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Notes

Gullies, polygons and mantles in Martian permafrost environments: cold desert landforms and sedimentary processes during recent Martian geological history

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Abstract: A range of cold desert landforms are found on the Martian surface that have been interpreted to indicate prevailing frozen and hyper-arid conditions for at least the past several million years. These cold desert conditions are punctuated by brief periods of localized surficial liquid water flow. Sediment transport pathways operate under these conditions of extreme cold and aridity and the processes involved generate permafrost landforms that are recognizable from spacecraft at local, regional and global scales. Thermal-contraction-crack polygons are associated with hemisphere-spanning mantle units that contain excess ice in the immediate subsurface. Sublimation is the dominant phase transition rather than melting under present Martian conditions. Evidence is presented for melting of near-surface snow, frost and/or ground ice in protected gully alcove microclimates during the most recent several million years.

Mars is a permafrost planet. The Martian surface supports a wide range of fluvial, volcanic and aeolian landforms analogous to features found on Earth (Chapman 2007; Carr & Head 2010). During most of the geological history of Mars (Laskar *et al.* 2002, 2004), the entire Martian surface and shallow subsurface have experienced mean annual temperatures well below 273 K (0 °C), commonly dipping below 220 K (Mellon & Jakosky 1993). Accordingly, the entire face of Mars meets the standard definition of a permafrost terrain (Gold & Lachenbruch 1973; Washburn 1973; French 2007). These permafrost conditions likely extend to a depth of several kilometres (Clifford & Parker 2001). Indeed, Mars may be considered a cryotic planet insofar as, at present, mean annual surface temperatures are below the melting temperature of several water-ice compounds and solutions (Yershov 1998).

Is the permafrost terrain of Mars similar to that of Earth? Although permafrost conditions persist over *c.* 20% of the Earth's land surface, much of Earth's permafrost is found in the continental and maritime regions of the North American and Eurasian Arctic (French 2007). In these warmer climate zones, permafrost commonly experiences summertime melting as the 0 °C (273 K) isotherm penetrates the frozen ground surface (Washburn 1973; Williams & Smith 1989; Yershov 1998; French 2007). This seasonally thawed portion of terrestrial permafrost is referred to as the 'active layer' and, when water-saturated ('wet'), it is the horizon in which many of the classic permafrost landforms arise (Williams & Smith 1989; Vliet-Lanoe 1991; Yershov 1998).

In contrast, Mars currently lacks a wet active layer, and has probably not experienced climate conditions permitting the widespread development of a wet active layer over at least the last 5–10 Ma (Kreslavsky *et al.* 2008). Interestingly, though, many of the most dramatic Martian permafrost landforms (Fig. 1) including gullies, thermal-contraction-crack polygons and the latitude-dependent mantle (LDM), all formed more recently than *c.* 5 Ma (Mustard *et al.* 2001; Head *et al.* 2003; Milliken *et al.* 2003; Kuzmin *et al.* 2004; Riess *et al.* 2004; Levy *et al.* 2009*a*; Schon *et al.* 2009*a*) Accordingly, it is essential to consider Martian permafrost from a cold desert climate perspective in which wet active layers are rare or absent (Anderson *et al.* 1972; Gibson 1980; Marchant & Head 2007).

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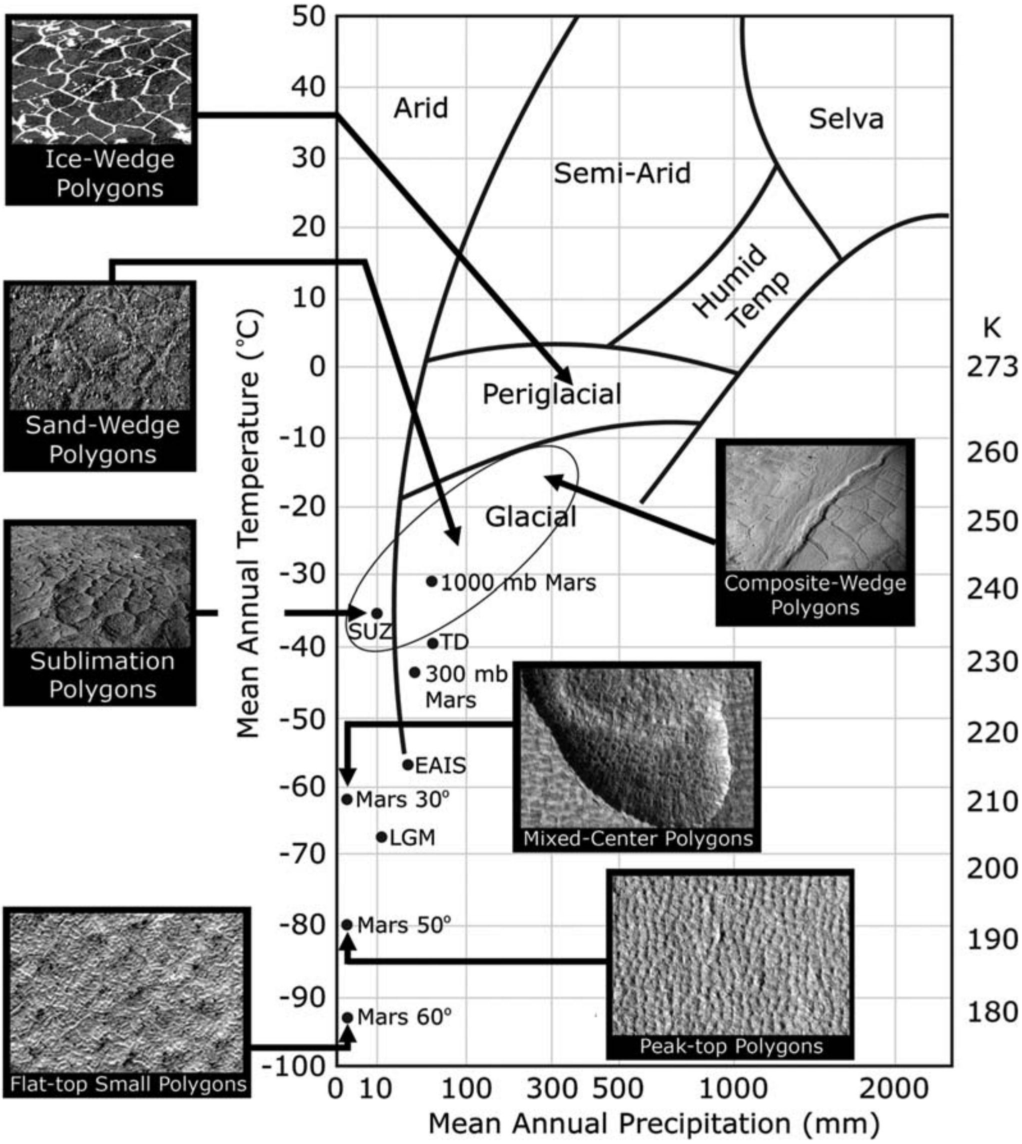

