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Thermal contraction crack polygons on Mars: A synthesis from HiRISE, Phoenix, and terrestrial analog studies

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ABSTRACT

Thermal contraction crack polygons are complex landforms that have begun to be deciphered on Earth and Mars by the combined investigative efforts of geomorphology, environmental monitoring, physical models, paleoclimate reconstruction, and geochemistry. Thermal contraction crack polygons are excellent indicators of the current or past presence of ground ice, ranging in ice content from weakly cemented soils to debris-covered massive ice. Relative to larger topographic features, polygons may form rapidly, and reflect climate conditions at the time of formation—preserving climate information as relict landforms in the geological record. Polygon morphology and internal textural characteristics can be used to distinguish surfaces modified by the seasonal presence of a wet active layer or dry active layer, and to delimit subsurface ice conditions. Analysis of martian polygon morphology and distribution indicates that geologically-recent thermal contraction crack polygons on Mars form predominantly in an ice-rich latitude-dependent mantle, more likely composed of massive ice deposited by precipitation than by cyclical vapor diffusion into regolith. Regional and local heterogeneities in polygon morphology can be used to distinguish variations in ice content, deposition and modification history, and to assess microclimate variation on timescales of ka to Ma. Analyses of martian polygon morphology, guided by investigations of terrestrial analog thermal contraction crack polygons, strongly suggest the importance of excess ice in the formation and development of many martian thermal contraction crack polygons—implying the presence of an ice-rich substrate that was fractured during and subsequent to obliquity-driven depositional periods and continually modified by ongoing vapor equilibration processes.

1. Introduction

Thermal contraction crack polygons are a key element of cold and polar landscapes on Earth and Mars (Figs. 1 and 2). A century of research by polar scientists has demonstrated the importance of polygonally patterned ground—and in particular, of thermal contraction crack polygons—in the interpretation of climate history, glacial and periglacial stratigraphy, polar hydrology, and cold desert ecology (Leffingwell, 1915; Lachenbruch, 1962; Black and Berg, 1963; Péwé, 1963; Washburn, 1973; French, 1976; Marchant et al., 1993, 2002; Sugden et al., 1993, 1995; Marchant and Denton, 1996; Murton, 1996; Yershov, 1998; Virginia and Wall, 1999; Bockheim, 2002; Doran et al., 2002; French, 2003; Fortier et al., 2007; Kowalewski et al., 2006; Kowalewski and Marchant, 2007; Marchant and Head, 2007; Murton and Bateman, 2007; Levy et al., 2008a; Kowalewski, 2009). Theoretical considerations of martian permafrost (Wade and De Wys, 1968; Mellon, 1997; Mellon et al., 2009), coupled with combined exploration of polygonally-patterned terrain by landers and orbiters has revealed Mars to be a permafrost-dominated planet with a rich surface history of thermal contraction cracking (Mutch et al., 1976, 1977; Malin and Edgett, 2001; Seibert and Kargel, 2001; Kuzmin and Zabalueva, 2003; Milliken et al., 2003; Mangold, 2005; Kostama et al., 2006; Smith et al., 2008, 2009; Levy et al., 2009c).

Several critical questions remain outstanding regarding the use of martian polygons as a paleo-climate indicator. (1) What has increasing image resolution revealed about the morphology of martian thermal contraction crack polygons and what does this indicate about polygon-formation conditions? (2) What is the distribution of martian polygons—are they present in random patches, or do zonal climate conditions affect their distribution? (3) What do polygons indicate about the origin, composition and ice content of the martian permafrost: is it dominated by snow and ice deposition producing massive ice or by cyclical vapor diffusion into regolith producing pore ice? (4) What does polygon morphology indicate about the presence, absence, and role of liquid water in recent martian permafrost terrain evolution? (5) What are the rates and duration of climate conditions permitting thermal contraction

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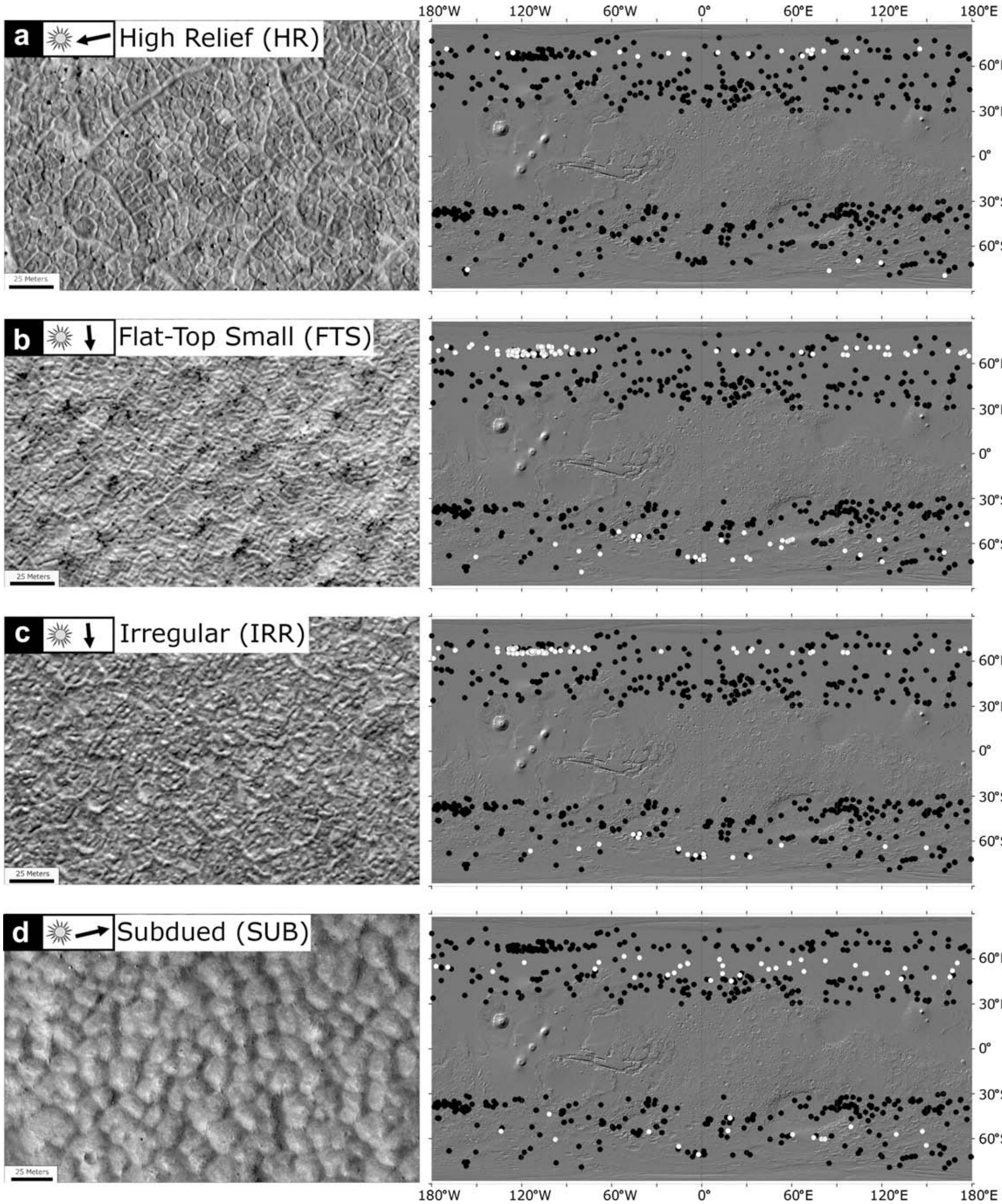

Fig. 1. (Left column) Thermal contraction crack polygons on Mars and (right column) distribution maps of HiRISE images containing the documented polygon morphological group (as defined by Levy et al., 2009c). White dots indicate HiRISE images in which the polygon group can be observed; black dots indicate HiRISE images not featuring the morphological group. North is to image top in all panels. (a) High-relief polygons. Portion of PSP_001474_2520. (b) Flat-top small polygons. Portion of PSP_001959_2485. (c) Irregular polygons. Portion of PSP_001959_2485. (d) Subdued polygons. Portion of PSP_003818_1360. Part B. (e) A gully–polygon system, or "gullygons". Portion of PSP_001357_2200. (f) Peak-top polygons. Portion of PSP_01737_2250. (f) Mixed-center polygons. Portion of PSP_002175_2210.

crack polygon formation on Mars—are polygons recent or relict features? (6) How extensively do polygon-forming processes rework the martian surface, altering our ability to date martian permafrost terrain? (7) How do polygons interact with larger ice-related features on Mars, such as glacial landforms (concentric crater fill, lineated valley fill), and what do these relationships indicate about primary ice precipitation and vapor diffusion over longer-term Amazonian climate? (8) How can terrestrial analogs be used to understand processes occurring in martian thermal contraction crack polygons, and how can terrestrial analog studies be useful in guiding the exploration of martian polygonally patterned ground?

Here, we address these outstanding issues, with the goal of building a synthesis on the basis of the current understanding of martian thermal contraction crack polygons. Conceptually, polygons are a powerful tool for linking several fields of cold regions research, including: (1) landscape geomorphology (Péwé, 1959, 1963, 1974; Berg and Black, 1966; Black, 1982; Bockheim, 2002; Sletten et al., 2003; Marchant et al., 2002, 2007; Marchant and Head, 2007), (2) remote sensing observations (Boynton et al., 2002; Feldman et al., 2002; Kuzmin et al., 2004; Bandfield, 2007), (3) physical modeling of the behavior of frozen materials (Lachenbruch, 1962; Mellon, 1997; Plug and Werner, 2001, 2002; Maloof et al., 2002; Arenson and Springman, 2005; Kowalewski et al., 2006; Kowalewski and Marchant, 2007; Swanger and Marchant, 2007; Kowalewski, 2009), (4) soil properties analysis and classification (Sugden et al., 1995; Bockheim, 2002; Bockheim et al.,

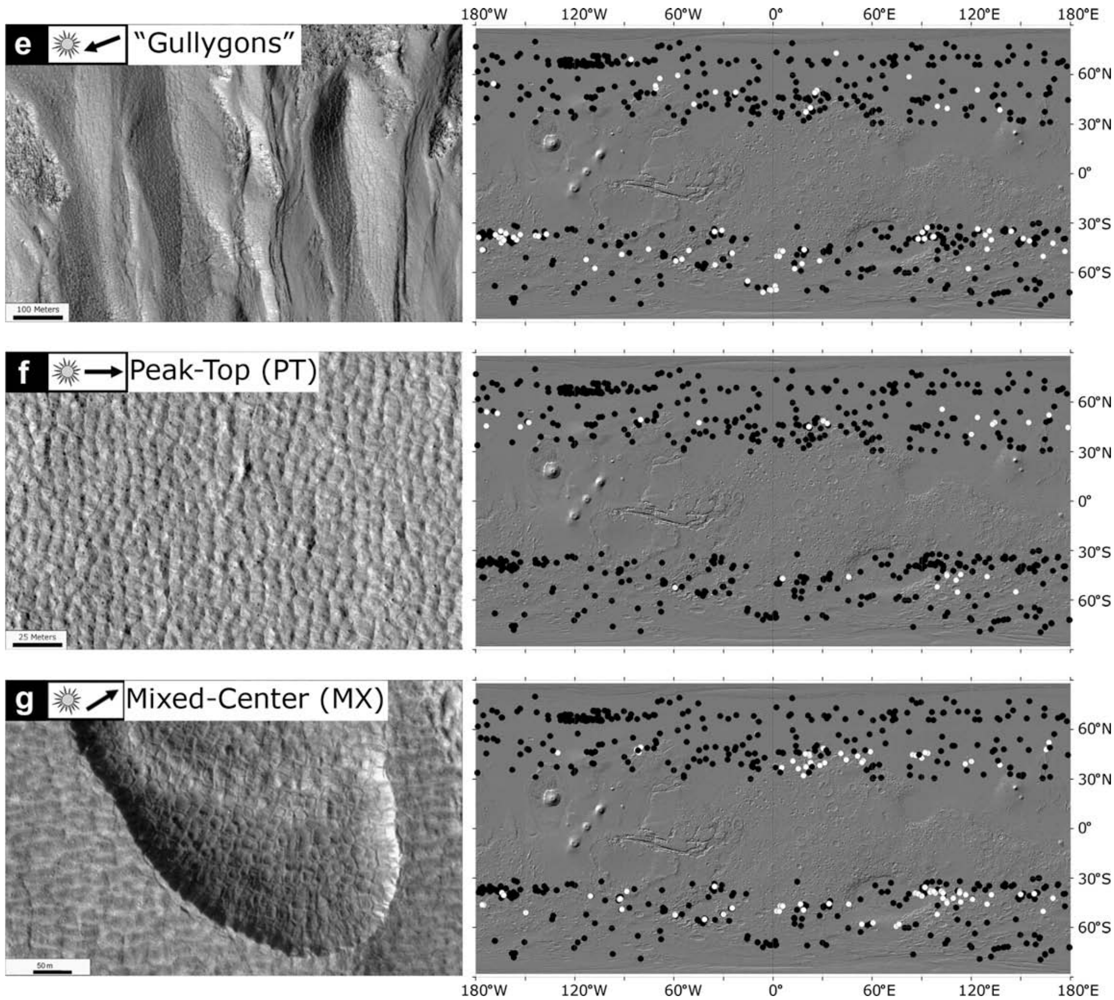

Fig. 1 (*continued*)

2007), and (5) climate record analysis on regional and local scales (Denton et al., 1993, 2002; Sugden et al., 1993; Doran, 1995; Marchant et al., 1996; Doran et al., 2002; Kowalewski et al., 2006; Kowalewski and Marchant, 2007; Swanger and Marchant, 2007) (Fig. 3). It is only through the combined research of workers from a range of permafrost-related disciplines that a complete understanding of the processes, properties, and significance of terrestrial thermal contraction crack polygons has begun to emerge. Accordingly, we first review recent advances in terrestrial thermal contraction crack polygon research in order to provide insight into the range of potential polygon processes observed on Mars. Next, we address the physical processes involved in thermal contraction crack polygon formation, and the efforts to extend our understanding of the critical mechanisms to martian conditions, before linking historical observations of polygonally patterned ground on Mars to recent results from the Mars Reconnaissance Orbiter HiRISE experiment (McEwen et al., 2007a). Finally, we address the future of martian polygon and permafrost investigations by outlining significant outstanding questions and strategies for connecting terrestrial analog research to ongoing planetary exploration.

2. Advances in terrestrial polygon research

Analysis of the morphology and development of polygonally patterned ground across a range of climates and land surfaces on Earth has led to the accumulation of a large body of information that is useful for understanding the origin, significance, and processes operating in martian thermal contraction crack polygons (e.g., Marchant and Head, 2007). Thermal contraction crack polygons represent one end-member landform in the range of "patterned ground" features observed in polar terrains on Earth (e.g., Washburn, 1973; French, 1976). In particular, thermal contraction crack polygons are a form of "non-sorted patterned ground," distinct from sorted circles, nets, and polygons that arrange sediments by surficial freeze–thaw processing (Washburn, 1973). Here, we address the basis for comparisons between thermal contraction crack polygons on Earth and Mars, provide an overview of polygon morphological groups, and address several principle problems in terrestrial polygon research: cracking mechanisms, resurfacing rates and processes, and water transport in polygonally-patterned terrains.

