Fluvial valley networks on Mars

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Introduction

The Earth, Mars and Saturn’s moon Titan are the only planetary surfaces known to have widespread, branching, fluid-carved channels or valley networks. Lava channels and collapsed lava tubes formed relatively few sinuous valleys with few to no tributaries on the volcanic plains of Venus (where the average air temperature is ≈ 450°C), the Moon (Figure 19.1(c)) and Jupiter’s moon Io (e.g. Wilhelms, 1987; Baker et al., 1997; McEwen et al., 2000). Venus also has some valley networks with rectangular, labyrinthic, pitted or irregular network structure, reflecting a joint volcanic and tectonic origin (Baker et al., 1997). Dense branching networks occur on Titan (Figure 19.1(a)), but liquid methane is the erosive fluid under its 1.6-bar nitrogen atmosphere at −180°C, and water ice comprises most of the bedrock and sediment (Tomasko et al., 2005). The current Martian atmosphere is too thin and cold to maintain water in liquid state, but the older terrains have been heavily eroded and incised by valley networks (Figure 19.1(b)), suggesting that past geologic or climatic conditions supported flowing water. These ancient networks are similar in some respects to their modern terrestrial counterparts, but they are immature if formed by runoff (e.g. Howard et al., 2005), and many authors attribute them primarily or exclusively to groundwater sapping (e.g. Pieri, 1980; Carr...
Figure 19.1  Sinuous valleys formed by flowing methane, water and lava respectively. (a) Networks on Saturn’s moon Titan debouched to dark plains (descent image from European Space Agency Huygens probe (credit: ESA/NASA/JPL/University of Arizona). (b) Evros Vallis on Mars (12.5°S, 14.5°E, THEMIS daytime infrared image mosaic, credit: University of Arizona). (c) Hadley Rille on the Moon (Apollo 15 image AS15-1135[M]).
and Clow, 1981; Carr and Malin, 2000; Gulick, 2001). The investigation of aqueous processes on early Mars is a major focus of NASA’s Mars Exploration Program, because environmental conditions suitable for liquid water may have supported life or prebiotic chemistry on Mars, and much of the geologic record from the first billion years of Earth’s history has been lost to erosion, metamorphism and subduction.

In this chapter, we describe the history and present state of Martian fluvial geomorphology, emphasizing quantitative analyses of drainage networks, watershed topography and hydrology. Most studies to date have focused on valley rather than channel characteristics, because aeolian processes and small meteorite impacts have slowly degraded and partially filled the valleys for \( \sim 3.7 \) Gyr since the epoch of widespread fluvial erosion (absolute, numerical age estimates herein follow Hartmann and Neukum, 2001), leaving few interior channels exposed (e.g. Carr and Malin, 2000; Irwin et al., 2005a). Moreover, the available orbital imaging has a resolution of \( \sim 0.3–250 \) m/pixel, with limited spatial coverage at high resolution, and robotic landers have not yet visited a valley network. Despite these limitations, investigators have used counts of superimposed impact craters, analyses of valley planform and topography, basic channel morphometry and theoretical modelling to constrain the age, formative processes, developmental history, hydrology, and climatic implications of Martian valley networks.

**Early observations**

Johannes Kepler’s laws of planetary motion, published in 1609 and 1618, were largely based on Tycho Brahe’s earlier observations of Mars. The seventeenth- and eighteenth-century astronomers Christian Huygens, Giovanni Cassini, Giacomo Miraldi and William Herschel recognized the Martian polar ice caps and estimated the planet’s orbit (1.52 AU), diameter (6792 km), rotation period (24 hours, 37 minutes) and axial tilt (25°) (modern data in parentheses are from Kieffer et al., 1992 and Smith et al., 2001). Formal telescopic maps succeeded early sketches during the nineteenth century, interpreting relatively bright and dark regions as continents and seas respectively (summarized by Flammarion, 1892, 1909). Even today, the best ground-based telescopes can resolve few other surface features. Popular conceptions of ‘Mars as the Abode of Life’ were also founded on perceived observations of dark-toned lineations on the surface, which Percival Lowell (1895, 1906, 1908) interpreted as bands of vegetation along artificial canals used to transport water from the ice caps to arid equatorial regions. Some contemporaries of Lowell who could not see the lineations were sceptical of these claims (e.g. Evans and Maunder, 1903; Wallace, 1907), and spacecraft data later showed that, with the possible exception of the Valles Marineris canyon system, the mapped ‘canals’ generally do not correspond to obvious valleys or other topographic features (Sagan and Fox, 1975). Nineteenth-century astronomers also observed the Martian atmosphere with clouds and global dust storms. Stoney (1898) correctly proposed that
the polar caps might contain frozen carbon dioxide (the wintertime caps at both poles and the top of the perennial southern cap), although most of their mass is water ice (Kuiper, 1952; Kieffer et al., 1976; Bibring et al., 2004). Theoretical calculations and infrared measurements suggested that Mars has an average surface temperature well below freezing ($-63^\circ$C, with extremes below the $-133^\circ$C freezing point of CO$_2$) (e.g. Wallace, 1907; Coblentz, 1925; Sinton and Strong, 1960). Kuiper (1952) made the first spectral identification of CO$_2$ in the atmosphere, although it was originally thought to be a minor component (now known to be 95 per cent by volume). The average atmospheric pressure was gradually revised below Lowell's (1908) 87-mbar estimate to $\sim$ 5.6 mbar, which is below the 6.1-mbar triple point where water first becomes stable as a liquid.