Fig. 1. Plot of permafrost landforms diagnostic of a range of morphogenetic climate regions on Earth and Mars (adapted from Baker 2001 and Marchant & Head 2007). Oval represents mean annual climate conditions typical of the Antarctic Dry Valleys. SUZ indicates the Antarctic Stable Upland Zone (Marchant & Head 2007). TD indicates Taylor Dome and LGM indicates the Last Glacial Maximum in interior Antarctica. Modern conditions at a range of latitudes on Mars and representative thermal-contraction-crack polygon populations typical of those latitudes are plotted, as are conditions modelled for ancient Mars at higher atmospheric pressures. For Martian polygons, field of view is *c*. 300 m in all cases (nomenclature from Levy *et al.* 2009*d*). Flat-top small polygons are excerpted from PSP_001959_2485; peak-top polygons from HiRISE image PSP_001737_2250 and mixed-centre polygons from PSP_002175_2210. The field of view in the illustration of sublimation polygons in Beacon Valley, Antarctica is *c*. 200 m wide. Oblique aerial view of sand-wedge polygons in lower Beacon Valley, Antarctica, has a field of view *c*. 50 m wide. Composite-wedge polygons are illustrated in Wright Valley, Antarctica, cross-cut by a gully channel with a field of view *c*. 75 m wide. Aerial view of ice-wedge polygons in Taylor Valley, Antarctica has a field of view *c*. 75 m wide.

On Earth, cold desert permafrost environments are more typical of the coldest Antarctic and Arctic environments than of warmer and more widely-studied (and inhabited) permafrost zones (Gibson 1980; Marchant & Head 2007; Levy *et al.* 2008). Extreme cold is a critical element for understanding terrestrial analogues for permafrost terrain on Mars; for example, the Phoenix lander (Smith *et al.* 2009) was sent to explore Martian permafrost near 68°N latitude and reported peak summer air temperatures of only *c.* 245 K with atmospheric water vapour pressures of *c.* 1.8 Pa (Whiteway *et al.* 2009). These conditions are comparable to, but still colder and more arid than, the coldest and driest permafrost microclimate in the Antarctic Dry Valleys. There, mean annual temperatures are *c.* 251 K and mean annual water vapour pressure is *c.* 40–50 Pa (Marchant & Head 2007). In extreme cold deserts, low-temperature, sublimation-driven processes dominate the geomorphologic record (Chinn 1981; Marchant *et al.* 2002; Marchant & Head 2007). This major difference, between wet and dry permafrost, guides much of the following discussion. Advanced studies of Martian permafrost incorporating future lander data may integrate other critical climate controls on permafrost morphology, such as annual positive degree-days and snow recurrence intervals (McKay 2008).

Sedimentary processes in Martian permafrost

Observations of the Martian surface suggest that rocky, regolith-surfaced landscapes abound, making Mars an ideal laboratory for considering sedimentary processes in cold desert permafrost environments (Mutch *et al.* 1976, 1977; Golombek & Rapp 1996; Wyatt *et al.* 2004; Golombek *et al.* 2008; Edwards *et al.* 2009). Particle sizes of Martian sediments range from boulders observable from orbit down to micron-scale dust particles observable with lander and rover microscopic imaging systems (Golombek & Rapp 1996; Pike *et al.* 2009). Boulders, cobbles, pebbles and finer sediments are common in Martian permafrost terrains, and the detailed analysis of sediment sorting or arranging is the subject of ongoing inquiry. Sorting by dry cryoturbation processes is suggested by Mellon *et al.* (2008), Heet *et al.* (2009) and Mellon *et al.* (2009*b*). Dry, non-churning permafrost processes are favoured by Levy *et al.* (2008, 2010*a*) (Fig. 2).

This paper discusses three sedimentary landforms typical of Martian permafrost environments, gullies, polygons and mantling units. We explore how sediments are transported in these landforms and interpret mantling units as primarily resulting from atmospheric emplacement of ice and sediment, polygons as resulting primarily from sublimation-driven modification of mantling units and gullies as resulting from the top-down melting of near-surface ice and entrainment of mantle-related sediments.

Global-scale sedimentary processes: the Martian latitude-dependent mantle

What is particularly striking about the distribution of permafrost landforms on Mars is the fact that, despite the global occurrence of permafrost climate conditions (surface temperatures <0 °C over inter-annual periods), ice-related landforms (gullies, thermal-contraction-crack polygons, etc.) have been shown to occur in latitude-dependent clusters (Kreslavsky & Head 2000; Mustard *et al.* 2001; Head *et al.* 2003; Milliken *et al.* 2003; Kostama

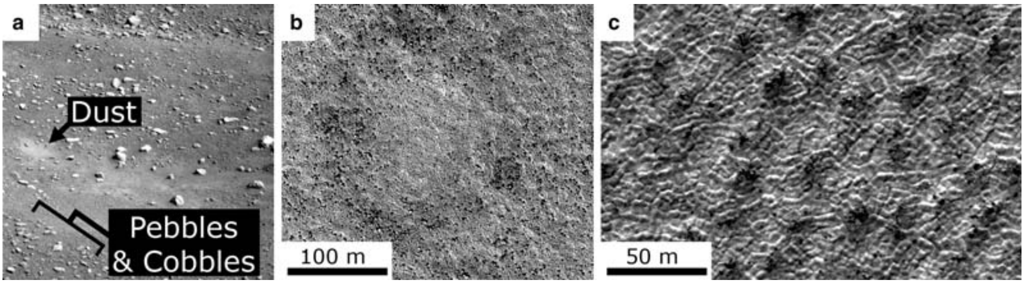

Fig. 2. Sedimentary clasts on the Martian surface. (**a**) Dust, pebbles and cobbles at the Phoenix landing site. Dust patches commonly accumulate in polygon troughs, while polygon interiors are typically armoured by desert pavements of pebbles and cobbles (portion of Phoenix lander Surface Stereo Imager frame SS051EFF900731785_15C28R2M1-b). (**b**) A 'boulder halo' indicating the location of a buried impact crater on the Martian northern plains (Levy *et al.* 2008) (portion of PSP_001477_2470). (**c**) Boulder piles accumulated on polygonally patterned knolls near the Phoenix landing site (portion of PSP_001959_2485).

et al. 2006; Soare *et al.* 2007; Levy *et al.* 2009*a*) (Fig. 3). For example, thermal-contraction-crack polygons (see next section) have been shown to form in geologically recent deposits that are crater-dated to less than several Ma. These deposits drape and smooth underlying terrain, and are present in stacked layers continuously from high latitudes equator-wards to *c.* 60°. The deposits grow patchier and show signs of degradation from *c.* 60° to *c.* 30° (Kreslavsky & Head 1999, 2000, 2002; Mustard