2.1. Advances in thermal contraction cracking

Failure of ice-cemented permafrost in response to stresses generated by thermal contraction has long been implicated in the formation of thermal contraction crack polygons: a process facilitated by the large coefficient of thermal expansion of ice ($\sim 5 \times 10^{-5}$/K) relative to rock ($\sim 5\text{–}10 \times 10^{-6}$/K) (Leffingwell, 1915; Péwé, 1959; Lachenbruch, 1962; Hobbs, 1974; de Castro and Paraguassu, 2004) and the relatively low tensile strength of ice (~ 2 MPa) (Hobbs, 1974; Mellon, 1997). Significant progress in modeling thermal contraction cracking was made by Lachenbruch (1962), work that implicated the rapid cooling of a visco-elastic (rather

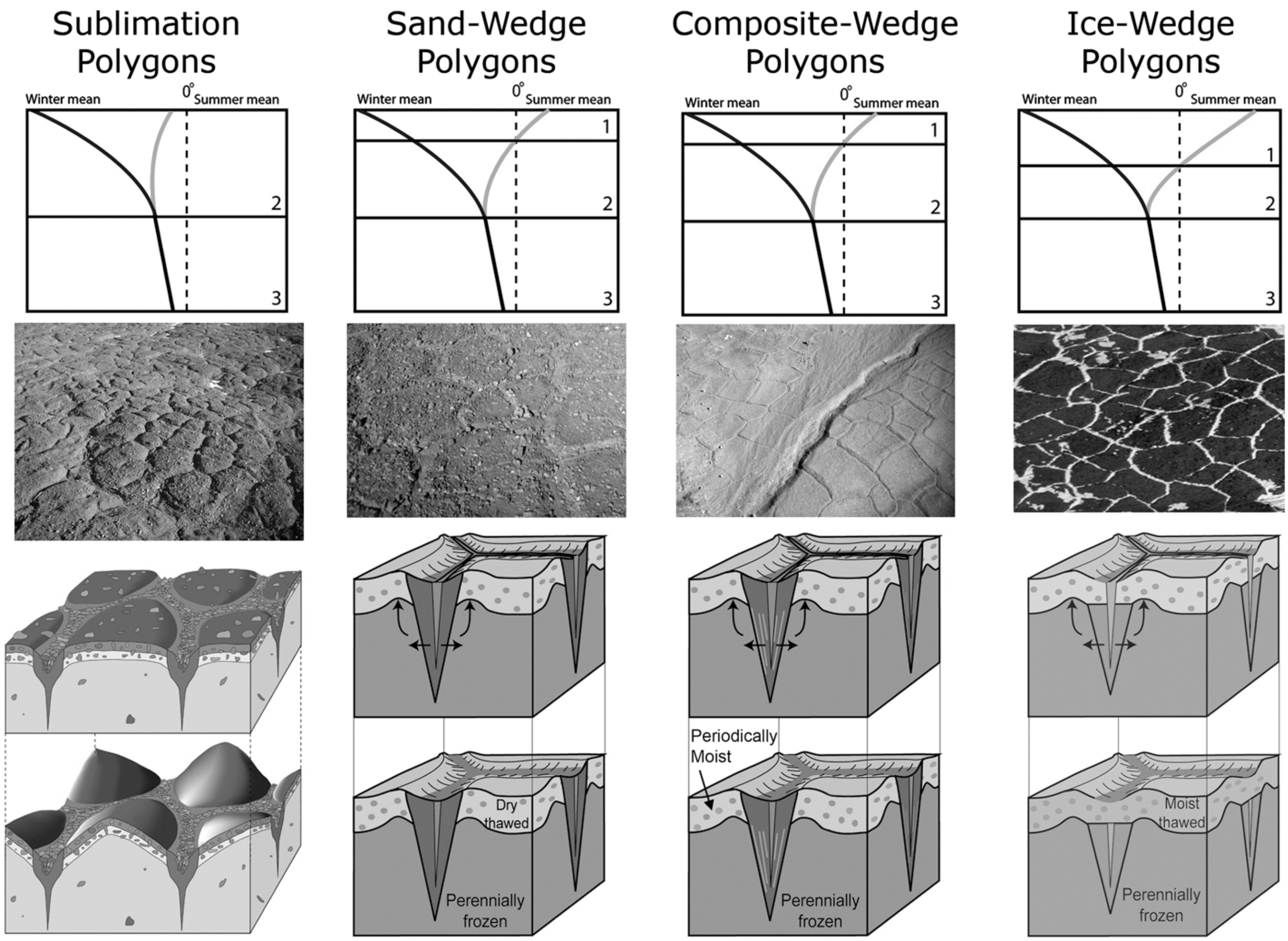

Fig. 2. Polygon morphology and surface thermal profiles as a function of microclimate zone in the Antarctic Dry Valleys (adapted from Marchant and Head, 2007). Top (row 1): schematic showing vertical thermal profiles for polygon-forming substrates. Dashed line represents 0 °C (273 K) baseline; dark and light lines show winter-mean and summer-mean soil temperatures as a function of depth, respectively. Numbered soil "horizons" are defined on the basis of temperature profiles. Horizon 1 is a surface layer that experiences summer temperatures above 0 °C (273 K). In the case of ice-wedge polygons in the coastal thaw zone (CTZ) (right), soils are seasonally moist and thus seasonal oscillation about 0 °C (273 K) produces a classic active layer (see text for discussion). For sand-wedge and composite-wedge polygons in the inland mixed zone (IMZ) (center two columns), soils are typically too dry to produce classic active-layer disturbance, even though summer soil temperatures rise above 0 °C (273 K). Horizon 1 is not present for sublimation polygons (Marchant et al., 2002) in the stable upland zone (SUZ) (left) because mean-summer soil temperatures fail to rise above 0 °C (273 K); this zone lacks a traditional active layer. Horizon 2 reflects the depth to which near-surface materials experience seasonal temperature change. Temperature oscillation results in material expansion/contraction and is responsible for the initiation of polygonal terrain. Horizon 3 reflects a zone of uniform temperature increase with depth; the base of the permafrost would occur where temperatures exceed 0 °C. Middle (row 2). Left: oblique-aerial view of sublimation polygons (Marchant et al., 2002) in the SUZ; field of view (FOV) is ∼200 m. Center-left: oblique-aerial view of sand-wedge polygons in lower Beacon Valley (Marchant et al., 2002); FOV is ∼50 m. Center-right: oblique view of composite-wedge polygons in the IMZ, cross-cut by a seasonally active gully; FOV is ∼75 m. Right, oblique-aerial view of ice-wedge polygons in CTZ; FOV is ∼75 m. Bottom (row 3): block diagrams illustrating the development of sublimation-type polygons (left), sand-wedge polygons (left-center), composite-wedge polygons (right-center), and ice-wedge polygons (right). Sublimation polygons form in an ice-dominated substrate (Marchant et al., 2002), while sand-wedge, composite-wedge, and ice-wedge polygons typically form in ice-cemented sediment (Marchant and Head, 2007).

than elastic) permafrost layer in the generation of thermal stresses sufficient to overcome the tensile strength of permafrost. This concept has been extensively expanded upon by Mellon (1997, 2008), who applied a refined visco-elastic cracking model to martian conditions (showing that thermal contraction cracking at a range of length scales is expected in martian permafrost under current climate conditions); by Plug and Werner (2001, 2002), who expanded on the details of network development, fracture propagation, and the importance of peak cooling events over mean conditions in establishing fracture network morphology; and by Maloof et al. (2002), who demonstrated the importance of a critical ice content in permafrost that drives a critical viscosity in permafrost necessary to propagate surficial fractures to depth. These theoretical advances underscore the concept that polygon morphology is strongly influenced by surface rheological properties responding to climate conditions (Fig. 2). For example, in Beacon Valley, Antarctica (Sugden et al., 1995; Marchant et al., 2002, 2007; Kowalewski et al., 2006; Kowalewski and Marchant, 2007; Kowalewski, 2009) rheological differences between debris-covered glacier ice (Fig. 4) and ice-cemented sediment produce significant differences in polygon morphology: small (∼10–15 m diameter), domical, high-center sublimation polygons (Marchant et al., 2002; Marchant and Head, 2003) form in a debris-covered glacier ice substrate (Sugden et al., 1995; Marchant et al., 2002), while larger (∼15–20 m diameter), low-relief sand-wedge polygons form in an ice-cemented sediment substrate (Fig. 4) under identical surface climate conditions (Marchant and Head, 2007).

In addition to enhancements in the modeling of thermal contraction cracking, additional field investigations have demonstrated that variable initiation and propagation behavior of

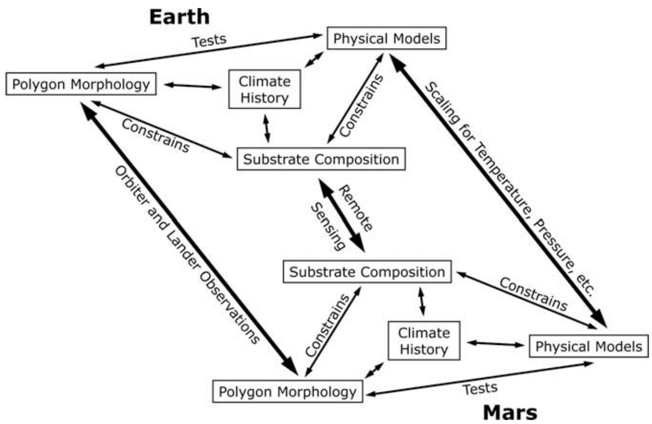

Fig. 3. Schematic illustration of the strategies and goals of thermal contraction crack polygon analysis. Exploration of polygonally patterned ground on Earth requires the combined expertise of geomorphologists, geocryologists, physical modelers, and climate modelers. Each element of polar science provides critical tests and predictions for the associated disciplines. The exploration of martian thermal contraction crack polygons is focused on the goal of interpreting Amazonian climate history. This goal can be accomplished by combined orbiter and lander geomorphological observations, remote sensing and in situ substrate analysis, and the application of physical models describing the response of martian permafrost to climate processes.

thermal contraction cracks is critical to assessing polygon development over time. Long-term studies of ice-wedge cracking have shown that fractures form on annual to inter-annual timescales, but that predicting whether a specific crack will open during a particular year, based on available climate data, remains an elusive goal (Berg and Black, 1966; MacKay, 1992). Further, crack propagation direction (i.e., from the top of buried ice-wedges up or from the ground surface down) has been shown to be a complex function of surface cover, wedge properties, climate, and age (MacKay, 1992, 2000). These observations, coupled, with thaw-slumping in ice-wedge polygons, and non-homogeneous removal of ice-cemented material by sublimation in sublimation and sand-wedge polygons (Marchant et al., 2002; Kowalewski and Marchant, 2007; Kowalewski, 2009), confound efforts to use polygon wedge dimensions or cracking frequency as a metric of polygon age or development (Berg and Black, 1966; MacKay, 1992, 2000; Sletten et al., 2003). One approach, crater counting on polygonally patterned martian surfaces (see Section 4.4) is a planetary solution to the lack of an intrinsic polygon chronometer, and is functionally comparable to terrestrial cosmogenic nuclide exposure age dating used to determine the age of Antarctic permafrost features (e.g., Brown et al., 1991; Brook et al., 1993; Schäfer et al., 2000; Stone et al., 2000; Smith et al., 2001; Marchant et al., 2002; Staiger et al., 2006; Swanger, 2009).

2.2. Polygon forms as a proxy for local microclimate conditions

Detailed geomorphological analysis of terrestrial thermal contraction crack polygons suggests that polygon morphology is strongly diagnostic of local microclimate conditions (e.g., air and ground temperature, atmospheric humidity, and ground/soil water content) (Marchant and Denton, 1996; Marchant and Head, 2004, 2005, 2007; Kowalewski et al., 2006; Kowalewski and Marchant, 2007; Kowalewski, 2009). As features typical of permafrost (ground that has a mean annual temperature <0 °C over inter-annual periods) (Washburn, 1973; French, 1976; Yershov, 1998), terrestrial thermal contraction crack polygon morphology is strongly influenced by the range of periglacial (e.g., non-glacial, cold climate, Washburn, 1973) processes acting over local, regional, and global scales (Washburn, 1973; Bockheim, 2002; Marchant et al., 2002; Kowalewski et al., 2006; Kowalewski and Marchant, 2007; Marchant and Head, 2007), which, in turn, are influenced by local microclimate conditions. Thus, well-developed varieties of contraction crack polygons can be considered proxies for local environmental conditions (Fig 2) and can be used to distinguish between polar climate zones (e.g., Antarctic Dry Valley microclimate zones (Marchant and Head, 2007) or High and Low Arctic environments).

Terrestrial thermal contraction crack polygons are commonly classified by the material properties of the subsurface wedges that form the polygonal network: criteria that are strongly dependent on subsurface rheological properties (including ice content), and climate-driven environmental factors, including the presence or absence of an active layer (the layer of a permafrost surface that is seasonally warmer than 0 °C) and soil moisture (Péwé, 1959;

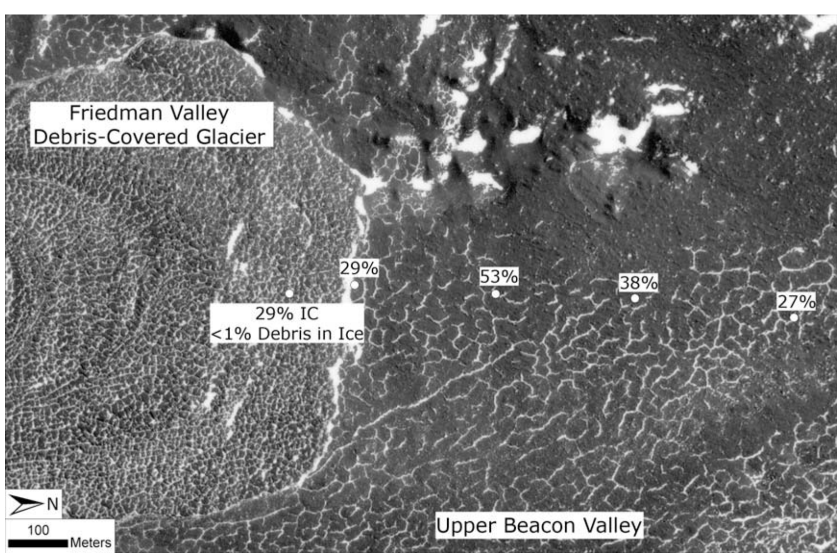

Fig. 4. Sublimation polygons on the Friedman Valley debris-covered glacier (left) and sand-wedge polygons in upper Beacon Valley (right). White circles indicate measurement locations of substrate ice content (typically ice-cemented sediment) made using gravimetric analysis. Ice content is given as the percent ratio of ice mass to sediment mass. Friedman Valley debris-covered glacier ice is nearly pure ice, with a lithic fraction of <1%. "IC" indicates a measurement of ice-cemented sediment capping glacier ice in an adjacent polygon. Cracking substrate rheological properties such as ice content have a strong effect on polygon morphology.

Lachenbruch, 1961; Maloof et al., 2002; Marchant et al., 2002; Kowalewski et al., 2006; Kowalewski and Marchant, 2007; Marchant and Head, 2007; Swanger and Marchant, 2007; Kowalewski, 2009) (Fig. 2). In relatively warm and wet climates, in which a typical, water-saturated active layer is seasonally present, ice-wedge polygons commonly develop as summertime meltwater seeps into seasonally open thermal contraction fractures and freezes (Berg and Black, 1966; Washburn, 1973). In cold and dry climates, in which soil conditions are too arid to develop traditional wet active layers (Bockheim et al., 2007), sand-wedge polygons may form as sand winnows into thermal contraction cracks (Péwé, 1959; Berg and Black, 1966). Composite-wedge polygons form in regions supporting a dry active-layer that receives localized additions of moisture (e.g., from gully-related flow, snowbanks, etc., Murton, 1996; Ghysels and Heyse, 2006; Levy et al., 2008a) and are characterized by subsurface wedges composed of alternating bands of sediment and ice or of ice-cemented sediment (Berg and Black, 1966). Sublimation polygons (Marchant et al., 2002) are a special type of sand-wedge polygon that form where excess ice (ice exceeding available pore space) occurs in the shallow subsurface and is preferentially removed along contraction cracks via sublimation to accommodate growing wedges of ice-free, fine-grained, and winnowed sediment (Marchant et al., 2002; Kowalewski et al., 2006; Levy et al., 2006; Kowalewski and Marchant, 2007; Marchant and Head, 2007; Kowalewski et al., 2008). This physical (climate) control of polygon morphology makes polygons useful indicators of current climate processes in varied polar terrains on Earth, as markers for paleoclimate reconstruction, and as analogs for polygonally-patterned landforms observed on Mars (Fig. 3) (Black, 1976; Gibson, 1980; Marchant and Head, 2007).

2.3. Thermal contraction crack polygon morphology

Each of the types of thermal contraction crack polygon outlined above has distinct morphological characteristics that can be identified during surficial or aerial observations, supported terrestrially by observations of polygon wedge morphology in excavated cross-sections (Figs. 2 and 5) (e.g., Black, 1952; Pewe et al., 1959; Berg and Black, 1966; Marchant et al., 2002; Marchant and Head, 2007). Ice-wedge, sand-wedge, composite-wedge, and sublimation polygons are all classified as "non-sorted polygons" (Washburn, 1973), meaning that wedge-filling material, wedge structure, and polygon micro-topography are more diagnostic of polygon-forming mechanical processes than of a re-distribution of grain sizes (from sand to boulders) by freeze–thaw cycling. Some sediment sorting may occur in ice-wedge and composite-wedge polygons (Péwé, 1959; Berg and Black, 1966; MacKay, 2000; Levy et al., 2008a), and boulders and cobbles may slump from polygon margins and collect in troughs associated with sand-wedge and sublimation polygons (Marchant et al., 2002) (see Section 2.4 for a more complete discussion of cryoturbation and resurfacing).

Ice-wedge polygons (Fig. 2) have a range of surface expressions that are tied intimately to the activity of the wedge, the thermal state of the host permafrost, and in arctic environments, the extent and variety of vegetation growing at the ground surface (MacKay, 2000). Incipient ice-wedges are commonly delineated by a network of fine surface fissures ("frost cracks") (Lachenbruch, 1962; MacKay, 2000). Active ice-wedge polygons are commonly outlined by troughs flanked by raised rims resulting from the lateral and upward displacement of soil or sediment by the growing ice wedge (Berg and Black, 1966; Black, 1982; Yershov, 1998), and are described as "low-centered" polygons accordingly (MacKay, 2000). Because ice-wedge polygons form towards the warmer limit of polygon-forming environments (Gold and Lachenbruch, 1973; Washburn, 1973; Yershov, 1998; Marchant and Head, 2007), the formation of thermokarst (surfaces modified by the melting of ice-rich permafrost) can alter the appearance of degrading ice-wedge polygons, forming unstable, high-centered polygons as wedge ice melts and drains into local streams and ponds, or dramatically walled polygons (e.g., "fortress polygons," Root, 1975) in response to melting and drainage of excess ice in polygon interiors (MacKay, 2000). High-centered, decaying ice-wedge polygons can form in shaley soils in the arctic, where fluvial drainage removes poorly consolidated sediments from melted ice-wedge casts (Jahn, 1983), and are also described in the Transantarctic Mountains, where relict ice-wedges have sublimated away, leaving hollowed, slumping casts behind (Matsuoka and Hiraka, 2006).