More detailed data from interplanetary spacecraft record conditions that are even less favourable for life. The first successful flyby mission to Mars, Mariner 4 in 1965, returned 22 images of a cratered landscape in the southern hemisphere, and the Mariner 6 and 7 flybys imaged a similar region at higher resolution in 1969. The preservation of ancient impact craters seemed to defeat the long-lived paradigm of a water-rich planet, but many of the Martian craters had been substantially modified (Leighton et al., 1965). Erosion rates on Mars were interpreted to be very low, but higher than on the Moon (Anders and Arnold, 1965; Baldwin, 1965; Hartmann, 1966; Öpik, 1966; Murray et al., 1971), and the low atmospheric pressure favoured wind over water as the erosive fluid (Sharp, 1968). Then in 1971, Mariner 9 became the first successful Mars orbiter, returning images of the entire surface at 1 km/pixel resolution with local imaging at 100 m/pixel scale. This mission revealed a planet much more diverse than the flyby missions had suggested, including giant volcanoes and canyons, circumpolar layered deposits, grabens, smooth plains, large outflow channels that resemble flood-carved features on the Earth and smaller valley networks (McCauley et al., 1972). Outflow channels are distinguished from valley networks by their large size (tens to $\sim$ 200 km wide and up to thousands of kilometres long), origin from a point source, anabranching path, streamlined islands and erosional bedforms (Figure 19.2). They are thought to have originated from very large, short-lived groundwater discharges of $\sim$ 10$^6$–10$^8$ m$^3$/s (Baker and Milton, 1974; Carr, 1979, 1996; Baker, 1982; Wilson et al., 2004), but overflow of enclosed basins also contributed to some large channels and valleys (e.g. Parker, 1985; Grant and Parker, 2002; Irwin et al., 2004).

Distribution, age, origin and morphology of valley networks

Geologic, topographic and regional distribution

Valley networks primarily occupy the most heavily cratered regions on Mars and are uncommon on younger surfaces (Pieri, 1976, 1980; Carr and Clow, 1981; Carr, 1996).
These ancient terrains are found on the southern highland plateau, whereas the northern lowland plains and western equatorial Tharsis volcanic province have been resurfaced since widespread valley development ceased (e.g., Tanaka, 1986) (Figure 19.3). Valleys are relatively sparse below about $-1500$ to $-1900$ m elevation, which may be attributable to thick volcanic and airfall mantling or contemporary base level control below that topographic level. Pieri (1976) and Carr and Clow (1981) showed that valley networks are concentrated in the dark-toned regions seen in telescopes, but that thick, brighter dust mantles may overlie valleys in the high-albedo areas. Valleys are evident but poorly developed in middle to high latitudes, owing to dust mantling (Soderblom et al., 1974;
Figure 19.3  (a) Global shaded relief map of Mars, with outlined location of (b). Valley networks are concentrated in the heavily cratered region that dominates the southern hemisphere. The northern lowland plains and the Tharsis and Elysium volcanic regions (∼0° N, 110° W and 20° N, 150° E respectively) have been completely resurfaced since the epoch of major fluvial activity. The circumference of Mars is 21,339 km at the equator, or 59.275 km per degree. (b) Valley networks mapped in an equatorial highland region by Carr and Chuang (1997) using Viking Orbiter imaging at 1:2-M scale (white lines), compared with major valley networks mapped by A. D. Howard using recent THEMIS imaging and MOLA topography (black lines). Martian valley networks are longer and more numerous than was evident in earlier imaging, but network integration does not approach that found on the Earth.
Pieri, 1976; Carr and Clow, 1981) or less favourable paleoclimates (Williams and Phillips, 2001). Many small valleys occur on intercrater plains, but some originate at crater rims or other sharp ridge crests where no upslope aquifers were available (e.g. Milton, 1973; Masursky et al., 1977; Craddock and Maxwell, 1993; Craddock and Howard, 2002; Irwin and Howard, 2002; Grant and Parker, 2002; Hynek and Phillips, 2003; Stepinski and Collier, 2004) (Figure 19.4). Most Martian valley networks are incised into darker layers of impact ejecta and sedimentary rocks (Malin, 1976a; Malin and Edgett, 2001), which were derived from basaltic igneous rocks (Christensen et al., 2001) during the epoch of heavy meteorite bombardment.

**Ages of valley networks**

The unexpected discovery of branching valleys on Mars (initially termed ‘small channels’, ‘runoff channels’ or ‘furrows’) raised the questions of when and for how long the valleys had been active. Unfortunately, we have no rocks from known locations on Mars that could provide absolute age estimates. Relative ages of planetary geologic units are determined using superposition relationships and impact crater populations, as surfaces accumulate craters with time (e.g. Carr, 1981, pp. 54–64). These crater counts are not easily converted to absolute ages, however, because the Martian impact cratering rate and how it declined over time are not precisely known (e.g. Hartmann and Neukum, 2001; Strom et al., 2005). Early workers concluded that the outflow channels and valley networks were relatively old and did not form during more recent cycles of lower-magnitude climate change (Hartmann, 1974; Sharp and Malin, 1975; Pieri, 1976). Malin (1976a, 1976b) dated the valley networks to the epoch of heavy meteorite bombardment, which declined about 3.85 Ga on the Moon and ~3.7 Ga on Mars, and later workers have confirmed this relative age (Masursky et al., 1977; Pieri, 1980; Carr and Clow, 1981; Baker and Partridge, 1986; Craddock and Maxwell, 1990; Maxwell and Craddock, 1995; Irwin and Howard, 2002; Ansan and Mangold, 2006). Fluvial activity primarily occurred within the Noachian Period at the base of Tanaka’s (1986) three-period stratigraphic scheme. Valley network activity was contemporary with the more rapid erosional modification and infilling of Noachian impact craters, whereas younger craters of the Hesperian and Amazonian Periods (<3.7 Ga) have a relatively fresh morphology (Craddock and Maxwell, 1990, 1993; Craddock et al., 1997; Forsberg-Taylor et al., 2004) (Figure 19.4(a)). Fresh crater populations constrain the end of the period of valley network activity. Short-lived, episodic outflow channel activity occurred over a much longer interval of time, from the Noachian and Hesperian to the Amazonian Periods (e.g. Mouginis-Mark, 1990; Zimbelman et al., 1992; Tanaka, 1997), and at least one channel potentially formed within the last 10⁸ yr (Burr et al., 2002; Berman and Hartmann, 2002).
Figure 19.4 The fluvial modification of impact craters included reworking the rough ejecta, eroding and widening the crater rim, gullying the interior wall and burying the central peak with material shed from the walls. (a) Overlapping fresh (top, left of centre) and modified impact craters in the Terra Sirenum region (25°S, 141.5°E). More heavily modified impact craters stratigraphically underlie fresher craters, unless the more degraded one is much smaller. (b) Impact crater with a densely dissected rim and a still-exposed central peak in Libya Montes, Terra Tyrrhena region (5°S, 72.6°E). (c) The more heavily modified Dawes crater in the Terra Sabaea region (9°S, 38.1°E). All are excerpts from the THEMIS daytime infrared mosaic.
Another issue is whether the highland valley networks were continuously active over a long period or whether they experienced multiple reactivations of regional to global extent early in Martian history (Grant, 1987; Grant and Schultz, 1990; Baker et al., 1991; Gulick et al., 1997). Some spatially localized valley networks dissect volcanoes and volcanic plains of Early to Late Hesperian age (Gulick and Baker, 1990; Scott et al., 1995; Mangold et al., 2004; Quantin et al., 2005), or they originate in the Noachian highlands and cross into younger terrains (Irwin et al., 2005b). Howard et al. (2005) and Irwin et al. (2005b) showed examples of late-stage valley entrenchment, overflow of previously enclosed basins and coarse sedimentation that occurred sometime between the Noachian/Hesperian boundary (~3.7 Ga) and the middle of the Hesperian Period (~3.6 Ga). This epoch of fluvial activity lasted in the order of ~10^5 yr and temporarily exceeded the erosion rates that prevailed before that time. Others have suggested that an early epoch of runoff declined to a regime dominated by groundwater sapping, which downcut the lower reaches of older valleys (Baker and Partridge, 1986; Harrison and Grimm, 2005). Some small impact craters and other steep slopes in the mid-latitudes have very fresh gullies, but these youngest features are localized and conveyed relatively little sediment (Malin and Edgett, 2000b).