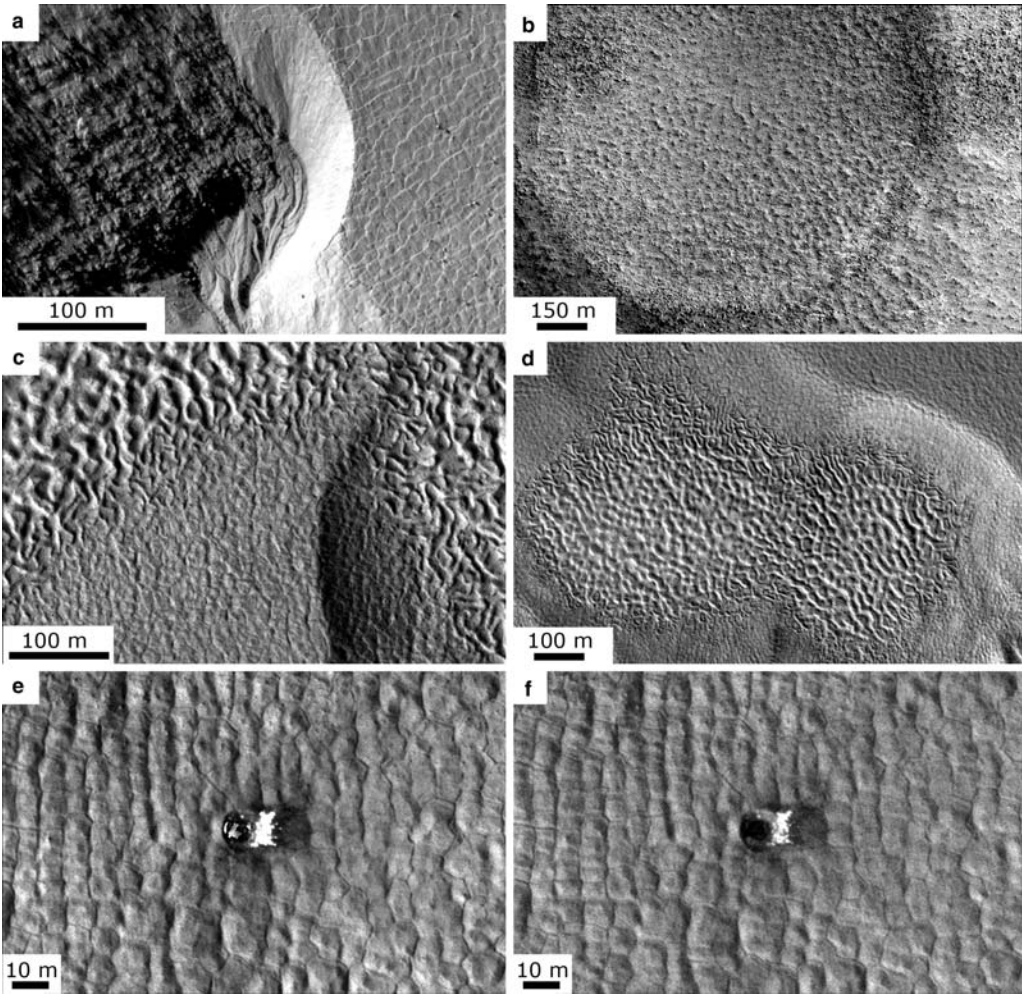

Fig. 3. Key properties of the Martian latitude-dependent mantle (LDM) indicating that it is composed largely of massive, atmospherically-emplaced excess ice. (**a**) Medium-toned, polygonally patterned LDM material (right) is easily eroded by gully activity, and can be distinguished from darker-toned bedrock units (left) (portion of PSP_006794_1420). (**b**) LDM material fills craters, smoothing topographic variation (portion of PSP_006931_2530). (**c**) LDM material drapes underlying topography and landforms, and can accumulate to tens of metres thickness as indicated by shadow measurements. See detailed discussion of draping morphologies in Levy *et al.* (2009*b*) (portion of PSP_002175_2210). (**d**) 'Windows' can be eroded through LDM deposits, revealing pristine underlying landforms preserved beneath tens of metres of LDM deposits (portion of PSP_002175_2210). (**e**) Fresh impact crater, formed between 2004 and 2008 exposing and ejecting bright excess water ice at 46°N latitude, beneath a lithic lag deposit (Byrne *et al.* 2009) (portion of PSP_010861_2265), with north to image top and illumination from the left. (**f**) Re-imaging of the crater shown in part (e), 127 sols (Mars days) after the image in part (e) was collected. Note darkening of crater bottom and of ejecta. Sublimation rate modelling by Byrne *et al.* (2009) indicates a sediment/ice ratio of *c.* 1% sediment to *c.* 99% water ice.

et al. 2001; Head *et al.* 2003; Milliken *et al.* 2003; Schon *et al.* 2009*b*) (Fig. 4). LDM deposits vanish equator-wards of *c*. 30° (Milliken *et al.* 2003).

This unusual spatial distribution of ice-related features is generally consistent with the predicted stability depth for ice in the upper *c*. 1 m of the

