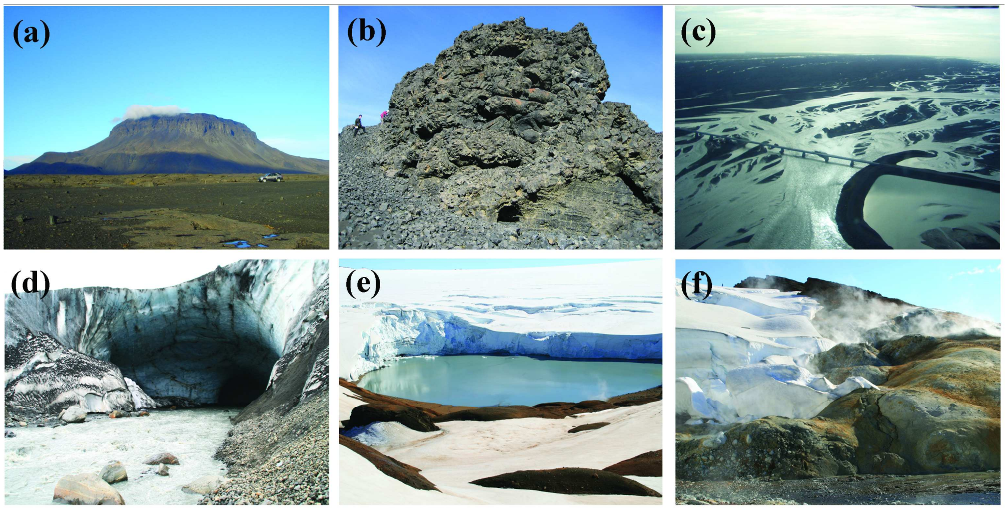

FIG. 1. Icelandic examples of subglacial volcanic products and environments. a) Tuya (Herðubreið) with car for scale; b) Pillow mound, in central Iceland (people on left for scale); c) Sandur subglacial outwash plain in south Iceland (road bridge for scale); d) Glacier cave at Kverkfjöll; e) Subaerial glacial meltwater lake (~500m across) above the subglacial volcano Kverkfjöll; f) fumaroles and hot springs interacting with the glacier surface at Kverkfjöll volcano.
159x80mm (488 x 488 DPI)

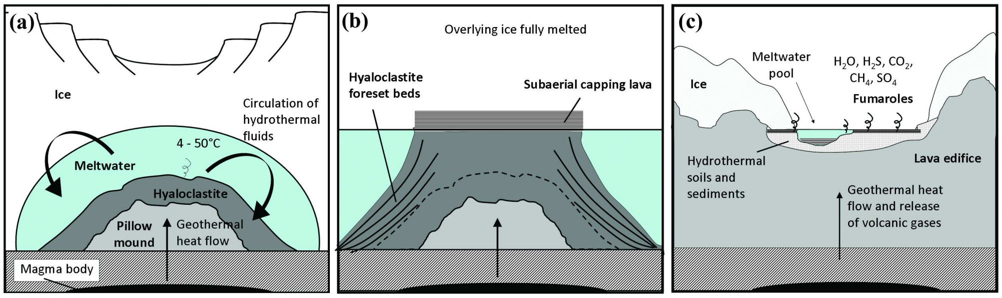

FIG. 2. Simplified diagram showing the processes of volcano – ice interaction: a) initial subglacial eruption into overlying ice, creating a meltwater lens and ice cauldron at the glacier surface; b) emergent eruption, whereby sustained volcanic activity eventually melts through the ice, resulting in subaerial capping lavas when the edifice becomes higher than the waterline, forming a tuya morphology; c) subaerial hot springs and fumaroles at the glacier surface, sourced by surface glacial meltwater and underlying geothermal activity.
159x46mm (488 x 488 DPI)

FIG. 3. Map of localities given in Table 1 (Image credit: National Geographic Society, MOLA Science Team, MSS, JPL, NASA).
159x99mm (488 x 488 DPI)