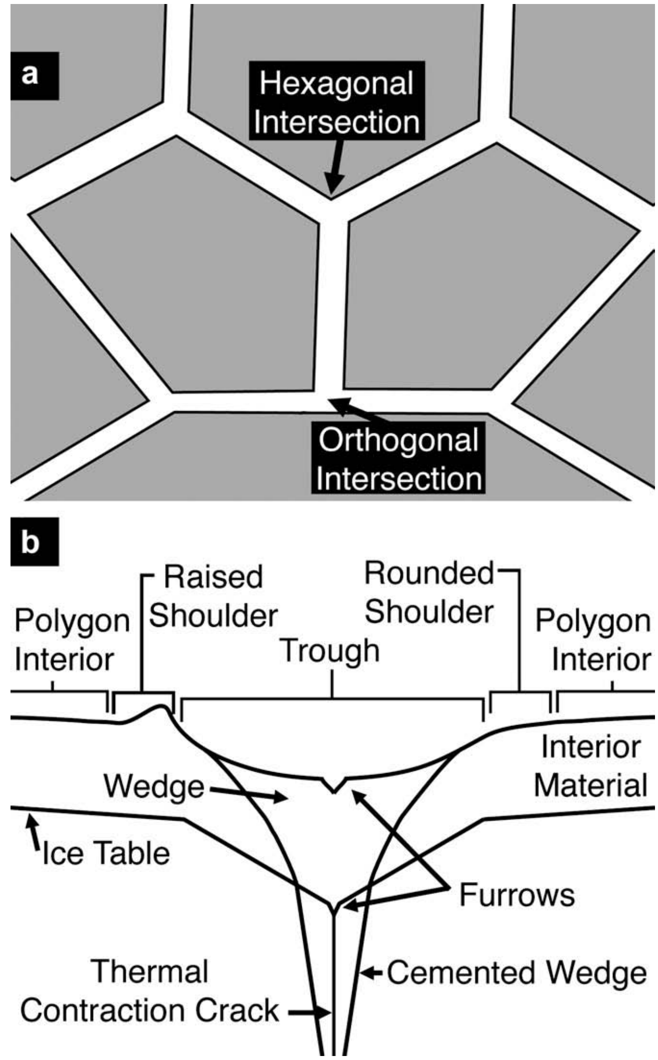

Fig. 5. Schematic illustration of polygon morphological components. (a) Map-view of polygon trough intersection angles. (b) Cross-sectional view of polygon morphological elements.

Active sand-wedge polygons (Fig. 2) can be difficult to distinguish from active ice-wedge polygons, as sand-wedge polygons are commonly characterized by the presence of raised shoulders flanking polygon troughs (Fig. 5) that form due to the re-expansion of warming permafrost pushing against dry sediments that have fallen into open thermal contraction cracks (Péwé, 1959; Berg and Black, 1966; Yershov, 1998; Murton et al., 2000). Infiltration of fines into open cracks commonly produces surface furrows or sand funnels as fines infiltrate from the dry active layer into the subsurface (Berg and Black, 1966; Péwé, 1959). In the subsurface, raised rims initiate as cuspate, upturned strata

(Péwé, 1959); at the surface, rims grow wider (from slumping) and taller (from additional volumetric expansion of the wedge) with increasing age (Berg and Black, 1966). There is considerable disagreement whether growing polygon shoulders can ever coalesce and aggrade into high-centered sand-wedge polygons. Berg and Black (1966) report no evidence of sand-wedge polygon shoulders coalescing into high-centered polygons, consistent with observations by Sugden et al. (1995), Marchant et al. (2002), and Marchant and Head (2007) identifying multiple generations of cross-cutting wedges in some sand-wedge and sublimation polygons. In contrast, Sletten et al. (2003) postulate the presence of a closed-cell cryoturbation cycle (see next section) that transfers wedge material into polygon interiors. Lastly, sand-wedge polygons can be distinguished from ice-wedge polygons, as sand-wedge polygons form in hyper-arid regions with dry active layers (Bockheim et al., 2007; Marchant and Head, 2007), meaning that they rarely form with accessory landforms that require seasonal thaw and active layer saturation, such as pingos, alases (thaw lakes), solifluction lobes, thaw caves, or complex fluvial systems (Leffingwell, 1915; Lachenbruch, 1962; Washburn, 1973; French, 1976; Fortier et al., 2007).

Composite-wedge polygons (Fig. 2) represent an intermediate landform between ice-wedge and sand-wedge polygons, and accordingly, display a range of surface morphologies (Berg and Black, 1966; Murton, 1996; Ghysels and Heyse, 2006; Levy et al., 2008a). Although composite-wedge polygons are commonly observed in exposures of inactive, buried permafrost in the Arctic (Murton, 1996; Ghysels and Heyse, 2006), Antarctic composite-wedge polygons are currently forming under present microclimate conditions, providing insight into active composite-wedge polygon development (Berg and Black, 1966; Levy et al., 2008a). Composite-wedge polygons include examples of entirely flat lying microtopography, slightly high-centered polygons, and low-center polygons with gently raised shoulders (Berg and Black, 1966; Levy et al., 2008a). An important aspect of Antarctic composite-wedge polygons is that they form in microenvironments in which seasonally produced meltwater is ephemerally present in locally concentrated, integrated reservoirs, such as snow packs, gullies, and hyporheic zones, providing a range of accessory landforms that can be used to distinguish composite-wedge polygons from sand-wedge or ice-wedge polygons (McKnight et al., 1999; Head et al., 2007a; Levy et al., 2008a).

Active sublimation polygons (Fig. 2) consist of high-centered to domical, mounds of sediment-covered ice, bounded by deep, non-raised-rim troughs (Berg and Black, 1966; Marchant et al., 2002; Marchant and Head, 2007). Sublimation polygons form in the absence of a liquid phase (Marchant et al., 2002). The presence of massive, sublimating ice in the polygon-forming substrate allows sublimation to balance or exceed lithic input to the growing sediment wedge (Marchant et al., 2002). Coupled with the trapping of coarse grains at the fracture surface, complete filling of thermal contraction cracks is minimized, preventing the formation of raised shoulders in sublimation polygons (Berg and Black, 1966; Marchant et al., 2002). Sublimation polygons may form on ice-cored moraines, atop stranded glacier ice, or on debris-covered glaciers (Berg and Black, 1966; Marchant et al., 2002). Sublimation polygon troughs may be oriented by debris-covered glacier flow-related stress (Levy et al., 2006), and commonly develop from fine surface fissures to depressed troughs tens of cm to >1 m deep (Marchant et al., 2002; Kowalewski and Marchant, 2007; Kowalewski, 2009). Surface furrows may be present in active sublimation polygon troughs, indicating the winnowing of fine material into open subsurface fractures (Marchant et al., 2002). More commonly, however, mature troughs are surfaced by gravitationally-slumped boulders, cobbles, or a coarse gravel lag too large to enter the wedge (Marchant et al., 2002) (Fig. 2).

2.4. Resurfacing and polygons

Modification of polygonally-patterned surfaces by polygon-forming processes such as continued fracturing, thermokarst formation, gravitational slumping, sublimation of subsurface ice, etc. all combine to continually change the initial record of thermal contraction crack processes in polar landscapes. Ongoing fracturing and ice melting/sublimation alter the composition and microtopography of polygon interiors, while gravitational slumping and thermokarst formation tend to alter the shape of polygon troughs. Understanding the mechanisms that degraded and rework polygonally-patterned surfaces is critical for assessing the age and duration of polygonally-patterned landscapes, a process complicated by the lack of a reliable chronometer intrinsic to the polygon-forming process (see previous section).

In regions with traditional (saturated) active layers, annual freeze–thaw cycles produce a range of archetypical landforms indicative of reworking of surface soils and rock clasts by ice-related processes (cryoturbation) (Washburn, 1973; French, 1976; Bockheim and Tarnocai, 1998), including frost-heaved/jacked blocks and complexly sorted sediments and rocks. Cryoturbation is a disruptive pedogenic process, and results in mixing of the ground surface by the displacement, rotation, and sorting of sediment particles (Fox, 1994). Cryoturbation is driven by inhomogeneities in the freezing, thawing, and thermal expansion of water in the near-surface (Fox, 1994). Long-term warming and thawing of ice-cemented permafrost in tundra environments generates water-saturated conditions that drive cryoturbation of wet active layers, resulting in the reworking of polygonally-patterned surfaces, such that much of the Earth's seasonally-active permafrost is less than several tens to hundreds of thousands of years old (Gold and Lachenbruch, 1973; Washburn, 1973; Denton et al., 1989, 1993; Yershov, 1998; Sletten et al., 2003) (although recent reports of ice-wedges approaching ~700 ka in age suggest that preservation of deeply buried permafrost, below saturated active layers, may extend through some glacial/interglacial cycles, e.g., Froese et al., 2008). This polewards retreat of terrestrial permafrost environments since the last glacial maximum is strikingly analogous to changes in the distribution of ice stability thought to have occurred due to obliquity variations in martian history (Jakosky and Carr, 1985; Mellon and Jakosky, 1995; Laskar et al., 2002, 2004; Head et al., 2003; Forget et al., 2007; Marchant and Head, 2007).

Traditional cryoturbation is a water- and ice-activated process, and accordingly, it is most strongly observed in poorly drained, silty or peaty soils supporting a wet active layer (Bockheim and Tarnocai, 1998; MacKay, 2000). Cryoturbation in saturated sediments is thought to arise from convective cycling of sediments driven by the propagation of freezing fronts through the active layer (from the surface downwards, and from the ice table upwards) (Vliet-Lanoe, 1991). Convective behavior accounts for the formation of a range of sorted soil and rock structures driven by ice-lensing (Kessler and Werner, 2003). A half-cell of convective cryoturbation behavior has been observed in peat-surfaced ice-wedge polygons (MacKay, 2000). MacKay (2000) shows that the long-term contraction of ice-rich active layer soils (with up to ten times greater ice mass than soil mass) is constrained by the polygon center, while the expansion of the active layer soil is unconstrained by the free surface at the polygon margin, resulting in a net transport of active layer material towards the polygon margin as unmelted permafrost continues to expand through the summer beneath the melting front. Laterally transported material accumulates in raised polygon shoulders or falls into polygon troughs, where it locally increases the height of the permafrost, resulting in the formation of secondary, syngenetic wedges (wedges than develop in a thickening permafrost sequence and are the same age as the host medium) (MacKay, 2000).

Can similar processes operate on sand-wedge polygons forming in the absence of a seasonally saturated active layer? A "dry active layer" is a permafrost soil horizon that is seasonally warmed to above 0 °C under arid climate conditions that do not result in the saturation of the warmed soil with meltwater (e.g., Marchant et al., 2002; Marchant and Head, 2007; Kreslavsky et al., 2008). Major differences between the ice-wedge case and the sand-wedge case include: (1) the sand-wedge polygons lack a wet active-layer that can be shuttled along on top of an expanding and melting ice table (the "piggyback" process of MacKay, 2000); (2) the sand-wedge polygons typically form in ice-cemented permafrost with a significantly lower ice content and thermal expansion coefficient than ice-wedge polygons in wet, peat-rich environments; and (3) a convective return flow (closed-cell cryoturbation) has been proposed for such sand-wedge systems, resulting in the elevation of the height of polygon centers (Sletten et al., 2003). In contrast, material transported wedge-wards in the ice-wedge case accumulates as raised shoulders or within the troughs, with any measured polygon center height or volume increases attributed to frost-heave of markers relative to subsidence of melting ice-wedges (MacKay, 2000).

The closed-cell cryoturbation process can be tested by examining cross-sections of active and relict sand-wedges. Excavations into active Antarctic sand-wedge polygons show a wedge cross-section characterized by massive or foliated sands and cuspate, upturned polygon margins, typical of the raised-shoulder morphology observed in ice-wedge polygons, rather than looped horizons generated by closed-cell convection (Péwé, 1959; Murton et al., 2000; Marchant et al., 2002; Marchant and Head, 2007; Murton and Bateman, 2007; Kowalewski, 2009). Well-defined sand-wedges, rather than fully mixed sediment horizons, are typical of active sand-wedge polygons found at high latitudes (e.g., Antarctic examples reported by Sugden et al., 1995; Marchant et al., 2002; Marchant and Head, 2007; Kowalewski, 2009; Swanger, 2009) as well as of relict sand-wedges present at low latitudes (Murton, 1996; Maloof et al., 2002; Murton et al., 2000; Kovacs et al., 2007).

In contrast to extensively disturbed sediments, some sublimation polygons form on extremely stable sediment surfaces. Although some lateral movement of sediments by slumping can occur at the margins of sublimation polygons (Berg and Black, 1966; Marchant et al., 2002), the vertical ablation of underlying massive ice results in a stable lowering of the permafrost surface dominating the pedogenic profile (Marchant et al., 2002; Bao et al., 2008). The un-churned nature of sediments atop sublimation polygons in Beacon Valley, Antarctica, is evidenced by profiles of cosmogenic ^{3}He obtained from cobbles in till overlying buried glacier ice; measured concentrations in subsurface clasts are deficient in ^{3}He relative to values expected by taking only shielding from overlying till into account (Marchant et al., 2002). This deficiency indicates that clasts have moved upwards relative to the ground surface as overlying ice sublimes; rates of ice sublimation calculated from these data (e.g., Schaefer et al., 2000) are on the order of ~10–50 m Ma^{-1}. Had cryoturbation occurred during this process, clasts at the surface and at depth would likely show similar ^{3}He abundances, reflecting the up and down vertical movement of sediment in a cryoturbated soil column (Marchant et al., 2002). Evidence for stable surfaces in regions with sublimation polygons also comes from the observation and radiometric dating of in situ volcanic ash trapped in polygon troughs—some of which has escaped cryoturbation for up to 8.1 Ma (Sugden et al., 1995; Marchant et al., 1996). Morphological evidence for stable surfaces may even apply to regions with sand-wedge polygons in Beacon Valley: evidence comes from the long-term preservation of meter-scale moraines of Quaternary and Pleistocene age (Fig. 6) emplaced atop sand-wedge polygons. Together, these lines of evidence suggest that the surfaces of sublimation polygons and even sediment-starved sand-wedge polygons may be exceptionally stable, and may not be rapidly reworked by "dry" cryoturbation processes (e.g., Marchant et al., 2002; Marchant and Head, 2007; Kreslavsky et al., 2008). We will return to this concept in analysis of ongoing research at the NASA Phoenix landing site (Section 5).

2.5. Polygons and polar hydrology

Although the occurrence of ice-wedge polygons in well-developed lacustrine and riparian systems has been extensively documented in the Arctic (Leffingwell, 1915; Hopkins, 1949; Lachenbruch, 1962, 1966; Plug and Werner, 2002), the role that polygons themselves play in the initiation of polar fluvial systems is only beginning to come to light (and is of particular interest on Mars, where geologically recent, fluvial gully systems are present) (e.g., Malin and Edgett, 2000; Dickson and Head, 2008; Head et al., 2008). Fortier et al. (2007) have documented the rapid growth of Arctic fluvial drainages or streams, in which ice-wedges melt to form ice caves, and then open channels, resulting in the incision of streams up to 750 m long over the span of four years. These channels may continue to expand, collecting and transmitting regional runoff as part of an integrated hydrological system (Fortier et al., 2007). Analogously, composite-wedge polygons in the Antarctic Dry Valleys have been shown to enhance the formation of gullies, promoting the accumulation of snow (that melts during peak summer heating), in addition to accumulation and transport of gully-related water (channeling melt through polygon troughs) (Levy et al., 2008a). In the Antarctic case, the melting of ground ice is not a significant contributor to gully stream flow (Head et al., 2007b; Morgan et al., 2007; Dickson et al., 2008; Levy et al., 2008a; Morgan, 2009). Rather, ice-cemented sediments provide a meters-thick impermeable substrate over which gully-related liquid water flow occurs during peak summer conditions (Head et al., 2007b; Levy et al., 2008a). This recent work suggests that thermal contraction crack polygons are not merely passive landscape elements, but rather, are important parts of developing cold desert landsystems, providing positive feedbacks to the formation of a range of features of geological, hydrological, and astrobiological interest (Levy et al., 2008a, 2009d).