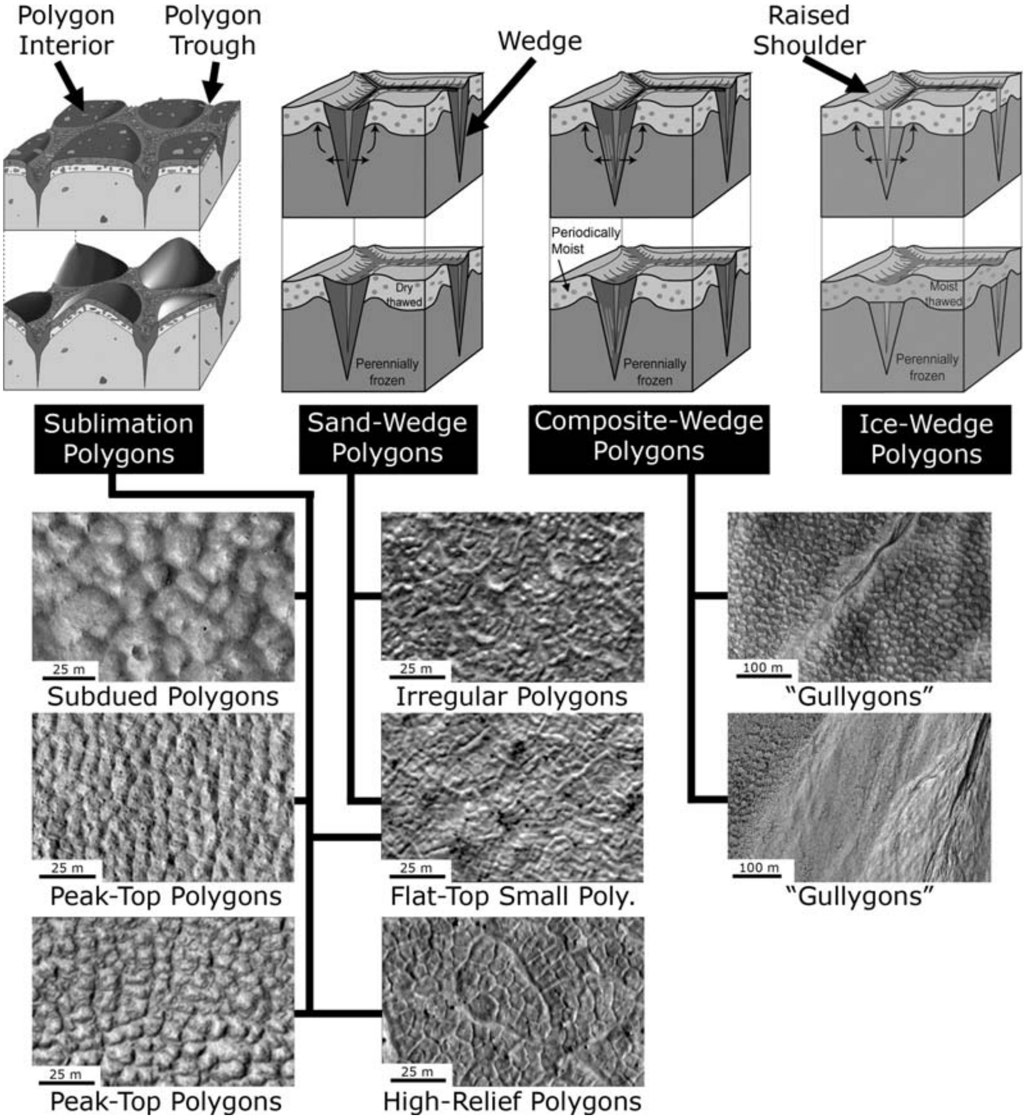

Fig. 4. Schematic illustration of thermal contraction-crack polygon types observed on Earth and comparisons with Martian landforms. Block diagrams are adapted from Marchant & Head (2007) and show key morphological properties of ice-wedge, sand-wedge, composite-wedge and sublimation polygons. Levy *et al.* (2010*a*) use a range of morphological characteristics to connect Martian thermal-contraction-crack polygons with genetic end-member types observed on Earth. Sublimation polygon variants are the dominant polygon type observed on Mars, grading into traditional sand-wedge polygons where excess ice is less abundant. Composite-wedge polygons may form in regions with occasional inputs of liquid water associated with gully activity. There is no definitive evidence of the presence of active ice-wedge polygons on the Martian surface. High-relief polygons are from a portion of PSP_001474_2520; flat-top small polygons are excerpted from PSP_001959_2485; irregular polygons are a portion of PSP_001959_2485; peak-top polygons are excerpted from PSP_01737_2250 and PSP_003217_1355; subdued polygons are a portion of PSP_003818_1360; and 'gullygons' are from PSP_002368_1275 and PSP_001846_2390.

Martian surface based on neutron-, gamma-ray- and thermal emission-spectrometer results coupled with insolation-driven thermal modelling. For thermal modelling, see Jakosky & Carr (1985), Mellon & Jakosky (1995), Mellon *et al.* (2004, 2009*a*), Bandfield (2007) and Vincendon *et al.* (2010). For geophysical results, see Boynton *et al.* (2002), Feldman *et al.* (2002), Mitrofanov *et al.* (2002) and Kuzmin *et al.* (2004).

The stability of subsurface ice alone cannot explain the presence of metres-thick, topographically high, crater-filling deposits that are surfaced by thermal-contraction-crack polygons and gullies. In light of global morphological observations, a number of authors have proposed the presence of a Martian latitude-dependent mantle (LDM), a metres-thick ice-and-dust layer which was deposited as atmospherically precipitated ice and lithic material during recent (*c*. 2–4 Ma) periods of high orbital obliquity (Mustard *et al.* 2001; Kreslavsky & Head 2002; Head *et al.* 2003; Laskar *et al.* 2004). The LDM model predicts the presence of massive, excess ice and dust beds that have undergone sublimation, allowing excess ice nearest the surface to be removed down to the depth of subsurface ice stability while simultaneously producing a thick, rocky, protective lag deposit at the surface (Head *et al.* 2003; Schorghofer & Aharonson 2005; Schorghofer 2007).

The presence of nearly pure-water ice beneath a lithic lag deposit at Martian middle to high latitudes and the mapped range of the LDM was confirmed by (Byrne *et al.* 2009) through the detection of fresh impact craters that exposed a bright substrate spectroscopically identified as water ice (Fig. 3e). The bright, spectroscopically-diagnostic ejecta and crater-bottom material faded to background brightness and spectroscopic parameters over a series of observations. Darkening time was shown to be consistent with an ice/rock mixing ratio of 99% ice to 1% sediment across the northern hemisphere study sites (Byrne *et al.* 2009). For comparison, terrestrial sublimation polygons in Beacon Valley, Antarctica, form in a buried glacier-ice substrate that is *c*. 97% water ice and *c*. 3% sand and rock (Marchant *et al.* 2002).

Typically, interest in the LDM focuses on the presence of the massive, excess subsurface ice deposits – an interest based on the astrobiological importance of ice as a potential source of water and as a microbial habitat on Mars (Lederberg & Sagan 1962; Gilichinsky *et al.* 1992, 2007; Dickinson & Rosen 2003). Estimates for the ice content of the LDM suggest that it represents a reservoir of *c*. 3.9×10^5 km^3 of ice (Head *et al.* 2003; Levy *et al.* 2010*a*). This is approximately one-tenth the volume of the current, residual polar caps (Smith *et al.* 1998; Zuber *et al.* 1998). Estimates of the volume of ice in the LDM are most strongly affected by the spatial extent of the deposit which is well constrained by surveys of image data, and are secondarily affected by estimates of LDM thickness and by the mixing ratio of ice to dusty debris (Levy *et al.* 2010*a*). Turning these ice-reservoir calculations around, an estimate can be made of the volume of the lithic (primarily dust) component of the LDM. Using values reported by Head *et al.* (2003) and Levy *et al.* (2010*a*), if LDM deposits span *c*. 5×10^7 km^2 of the Martian surface, are 10 m thick and have a ratio of ice to lithic fines of 4 (80% ice to 20% dust, assuming that some regions consist of pore-ice permafrost rather than the nearly pure ice observed by Byrne *et al.* 2009), then the LDM represents a global deposit of *c*. 1×10^5 km^3 of dust. This is a global layer over half a metre thick and suggests that the LDM may represent a major sedimentary deposit on Mars. The thickness of this unit is particularly interesting given the extremely slow erosion rates observed at the Mars Pathfinder landing site: as little as 0.01–0.04×10^{-9} m per year (Golombek 1999).

In summary, the Martian latitude-dependent mantle (LDM) may represent a truly unique cold-climate, sedimentary landform in planetary permafrost science. This young, massive ice deposit is globally distributed at high latitudes and is the substrate in which a wide range of thermal-contraction-crack polygons form (Mustard *et al.* 2001; Head *et al.* 2003; Milliken *et al.* 2003; Mangold 2005; Levy *et al.* 2010*a*). The LDM is the permafrost layer underlying the eroded surface of Martian gullies (Levy *et al.* 2009*b*), and may be the source of gully sediments and even of some gully meltwater (Dickson & Head 2009; Levy *et al.* 2010*b*). The latitude-dependent mantle is the unifying substrate in which recent Martian permafrost landforms develop.