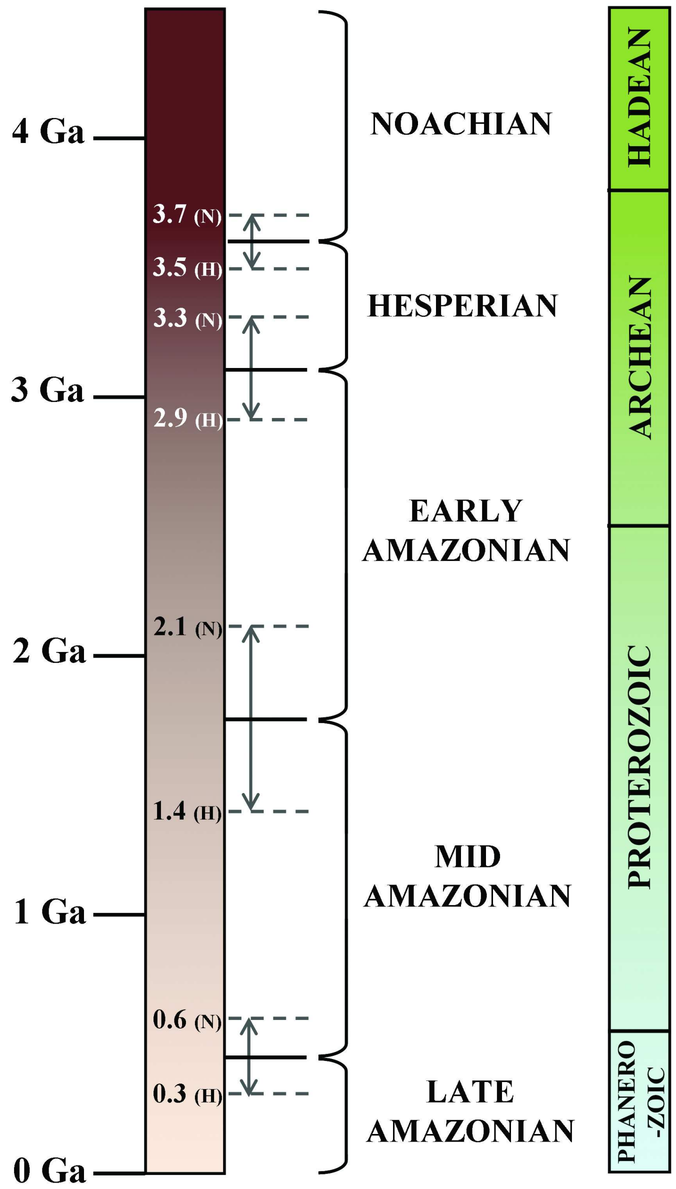

FIG. 4. Geological time scales for Mars (left) and the corresponding divisions for Earth (right). Boundary variations for both the Neukum ('N') or Hartmann ('H') chronology models (Hartmann & Neukum 2001) are shown.
85x149mm (522 x 522 DPI)