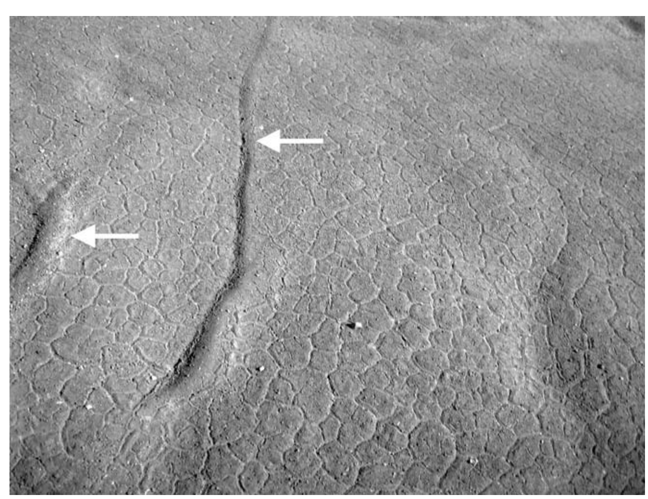

Fig. 6. Moraines (arrows) and sand-wedge polygons in lower Beacon Valley. Moraines are Pliocene and younger (Marchant et al., 2002). Field of view ~200 m wide.

2.6. Applying terrestrial polygon knowledge to Mars

Before proceeding further with discussion of martian thermal contraction crack polygons, it is important to take full stock of the limitations inherent in applying terrestrial analog analyses to features observed on Mars (e.g., Marchant and Head, 2007); by fully understanding the differences between terrestrial analogs and planetary subjects, we gain a greater appreciation of many similarities between the two (Fig. 3). Field access to terrestrial Arctic and Antarctic thermal contraction crack polygons provides considerable accumulated experience with, and insight into, polygon surface morphology, subsurface structure, and development over time. In contrast, analysis of martian polygonally patterned ground is limited to remote observations (Mangold et al., 2004; Mangold, 2005; Levy et al., 2009c) and to a small number of widely spaced surface experiments with capabilities to trench and explore the upper ~20 cm of dessicated surface soils, but with limited ability to vigorously excavate through entire polygon structures (Mutch et al., 1976, 1977; Smith et al., 2008, 2009).

Extrapolating from terrestrial examples to martian conditions comes with the inherent risks associated with applying insights from familiar environments into alien conditions. This challenge is compounded by the possibility of observing preserved relict landforms on the martian surface that formed under conditions different from those currently prevailing (e.g., Marchant and Head, 2007). Even under current conditions, the precise role of seasonal CO_2 frost deposition (e.g., Smith et al., 1998) in polygon-related processes is only beginning to be understood, and may have profound effects on soil microphysics (compaction or agitation of clasts), water vapor diffusion (limiting fluxes in or out of soils), and seasonal thermal response (insulating surfaces until CO_2 frost is removed).

Beyond exploration constraints, the differences in the physical environment between terrestrial and martian thermal contraction crack terrains necessitates careful application of models—particularly models containing any empirically-derived constants that may be applicable under terrestrial polar conditions (e.g., Hobbs, 1974) but which lose accuracy under the extreme cold and low pressures of martian circumpolar terrains. For example, Mellon (1997) is judiciously conservative in extrapolating from terrestrial to martian conditions, taking account of factors requiring extrapolation, such as lower mean annual temperatures on Mars leading to more brittle permafrost (less thermally-activated viscous relaxation) that generates higher thermal contraction stress per Kelvin cooling than would be expected on Earth. Likewise, recent improvements to vapor diffusion modeling have expanded the treatment of the effects of Knudsen diffusion of water vapor through martian till, which is critical for constraining rates of sublimation of water ice present beneath martian regolith—effects that are more important in martian permafrost modeling than on Earth (where Fickian diffusion dominates, e.g., Kowalewski et al., 2006; Kowalewski and Marchant, 2007; Kowalewski, 2009) due to low martian atmospheric pressure (Helbert et al., 2009; Ulrich, 2009). Additionally, detailed physical treatment of the mechanical properties of a range of martian polygon-forming substrates is only just beginning to expand from models of regolith cemented with pore ice (Mellon et al., 2009) into understanding of the interactions between pore-ice-cemented sediments, potential massive ice deposits, etc. (e.g., Head et al., 2003; Schorghofer, 2007). Even when modeling and observational work are done with the utmost care, terrestrial field experience cautions that microclimatic and even meter-scale heterogeneities in permafrost ice content, soil composition and origin, ice distribution, proximity to related landforms (e.g., streams, lakes), and age can have pronounced effects on polygon morphology and development (Bockheim, 2002; Marchant et al., 2002; Bockheim et al., 2007; Gooseff et al., 2007; Marchant and Head, 2007; Levy et al., 2008a). Many of these properties are not easily detected in orbital observations, underlining the importance of correlating observations of polygonally patterned ground on Mars between multiple high-resolution datasets.

An alternative approach to investigating landform processes occurring under current Mars conditions (e.g., exceptionally low mean annual temperature and low atmospheric water vapor pressure) is to use observed landforms to drive interpretations of past climate conditions (e.g., Seibert and Kargel, 2001; Page, 2006; Soare et al., 2008). This relict morphology approach is particularly effective when combined with detailed analyses of martian orbital-element-driven climate history (e.g., Laskar et al., 2002; Forget et al., 2006) and can be used to infer the past occurrence of surface conditions different from the present over 10–20 Ma timescales, such as active layer processes (Kreslavsky et al., 2008).

3. A brief summary of patterned ground on Mars

Exploration of martian permafrost, and martian polygonally patterned ground in particular, has occurred at an ever-accelerating rate as spacecraft observations have improved the resolution, coverage, and variety of observations of the frozen martian surface. Beginning with early theoretical, telescopic, and terrestrial analog investigations (e.g., Lowell, 1907; Wade and De Wys, 1968; Ingersoll, 1970; Anderson et al., 1972; Morris et al., 1972), research has focused on assessing the age, origin, distribution, and ice-content of martian permafrost— in order to assess its geological, hydrological, and biological implications (Lederberg and Sagan, 1962; Gilichinsky et al., 1992, 2007; Mellon, 1997; Feldman et al., 2002; Dickinson and Rosen, 2003; Mitrofanov et al., 2007a,b; Kreslavsky et al., 2008; Levy et al., 2009c, 2009d; Marchant and Head, 2010).

Early results, driven largely by the Viking landers and orbiters, established the presence of a range of permafrost and polygonally-patterned landforms on Mars, including potential frost cracks, ground ice, sorted sediments, gelifluction lobes, and glacial features (Mutch et al., 1976, 1977; Squyres, 1978, 1979; Lucchitta, 1981; Squyres and Carr, 1986). Large-scale fracture patterns, such as those observed in Utopia Planitia, hinted at the possibility of widespread thermal contraction crack activity on the martian surface, however, some large-scale (hundreds of meter) fracture features have been shown to not be ice-related (Hiesinger and Head, 2000; Lane and Christensen, 2000), while others remain as candidate frost features (Abramenko and Kuzmin, 2004).

The arrival of the Mars Global Surveyor (MGS) (1997–2006), Mars Odyssey (MO) (2001-present), and Mars Express (MEx) (2003-present) spacecraft in martian orbit initiated a revitalized age of exploration of martian permafrost. High resolution image data from the Mars Orbiter Camera (MOC) (Malin and Edgett, 2001) produced unprecedented views of martian polygonally patterned ground (Malin and Edgett, 2001; Seibert and Kargel, 2001; Milliken et al., 2003; Mangold, 2005; Kostama et al., 2006), supported by a range of topographic (Zuber et al., 1998; Jaumann et al., 2007) and geophysical datasets such as neutron spectrometer/gamma-ray spectrometer water-equivalent hydrogen measurements (Boynton et al., 2002; Feldman et al., 2002; Kuzmin et al., 2004; Mitrofanov et al., 2007a,b) that directly implicated the presence of water ice in the upper meter of martian polygonally-patterned surfaces (Mangold et al., 2004). Global-scale water–ice distribution has been refined using thermal inertia and surface temperature data (Bandfield, 2007; Bandfield and Feldman, 2008; Feldman et al., 2008), broadly confirming the detection of near-surface water ice (ice-rich permafrost) and permitting the accurate prediction of ice table depths at polygonally patterned sites (Sizemore and Mellon, 2006; Bandfield, 2007; Mitrofanov et al., 2007a; Mellon et al., 2008a).

Perhaps the most striking result to arise from MGS/MO/MEx observations of martian permafrost terrain was the observation that ice-related landforms on the martian surface (e.g., polygons, scalloped depressions, pits) are not randomly distributed, but rather are found concentrated in latitude-dependent deposits (Kreslavsky and Head, 2000; Mustard et al., 2001; Head et al., 2003; Milliken et al., 2003; Kostama et al., 2006; Soare et al., 2007). Polygonally-patterned surfaces on Mars are typically geologically recent units (less than several Ma) that drape underlying terrain, smoothing topography, and typically span continuously from high latitudes equatorwards to ~60°, growing patchier and more dissected from ~60° to ~30°, equatorwards of which they vanish (Kreslavsky and Head, 1999, 2000, 2002; Mustard et al., 2001; Head et al., 2003; Milliken et al., 2003; Mangold, 2005; Kostama et al., 2006; Levy et al., 2009c). These collected observations have been interpreted to indicate the presence of a martian latitude-dependent mantle (LDM) (Kreslavsky and Head, 2000; Mustard et al., 2001; Head et al., 2003).

The unprecedented increase in the coverage and resolution of spacecraft image and topographic data provided by MGS/MO/MEx led to multiple opportunities to test hypotheses regarding the physical state of martian polygonally-patterned permafrost, addressing critical questions regarding the role of a martian active layer (the portion of permafrost that is seasonally warmer than 0 °C). Although observations of degraded martian craters and slopes strongly suggest the presence of a martian active layer during obliquity highs more than 5–10 Ma before the present (Kreslavsky et al., 2008), climate and surface temperature balance models (Ingersoll, 1970; Haberle et al., 2001; Forget et al., 2007) indicate currently frigid global conditions (Kreslavsky et al., 2008) and appear at odds with reports of recent (<~2 Ma) active layer or thermokarst processes in LDM-related terrains (e.g., Page, 2006). Isolating the evidence for or against active layer processes in the polygonally-patterned portions of the martian LDM is critical for determining the role of liquid water in geologically recent surface processes.

Combined analysis of polygon morphology (Mangold, 2005; Levy et al., 2009c), water-equivalent-hydrogen (Boynton et al., 2002; Feldman et al., 2002; Kuzmin et al., 2004; Mitrofanov et al., 2007a,b), surface thermal response (Bandfield, 2007; Bandfield and Feldman, 2008; Feldman et al., 2008), and Phoenix lander soil data (Smith et al., 2008, 2009) all strongly indicate the presence of ice in the martian latitude-dependent mantle (LDM). But what is the origin of the martian LDM and how did it become so ice-rich? Morphological and stratigraphic evidence strongly indicates that the LDM formed through the atmospheric deposition of dusty ice during recent "ice age" periods of high obliquity, producing beds of potentially massive, dirty ice (Mustard et al., 2001; Head et al., 2003; Laskar et al., 2004; Schon et al., 2009b) that have subsequently partially sublimated to produce protective lag deposits (Head et al., 2003; Schorghofer and Aharonson, 2005; Schorghofer, 2007). In contrast, ongoing modeling of average-climate-driven martian water/ice stability indicates that global water ice deposits observed by MGS/MO are broadly (although not entirely) in diffusive equilibrium with the martian atmosphere, raising the possibility that martian permafrost is dominated by diffusively emplaced pore ice (Jakosky and Carr, 1985; Mellon and Jakosky, 1995; Mellon et al., 2004; Hudson et al., 2009a; Mellon et al., 2009), or possibly near-surface shells of diffusively-emplaced excess ice (ice volume exceeding dry pore space) (Fisher, 2005). Distinguishing between these two end-member ice-depositional modes (primary precipitation versus diffusive emplacement) is critical for understanding the magnitude and significance of martian ice-rich, polygonally-patterned terrains.

4. New observations and results from MRO/HiRISE

The High Resolution Imaging Science Experiment (HiRISE) has provided enhanced viewing of martian polygonally patterned ground at a spatial resolution up to an order of magnitude greater than that of previous instruments (McEwen et al., 2007a,b), permitting viewing of polygonally patterned ground with a clarity previously found only in lander-based image data (Mutch et al., 1976; Smith et al., 2008, 2009). Meter-scale resolution data provide links among local sites observed around landers to broader-scale martian environments, and permit unprecedented comparison of martian terrain with terrestrial analog field observations. HiRISE image analyses have addressed several of the most pressing questions raised by previous studies of martian polygonally patterned ground. Here we address advances in the current understanding of small-scale (<~25 m) martian thermal contraction crack polygon morphology, the nature of the ice component in martian permafrost, the role of water in martian polygon development, the age and distribution of polygonally-patterned surfaces on Mars, and insights derived from analysis of interactions between polygons and other ice-related martian landforms.

4.1. Polygon morphology

The high spatial resolution of HiRISE image data (25 cm/pixel) (McEwen et al., 2007a) has greatly increased the detail discernable in images of martian thermal contraction crack polygons (Fig. 1). Not all polygons observed in HiRISE images have been interpreted to form from the thermal contraction of ice-rich permafrost. For example, Page (2006) interprets Marte Valles patterns to indicate sorted polygon formation, while Osterloo et al. (2007) interpret chloride-related fractures as desiccation cracks. All the features subsequently identified as martian thermal contraction crack polygons in this paper meet the criteria outlined by Levy et al. (2009c) for distinguishing thermal contraction crack polygons from other natural network features (such as desiccation cracks, lava-related cooling cracks, or bedrock jointing) (see also Fig. 7).

A variety of morphological components of thermal contraction crack polygons, including surficial troughs, mounded centers, and raised shoulders (where present), are discernable in HiRISE images (Figs. 1 and 5), permitting detailed analysis of polygon characteristics (McEwen et al., 2007b; Mellon et al., 2007; Levy et al., 2008c, 2009b,c; Mellon et al., 2009). Polygonally patterned ground featured in HiRISE data is more clearly resolved than in MOC data, permitting the identification of a range of small-scale thermal contraction crack features <~25 m in diameter. For example, polygons typical of northern polar regions near the Phoenix lander average ~3–6 m in diameter as measured by GIS (Levy et al., 2009c; Mellon et al., 2009) and automated methods (Pina et al., 2009). These features are very similar in size to terrestrial thermal contraction crack polygons (Lachenbruch, 1961; Marchant et al., 2002; Marchant and Head, 2007), and represent the smaller end of the range of polygon diameters observed at MOC resolution (e.g., Mangold (2005) reports a size range for "small" polygons of 10–50 m). Martian polygon troughs have been shown through HiRISE stereo image analysis (and confirmed by Phoenix lander data) to be typically less than several tens of cm deep (Arvidson et al., 2008; Smith et al., 2008, 2009).