Weathering and erosion rates declined severely when fluvial activity ended (Carr and Clow, 1981). Modified impact craters record denudation rates of ~0.1–10 m/Myr during the time of fluvial activity, whereas later Hesperian and younger craters (~3.6 Ga) are little modified (Craddock and Maxwell, 1993; Carr, 1996; Craddock et al., 1997). Small contemporary valleys on intercrater slopes (Figure 19.5) are typically 0.5–4 km wide (median 1.6 km) by 20–250 m deep (median 80 m), but they have experienced only ~20 m of infilling in 3.6 Gyr since the mid-Hesperian (Goldspiel et al., 1993a; Williams and Phillips, 2001). Weathering and the aeolian modification of plains have also been extremely slow. The infilling of impact craters limits the total redistribution of plains material since the Noachian to <0.01 m/Myr (Arvidson et al., 1979; Carr, 1996). Lander observations suggest aeolian denudation rates of plains basalt on the order of centimetres per billion years since the Noachian Period (Golombek and Bridges, 2000; Golombek et al., 2006), although the wind has deeply eroded less resistant, presumably fine-grained, layered sedimentary rocks or tephra in some areas (Ward, 1979; Malin and Edgett, 2000a). The water cycle appears to have declined suddenly, as interior channels do not record gradual declines in discharge, some valleys have V-shaped cross-sections that suggest active downcutting with little subsequent modification and most delta surfaces were not entrenched with declining water level in lake basins (Irwin et al., 2005b) (Figure 19.6).

**Water-source hypotheses and implications for paleoclimate**

The most contentious issue regarding valley networks is their water source. The first papers on Martian fluvial landforms compared the common theatre-headed valley
Figure 19.5 Classification of Martian valley networks. (a) Nirgal Vallis, a large stem valley with entrenched meanders and relatively few, stubby tributaries (28.3° S, 41.4° W). (b) Paraná Valles, a typical valley network with poorly dissected interfluve surfaces, a relatively constant width downslope and tributary valleys that are similar in width to the stem (22.4° S, 10.4° W). (c) Durius Vallis, a large stem valley with increasing width downslope and much smaller tributaries (16.6° S, 172° E). (d) Dense slope valleys on the southern rim of Schiaparelli crater (6.6° S, 16° E). Note origin of valleys near sharp drainage divides. All are excerpts from the THEMIS daytime infrared mosaic.

networks to box canyons with headwall springs in the south-western United States and Hawaii (Milton, 1973; Sharp and Malin, 1975; Malin, 1976a, 1976b; Pieri, 1976, 1980; Masursky et al., 1977; Carr and Clow, 1981; Baker, 1982; Mars Channel Working Group, 1983; Laity and Malin, 1985; Kochel and Piper, 1986; Howard et al., 1988). Groundwater sapping depends on spring discharge, which weathered the aquifer material, undermines the surface and extends a valley headward (Dunne, 1980). The hypothesis that groundwater alone carved the valley networks gained wide acceptance during and after the Viking missions (1976–1982). Later workers developed this concept, suggesting that prolonged groundwater sapping could occur without atmospheric recharge, if volcanic
intrusions caused sub-surface heat fluxes to vary over time. The geothermal heating of groundwater would encourage upward diffusion of vapour, which would accumulate near the surface as permafrost (Clifford, 1991, 1993; Clifford and Parker, 2001). This ice would later melt when a magma body was intruded (Wilhelms and Baldwin, 1989; Gulick and Baker, 1989, 1990; Brakenridge, 1990; Gulick, 1998, 2001; Harrison and
The resulting spring discharge might flow for long distances on the surface under an ice cover, as long as the ice remained intact and adequate heat was carried from the aquifer to balance heat lost to the atmosphere (Wallace and Sagan, 1979; Carr, 1983; Goldspiel and Squyres, 2000). Other authors suggested that impact heat would melt ground ice and release water from impact crater rims, forming gullies at higher elevations (Maxwell et al., 1973; Brakenridge et al., 1985). Investigators pointed to the primary igneous rather than weathered or evaporitic spectral signature of most regions (Bandfield et al., 2000, 2003; Christensen et al., 2001; Gaidos and Marion, 2003), and the incomplete dissection of the highlands by theatre-headed valleys as evidence against a warmer, wetter paleoclimate with widespread precipitation (e.g. Carr and Malin, 2000).