Regional-scale sedimentary processes: thermal-contraction-crack polygons

Striking networks of tessellated, patterned ground are abundant at Martian middle and high latitudes (polewards of *c*. 30°) (Mellon 1997; Malin & Edgett 2001; Seibert & Kargel 2001; Mangold 2005; Kostama *et al.* 2006; Mellon *et al.* 2008; Levy *et al.* 2009*a*) (Fig. 3). The relative importance, or even presence, of periglacial (freeze-thaw) sorting of sediments in Martian permafrost terrains is a subject of vigorous and ongoing debate. Some form of wet active-layer sorting is suggested by Balme *et al.* (2009), while a dry cryoturbation mechanism is preferred by Mellon *et al.* (2008, 2009*b*) and Heet *et al.* (2009). Dry, stable and minimally sorting processes are preferred by Levy *et al.*

(2008, 2009*c*, 2010*a*). Here, we focus on small-scale (<*c*. 25 m diameter) thermal-contraction-crack polygons, a class of unsorted permafrost features diagnostic of ice-rich permafrost in terrestrial polar environments. By virtue of the processes involved in their formation, thermal-contraction-crack polygons represent a unique depositional environment for sediments in Martian permafrost terrains.

Thermal-contraction-crack polygons form through climate- and substrate-dependent mechanisms. As a result, they can be used as markers of microclimate history and permafrost thermal conditions (Black 1976; Marchant & Denton 1996; Marchant & Head 2007; Levy *et al.* 2010*a*). Thermal-contraction-cracks form in ice-rich permafrost as it undergoes thermal contraction in response to cooling temperatures. When thermal tensile stresses at or near the ground surface exceed the tensile strength of the frozen ground fractures form orthogonal to the ground surface (the cooling plane) (Lachenbruch 1961, 1962; Mellon 1997; Plug & Werner 2001, 2002; Maloof *et al.* 2002). As fractures propagate parallel to the frozen ground surface they intersect to form the eponymous 'thermal-contraction-crack polygons' (Lachenbruch 1961, 1962; Plug & Werner 2001), forming closed polygonal shapes in map view. The size and shape of thermal-contraction-crack polygons is determined by complex interactions between ice content, cooling history and other mechanical properties of the soil, and is the subject of ongoing investigations (Lachenbruch 1961, 1962; Plug & Werner 2002; Mellon *et al.* 2008, 2009*a*). What makes thermal-contraction-crack polygons interesting as sedimentary features is the next step in polygon formation.

Once fractures open in a frozen ground surface, infilling of fractures may occur as overlying material enters the fracture. Infilling processes are diagnostic of the climate conditions in which the fracture formed (Marchant & Head 2007). Repeated fracturing along the same plane of weakness in the frozen ground, coupled with repeated infilling, can lead to the formation of wedges of material underlying polygon troughs. Different permafrost climate conditions leave unique wedge structures in the stratigraphic record (Pewe 1963, 1974; Murton 1996; Murton & Bateman 2007). In warmer and wetter permafrost environments, in which a seasonally saturated active layer forms, meltwater can percolate through overlying peat, vegetation or regolith, filling thermal-contraction cracks with relatively pure liquid water that subsequently freezes, forming ice-wedge polygons (Leffingwell 1915; Lachenbruch 1962; Berg & Black 1966; Black 1982; Washburn 1973; Sletten *et al.* 2003; French 2007; Marchant & Head 2007). In cold and arid environments, in which either an active layer does not form or in which the active layer is water-free ('dry') (Bockheim *et al.* 2007), sand particles and other fines can winnow into open fractures from above, forming sand-wedge polygons (Pewe 1959; Berg & Black 1966; Murton *et al.* 2000; Sletten *et al.* 2003; Marchant & Head 2007; Murton & Bateman 2007). Some polygon-forming environments are too cold or too arid to regularly experience typical, widespread and saturated active-layer conditions, but do experience occasional, localized inputs of liquid water to the subsurface. This can occur, for example, due to ephemeral snowbank accumulation and melting. Alternating inputs of water and dry sediment to thermal-contraction cracks form composite-wedge polygons (Berg & Black 1966; Murton 1996; Ghysels & Heyse 2006). Finally, in select permafrost environments that are too cold to generate a seasonal active layer and that have abundant excess ice (ice exceeding available pore space) in the subsurface, sublimation polygons may form as ice sublimates preferentially along thermal-contraction cracks and is partially replaced by sieved fines winnowed from overlying tills (Marchant *et al.* 2002; Kowalewski *et al.* 2006; Levy *et al.* 2006; Kowalewski & Marchant 2007; Marchant & Head 2007). Sublimation polygons forming in buried or stranded glacier ice are most common in the Antarctic Dry Valleys (Marchant *et al.* 2002).

The response of permafrost to wedge growth is diagnostic of polygon type (Marchant *et al.* 2007). Active ice-wedge and sand-wedge polygons commonly form broad, raised shoulders and low-lying centres as the ice-cemented soil adjacent to the wedges re-expands as the permafrost warms in summer. The increased subsurface volume (the ice- or sand-wedge) is accommodated by the wedge-adjacent permafrost deforming upwards towards the free surface at the ground–atmosphere interface (MacKay 2000). Ice-wedge polygons may become high-centred in response to thermokarst (melting) modification of the ice-wedge and rapid drainage of surrounding soils (MacKay 2000). Composite-wedge polygons may have slightly raised shoulders or may be flat-lying (Berg & Black 1966; Murton 1996; Ghysels & Heyse 2006). Active sublimation polygons are characteristically convex-up with high, domical centres that are underlain by relatively stable ice, surrounded by depressed troughs that lack raised shoulders (Marchant *et al.* 2002; Kowalewski *et al.* 2006; Kowalewski 2008). In the case of sublimation polygons, the addition of winnowed sediment to the subsurface is balanced (and often exceeded) by the preferential sublimation of buried ice, resulting in low-troughed and high-centred polygons (Marchant *et al.* 2002; Kowalewski *et al.* 2006).