Beyond yielding single morphometric properties such as size, low sun-angle HiRISE image data have revealed the complex three-dimensional morphology of martian polygons in exquisite detail (McEwen et al., 2007a), comparable to air photographs used to map terrestrial polygons (Black, 1952). Across the latitude-dependent mantle (Head et al., 2003), martian thermal contraction crack polygons are overwhelmingly high-centered (Fig. 1) (Levy

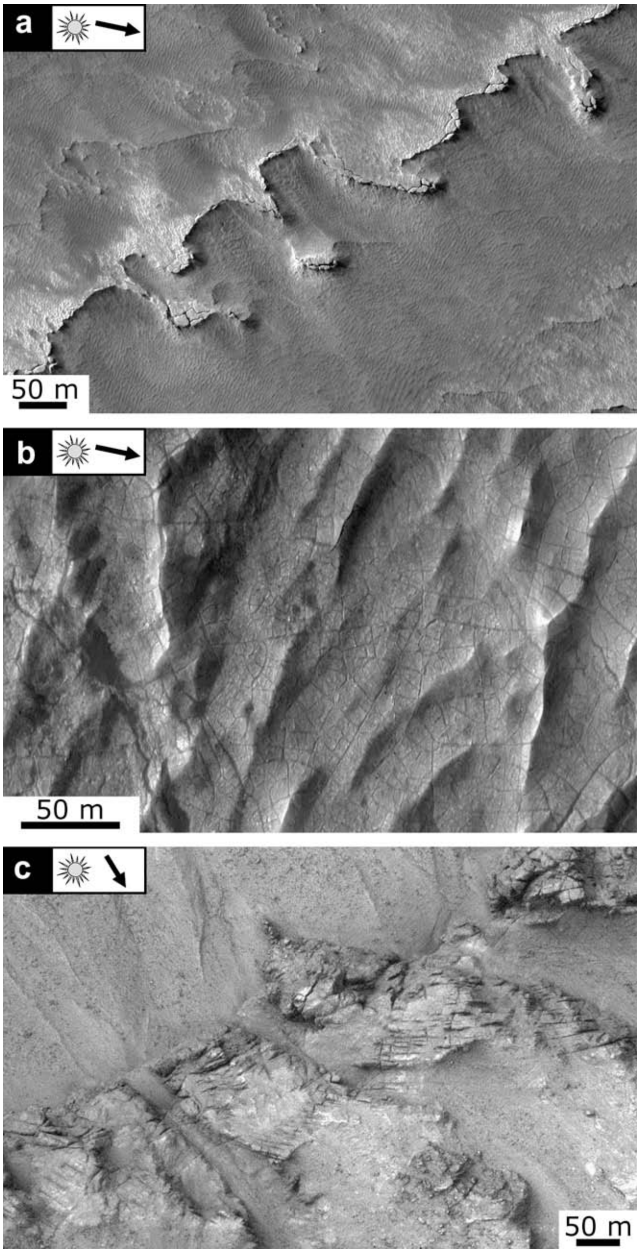

Fig. 7. Fracture networks that do not meet the criteria outlined by Levy et al. (2009c) for identification of permafrost-related thermal contraction crack polygons. (a) Desiccation-like fractures in chloride-bearing deposits (Osterloo et al., 2007). Portion of HiRISE image PSP_003160_1410. (b) Bedrock jointing located neat the Argyre Basin. Portion of HiRISE image PSP_003934_1275. (c) Fractures identified as columnar joints by Morgan et al. (2009) in Asimov Crater. Portion of PSP_010710_1320.

et al., 2009c), ranging from flat-topped polygons with depressed bounding troughs, to dramatically pitched, nearly conical peak-topped polygons. Polygons vary in trough intersection angle and the degree to which troughs are networked over hundreds of meter length scales. In addition, morphologically different groups of polygons (e.g., Levy et al., 2009c) are commonly present in a single HiRISE image, suggesting that local heterogeneities in the polygon-forming substrate are important controls on martian polygon morphology (as on Earth, e.g., Fig. 4). Conversely, morphological groups of polygons can be mapped across martian mid-to-high latitudes in both northern and southern hemispheres (Fig. 1), suggesting that global scale climate properties such as surface temperature (and accordingly, insolation) strongly affect the distribution and development of martian polygons (analogous to climate controls on transitions between polygon types on Earth) (e.g., Marchant and Head, 2007). Alternatively, transitions from high latitude, flat-topped polygons, to mid-latitude peak-top polygons, to lower-mid-latitude subdued polygons (Fig. 1) may reflect a maturation of polygon micro-topography due to ice loss proceeding from polygon margins to polygon interiors over time (Fig. 2, left column, and Fig. 8) (Marchant et al., 2002; Mangold, 2005). Such a maturation sequence is broadly consistent with increasing polygon network age from the poles to the equator (Levy et al., 2009c), and is consistent with more recent latitude-dependent mantle emplacement at higher latitudes (Kostama et al., 2006; Kreslavksy, 2009; Schon et al., 2009b).

4.2. Ice distribution and origin in martian polygons

What do the morphological characteristics of martian polygons observed with HiRISE indicate about the amount, distribution, and origin of ice in polygonally-patterned portions of the martian LDM? (Lefort et al., 2009; Levy et al., 2009c,d; Mellon et al., 2009). In particular, can martian polygon morphology be used to

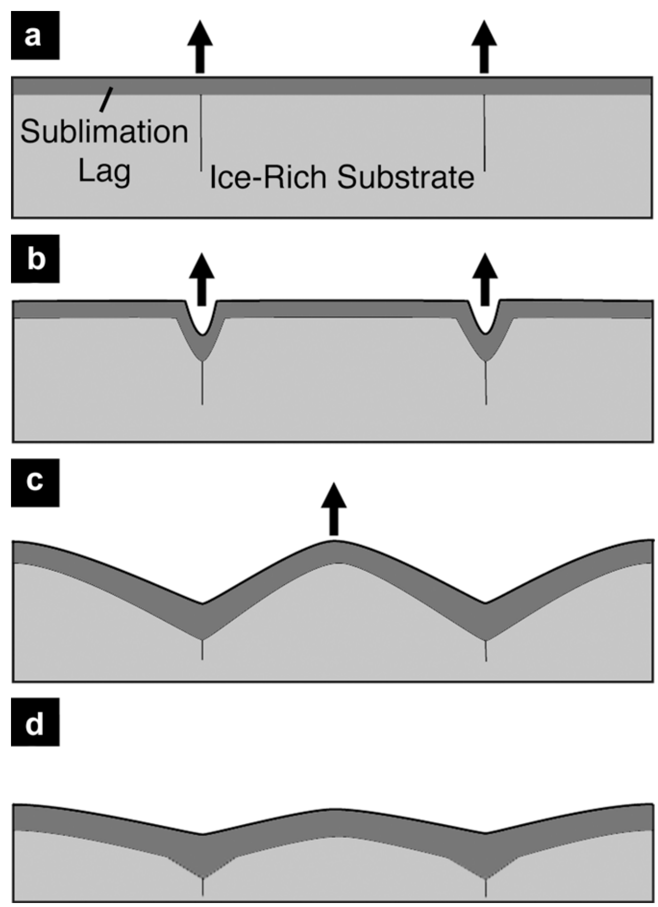

Fig. 8. Schematic illustration of a potential sequence for martian polygon micro-topography development, after Mangold (2005). Arrows indicate loci of increased sublimation due to disrupted and thinned till cover, and/or increased ice surface area in contact with a dry atmosphere. (a) An ice-rich substrate (light gray) covered by a sublimation lag deposit (dark gray) is fractured by thermal contraction (dark lines), focusing sublimation along cracks. (b) Flat-topped, small polygons develop as ice is preferentially removed along polygon margins. (c) Continued ice removal along polygon margins produces peak-topped and highly domical sublimation polygons. In response to thickened till cover at polygon margins, and slumping of till on steep ice slopes, sublimation shifts towards the polygon interior. (d) Subdued polygons result from loss of polygon core ice, producing little microrelief as the ice surface approaches an equilibrium level across the polygonally patterned surface.

distinguish meters-thick accumulations of massive ice (Head et al., 2003) from pore-ice-cemented terrain (Mellon et al., 2008a,b)? From terrestrial examples, high-centered polygons are strongly indicative of the removal of excess ice at polygon margins (Jahn, 1983; MacKay, 2000; Marchant et al., 2002; Matsuoka and Hiraka, 2006; Levy et al., 2008c, 2009c). There are two models that can produce the relief typical of high-centered polygons, and each requires very different subsurface-ice concentrations and melting or sublimation histories. For an extremely cold and dry Mars, e.g., without significant melting, high-centered polygons can form in regions with widespread excess ice if sublimation preferentially removes ice at polygon margins/thermal-contraction cracks (a model initially described for Beacon Valley, Antarctica, by Marchant et al., 2002). In contrast, high-centered polygons can also form in regions with limited excess ice in which ice is locally concentrated at the margin of ice-wedge polygons (e.g., MacKay, 1990). In this case, local melting of the ice at polygon margins can create deep troughs that surround high-centered polygons. Localized thermokarst of this type has been used to infer recent warming for some sites in the Arctic (MacKay, 2000; Matsuoka and Hiraka, 2006). Can these two alternatives be distinguished with HiRISE image data?

The morphology of clusters of small, high-centered and flat-topped polygons (the "flat-top small" group of Levy et al., 2009c) (Fig. 1) can help distinguish excess ice in the polygon-forming substrate, from excess ice concentrated at polygon troughs. Between ~60° and ~70° (a latitude range containing the Phoenix lander) in both hemispheres, flat-top small polygons are commonly present in clustered groups on topographically high mounds or knolls (~20–30 m in diameter) (Levy et al., 2008c, 2009c; Mellon et al., 2009). HiRISE stereo image topographic profiles (Kirk et al., 2007) across these polygonally-patterned knolls indicate that the knolls have pole-facing (north-facing) slopes that are approximately twice as steep as their equator-facing (south-facing) slopes (Fig. 9), an asymmetry typical of individual sublimation polygons and "megagon" groups of sublimation polygons present in Beacon Valley, Antarctica (Figs. 2 and 10) (Marchant et al., 2002). Levy et al. (2008c) interpreted the observed slope asymmetry in the martian polygonally-patterned knolls to indicate the preferential sublimation of subsurface ice on warmer, equator-facing slopes (resulting in slumping and shallowing of the knoll slopes) relative to enhanced preservation of shallower ice on colder, pole-facing slopes (cementing the knoll slopes and maintaining the relatively steep profiles). In particular, Levy et al. (2008c) conclude that unless the polygonalized knolls contain uniformly distributed excess ice within the mound (and by extension, within the polygon interiors) no topographic asymmetry would develop (i.e., the knolls would be sediment-supported). This, coupled with the high-centered morphology typical of flat-top small polygons, suggests that the martian flat-top small polygons are analogous to terrestrial sublimation polygons (Marchant et al., 2002), and form in a substrate containing distributed excess ice rather than excess ice concentrated along fractures or wedges.

Not all polygons forming in latitude-dependent mantle material are high centered, however; what do polygons with raised shoulders indicate about local differences in permafrost ice content and polygon development? "Mixed-center" polygons (polygons transitioning between high- and low-centered examples over horizontal length scales of 1–5 polygon diameters, Fig. 1) (Levy et al., 2009c) are common at martian mid-latitudes (~45° in both hemispheres) and are particularly common on the walls and floors of "scalloped terrain" in Utopia Planitia, and circum-Hellas (Lucchitta, 1981; Seibert and Kargel, 2001; Lefort et al., 2007, 2009; Soare et al., 2008; Zanetti et al., 2008), and in thick (tens of meters) mantle deposits atop concentric crater fill (Levy et al., 2009b). Some workers have suggested that scalloped, polygonally patterned units represent regional thermokarst-like melting prior to LDM emplacement (Costard and Kargel, 1995; Soare et al., 2007, 2008). Alternatively, analyses combining HiRISE image and topography data (HiRISE stereo and HRSC) (Zanetti et al., 2008; Lefort et al., 2009), have demonstrated that the formation of scallops (and the thermal contraction crack polygons present in and around the scalloped depressions) is more consistent with locally enhanced sublimation of near-surface ice on slopes receiving heightened insolation and subsequent slumping of desiccated sediments. In particular, Lefort et al. (2009) suggests that polygons in Utopia Planitia scalloped depressions (mixed-center polygons) formed contemporaneously with the scallops by sublimation-driven geo-

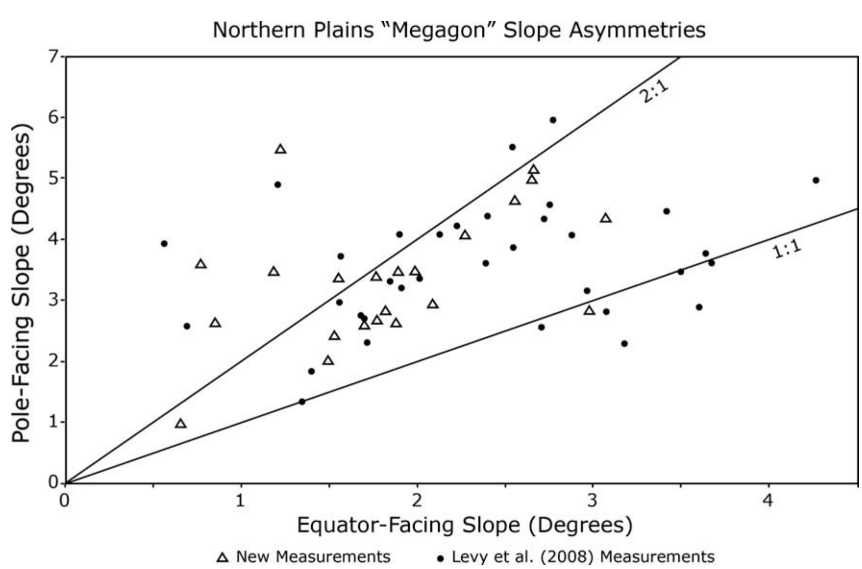

Fig. 9. Slope asymmetries measured for pole-facing and equator-facing surfaces on rubble-pile-topped, polygonally-patterned knolls in the martian northern plains. Martian knolls may be analogous to terrestrial "megagon" clusters of sublimation polygons (Levy et al., 2008c). Closed circles represent measurements reported by Levy et al. (2008c) on HiRISE image PSP_001958_2485, using HiRISE stereo topography produced by the USGS Astrogeology Branch. Open triangles indicate new measurements made on HiRISE image PSP_001919_2470 using HiRISE stereo data produced by the USGS Astrogeology Branch. Pole-facing knoll slopes are approximately twice as steep as equator-facing knolls slopes, interpreted to indicate removal of excess ice on warm, equator-facing slopes.

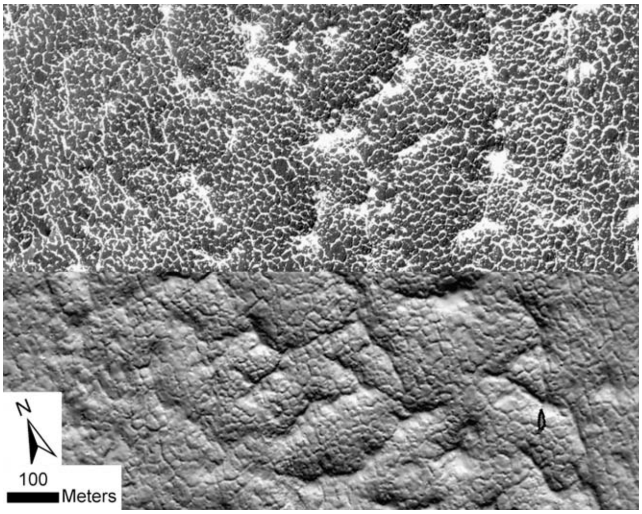

Fig. 10. Sublimation polygons and "megagons" (larger, ~100 m, topographically high mounds covered in sublimation polygons) in Beacon Valley, Antarctica (Marchant et al., 2002). Image top: air photograph of polygons and megagons. Image bottom: LIDAR digital elevation model hillshade (illumination from image top) of adjacent groups of polygons and megagons. Antarctic air photographs and LiDAR topography data kindly made possible in part through a joint effort from the NSF and the USGS.

morphic processes. Likewise, raised shoulders on some mixed-center polygons have been accounted for by a two-stage sublimation process occurring in very ice-rich LDM deposits. In this two-stage process, sublimation is initially focused at thermal contraction cracks (Marchant et al., 2002; Kowalewski and Marchant, 2007; Kowalewski, 2009), but shifts to polygon interiors as dry (desiccated and/or windblown) sediment accumulates at the polygon trough, sealing off underlying ice and building sediment-rich wedges that remain topographically high relative to the collapsed polygon centers (Levy et al., 2009b). Thus, rather than indicating recent, wet active layer or thermokarst modification (surface depression caused by melting of ground ice, Washburn, 1973) of the latitude-dependent mantle, these mixed-center polygon surfaces form in extremely dry conditions (no melting) and also strongly indicate the presence of abundant excess ice in LDM deposits (interpreted to originate from a primary precipitation source, e.g., Head et al., 2003). This ground ice is of sufficient purity to leave depressions tens of meters deep in locations experiencing locally intense sublimation.

Another polygonally patterned feature present on the martian surface that may provide information about ice concentration in the cracking LDM medium is "basketball terrain" rubble piles (Kreslavsky and Head, 2000; Malin and Edgett, 2001; Levy et al., 2008b, 2009c; Mellon et al., 2009). Basketball terrain rubble piles are accumulations of <~1 m boulders, locally concentrated in piles ~3–5 m in diameter, that are commonly perched atop topographically high mounds surfaced by smaller-scale thermal contraction crack polygons (Levy et al., 2008c, 2009c; Mellon et al., 2009). Rubble piles are typically spaced 20–30 m apart, in staggered or linear patterns, and are connected by sparse strings of boulders in some locations (Levy et al., 2008c, 2009c; Mellon et al., 2009). Levy et al. (2008b) proposed that many of the observed boulders may have been excavated by impacts (e.g., Heimdall crater near the Phoenix landing site, Arvidson et al., 2008) from the Hesperian ridged plains surfaces interpreted to underlie the LDM surfaces and the Vastitas Borealis Formation (Fig. 11) (Head et al., 2002; Kreslavsky and Head, 2002); however, the possibility remains that some boulders were also brought to the surface by geographically-limited and ancient (>5–10 Ma old) active layer processes (particularly on steep slopes) described by Kreslavsky et al. (2008). Levy et al. (2008b) and Mellon et al. (2008b) have proposed that the arrangement of boulder piles is consistent with a previously extant network of larger (~30 m diameter) thermal contraction crack polygons. Mellon et al. (2008b) suggest several mechanisms that could concentrate boulders in paleo-polygon troughs, ranging from gravitational or aeolian sorting and slumping (e.g., Rosato et al., 1987), to surface creep, to a complex (water-free) dry cryoturbation process cycling boulders through troughs to polygon interiors (see Section 2.4 for a discussion of closed-cell cryoturbation).