The groundwater-sapping hypothesis suffers from a number of weaknesses, summarized by Craddock and Howard (2002): (1) even flows with high sediment concentrations would require recharge to transport sediment equivalent to the measured volumes of valley networks (Howard, 1988; Gulick and Baker, 1990; Goldspiel and Squyres, 1991; Goldspiel et al., 1993b; Grant, 2000; Gulick, 2001). Storm runoff is responsible for much of the aquifer recharge and sediment transport in terrestrial desert rivers (e.g. Howard et al., 1988). (2) Geothermal mechanisms for near-surface ice accumulation and melting in equatorial regions are theoretical and lack empirical support, particularly as most valley networks formed in cratered rather than volcanic terrains. Reasonable geothermal heat fluxes might have supported liquid water at least 300 m below the surface (Goldspiel and Squyres, 2000; Travis et al., 2003), but most valley networks are not, and never were, that deep (Goldspiel et al., 1993a; Williams and Phillips, 2001; Howard et al., 2005). (3) Flowing water must have been spatially ubiquitous, temporally long-lived (at least episodic over \(2 \times 10^8\) yr) and volumetrically abundant to modify impact craters. Where adjacent Noachian craters of similar size are observed, the stratigraphically older ones are more heavily modified (Figure 19.4(a)). Fluvial erosion and sedimentation are required to explain the concave profile at the transition between a crater wall and floor (Craddock et al., 1997; Forsberg-Taylor et al., 2004). (4) Theatre headwalls are not unique to groundwater sapping but are found in many valleys where waterfalls sap (i.e. undercut) a resistant caprock and erode a weaker basal layer. (5) The lack of massive carbonate deposits on Mars may reflect an acidic environment rather than a lack of surface water (Fairén et al., 2004). Adequate impact gardening and aeolian erosion have occurred to expose fresh basaltic surfaces to orbiting sensors (Mars Exploration Rover findings have recently supported both of these claims, e.g. Squyres et al., 2004, Golombek et al., 2006). (6) Climatic models and morphometric analyses often yield ambiguous results, as discussed below.

To carve valleys by overland flow, Mars would require a thicker, warmer atmosphere capable of supporting more intense rainfall or snowmelt and long-distance flow without freezing (Sagan et al., 1973; Pollack et al., 1987). Climate modellers have raised the main
objection to this concept, with most finding that a thick CO\textsubscript{2} greenhouse alone could not warm Mars above a globally averaged 0\textdegree C (Kasting, 1991; Squyres and Kasting, 1994; Haberle, 1998; Colaprete and Toon, 2003), because changes in the Sun’s elemental composition through time imply that it was only $\sim$ 75 per cent as bright at 3.7 Ga (Gough, 1981). Recent studies have suggested that the early Sun may have been a few percent more massive than at present, making it less dim than otherwise expected (e.g. Boothroyd \textit{et al.}, 1991; Graedel \textit{et al.}, 1991; Whitmire \textit{et al.}, 1995). Other greenhouse gases (Sagan and Chyba, 1997), including water vapour excavated by large impacts (Segura \textit{et al.}, 2002; Colaprete \textit{et al.}, 2005) and volcanism (e.g. Baker \textit{et al.}, 1991), would also contribute to greenhouse warming. This issue has not yet been resolved, but a combination of the above factors may have contributed to a long-lived or episodic water cycle on early Mars.

Valley morphology and diversity

Published classification schemes have differentiated Martian valleys by size, morphology and network planform. Excluding troughs related to volcanoes, crustal extension and collapse, three basic categories include: (1) ‘fretted’ valleys with a gridded planform and little evidence of through-flowing water (the origin of these valleys remains unclear, Carr, 2001), (2) the monolithic outflow channels (Figure 19.2) and (3) smaller valley networks. Early authors also subdivided the latter category in a qualitative but broadly consistent manner (Figure 19.5), summarized as: (3a) large, widely spaced, sinuous stem valleys that are $\sim$ 10 km wide and $\sim$ 1 km deep ($\pm$ 50 per cent), increase in width downslope and have tributaries much smaller than the stem; (3b) small valley networks with more closely spaced tributaries, which have similar width to higher-order segments downslope; and (3c) dense, sub-parallel slope valleys (Masursky, 1973; Sharp and Malin, 1975; Pieri, 1976; Masursky \textit{et al.}, 1977). Valley spacing and length decrease from category 3a to 3c. Pieri (1980) refined his earlier work to include eight classes of network planform, including digitate (fanned), stem (category 3a), parallel (category 3c), rectilinear, radial centrifugal (away from central highs) and two types of radial centripetal planform (exterior and interior drainage into central basins). True dendritic patterns reflecting the full development of network structure under homogenous geological conditions have not been seen on Mars, although the term has been casually applied in planetary literature.

Most Martian fluvial valleys have either flat-floored or V-shaped cross-sectional profiles. The former category (including much of 3b above) includes a trapezoidal cross-section with sidewalls near the angle of repose, a nearly constant valley width downstream, an amphitheatre headscarp and poorly dissected interfluve areas between major tributaries (e.g. Pieri, 1980; Baker, 1982; Mars Channel Working Group, 1983). This morphology is most common on low-gradient intercrater plains. Most valleys
with V-shaped cross-sections occur on steeper headwater slopes (Figure 19.5(d)) or at downstream sites where a valley incised a convex break in slope, such as a crater rim (Figure 19.6(c)) (Baker and Partridge, 1986; Williams and Phillips, 2001). Fully dissected surfaces, leaving sharp divides between tributaries, are uncommon and are usually restricted to steep interior walls of impact craters and other slopes (Figures 19.5(d) and 19.6(a)) (e.g. Moore and Howard, 2005; Quantin et al., 2005).