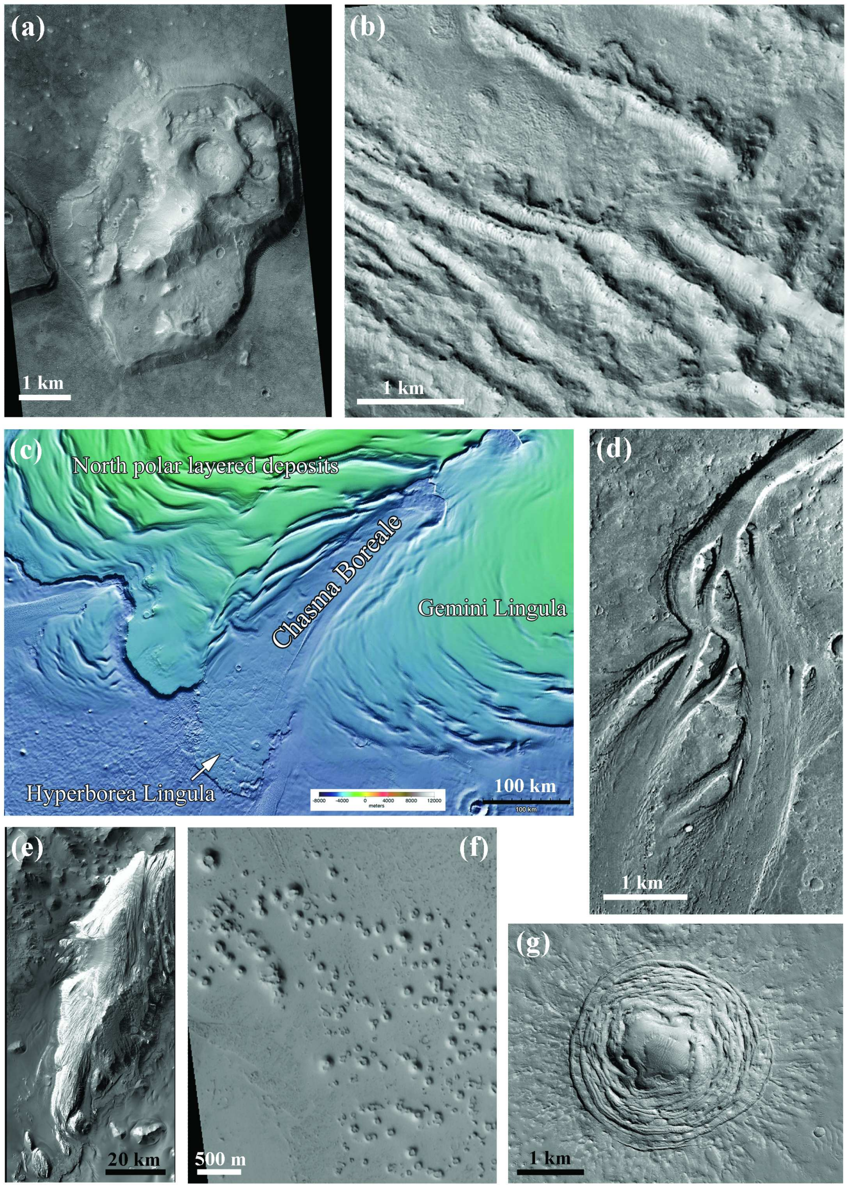

FIG. 5. Examples of putative glaciovolcanic features on Mars as described in the text and given in Table 1. a) Mesa at Acidalia Planitia interpreted as a possible tuya (Martinez-Alonso et al., 2011; HiRISE image PSP_009497_2210_RED) ; b) putative subglacially erupted dykes near the Elysium Volcanic Province (Pedersen et al., 2010; HiRISE image PSP_006591_2165); c) Chasma Boreale (Fishbaugh & Head 2002; MOLA shaded relief map overlain on a THEMIS IR day 100 m mosaic); d) large drainage channels in the Athabasca Valles (Burr et al., 2002; MOC image M2101914); e) Interior Layered Deposit in Juventae Chasma (Chapman 2003; MRO CTX image B18_016712_1762_XN_03S061W); f) pseudocraters or "rootless cones" north of the Cerberus plains (Fagents et al., 2002; MOC image M08/01962); g) ice-cauldron morphology at Galaxias Fossae (Levy et al., 2010; HiRISE image PSP_005813_2150).
160x223mm (349 x 349 DPI)

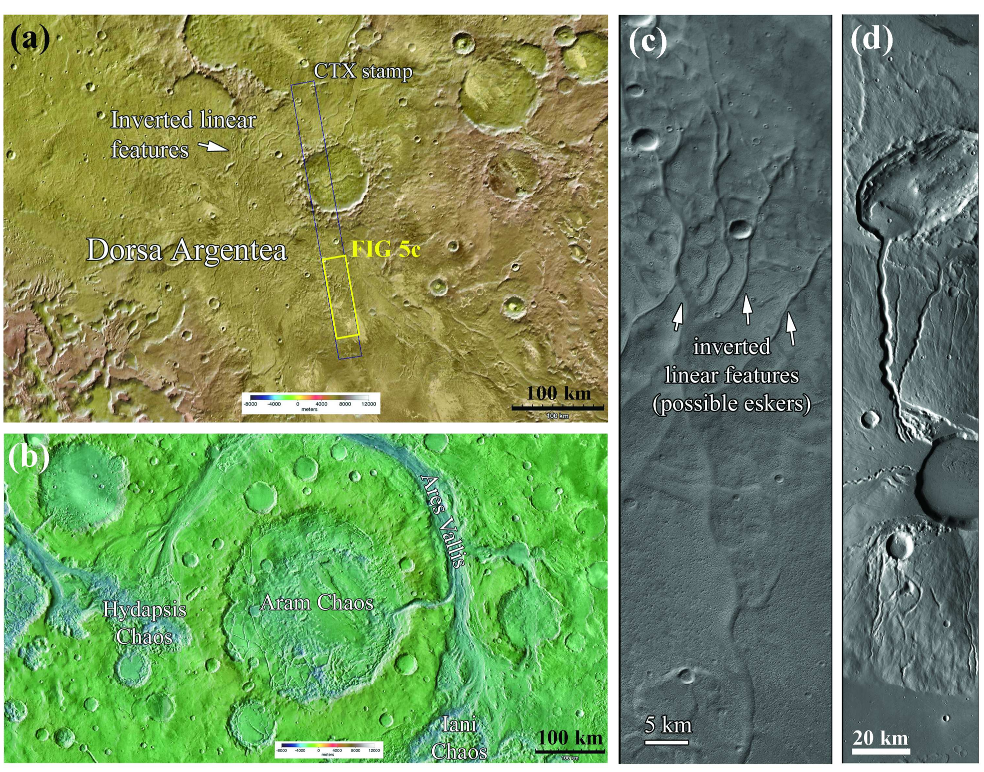

FIG. 6. More examples of putative glaciovolcanic features on Mars as described in the text and given in Table 1. a) broad view of the Dorsa Argentea Formation with numerous inverted linear features, shown in more detail in Figure 5c (Ghatan & Head 2004; (MOLA shaded relief map overlain on a THEMIS IR day 100 m mosaic); b) Chaos terrain and drainage channels at Xanthe Terra (Chapman & Tanaka 2002; MOLA shaded relief map overlain on a THEMIS IR day 100 m mosaic); c) Inverted linear features interpreted as eskers (Ghatan & Head 2004, MRO CTX image P13_006282_1046_XN_75S043W); d) volcano Ceraunius Tholes showing evidence of floodwater originating at the caldera rim (Fassett & Head 2007; MRO CTX image B04_011399_2045_XN_24N097W).
159x124mm (488 x 488 DPI)