In the Antarctic Dry Valleys, boulders commonly accumulate in sublimation polygon troughs due to gravitational sliding, rolling and slumping due to over-steepening of polygon troughs by enhanced sublimation of buried ice along polygon margins (Marchant et al., 2002; Kowalewski and Marchant, 2007; Kowalewski, 2009). Polygon troughs are typically at their steepest and deepest at junctions between troughs, and accordingly, act as loci of boulder accumulation, concentrating slumped cobbles and boulders from the corners of several polygons (Levy et al., 2006). Boulders accumulated in old polygon troughs and at polygon junctions would overlie and armor finer, winnowed sediments (Marchant et al., 2002), eventually burying ice deeper, and thereby protecting it from enhanced sublimation. If ice removal continued from the centers of these large polygons, topographic inversion similar to that described above (for mixed-center polygons) could occur, resulting in accumulations (at former polygon junctions) of boulders on topographically high piles of sediment overlying protected ice, linked by chains of boulders localized to former polygon troughs. This potential sequence of events is illustrated schematically in Fig. 12.

Mellon et al. (2008b) propose aeolian deflation of sediments that have lost cementing pore ice, concentrated between rubble piles, as a mechanism for raising rubble pile mounds relative to surrounding terrain. In contrast, we point out that any inversion process responsible for removing polygons large enough to shed boulders by gravitational slumping would require the removal of several meters of excess subsurface material from beneath the polygon interiors, without producing noticeable drifts of sedimentary material. Accordingly, we interpret this hypothetical series of events to imply the vertical removal of several meters of excess ice from the ancient polygonally-patterned substrate by sublimation. If this older polygonal surface is comparable to the current polygonally-patterned LDM, significant volumes of excess ice are strongly implied, distributed uniformly throughout the medium (e.g., not in a "cryo-shell" as described by Fisher, 2005). This process strongly favors precipitated ice deposits for constituting LDM surfaces (Head et al., 2003), rather than vapor-diffusion-emplaced ice (Mellon et al., 2008a,b), and is consistent with the infilling of "boulder halo" craters (Fig. 11) with LDM material. Accordingly, the presence of "rubbles piles" on Mars suggests that similar sublimation polygon-like processes (e.g., gravitational sorting of clasts into polygon troughs and fractures) may have occurred in the recent geological past. Polygons resulting in "rubble pile" formation may be strongly analogous to terrestrial sublimation polygons, in which minor changes in sediment sorting within and above the polygon wedge drive enhanced sublimation along polygon troughs; alternatively, high-centered polygons for which minimal sediment transport is observed may represent enhanced sublimation along polygon fractures due to increased ice surface area contact with the martian atmosphere (e.g., Marchant et al., 2002; Kowalewski and Marchant, 2007; Kowalewski et al., 2006).

Is there other morphological evidence for high ice content of the polygonalized portion of the latitude-dependent mantle? At both high (~68°N) and middle (~45°N) latitude locations, terrace- or step-like layering in LDM deposits is visible (Fig. 13) (Arvidson

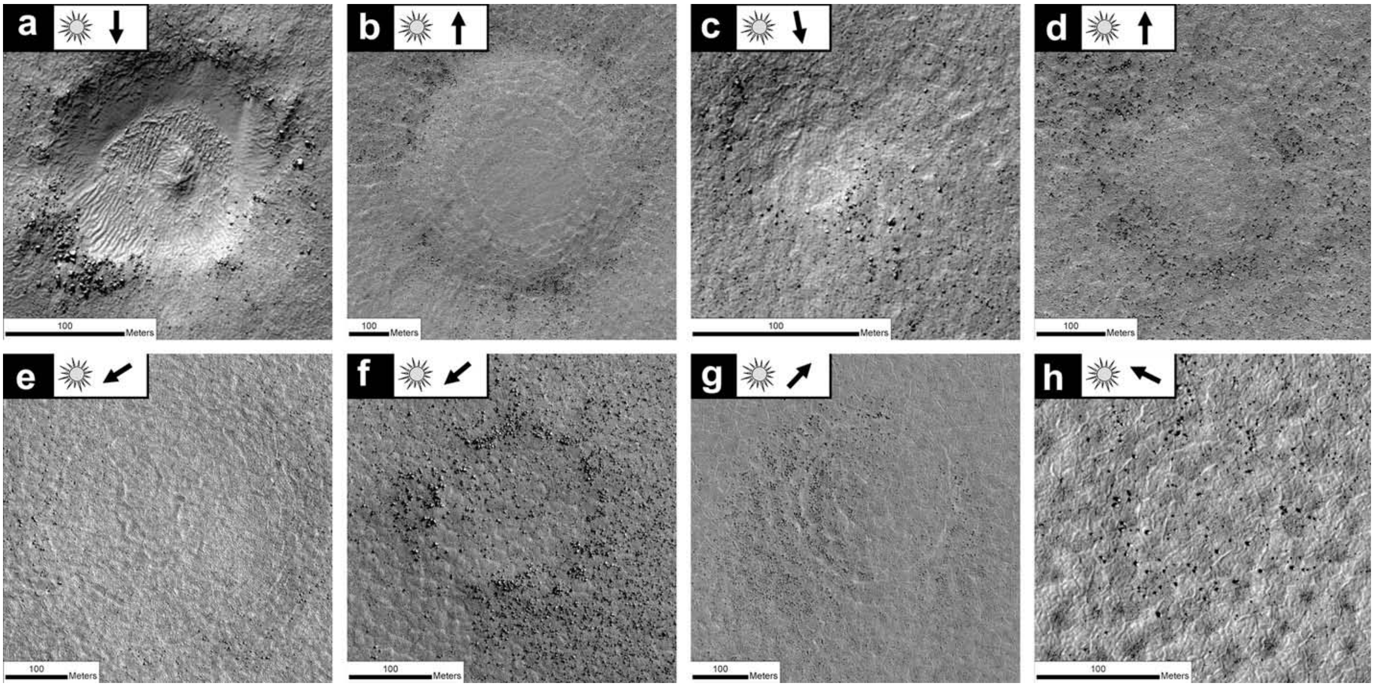

Fig. 11. "Boulder halo" craters (after Levy et al., 2008a) on the northern plains of Mars. Note the presence of meter to multi-meter boulders in proximity to impact craters (and crater-like forms), overlying thermal contraction crack polygons. Some boulders are concentrated in topographic lows between polygonally-patterned high ground, or in troughs between large polygons. Other boulders are present on top of polygon interiors. Dry cryoturbation processing of meter-scale boulders through meter-scale polygons is improbable. Instead, the observations suggest surface slumping, rolling, or sliding of clasts into their present positions. (a) Portion of PSP_001985_2470. (b) Portion of PSP_001477_2470. (c) Portion of PSP_001576_2460. (d) Portion of PSP_001477_2470. (e) Portion of PSP_001803_2485. (f) Portion of PSP_002417_2420. (g) Portion of PSP_001413_2495. (h) 1343_2510.

et al., 2008; Levy et al., 2009b,c; Schon et al., 2009b). At such locations, contacts between surfaces displaying different polygon morphological groups, for example, between "flat-top small" and "irregular" polygons (Fig. 13a) or between mixed-center polygons and underlying (non-polygonally patterned) "brain terrain" surfaces (Fig. 13b) are visible across topographic steps. Changes in elevation between LDM layers are typically on the order of a few meters at high latitude (Arvidson et al., 2008; Levy et al., 2009c) or up to tens of meters in thick LDM accumulations at lower latitudes (Levy et al., 2009b). Similar changes in polygon morphology between surface levels in scalloped terrain are observed in depressions several meters deep (Lefort et al., 2009). Terrace-like layering in polygonally-patterned LDM material typically outcrops as rounded or sinuous exposures of high terrain, hundreds of meters across, stepping down to more broadly distributed low-lying terrain, or as rounded "windows" into the lower mantle levels (Fig. 13a), possibly associated with underlying crater topography. The removal of several meters of LDM material from such windows strongly argues that the polygonally-patterned substrate has a high enough ice content to sublimate extensively on local scales, leaving little evidence of mantle cover behind.

Smaller-scale processes than those producing large windows or terraces through LDM layers provide unique access to the subsurface of the LDM. Small, fresh impact craters have been shown to have formed on Mars within the past decade (Malin et al., 2006; Byrne et al., 2009). Small impacts into mid-latitude (~45–55°N) polygonally-patterned LDM terrain have exposed bright material buried beneath a lower albedo surface cover; CRISM observations confirm that the bright material is water ice (Byrne et al., 2009). Over time, the exposures of buried ice darken, achieving a color and albedo comparable to the surrounding, polygonally-patterned terrain, suggesting that the patterned LDM substrate is exceptionally ice-rich, but contains a sufficient percentage of lithic fines to produce a sublimation lag deposit along the surface in contact with the martian atmosphere (Head et al., 2003; Schorghofer and Aharonson, 2005; Schorghofer, 2007; Byrne et al., 2009; Levy et al., 2009c).

4.2.1. The martian polygonally-patterned permafrost ice reservoir

How much ice is actually present in the latitude-dependent mantle? The morphological lines of evidence presented here strongly argue for an ice-rich latitude-dependent mantle, containing intermixed dust or lithic fines. This composition for the LDM is consistent with an atmospheric origin (condensation of ice and precipitation of snow and frost) of massive beds of ice that surficially exchange water vapor with the present atmosphere (Kreslavsky and Head, 2000; Mustard et al., 2001; Head et al., 2003; Schorghofer and Aharonson, 2005; Schorghofer, 2007). The formation of overlying sublimation lag deposits implies that the lithic component of the LDM ice is greater than 0% (Marchant et al., 2002), however, placing a firm upper bound on the ice content required to form windows or scallops is difficult (and is largely controlled by the efficiency of aeolian removal of lag deposits).

Taking the presence of thermal contraction crack polygons as a proxy for the presence of subsurface ice, a conservative estimate of the magnitude of the martian ice reservoir locked in LDM deposits can be made by taking the product of the surface area dominated by polygonally patterned ground, the depth of the icy layer, and a mixing ratio between ice and the lithic fraction. Surface area was mapped in a sinusoidal equal-area projection, including the spaces between HiRISE images containing polygonally patterned ground (Levy et al., 2009c) as also "permafrost-covered," and avoiding large physiographic features that appear non-patterned, such as the crest of the dichotomy boundary, Hellas basin, etc. This yields a total spatial extent of 4.9×10^7 km^2 (2.5×10^7 km^2 in the northern hemisphere, and 2.4×10^7 km^2 in the southern hemisphere), confined between 30° and 80° in both hemispheres. We take a conservative depth of the icy portion of the LDM at 10 m (Head et al., 2003; Mellon et al., 2009). We set the ice/rock mixing

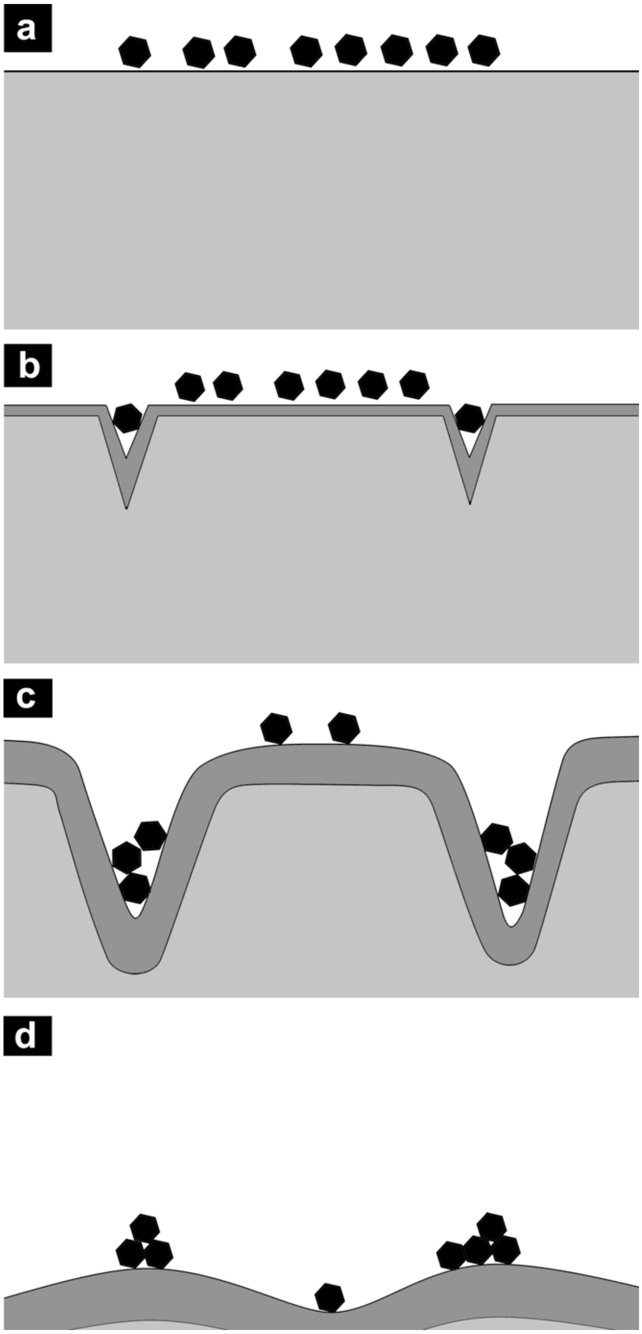

Fig. 12. Schematic illustration of possible boulder positioning sequence on LDM surfaces to create rubble piles. (a) Boulders (dark hexagons) are emplaced on the ice-rich LDM (light gray) by impact excavation of rocky basement. (b) Thermal contraction cracks open in the LDM and are widened by enhanced sublimation along fractures. A sublimation lag deposit (dark gray) begins to develop. Boulders present over growing troughs may be caught in the troughs, although they may be too large to be winnowed into underlying wedge material. (c) Larger, more steeply dipping troughs develop as enhanced sublimation at polygon fractures removes ice from polygon margins, causing more boulders to slump into troughs (particularly concentrated at junctions between troughs). (d) Continuing removal of subsurface ice by sublimation lowers the LDM surface, however, concentrations of wedge sediments and slumped boulders along former troughs protects underlying ice-rich mantle, resulting in preservation of high topography along former troughs (now rubble-pile-surfaced, polygonally-patterned mounds).

ratio at 80% ice to 20% lithic fines by volume. This implies ice volume exceeding pore space (McKay et al., 1998; Fisher, 2005; Sizemore and Mellon, 2006, 2008), and is sufficient to produce sublimation subsidence of polygonally-patterned features. This yields a global permafrost ice reservoir of ~3.9×10^5 km^3, comparable to ice volume estimates of the martian LDM made by Head et al. (2003). Taking a 90% volumetric conversion from ice to liquid water yields a globally averaged water layer of ~2.5 m deep for the martian LDM ice reservoir. This estimate is most strongly sensitive to the spatial extent of polygonally patterned, ice-rich mantle material, which is well-constrained by global thermal contraction crack polygon surveys (e.g., Mangold, 2005; Levy et al., 2009d), rather than to the precise ice/lithic mixing ratio. For comparison to other ice reservoirs, the ice volumes of the north and south polar caps are 2–3 $\times 10^6$ km^3 and 1.2–1.7 $\times 10^6$ km^3, respectively (Smith et al., 1998; Zuber et al., 1998), yielding a globally averaged water depth of ~30 m. Compared to known global ice reservoirs, the volume of ice required to produce polygonally-patterned LDM deposits with a mixing ratio approaching 100% ice (e.g., 98% pure water ice observed at Phoenix, Smith et al., 2008, 2009) is quite small, and is consistent with the mid-latitude deposition of destabilized polar ice during periods of high obliquity, formation of protective sublimation tills as surface soils desiccated and ice returned polewards (Head et al., 2003), and subsequent modification of the remaining ice-rich LDM into its present, polygonally patterned state (Head et al., 2003; Schorghofer, 2007).