On Mars, thermal-contraction-crack polygons can be identified based on a range of morphological characteristics using both orbital and lander image data. Lander data is analysed by Heet *et al.* (2009), Mellon *et al.* (2009*a*), Levy *et al.* (2009*a*) and Smith *et al.* (2009). Orbital data analysis can be found in Mellon (1997), Malin & Edgett (2001), Seibert & Kargel (2001), Mangold (2005), Kostama *et al.* (2006), Levy *et al.* (2008, 2009*a*) and Mellon *et al.* (2008). Levy *et al.* (2009*a*) use multiple characteristics to identify thermal-contraction-crack polygons on Mars and to distinguish them from other polygonal landforms. These criteria include: (a) network morphology (indicating multiple episodes of fracturing), (b) polygon microtopography (showing raised rims or high, domical centres), (c) diameter (*c.* 25 m or smaller and comparable to terrestrial examples), (d) presence in latitude bands where active thermal-contraction cracking is modelled to presently occur (Mellon 1997), (e) presence on preferentially oriented slopes (which affects the depth and stability of ground ice), (f) surface age (most polygon networks on Mars are very young, <*c.* 2 Ma), (g) particle size and distribution (indicative of sublimation-driven rolling or slumping), (h) bedrock presence (permafrost-related polygons form in unconsolidated sedimentary units and not in bedrock), (i) associated landforms (suggesting permafrost processes, for example, scalloped terrain and mantling units) and (j) albedo (polygons tend to form in relatively low-albedo units). On the basis of multiple surveys of high-latitude datasets, thermal-contraction-crack polygons have been shown to be ubiquitous polewards of *c.* 50° latitude and to be very common polewards of *c.* 30° (Milliken *et al.* 2003; Mangold 2005; Levy *et al.* 2009*a*).

What does the morphology of Martian thermal-contraction-crack polygons suggest about polygon type? Overwhelmingly, Martian polygons are high-centred with depressed boundary troughs and flat or domical interiors (Levy *et al.* 2009*a*). They (a) show evidence of possible orientation-dependent slope asymmetry indicating massive (structureless) excess ice (exceeding pore space) (Mangold 2005; Levy *et al.* 2009*a*), (b) are observed along with a range of landforms suggesting stable (unchurned) and sublimation-driven surface processes, and (c) do not commonly show morphological indications of melting typical of thermokarst-modified ice-wedge polygons (Levy *et al.* 2010*a*). When considered together, this evidence suggests the dominance of active sublimation polygons or sediment-starved sand-wedge polygons on Mars (Fig. 5) (Mangold 2005; Mellon *et al.* 2009*a*; Levy *et al.* 2010*a*).

Polygons and fractures with raised shoulders are present in some locations on Mars (Lefort *et al.* 2009; Levy *et al.* 2009*d*). However, the lack of definitive, accessory landforms suggesting

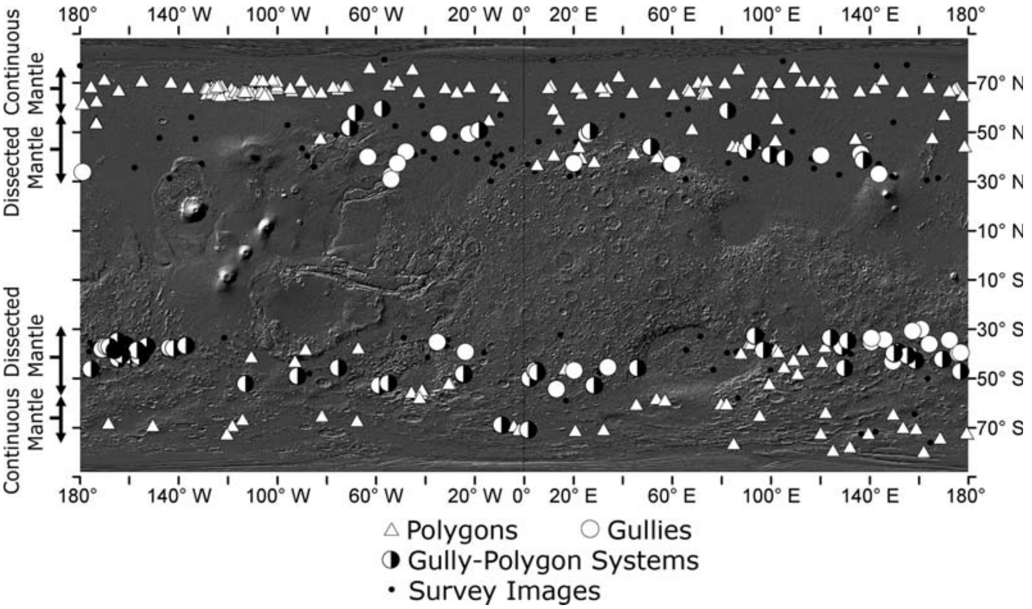

Fig. 5. Map showing the distribution of gullies, thermal contraction-crack polygons and gully-polygon systems across the Martian surface. Despite the global persistence of permafrost conditions on Mars, permafrost landforms such as gullies and polygons are latitudinally clustered. Both gullies and polygons are largely confined to the latitude-dependent mantle. Gullies are more common in the dissected mantle and polygons are common in both the dissected and the continuous portions of the LDM.

active-layer conditions (Dundas & McEwen 2009; Lefort *et al.* 2009) and a climate history incompatible with recent active layer formation (Kreslavsky *et al.* 2008) suggest that these polygons are not ice-wedge polygons. Rather, evidence suggests that they are either sand-wedge polygons or, more likely, sublimation polygons. In the sublimation polygon case, they may have undergone topographic inversion as the once-stable ice in the high centre of the polygon collapsed due to ongoing sublimation (Levy *et al.* 2009*d*). Liquid water volumes sufficient to produce ice-wedges seem not to be a major agent of geomorphologic work in recent thermal-contraction-crack polygon terrains. Where both gullies and polygons are present, the abundance of high-centred polygons suggests that, as modelled by Heldmann *et al.* (2005), liquid water involved in gully formation freezes and/or evaporates and does not initiate water-driven cryoturbation. This may produce composite-wedge polygons (Levy *et al.* 2009*b*). Rather than freeze-thaw phase transitions, sublimation appears to be the dominant player in determining the depth, stability and morphology of ice-related landforms on the Martian surface (Schorghofer 2007; Mellon *et al.* 2009*a*; Levy *et al.* 2010*a*; Sizemore *et al.* 2010).

Local-scale sedimentary processes: gullies

Gullies are not only one of the most interesting features of the Martian surface, but also one of the most enigmatic (Fig. 6). Gullies are a class of young Martian landform that is typically composed of a recessed alcove, one or more sinuous channels and a fan or apron downslope of the channel mouth (Malin & Edgett 2000, 2001). Gullies are typically *c.* 1–2 km long from alcove apex to fan (Malin & Edgett 2000, 2001). Erosion, transport and deposition of particulate material are hallmarks of sedimentary processes, but how did the Martian gullies form given the extreme cold and aridity of the Martian surface?

Several mechanisms have been proposed for the formation of gullies, ranging from water-free sediment flows (Treiman 2003; Shinbrot *et al.* 2004; Pelletier *et al.* 2008) to water-lubricated debris flows (Malin & Edgett 2000; Costard *et al.* 2002; Hartmann *et al.* 2003; Pelletier *et al.* 2008; Levy *et al.* 2010*b*), to water-rich sediment transport (fluvial or hyperconcentrated flow) and alluvial deposition (Heldmann & Mellon 2004; Heldmann *et al.* 2005; Dickson *et al.* 2007; Head *et al.* 2008; Dickson & Head 2009; Levy *et al.* 2009*b*). Given that mass movement of sediment appears to have occurred in gullied regions, or to be occurring (Malin *et al.* 2006), gullies represent an important sedimentary process in the Martian permafrost system.