The two types of cross-sections may represent different formative processes or lithologic controls, or they may be gradational forms. The V-shaped valleys probably represent late, rapid downcutting by runoff along a steep gradient, with little subsequent modification. Valley measurements show that width increased in proportion to depth of incision until the longitudinal profile stabilized, and valleys continued to widen afterwards to produce the flat-floored shape (Williams and Phillips, 2001). Rapid headward erosion or downcutting, possibly due to runoff bottlenecks in a cratered landscape, with little time afterwards for widening would produce a nearly constant valley width downstream in a runoff-dominated regime. Alternatively, if headward extension was due to slow groundwater sapping, the valley width and cross-section indicate that nearly all water originated at the valley head (e.g. Goldspiel et al., 1993a; Grant, 2000). The common valley headscarps are likely attributable to sapping (i.e. undercutting) in layered rocks, either by springs or waterfalls.

Morphometry

A number of investigators have tested the overland flow and groundwater hypotheses by comparing the morphometry (length, sinuosity, drainage density, cross-sectional profiles, network planform and longitudinal grading) of Martian valley networks to mature terrestrial networks. These studies have all shown significant differences between Martian drainage basins and the ideal quasi-equilibrium condition, but it remains uncertain to what degree these differences represent immaturity or a different water source.

Between 1997 and 2001, the Mars Orbiter Laser Altimeter (MOLA) on the Mars Global Surveyor (MGS) orbiter returned the first precise topographic map of Mars at < 0.5-km resolution. Previously, global-scale topography had very poor resolution and incorporated vertical errors up to a kilometre, as elevation estimates were based on the topographic and atmospheric occultation of a spacecraft’s radio signal as it passed behind the planet (Kliore et al., 1973; Smith et al., 2001). Studies of valley network development were thus restricted to the plan view of orbital imaging, although less precise local measurements of valley slopes and depths were made using stereo imaging, brightness contrasts across an image (photoclinometry), Earth-based radar tracks and shadows (e.g. Thornhill et al., 1993; Goldspiel et al., 1993a, 1993b; Lucchitta and Dembosky, 1994).
Planimetric measurements

Quantitative studies based on \( \sim \) 230-m/pixel orbital imaging showed that Martian valley networks are relatively short and discontinuous, with common lengths of tens to hundreds of kilometres (Carr and Clow, 1981; Baker and Partridge, 1986; Carr, 1995; Cabrol and Grin, 2001). Martian watersheds are poorly integrated, and the many enclosed impact craters and intercrater basins include sites where cratering disrupted earlier fluvial pathways. Some basins (particularly craters that formed on pre-crater slopes) are infilled or breached, but most larger basins drained internally (Grant, 1987; Goldspiel and Squyres, 1991; Maxwell and Craddock, 1995; Irwin and Howard, 2002; Kramer et al., 2003). Some significant exceptions have lengths of 1000–4700 km, including larger stem valleys and other networks that crosscut multiple basins on long regional slopes (e.g. Carr and Clow, 1981; Irwin et al., 2005b) (Figure 19.3(b)).

In a study of 71 typical valley networks, Cabrol and Grin (2001) found that most were of Horton (1945) order 3–4, reflecting short length with limited tributary development. In 14 large networks with a maximum Strahler (1952) order of 4, Carr (1995) found bifurcation ratios of 2.9 to 7.6 (average 4.3), similar to terrestrial networks. Length ratios were also comparable but had a relatively large range of 1 to 6.9 (average 2.9), possibly reflecting underdeveloped drainage basins. Using higher-resolution imaging, Ansan and Mangold (2006) report similar results but a higher network order of 5–7 for Warrego Valles. On some volcanoes with dense valleys, Gulick and Baker (1990) measured Shreve (1966) network magnitudes from 2 to 34.

Most Martian valleys have low sinuosity (Grant, 2000), as do their interior channels where evident (Irwin et al., 2005a). Many investigators have attributed this relative straightness to a structural control of groundwater flow (e.g. Pieri, 1980; Brakenridge, 1990); however, new topographic data show that nearly all Martian valley networks follow the steepest topographic gradient, regardless of the local structure. Straight or braided (as opposed to meandering) reaches of terrestrial streams occur where stream power, bank erodibility and a relative supply of bedload are all very low or very high respectively (Knighton, 1998). Few meandering alluvial channels are evident on Mars, either because these channels were too shallow to be preserved or because the required sets of conditions were not often met, but some Martian stem valleys have entrenched meanders (Figure 19.5(a)). These features record meandering surface flowpaths, which can develop over \( \sim \) 100- to 1000-year timescales, that experienced longer-term downcutting due to excess transport capacity (i.e. stream power) relative to sediment supply. Tectonic uplift does not appear to have been important on Mars, but most valleys with entrenched meanders extend from a low-gradient plain onto a steeper surface, such as an impact crater rim or the wall of a deep stem valley, encouraging headward incision.
Many investigators have measured drainage density, the total length of valleys per unit area, in local to regional study areas. Most studies have found values in the order of $10^{-2}$ to $10^{-1}$ km/km$^2$ on dissected surfaces, one to three orders of magnitude less than typical terrestrial values (Grant and Boothroyd, 1985; Baker and Partridge, 1986; Grant, 1987; Grant and Schultz, 1993; Tanaka et al., 1998; Grant, 2000; Cabrol and Grin, 2001; Gulick, 2001; Irwin and Howard, 2002; Craddock and Howard, 2002; Hynek and Phillips, 2003; Stepinski and Collier, 2004; Ansan and Mangold, 2006; Luo and Stepinski, 2006). Drainage densities above 1.0 have been measured only in Valles Marineris (Mangold et al., 2004) and on some volcanoes (Gulick and Baker, 1990), and the reasons for this variability across Mars remain uncertain (Luo and Stepinski, 2006). Carr and Chuang (1997) made the first effort to quantify Martian drainage densities on a global scale (Figure 19.3(b)). They compared valley networks digitized on Viking Orbiter imaging (~230 m/pixel, 1:2-M scale) with Landsat images of Arizona, Nebraska, New York, Texas and Washington that were degraded to a similar resolution. They found that the average drainage density on Noachian plains is approximately 0.0032 km/km$^2$, but that terrestrial values were 0.065–0.209 km/km$^2$ over the range of climates studied. Several issues complicate such direct comparisons. (1) Viking Orbiter images were taken with different viewing geometries, times of day and atmospheric conditions, so that a feature visible in one image is often difficult to distinguish in the adjacent image. (2) The Landsat spectral bandpasses were selected primarily to monitor vegetation, which is often concentrated around stream channels, whereas images of Mars show little contrast except on steep slopes or compositionally distinct geological units. (3) Terrestrial rivers have been recently active, whereas Martian valleys have experienced 3.7 Gyr of degradation by wind, mass wasting and small impacts. Martian valleys may be evident only where they were deeply incised and not deeply buried, particularly in low-resolution imaging. (4) Regional measurements of drainage density incorporate recent deposits as well as ancient depositional basins, where a shallow channel network would have been easily buried or erased.