4.3. The role of liquid water

Liquid water has clearly played a role in the physical and chemical development of the martian surface, particularly in its early history (Squyres, 1989; Baker et al., 1991; Carr, 1996; Baker, 2001; Bibring et al., 2006). But how recently has liquid water interacted with polygonally-patterned surfaces on Mars, in what volumes, over how extended an area, and with what observable results? The presence of geologically recent gully–polygon systems on Mars (Head et al., 2008; Levy et al., 2008a, 2009d) suggests that very locally derived liquid water may have interacted with polygonally patterned ground ("gullygons") as recently as ~1–2 Ma, the age of the youngest dated gullies on Mars (Riess et al., 2004; Schon et al., 2009a). In these gully–polygon systems, polygonally patterned ground pre-dates the presence of gullies, appears to have modified gully formation, and has subsequently disrupted gully fan deposits through ongoing thermal contraction cracking (Levy et al., 2008a, 2009d). In gully–polygon systems, channel-like features are observed that are (1) continuous and sub-linear; (2) present in widened, curved, and down-slope-oriented polygon troughs; and (3) present in proximity to larger, traditional gully channels. Such "annexed channels" are interpreted to be remnants of polygons troughs through which liquid water flowed (Marchant and Head, 2007; Levy et al., 2009d). The propagation of fractures upward through younger fan deposits emplaced on top of these polygons suggests that some gullygons may have actively undergone seasonal thermal contraction and wedge growth during periods of active gully flow, channel erosion, and fan deposition (Levy et al., 2009d). The reemergence of polygons through some gully fans suggests that periods of gully activity were microclimatically warm enough to permit localized water ice melting in gully alcoves and surface flow through gully channels, but not so warm as to produce widespread saturated active layer activity (Kreslavsky et al., 2008), or complete destruction of previously emplaced polygons. On the basis of these observations, polygons present in martian gully–polygon systems have been interpreted to be analogous to composite-wedge polygons (Fig. 2) present in the relatively clement inland mixed zone of the Antarctic Dry Valleys (Marchant and Head, 2007; Levy et al., 2009c,d). Composite-wedge polygons are characterized by localized, peak-seasonal contact between meltwater and the ice table of the polygonally patterned surface. However, in the martian case, rapid sublimation of frozen meltwa-

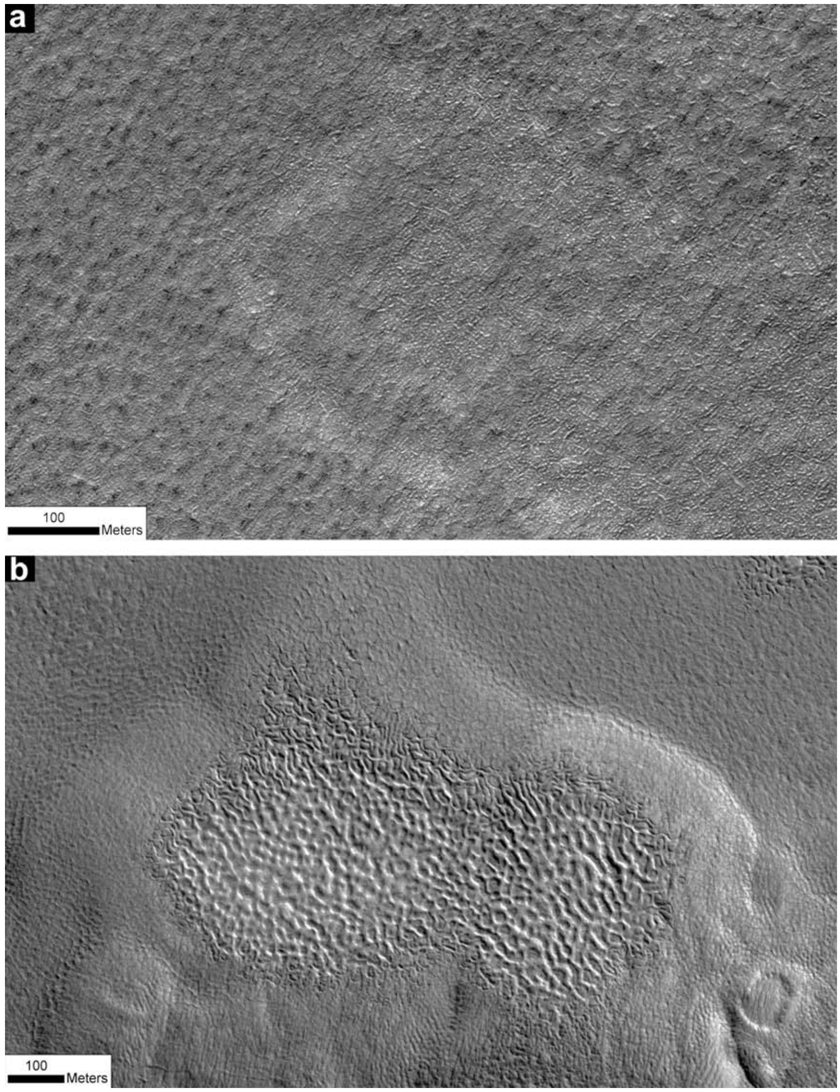

Fig. 13. Contacts between higher and lower lying polygonally-patterned LDM units, showing "windows" into underlying terrain. (a) Flat-top small polygons (perimeter of image) overlie irregular polygons (center and lower right of image). Portion of PSP_001880_2485. Illumination from image top. (b) Mixed-center polygons in a tens of meters thick layer of LDM (perimeter of image) overlie an exposed "brain terrain" concentric crater fill surface (image center). Portion of PSP_002175_2210. Illumination from lower left.

ter, rather than drainage and active layer processing, may be the ultimate fate of gully-related meltwater.

Some models for the formation of martian gullies suggest that the wholesale melting of "pasted-on terrain"—the morphological expression of LDM material present on many steeply sloping martian surfaces (Head et al., 2003, 2008)—is required for the formation of gullies (Costard et al., 2002; Christensen, 2003). Although the melting of pasted-on terrain is strongly implied in select locations showing dramatic evidence of wet debris-flow-like gully formation (Levy et al., 2009a), the presence of intact thermal contraction crack polygons both pre- and post-dating activity in many geologically-recent martian gullies suggests that widespread ground ice melting may not be required to produce all gully activity (Dickson et al., 2007; Head et al., 2008; Levy et al., 2009d). Given the association of some martian gullies with polygonally-patterned terrain (Levy et al., 2009d), and of many martian gullies with dissected latitude-dependent mantle material (Mellon and Phillips, 2001; Heldmann et al., 2005, 2007; Dickson et al., 2007; Dickson and Head, 2008), it is easier to reconcile the possibility that recent gully-related fluid flow is more closely linked to top-down melting of localized surface snow and ice and transport of desiccated surface sediments, than to widespread ground ice melting and thermokarst production derived from degrading LDM surfaces (Head et al., 2008).

Are there other polygonally-patterned landforms on Mars that may require larger or more widely distributed volumes of liquid water to form? A variety of landforms have been identified in polygonally-patterned portions of Utopia Planitia that have been interpreted to be pingos (Dundas et al., 2008; Burr et al., 2009a; de Pablo and Komatsu, 2009). These features are geographically distinct from equatorial landforms interpreted as pingos (Burr et al., 2005, 2009a; Page and Murray, 2006), the analysis of which is complicated by volcanic landforms also observed in Athabasca Valles (Jaeger et al., 2007) and Cerberus Fossae. Pingos form in tundra environments on Earth, and require saturated active layers, or even standing bodies of liquid water (and/or pressurized groundwater) to form (Washburn, 1973; Burr et al., 2009a,b). Pingos are of interest to thermal contraction crack studies in that pingos on Earth are commonly polygonally patterned, and because the putative martian pingos form in terrains surfaced by a range of young thermal contraction crack polygons and ancient glacial landforms. Although dramatic fracturing, and in some cases, organized cracking is observed on fractured mounds in Utopia Planitia (Dundas et al., 2008), many mound fracture patterns are similar to cracking patterns observed on inverted "oyster shell" (Mangold, 2003) or "ring-mold" (Kress and Head, 2008) craters, that are in some places surfaced by polygonally-patterned LDM, and which are not interpreted as liquid–water-related features (Dundas and McEwen, 2009). Additionally, the possibility of late Amazonian, widespread active layer or talik conditions in proximity to Utopia Planitia fractured mounds is inconsistent with the preservation of underlying glacial surface textures ("brain terrain") on Utopia Planitia concentric crater fill (that has been dated to 10–100 Ma) (Levy et al., 2009b). Widespread thermokarst formation would likely have disrupted these fine-scale textures, which are observed to emerge pristine from beneath eroded exposures of, or windows through, polygonally-patterned LDM material. Lastly, the preservation of tens of meters of ice-rich material within preserved pedestal craters concentrated in Utopia Planitia, argues for the protection of ice deposits at the martian surface over tens to hundreds of Ma under cold and arid conditions resulting in the formation of marginal sublimation pits but no melt features (Kadish et al., 2008). Thus, both morphological and stratigraphic arguments suggest that the Utopia Planitia region has remained cold and dry over at least the past ~100 Ma, inconsistent with the formation of active layer landforms. Accordingly, recent observations of fractured mounds across a range of latitudes by Dundas and McEwen (2009) suggest that ice cores produced by segregation ice formation are not detectable in potential pingo landforms and that caution should be exercised in interpreting these landforms as pingos. Rather than invoking a recent and shallow integrated hydrological system in Utopia Planitia, a complex, sublimation-driven erosion of ice-rich layered material may better account for the observed morphologies (Dundas and McEwen, 2009).

To summarize, analysis of a range of polygon morphological properties and interactions with nearby landforms indicates that geologically young polygons present in the LDM have not formed in conditions permitting the widespread presence of standing or flowing liquid water. Instead, the observed relationships between polygonally patterned ground and associated landforms may be plausibly explained by primarily solid–vapor phase transitions under cold and dry conditions. Globally distributed polygonally patterned ground morphology (Fig. 1) reflects zonally averaged, cold and dry climate conditions within at least the past 1–2 Ma, and likely extending back several tens to hundreds of Ma to preserve glacier-like surface textures. Model results predicting ground temperatures sufficient to produce an active layer (>0 °C) during periods of high (>35–40° obliquity) (e.g., Mangold, 2005; Kreslavsky et al., 2008) provide critical insight into the limited magnitude of surface reworking accomplished by wet and dry active layers under martian surface conditions. Gully–polygon interactions may reflect peak surface and atmospheric temperature conditions occurring within short-lived, localized microclimate zones. The morphology of a wide range of small thermal contraction crack polygons on Mars is most consistent with widespread sublimation polygon formation, exposures of sand-wedge polygons, and rare (gully-related) occurrences of composite-wedge polygons. No active ice-wedge polygons are likely to be forming under current martian conditions (Kreslavsky et al., 2008). In Fig. 14, the Levy et al. (2009c) morphological classification of martian polygons is mapped onto an inferred terrestrial classification on the basis of martian polygon morphology.

Has liquid water played any role in the modification of polygonally patterned ground at any scale during the most recent Amazonian? In contrast to geomorphologically dry conditions (insufficient surface water to produce characteristic saturated active layer features) at the meters scale, recently acquired spectroscopic and geochemical data suggests that at the micron scale, thin-film water alteration of martian minerals may have occurred over much of the late Amazonian (Wyatt and McSween, 2007; Boynton et al., 2009; Hecht et al., 2009; Poulet et al., 2009). Microphysical processes producing water-film alteration conditions on the surfaces of rocks in the Antarctic Dry Valleys are well-described (Staiger et al., 2006; Marchant and Head, 2007) and may be analogous to processes operating on Mars (Head and Kreslavsky, 2006). Thin-film altered martian rocks and minerals may represent the maximum extent of widespread liquid water-related processing in martian thermal contraction crack polygon terrains (Wyatt et al., 2004).

4.4. Determining the age of martian polygon networks

Polygonally patterned ground on Mars has been shown to be an exceptionally young surface feature (Head et al., 2003), characterized by ongoing thermal contraction cracking (Mellon, 1997). A baseline for the age of the martian LDM of ~0.4–2.1 Ma was developed by Head et al. (2003) based on correlation of mantle-related deposits with well-dated, obliquity-driven climate models. Given that polygonally-patterned surfaces on Mars may currently be undergoing modification, can crater counting be used to give a reliable estimate of the age of the polygonally patterned surface of the LDM? Previous crater-count age estimates of the LDM have

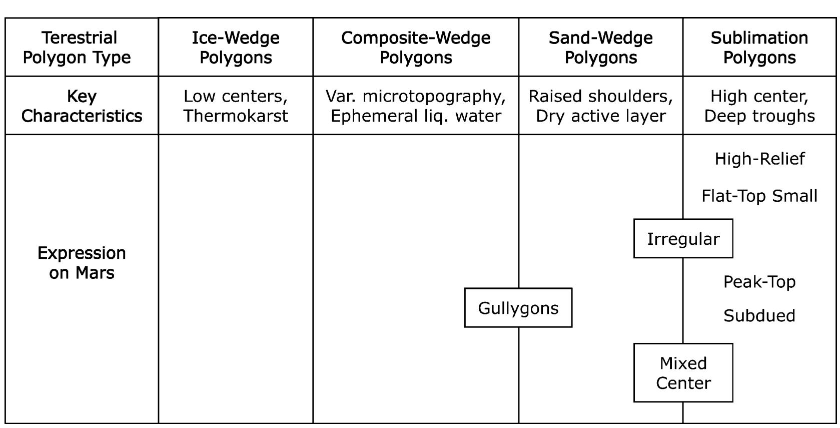

Fig. 14. Top row: polygon types on Earth (e.g., Marchant and Head, 2007) (see Fig. 2). Second row: key morphological characteristics of terrestrial polygon types. Third row: interpreted classification of martian polygon morphological groups defined by Levy et al. (2009c) (Fig. 1) into terrestrial polygon types, based on morphology and microtopography. Martian groups spanning more than one terrestrial polygon type show characteristics of both types. Clear examples of active ice-wedge polygons have not been documented on Mars in high-resolution studies.

attempted to count "fresh" craters that show minimal or no signs of thermal contraction cracking, using MOC image data to derive ages ranging from ~0.1 Ma (Kostama et al., 2006) to ~1 Ma (Mangold, 2005). Recent counts (Kostama et al., 2006; Levy et al., 2009c) have shown a latitude dependence of the age of polygonally-patterned surfaces, ranging from ~1 to 2 Ma at middle latitudes (~45°), to ~100–300 ka at higher latitudes (~55°), and finally to < 100 ka near 65°. Counts on HiRISE images between 60°S and 70°S have shown an exceptionally young surface age: 1–3 ka (Kreslavsky, 2009). The former counts (~2 Ma to <100 ka) have been interpreted to indicate a reduction in the deposition age of the LDM (younging) with increasing latitude (Levy et al., 2009c), while the exceptionally young, high-latitude counts may require a resurfacing process to remove small craters at polar latitudes (Kreslavsky, 2009). Determining the factors responsible for altering the apparent age of polygonally patterned units on Mars is critical for evaluating the age and origin of LDM deposits.

Crater counting on polygonally patterned mantle terrain is complicated by the obscuration and disruption of craters by polygons (Mangold, 2003; Kostama et al., 2006; Levy et al., 2009b). For mantle surfaces displaying small thermal contraction crack polygons, a reduction in the abundance of ~20 m diameter craters on mantle surfaces is typical, and may result from the removal or obscuration of craters with diameters comparable to polygon diameters (Levy et al., 2009c).

Thermal contraction crack polygons can form during permafrost development, forming "syngenetic wedges" that grow upwards towards an aggrading permafrost surface (MacKay, 1990). Alternatively, polygons can develop in stable permafrost that is undergoing little surface change (forming "epigenetic" polygons), or on degrading surfaces or slopes, forming "anti-syngenetic wedges" that propagate downwards into the permafrost at a rate controlled by surface ablation (MacKay, 1990). Thus, counting only fresh, unfractured craters yields a minimum age for the cessation of mantle emplacement (emplacement of additional mantle material would likely resurface existing craters, and promote syngenetic thermal contraction cracking). Accordingly, crater counts on polygonally patterned mantle surfaces indicate that many high-latitude fracture networks began forming as recently as several ka to several hundred ka, while lower (mid-latitude) networks have not undergone significant aggradation in 1–2 Ma, consistent with "ice age" emplacement of mid-latitude LDM deposits (Kreslavsky and Head, 2000; Mustard et al., 2001; Head et al., 2003). Even the youngest populations of polygons (e.g., polewards of 60°) form on timescales comparable to the most rapidly forming terrestrial polygons (Lachenbruch, 1962; Washburn, 1973; Sletten et al., 2003); however unlike quick-forming (years to decades) (Lachenbruch, 1962) terrestrial ice-wedge polygons, such high-latitude martian polygons have been interpreted to be analogous to terrestrial sublimation polygons (Levy et al., 2008c, 2009c) and are more likely to represent rapid trough expansion due to fracture-focused sublimation (Marchant et al., 2002; Kowalewski and Marchant, 2007; Kowalewski, 2009) than present-day, liquid–water processing (Baker, 2001). This rapid emplacement of mantle terrain and growth of polygons is inconsistent with models for excess ice emplacement by vapor diffusion (e.g., Fisher, 2005). Such vapor-diffusion ice emplacement models may require several hundred thousand (to over one million) years to generate the excess ice sufficient to produce the observed surface features (Fisher, 2005).

4.5. Are martian thermal contraction crack polygons equilibrium landforms?

In considering permafrost landforms, Marchant and Head (2007) define equilibrium landforms as, "landforms produced by specific geomorphic processes endemic to, and in balance with, local microclimate conditions." Are martian thermal contraction crack polygons active or relict landforms? Of the active landforms, can martian polygons be considered equilibrium landforms?

Several processes occurring at present on the martian surface are endemic to ice-rich substrates and are occurring in balance with current climate conditions. Notable examples of endemic processes include thermal contraction cracking due to seasonal thermal stresses polewards of ~30° latitude (Mellon, 1997; Mellon et al., 2008b) and seasonal exchange of water vapor from the ice table to the atmosphere (Mellon et al., 2008b). Examples of polygon-related phenomena in balance with current climate conditions include the widespread concurrence of martian ice table depth and subsurface ice stability as a function of latitude, insolation, albedo, thermal inertia, atmospheric temperature, and atmospheric relative humidity (Mellon and Jakosky, 1993, 1995; Mellon et al., 2004; Hudson et al., 2009b); frost accumulation and snowfall in the vicinity of the Phoenix lander (Searls et al., 2009; Whiteway et al., 2009); and frost accumulation in gully environments (Gulick et al., 2007; Head et al., 2008; Levy et al., 2009d). Conversely, observations indicate that some geomorphic processes are not currently occurring on the martian surface due to insufficiently warm and wet climate conditions. Notable among these phenomena are widespread liquid water surface stability and flow (which might produce regional gully activity or cause ice wedge formation) (Kreslavsky et al., 2008).