Since their discovery, evidence has continued to accumulate that implicates a water-related origin for Martian gullies. Malin & Edgett (2001) originally suggested that liquid water played a role in the formation of Martian gullies. Malin & Edgett (2001) argued for gully formation through a combination of overland flow, sapping and wet debris flow on the basis of channel morphology. They noted that gully channels are commonly sinuous, are branched or show anastomosing relationships, are commonly

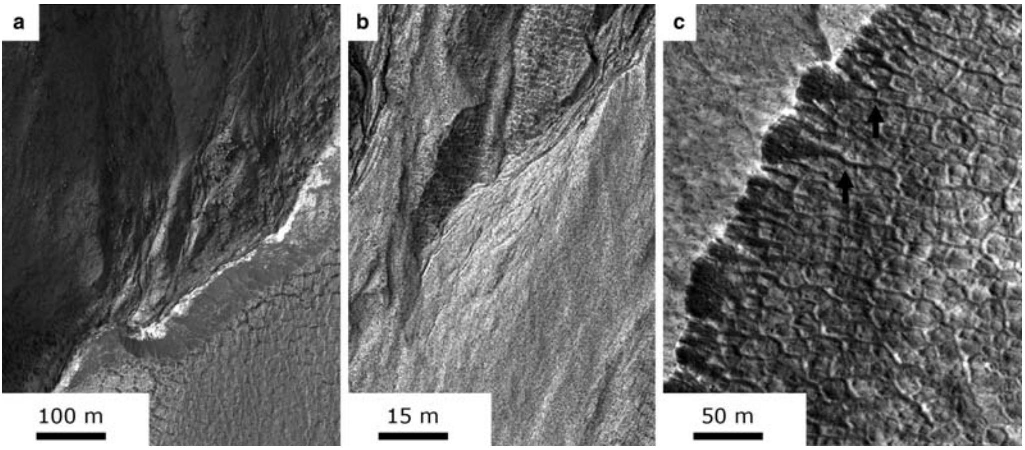

Fig. 6. Relationships between gullies and thermal-contraction-crack polygons in Martian 'gully-polygon systems.' (**a**) Gully channels and alcoves incise polygonally patterned mantle material (portion of PSP_002054_1325). (**b**) Gully fan deposits overprinting thermal contraction-crack polygons. Note fine-scale polygonal patterning of the fan surface (portion of PSP_002368_1275). (**c**) Gully channels 'annexing' thermal contraction crack polygons (portion of PSP_001938_2265).

flanked by levees and commonly display super-elevated banking and incision. Likewise, Malin & Edgett (2000) report that gully fan morphologies are strikingly similar to alluvial fan morphologies on Earth, showing evidence of diverging lineations radiating from the channel mouth, lobate margins and distal thinning. More recent observations of gully morphology using image data of higher spatial resolution reveal additional features of some gullies that are more compatible with sediment transport by liquid water than by dry, granular flow. For example, Mangold (2010) used photogrammetric techniques to determine that asymmetries in levee height at bends in Martian gully channels are most consistent with a fluid viscosity typical of terrestrial water-lubricated debris flows, and would not be likely to emerge due to dry mass wasting. Likewise, high-resolution analyses of other gully deposits have revealed the presence of cutbanks, terraces, cut-off channels, incised fans and channel-fill deposits (Schon & Head 2009; Schon *et al.* 2009*a*) – all features largely consistent with alluvial-fan processes involving multiple episodes of fluvial activity (Fig. 6).

If Martian gullies formed as a result of sediment transport by liquid water, what is the source of that water? Surface temperatures on Mars have been shown to only rise above the triple-point temperature of water (273 K) for a few tens of days of every Martian year and only at latitudes <*c.* 30° (Haberle *et al.* 2001). In an apparently paradoxical relationship to this climate pattern, gullies are most abundant between *c.* 30–55° latitude (Heldmann & Mellon 2004; Balme *et al.* 2006; Dickson *et al.* 2007; Dickson & Head 2009). Malin & Edgett (2000, 2001) hypothesized that liquid water involved in gully formation was stored in confined, subsurface, geothermally warmed aquifers that periodically ruptured the overlying permafrost, allowing water to flow over the surface. The flow of erupted, solute-free water entraining clastic materials was modelled by Heldmann *et al.* (2005). They were able to demonstrate that, while liquid water is not presently stable on the Martian surface, liquid water could flow a similar distance to the length of gully channels while undergoing evaporation and freezing. That is, liquid water is meta-stable on Mars (Hecht 2002).

However, several observational idiosyncrasies of Martian gullies appear to be at odds with the confined aquifer model (for example, Fig. 6). These are: (a) gullies form at a range of elevations along Martian slopes and not always along an exposed bedrock (confining) layer; (b) gully channels and alcoves typically reach the apex of the slope on which they form, commonly meeting neighbouring gully channels and alcoves across a narrow topographic divide or at the apices of crater central peaks; (c) gullies are absent from the lowest regions of the Martian surface, such as the Hellas Basin, where groundwater would be most likely to outcrop (Balme *et al.* 2006); (d) gullies are exclusively present polewards of 25° latitude and are most common between *c.* 30–55° latitude (Dickson *et al.* 2007; Dickson & Head 2009); (e) gullies show a strong slope orientation preference, appearing on polewards-facing slopes at low latitudes (*c.* 25–40°), equator-facing slopes at middle latitudes (*c.* 40–55°) and polewards-facing slopes at high latitudes (>55°) (Christensen 2003; Dickson & Head 2009); and (f) radar observations of gully sites originally inferred to have formed from groundwater release show no strong subsurface radar reflections indicative of the presence of liquid water reservoirs (Nunes *et al.* 2010).

In light of these observations, a consensus is emerging among some Martian gully researchers that a top-down melting of near-surface ice and/or surface frost and snow may better account for the generation of gully meltwater than a groundwater-release mechanism (Costard *et al.* 2002; Christensen 2003; Dickson & Head 2009; Williams *et al.* 2009). This surface-ice melting process is consistent with recent modelling results showing that in some microclimates, during periods of high Martian orbital obliquity, water ice can accumulate by both atmospheric deposition (including frost emplacement and/or snowfall) and melt at the Martian surface. This emplacement and melting of surface ice occurs in protected gully alcoves, and can produce ephemerally present surface runoff sufficient to erode gullies at the precise latitudes, elevations, slopes and orientations at which they are observed (Costard *et al.* 2002; Hecht 2002; Williams *et al.* 2008, 2009).