In general, new imaging at 1–100 m/pixel from the Mars Orbiter Camera (MOC) on MGS and the Thermal Emission Imaging System (THEMIS) on Mars Odyssey revealed more tributaries, a higher drainage density and better integration than was previously evident (e.g. Hynek and Phillips, 2003; Figure 19.3(b)), but Martian valley networks still appear underdeveloped relative to their terrestrial counterparts. All previous studies have concluded that poor development (i.e. formation, incision or preservation) of headwater tributaries is the main cause of low drainage density on Mars. If widespread precipitation was available, high infiltration capacities maintained by cratering may have impeded runoff production (Baker and Partridge, 1986; Gulick and Baker, 1990; Grant and Schultz, 1993; Carr and Malin, 2000). Alternatively, an Earthlike climate may have prevailed for a limited period.
Network-junction angles and drainage-basin topography

One of the few characteristics of valley networks that can be quantitatively assessed using Mariner 9 and Viking images is the angular structure of the network, particularly the angles of valley junctions. Early geomorphic literature (e.g. Horton, 1932) suggested that the planimetric form of valley networks was inherited from the topography at the time of initial channel incision (this may be largely true for Mars). Howard (1971) suggested, however, that junction angles dynamically adjust as topography evolves and proposed geometric and minimum power criteria for junction angles. A consequence of these models is that mean junction angles increase with concavity of the drainage network (e.g. Howard, 1990; Sun et al., 1994) and small tributaries merge with large rivers at high junction angles, relative to smaller angles between tributaries of equal size and order. Pieri (1980) showed that Martian junction angles tend towards small values and a high degree of irregularity, which he suggested was due to the immature state of the valley network, with strong structural controls and a lack of sufficient net erosion to develop a highly concave profile. The structure of the network also varies with the scale at which it is observed, a feature not seen in true dendritic networks on Earth.

Several investigators have used the D8 algorithm (Tarboton et al., 1991) to extract information on drainage basins (Hynek and Phillips, 2003; Stepinski and Collier, 2004). In this method, the surface flow direction is the steepest downward slope from the centre of a given pixel to the centre of the eight surrounding it. Flow direction is then integrated to determine the most probable flowpaths for surface water over the given DEM. Streams of different order and magnitude are also identified following several conventions (Horton, 1945; Strahler, 1952; Shreve, 1967). This information can then be used to characterize a number of parameters useful for describing a valley network system (e.g. contributing area, relief and steam order). However, an uncritical application of D8-based methods to MOLA topography leads to the artificial generation of drainage patterns on both dissected and undissected surfaces. Fresh craters have modified the surface since the time of valley network activity, and the derived order and drainage density of a network are functions of the DEM’s resolution. Manual editing of computationally identified networks is therefore required.

Although valley networks do not fully dissect the Martian surface, topography can be used to evaluate the cratered landscape’s adjustment to hypothetical fluvial processes. Stepinski et al. (2002, 2004) showed that runoff on Mars would organize with fractal planar characteristics similar to terrestrial networks. However, drainage basin length ($L$), area ($A$), slope and drainage-density characteristics reflect significant influences from both fluvial erosion and contemporary impact cratering. More densely dissected surfaces show a better adherence to Hack’s (1957) Law ($L \propto A^{0.6}$) than other highland areas with similar crater populations, but Martian drainage basins tend to be more
elongated ($L \propto A^{0.73}$). The poor longitudinal grading on Mars caused more energy to be dissipated in high-order segments downstream, a likely indicator that the networks were still growing headward and incising when the water supply declined. They also found that latitude and elevation have no net influence on watershed development (see also Luo and Stepinski, 2006). Luo (2000, 2002) compared the hypsometry of Martian and terrestrial drainage basins, finding that some have deeply incised stem valleys that are often attributed to groundwater sapping, whereas others have hypsometric characteristics more similar to graded watersheds. Topographic analyses using a circularity function (a plot of changes in a drainage basin’s shape at different topographic levels) suggest that Martian valley networks are entrenched below a precursor surface that has not been fully regraded by prolonged fluvial erosion (Stepinski and Coradetti, 2004; Stepinski and Stepinski, 2005; Luo and Howard, 2005). These studies favourably compared Martian drainage basins to terrestrial analogues in hyper-arid climates. Valley longitudinal profiles are commonly irregular, reflecting modest total erosion (e.g. Aharonson et al., 2002; Howard et al., 2005; Irwin et al., 2005b; Kereszturi, 2005). An arid climate with ephemeral runoff or a short duration of conditions favourable to precipitation may be responsible for low fluvial incision.

Alluvial deposits

Both alluvial fans and likely deltas have been recognized along the margins of Martian basins. These landforms provide a depositional record of past fluvial activity that is broadly similar in magnitude but shorter in duration relative to terrestrial desert environments.