Thus, to the extent that many kinds of small martian polygons are likely still undergoing seasonal thermal contraction cracking driven by the endemic response of ice-rich permafrost to seasonal cooling, they are active, equilibrium landforms. To the extent that martian ice table depth in the latitude-dependent mantle is set by current climate conditions, the LDM is an equilibrium terrain. To the extent that sublimation polygons and sand-wedge polygons are interpreted to be the dominant polygon type on Mars (to the exclusion of ice-wedge polygons (e.g., Levy et al., 2009c)), martian thermal contraction crack polygons are equilibrium landforms. Likewise, some landforms not in equilibrium with the current martian climate are observed to be undergoing modification towards equilibrium conditions. For example, observations that different sections of polygons at the Phoenix landing site (see next section) have different ice table depths over tens of cm length scales suggests that individual polygons are responding to changes in ice table depth caused by slumping of sediments, rock presence (e.g., Sizemore et al., 2009), or local micro-topography (Levy et al., 2009e). Likewise, exposed ice produced by impacts through overlying lag deposits (e.g., Byrne et al., 2009) undergoes sublimation to produce surficial lag deposits, suggesting that where disruptions occur, many ice-related martian landforms are undergoing modification in order to achieve equilibrium with prevailing climate conditions. Accordingly, many portions of the polygonally patterned latitude-dependent mantle may be well described as equilibrium landforms.

In contrast, some polygon-related landforms do not show evidence of current activity. "Rubble piles" (Fig. 12) and windows through mantle layers (Fig. 13) are interpreted as evidence of the removal of large volumes of LDM material (see previous sections). Accordingly, given the general concurrence of martian ice table depth and ice stability prevailing at present (Jakosky and Carr, 1985; Mellon and Jakosky, 1995; Mellon et al., 2004, 2009) these landforms may be considered relict landforms—features preserved by the cold and arid martian climate, but not currently developing. Likewise, some gully–polygon systems appear to be relict landforms, preserved pristinely on pole-facing slopes, but undergoing slow degradation and removal on warmer, equator-facing slopes (Levy et al., 2009d). Although polygons are common in craters containing recently formed bright gully deposits (e.g., Kolb et al., 2008; Pelletier et al., 2008) these deposits cannot be linked to current polygon-mediated activity.

Thus, polygonally-patterned LDM surfaces on Mars may consist of both equilibrium landforms and relict landforms. Current ice table depth is set primarily by climate and salts (Jakosky and Carr, 1985; Mellon and Jakosky, 1995; Mellon et al., 2004, 2009; Hudson et al., 2009a). Accordingly, the deposition of meters-thick ice deposits cannot be readily accounted for by currently operating geomorphic processes, and implies an older, slightly different, set of "ice age" climate conditions (e.g., Head et al., 2003). This difference between martian depositional and modification conditions over time and space is analogous to microclimate processes governing the growth and development of sublimation polygons forming atop debris-covered glacier ice in Antarctica (Marchant et al., 2002). In this terrestrial case, ice involved in polygon formation is deposited in equilibrium with the glacial accumulation zone microclimate conditions (e.g., a depositional environment) and is transported by glacier flow into warmer conditions at lower elevations, resulting in a gradual deepening of ice table depth (Marchant et al., 2002). At each step, the polygonally patterned frozen substrate is approaching local equilibrium with prevailing microclimate conditions.

Geomorphological response of the landscape to changing microclimate conditions over time and space is a hallmark of equilibrium landform processes. Accordingly, polygonally-patterned surfaces on Mars represent a range of cold desert processes, of which some are active and some are relict. Active, developing landforms forming under prevailing martian climate conditions over the past ~1–2 Ma can be considered potential martian equilibrium landforms. Other, relict, landforms provide insight into past climate conditions that have been preserved by the cold and hyperarid surface conditions acting over the past millions to hundreds of millions of years.

5. Outstanding questions in martian polygon studies: integrating lander data

Despite significant progress over the past decade in decoding the geological, climatological, and astrobiological significance of martian thermal contraction crack polygons, several questions remain points of continuing investigation. Lander-scale chemical and morphological analysis of LDM surfaces (Fig. 15) (e.g., Bonitz et al., 2008; Boynton et al., 2009; Hecht et al., 2009; Smith et al., 2008, 2009) raises questions regarding the local microstructure of LDM deposits.

Following the Viking Lander 2 (Mutch et al., 1976; Marchant and Head, 2007), the Phoenix lander has provided a unique first-look at permafrost processes operating on the surface of the martian polar plains (Smith et al., 2008, 2009). Critical results from the mission include the detection of an icy surface present beneath thin, overlying, dry sediments (Smith et al., 2008, 2009; Lemmon et al., 2008; Mellon et al., 2009); complex soil chemistry comparable to that observed in terrestrial cold and arid terrains (Boynton et al., 2009; Hecht et al., 2009; Kounaves et al., 2009); and weather/climate conditions consistent with the generation of a range of cold-desert landforms (Smith et al., 2008, 2009; Lemmon et al., 2008; Mellon et al., 2009). Questions remain, however, as to the abundance, origin, and subsurface distribution of the observed ice, the history of liquid water at the landing site, the age and stability of the landing site surface, and the relationship between the landing site and the martian latitude-dependent mantle (LDM).

Application of geomorphological techniques and principles developed in terrestrial permafrost settings (e.g., the Antarctic Dry Valleys, Berg and Black, 1966; Sugden et al., 1995; Marchant et al., 1996, 2002; Sletten et al., 2003; Kowalewski et al., 2006; Marchant and Head, 2007; Swanger and Marchant, 2007; Kowalewski, 2009; Swanger, 2008) can be usefully applied to lander-scale exploration of martian polar terrains (e.g., Levy et al., 2009e). For example, stable (non-churning) and vertically ablating surfaces are typical of extremely ice-rich permafrost surfaces in some portions of the Antarctic Dry Valleys (see Section 2.4)—particularly those in which sublimation is the dominant water phase transition. Terrestrial permafrost geomorphology experience provides a guide to geomorphic features observed on the martian surface that may suggest stable, vertically ablating, ice-rich permafrost (Fig. 15a and b) (Brook et al., 1993; Sugden et al., 1995; Marchant et al., 2002; Marchant and Head, 2007). Pitted rocks are present at the Phoenix site (Fig. 15c), and have been interpreted as vesicular boulders (Arvidson and Mellon, 2008). Alternatively, these boulders may have been pitted by a range of surficially-acting physical processes, such as salt weathering, that typically forms mm-to-cm-scale pits on static rocks in Antarctic environments on timescales from ~1 to 4 Ma (Matsuoka, 1995; Staiger et al., 2006; Head and Kreslavsky, 2006; Marchant and Head, 2007). This age for pitted rocks at the Phoenix landing site would be consistent with predictions of geologically recent coeval development of LDM and boulder piles (see Section 4.2). Next, although preliminary cobble and boulder counts at the Phoenix site have been interpreted to indicate thorough reworking of the surface by dry cryoturbation (Heet et al., 2009; Mellon et al., 2008b), the presence of cobbles and boulders overlying fine sediments, even in polygon interiors, strongly suggests the formation of "desert pavement" surfaces, typical of vertically ablating arid surfaces (Fig. 15a and b) (McFadden et al., 1987; Sugden et al., 1995; Marchant and Denton, 1996; Marchant et al., 2002). Pavement surfaces are ablational, consistent with the paucity of aeolian depositional ripples at the site reported by Arvidson and Mellon (2008)—although light-toned, smooth "dust pools" are abundant around the lander, and may be analogous to flat-lying sand accumulations typical of Antarctic polygonally-patterned terrains (Marchant et al., 2002; Levy et al., 2006). Additionally, linear groups of boulders and cobbles present at the Phoenix site appear analogous to the surface expression of relict polygon trough locations found in vertically-ablating Antarctic tills (Marchant et al., 2002). Surface manifestations of relict polygon wedges would not be preserved in a well-mixed, cryoturbated sediment column. Lastly, tightly clustered groups of cobbles and boulders at the landing site appear analogous to disintegrated boulders (or "puzzle rocks") common in Antarctic, arid terrains—clasts that are weathering in situ with minimal disarticulation by cryoturbation (Fig. 15d) (Staiger et al., 2006; Marchant and Head, 2007, Fig. 10). Taken together, these landforms provide strong evidence for a stable, non-churning permafrost surface at the Phoenix landing site.

As with sediment and rock analyses, observations of thermal contraction crack polygons resolved in exquisite detail by landers such as Viking Lander 2 and Phoenix provide insight into permafrost-related processes occurring in the upper centimeters of the martian LDM. Troughs emerging from polygon interiors that orthogonally intersect adjacent polygon troughs indicate that the Phoenix landing site surface has undergone more than one generation of polygon fracture formation, consistent with models predicting ongoing thermal contraction cracking (Fig. 15e) (Mellon, 1997). Ongoing fracturing, preferential sublimation of buried ice at fracture locations, winnowing of dry sediment into surface furrows (Fig. 15f), and periodic slumping of clasts into polygon troughs (typical of terrestrial sublimation polygons, e.g., Marchant et al., 2002) may account for both polygons observed at the site with both cobble/boulder-covered interiors overlying fines, as well as for troughs observed with cobble cover. If sublimation polygons are present at the Phoenix landing site (e.g., Levy et al., 2008c), a surface history dominated by vertical ablation of ice-rich subsurface material, rather than by cyclic cryoturbation, is strongly implied.

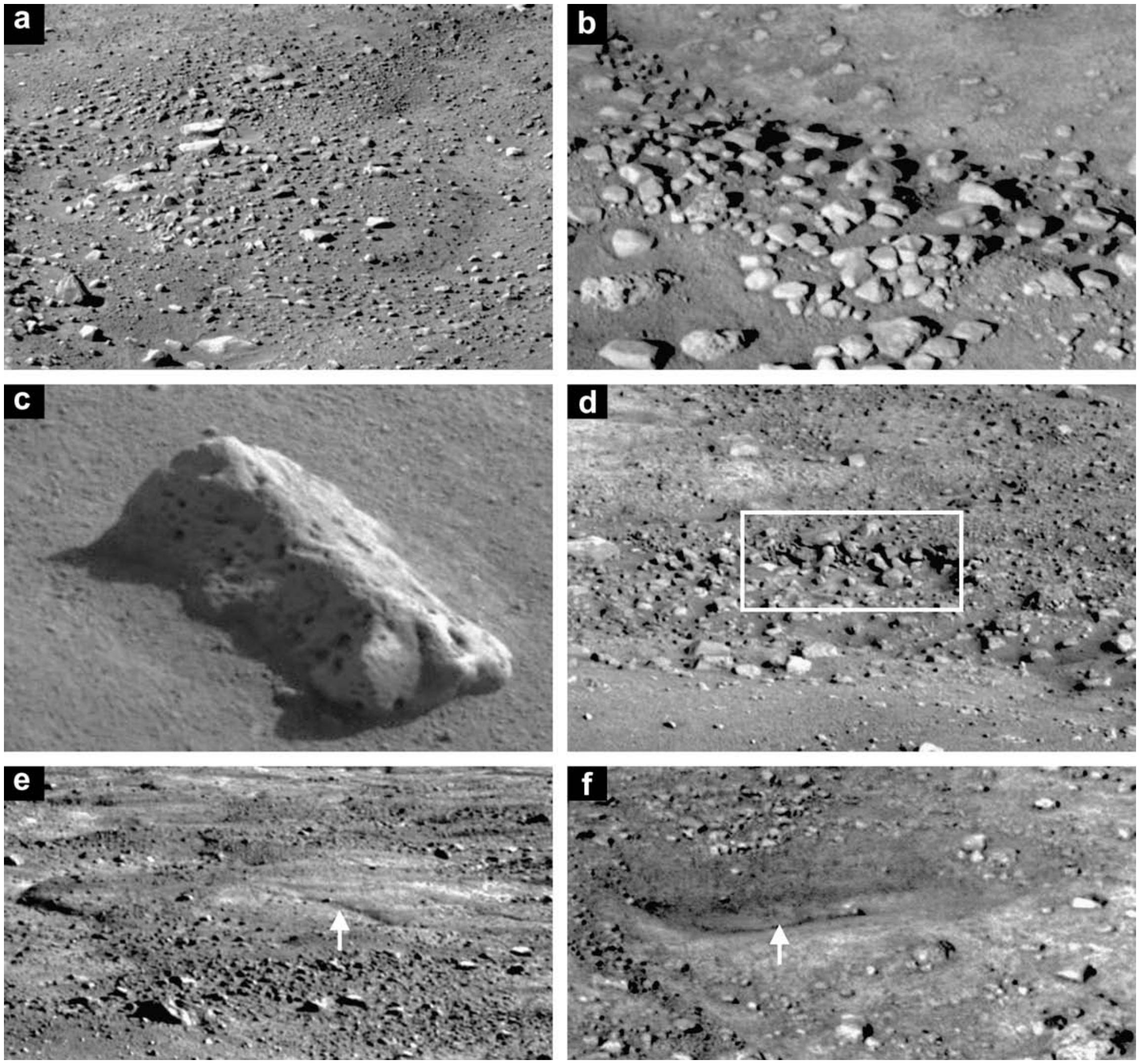

Fig. 15. Surface Stereo Imager views of the Phoenix landing site. All images excerpted from Phoenix image 17106 (see Smith et al., 2008, 2009). (a) Polygon with dense interior clast coverage, surrounded by troughs with reduced clast coverage, suggesting ongoing shedding of interior cobbles/boulders by gentle processes such as slumping and transport of fines by aeolian activity, typical of terrestrial sublimation polygons. (b) Cobbles overlying finer-grained sediment on a polygon interior. Superposition of coarse clasts over finer sediments suggests desert pavement formation, rather than upthrusting from within the polygon or frost-jacking. (c) Pitted boulder in near-foreground of PHX image 17106. Pits may be vesicles, or salt-weathering pits typical of boulders on stable surfaces in Antarctica (Marchant and Head, 2007). (d) A tightly clustered group of cobbles (box) similar in appearance to terrestrial "puzzle rocks" that form by in situ disintegration (e.g., Marchant and Head, 2007). (e) High-centered, round-shouldered polygon bounded by depressed troughs, and cross-cut by a narrower trough (arrow) connecting the polygon center to the bounding troughs. Multiple generations of fractures suggests ongoing thermal contraction cracking. (f) Possible furrow (arrow) in surface of trough between polygons, potentially indicating active winnowing of fines into open thermal contraction cracks.

Ongoing analysis of the Phoenix landing site will help localize global-scale observations of martian thermal contraction crack polygon terrain, extending observational evidence suggesting a history for permafrost at the site and environs characterized by the recent dominance of ice removal by sublimation, ongoing thermal contraction cracking, and very limited cryoturbation by either wet or dry processes. These observations are consistent with global observations of the martian latitude-dependent mantle (Head et al., 2003) suggesting the vertical ablation of excess ice in the martian subsurface (e.g., Schorghofer, 2007)—ice that is originally of primary atmospheric-deposition origin. With regional characterizations of the polygonally-patterned portions of the LDM in hand (Mangold, 2005; Levy et al., 2009c), local analyses, such as those performed at the Phoenix landing site, are the logical next step in documenting the processes operating on the martian LDM.

6. Conclusions

Thermal contraction crack polygons on Earth are a complex landform that has begun to be decoded by the combined investigative efforts of field geologists, physical modelers, climate modelers, and geochemists. Understanding the full history and significance of thermal contraction crack polygons on Mars will require a similarly coordinated effort to disentangle the complexities of local substrate and climate processes. Terrestrial thermal contraction crack polygons are excellent indicators of the current or past presence of ground ice, ranging in composition from massive ice to weakly cemented soils. Accordingly, polygons on Mars provide a unique perspective on climate processes that have dominated the martian surface over ka to Ma timescales. Polygons indicate overwhelmingly cold and dry conditions for at least the past several million years on Mars, characterized by ice-age snow and ice precipitation and seasonal vapor-phase exchange, and punctuated by isolated occurrences of melting in some microenvironments. This climate is consistent with hydration and weathering of surface sediments by ephemeral thin films of water. Accordingly, hyper-arid, terrestrial, polar biological analog environments should be more extensively explored to characterize potential and historical martian habitats. Thermal contraction crack polygons on Mars have been shown to be overwhelmingly a modification process affecting the latitude-dependent mantle (LDM), indicating the spatial dominance of an exceptionally ice-rich substrate in most polygonally-patterned terrains. Pore ice emplaced by vapor diffusion appears incapable of producing excess ice rapidly enough, or in thick enough accumulations to produce the observed polygonal landforms, while primary precipitation of ice, snow, and dust is a plausible mechanism. Lastly, ongoing analysis of local outcrops of the martian LDM will provide key insight into its mode of emplacement and modification, and the history of water and ice in a range of martian microenvironments.

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