It follows that morphological observations in permafrost regions can be used to differentiate between the two primary gully formation models: top-down melting and confined aquifer. Martian gully deposits are commonly found on surfaces modified by thermal-contraction-crack polygons (see previous section and Fig. 7). Thermal-contraction-crack polygons form in soil surfaces that are not merely frozen (mean annual temperature <273 K) but that are ice-rich. As a result, they are effectively impermeable on the timescales of water freezing/evaporation at the Martian surface (Heldmann *et al.* 2005; Levy *et al.* 2009*b*). Gully channels and alcoves commonly cross-cut thermal-contraction-crack polygons. In some locations gully fans overprint polygons and in other locations gully fans are cross-cut by thermal-contraction-crack polygons (Fig. 7). Levy *et al.* (2009*b*) interpret these stratigraphic relationships to indicate that widespread ice-cemented permafrost pre-dated the formation of the gullies, persisted through

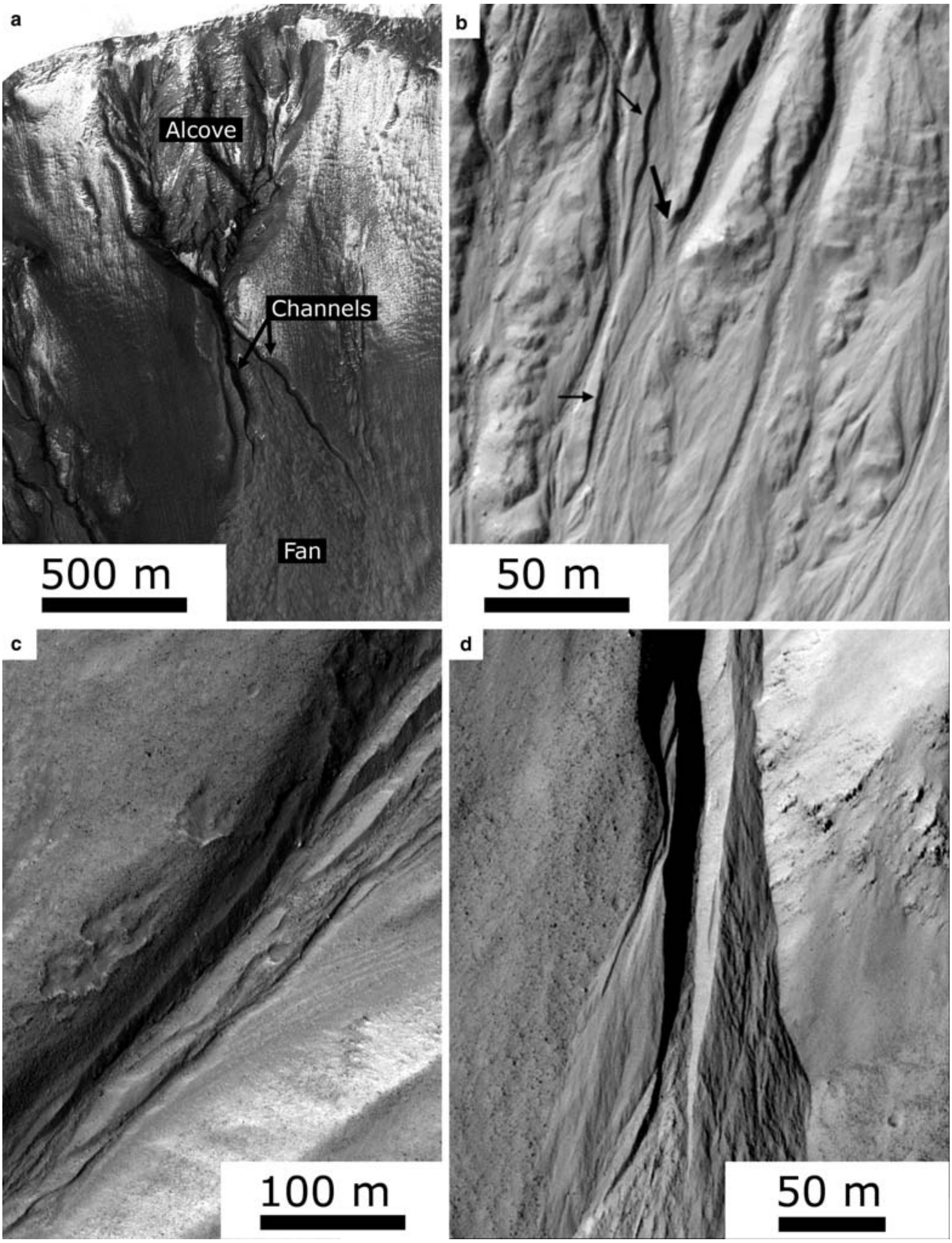

Fig. 7. Characteristic examples of Martian gully morphology indicative of multiple episodes of fluvial transport (modified from Schon & Head 2009). (**a**) Classic Martian gully on the interior slope of a crater. Gully elements – alcove, channel and fan – are illustrated. The gully extends all the way up to the crater rim. The sunlit side of the topographic divide is saturated with bright pixels in this contrast stretch (portion of PSP_001882_1410). (**b**) Sinuous and anastomosing gully channels (small arrows) and an eroded longitudinal bar (long arrow) downslope from a spur (modified from Schon & Head 2009, fig. 1). (**c**) Cut-banks, channel terraces and braided channels (portion of PSP_006593_1470). (**d**) A gully fan eroded by channels formed in subsequent flow events. A smaller channel has been abandoned and stranded at a higher topographic level from the large channel featured at image centre (portion of PSP_002292_1490).

the period of gully formation and endures to the present. This suggests the continuous presence of metre-thick effectively impermeable material underlying gullies in these locales. Water pressures exceeding the *c.* 2 MPa required to fracture ice-rich permafrost (Mellon 1997) would result in catastrophic eruption of water sourced by a confined aquifer, likely producing dramatic scouring in gullies that is not observed. Accordingly, Levy *et al.* (2009*b*) favour a top-down melting mechanism. The erosion of ice-rich, thermal-contraction-cracked permafrost during gully alcove formation suggests that some of the sediments involved in gully fan deposition are sourced in the underlying permafrost substrate (the LDM). This implies that ice-cemented, polygonally patterned permafrost may represent another critical element in the sedimentary system operating in the Martian cold desert.

Conclusions

The above examples provide an introduction to the range of permafrost landforms currently being explored on the surface of Mars and an illustration of some of the key processes in the geological development of Martian cold-desert landforms. Martian permafrost terrains represent the extreme cold and dry end of the wet-to-dry permafrost landform spectrum on planetary surfaces. The development of thermal-contraction-crack polygons on Mars appears to be largely incumbent on the presence of LDM deposits that feature excess ice in the shallow Martian subsurface and on cold, dry conditions under which sublimation is the dominant phase transition, rather than melting. Melting of near-surface snow, frost and/or ground ice during the most recent several million years on Mars has been largely confined to protected microclimates in gully alcoves, from which flows of water-borne sediment have been transported into their present configuration, forming gullies. Connecting these exceptionally young permafrost deposits to the longer term rhythms of climate change on Mars remains a topic of great interest to the geomorphologists, climate modellers and astrobiologists who will help guide the next generation of exploration in Martian polar regions.

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