Fans

Noachian impact craters with diameters of 10–70 km typically have 500 to 1000 m of sedimentary fill, and many have lost their well-defined rims to erosion (Figure 19.4) (Craddock et al., 1997; Craddock and Howard, 2002; Forsberg-Taylor et al., 2004). These crater floors typically decline towards the centre with slopes of about 0.5–1°, suggesting that the floor materials are fluvial bajadas supplied by parallel gullies on the craters’ interior walls. Lava eruptions or intrusions as well as airfall deposition may also have contributed to basin infilling, particularly in craters above this size range that often have flatter floors (Craddock and Howard, 2002). Well-developed, cone-shaped alluvial fans, with lengths of tens of kilometres and gradients of a few degrees, occur in some deep craters that formed late in the period of fluvial erosion on Mars (Moore and Howard, 2005). These fans typically radiate from deep, thoroughly dissected alcoves in the crater walls (Figure 19.6(a)). The gradients, size and concavity
of the alluvial fans quantitatively relate to the size and slope of the eroded alcoves in a manner that closely approximates relationships for large terrestrial alluvial fans. On some of the fans, the selective aeolian erosion of fines has revealed distributary channels in inverted relief (Figure 19.6(b)). The distributary network structure, fan gradients and channel-width suggest fluvial sedimentation rather than debris flows (Moore and Howard, 2005).

**Deltas**

Most valley networks debouch into impact craters or enclosed intercrater basins, but valley floors are usually graded to the terminal basin floor, with no positive-relief fan or delta at the valley mouth. This characteristic suggests that deep paleolakes were rare or short-lived. Either water was delivered to most basins less rapidly than evaporation and infiltration removed it or water levels fluctuated widely across the basin floors, keeping thick sedimentary deposits from accumulating at the basin margins. These comments are speculative, as the environmental and physiographic conditions that favoured paleolake development have not yet been constrained.

Irwin et al. (2005b) reviewed the literature on Martian paleolakes and listed 33 scarp-fronted deposits where valleys debouch into impact craters or other basins (e.g. Figure 19.6(c), (d)). Many of the putative deltas recognized by Cabrol and Grin (1999) could not be relocated in new, higher-resolution imaging. The deposits resemble deltas due to the steep scarp along their outer margins, although the aeolian deflation of fine sediments around an alluvial lag might yield a similar form. Several of these deposits have distributary channels, occasionally in inverted relief (Figure 19.6(d)). In other cases, distributary channels have been mantled, reworked or did not form (although the latter case would imply that these are not subaerial deltas). Other likely deltas have since been discovered, reflecting multiple lake levels (e.g. Di Achille et al., 2006; Weitz et al., 2006).

Few of the putative Martian deltas have been studied in detail, as most were discovered in the last several years when decametre- to metre-scale imaging became available. Malin and Edgett (2003) and Moore et al. (2003) described the most spectacular fluvial deposit on Mars, an 11- by 13-km (6–13 km$^3$) distributary fan in the 64-km Eberswalde crater (Figure 19.6(d)). This deposit has at least three lobes at different elevations, suggesting two stands of lake level and a complex network of meandering distributary channels that show evidence of lateral migration, vertical aggradation and avulsion. Fassett and Head (2005) described a broadly similar pair of deltas on the opposite side of the planet.

Well-developed alluvial fans and deltas appear to have formed during a terminal epoch of relatively intense fluvial erosion on Mars (Howard et al., 2005; Irwin et al., 2005b; Moore and Howard, 2005). There is little evidence for similar degraded, cone-shaped, gravelly alluvial fans and deltas dating from earlier in the Noachian, but the
significance of this absence is uncertain. Such features may have been degraded by wind erosion or buried as the craters filled with sediment; alternatively, the earlier erosional and depositional environment may have been less intense than the later period when well-developed valleys, fans and deltas formed.

Hydrology

The dominant discharge, annual runoff volume runoff per unit area, and flow longevity of Martian valley networks are all poorly constrained at present. The few recognized interior channels have experienced prolonged dry conditions with aeolian infilling and some modification of the channel banks, and only one basin has been used for an input/evaporation balance that loosely constrains the annual water budget (Irwin et al., 2005a, 2005b). However, where channel and basin dimensions can be measured, quantitative techniques applicable to alluvial channels are available to constrain the hydrology of ancient Martian rivers.

Scaling equations to Martian gravity

Both theoretical and empirical fluid flow equations can be adjusted for 0.38 times the terrestrial gravitational acceleration to estimate discharge, flow velocity, particle-settling velocity, bed shear stress, critical shear stress for entrainment and sediment-transport rates on Mars. The discharge $Q$ (m$^3$/s) of a channel can be calculated using a combination of the unit-balanced continuity and Darcy-Weisbach equations (the latter is similar in form to the Manning equation):

$$Q = HWV = H^{1.5}W(8gS/f)^{0.5}$$

(19.1)

where $H$ is mean flow depth (m) for channels with high width/depth ratio, $W$ is channel width (m), $g$ is gravitational acceleration (m/s$^2$), $S$ is slope and $f$ is the Darcy-Weisbach friction factor. Equation (19.1) is only useful where depth is known, and measurement errors are significant as depth is the most important contributor to discharge. Measuring the depth of a channel with stereo imaging or shadows and assuming bankfull conditions may yield large errors, since the channel may be either deeply entrenched below a terrace or partly filled with sand. The friction factor must also be estimated, but this is a smaller source of error given the natural range of values and its exponent of 0.5.

A simpler method uses only channel width and assumes that a dominant discharge controls this and other channel dimensions, which has been demonstrated in a variety of terrestrial settings. In humid regions, this flood has a recurrence interval of one to two years (Knighton, 1998), but less frequent floods often dominate arid-zone rivers.
Meander wavelength $\lambda$ scales with channel width as:

$$\lambda = k_\lambda W$$

(19.2)

on both Earth and Mars (Moore et al., 2003; Irwin et al., 2005a), and $W$ scales with discharge as:

$$W = k_w Q^{0.5}$$

(19.3)

where the coefficients $k_\lambda$ and $k_w$ are $\sim 10^{-14}$ and $\sim 3-5$ respectively. Solving for $Q$ yields:

$$Q = (W/k_w)^2 = (\lambda/k_\lambda k_w)^2$$

(19.4)

The coefficients depend on the contemporary resistance of the channel banks, as more resistant banks yield narrower and deeper channels with a somewhat smaller wavelength, but bank strength is unknown for Mars. Equation (19.1) shows that decreasing gravity reduces flow velocity, so, if slope and roughness are held constant, width and depth must be greater per unit discharge on Mars. The greater depth increases velocity above the factor 0.62 that would result from reducing gravity with a channel of the same dimensions (Pieri, 1980; Komar, 1980b). Empirical data show that $H \propto W^{0.69}$ (Williams, 1988), a relationship that approximately yields Equation (19.4) if substituted into Equation (19.1). In this case, width, depth and velocity on Mars would be 1.27, 1.18 and 0.67 times their value on Earth, and the discharge resulting from Equation (19.4) would be multiplied by 0.62 ($1.27^{-2}$).

The algebraic manipulation of regression equations introduces significant errors, however, so a function determined with $Q$ as the dependent variable is favoured (Williams, 1988). For a conservative estimate of discharge, Irwin et al. (2005a) measured the channel floor width rather than the bank-to-bank width (in case later mass wasting had modified the banks), and they applied an equation for sand-bed/sand-bank channels that are relatively wide per unit depth (Osterkamp and Hedman, 1982):

$$Q = 1.9 W^{1.22}$$

(19.5)

If $H \propto W^{0.15}$, an extreme case that may apply to some sand-bank channels, the width, depth and velocity scaling would be 1.48, 1.06 and 0.64 the terrestrial values respectively. To scale Equation (19.5) to Martian gravity, the result must also be multiplied by 0.62 ($1.48^{-1.22}$). Based on this derivation, the channel-forming discharges calculated by Irwin et al. (2005a) could be reduced by 18 per cent (they used a scaling coefficient of 0.76 assuming a smaller ratio of width to depth) to provide as conservative an estimate as the data and regression functions could reasonably support. If the channel banks were
more cohesive, the dominant discharges for the larger drainage basins would be higher than they reported.

Nummedal (1977), Komar (1980a, 1980b) and Pieri (1980) compared a similarly sized channel rather than a similar discharge between Earth and Mars, but the same discharge would form a wider and deeper channel on Mars, increasing a flow’s erosional efficiency beyond the effect of lower gravity. The critical shear stress $\tau_c$ (N/m$^2$) needed to mobilize a particle scales directly with gravity as:

$$\tau_c = 0.06(\rho_s - \rho_w)gD$$

(19.6)

where $\rho_w$ is fluid density, $\rho_s$ is particle density (kg/m$^3$) and $D$ is the particle diameter (m). Bed shear stress $\tau$ also scales with gravity as:

$$\tau = \rho_w g HS$$

(19.7)

However, for a given discharge and slope, the shear stress applied to a channel bed would be 0.40–0.45 or more of the terrestrial value, depending on the ratio of channel width to depth that is controlled by the resistance of channel-bank material. The gravitational force applied to a suspended particle scales with $g$, thus the settling velocity for coarse particles is multiplied by $\sqrt{0.38}$ and smaller factors for smaller particles, but flow velocity is at least 0.64–0.67 times the terrestrial value. These relationships impart a slight efficiency to gravel transport on Mars, but, once mobilized, smaller particles would remain suspended much longer in Martian rivers (Nummedal, 1977; Komar, 1979, 1980b; Pieri, 1980), increasing their transport rate by a factor of $\sim$ 1.5.

In contrast, the lower bed shear stress and particle settling velocities would reduce the abrasion rate of channel bedrock. Corrosion depends on pH, which is unknown but likely more acidic in an atmosphere rich in CO$_2$ and SO$_2$. The onset velocity for cavitation scales with $g^{0.5}$, so lower gravity provides little benefit, but a hypothetically lower atmospheric pressure would enable the process at lower mean velocities (Baker, 1979). For example, in a 100-mbar atmosphere, flows 1- and 10-m deep would undergo cavitation at velocities of 3 and 6 m/s respectively. These values correspond to fairly steep threshold slopes of 0.03 (1.64°) and 0.005 (0.3°) at downslope locations where contributing area is sufficient to accumulate flows of that depth. Considering these environmental effects, the incision of small headwater tributaries would take longer for a given flow rate, requiring a longer period of erosion (and more total water) or more rapid weathering for the same amount of bedrock erosion. Moreover, runoff production per unit precipitation is low where tributaries are poorly developed, reducing the effectiveness of flash floods (Patton and Baker, 1976). Scaling arguments suggest that Martian channels should have more bedrock- and gravel-floored reaches and distal fine-grained deposits with lower gradients.
Applications

Weihaupt (1974) was the first to apply paleohydrologic methods to a Martian fluvial system, Nirgal Vallis, which has well-developed, entrenched meanders along its main stem (Figure 19.5(a)). Meandering is not a recognized attribute of valleys carved by groundwater sapping, but the much smaller, straighter, theatre-headed tributaries to the main valley have no apparent overland drainage. Weihaupt (1974) estimated the mean annual flood based on the meander wavelength of the bedrock valley, and the bankfull discharge was based on the width of the valley floor, taking that to be a channel bed rather than a floodplain or terrace. These methods suggested discharges of 2700 m$^3$/s and 100 000 m$^3$/s respectively, which were not scaled for gravity. Malin and Edgett (2001) identified an interior channel that is locally exposed on the valley floor. This channel’s width is 12 per cent of the meander wavelength, as expected from Equation (19.2), so the discharge estimate is reduced to 4800 m$^3$/s, scaled for gravity and assuming poorly resistant channel banks (Irwin et al., 2005a).

Moore et al. (2003) found that the channel width on the Eberswalde crater delta was 14.5 per cent of the meander wavelength and that bankfull discharge was about 700 m$^3$/s, using Equation (19.4). Jerolmack et al. (2004) calculated the discharge at 410 m$^3$/s, using an equilibrium model for channelized alluvial fans. Given that discharge, the channel-bed materials would be too coarse for wind to remove, but the suspended load deposited on the floodplain would be susceptible to aeolian erosion. This feature explains why the channel beds were preserved as ridges while much of the floodplain had been blown away. The development of meanders by lateral accretion was also consistent with the estimated stream power (Irwin et al., 2005b). Based on reasonable sediment yields and timescales for meander development and avulsion, the longevity of the contributing valley network ranges from $10^5$–$10^6$ yr (Moore et al., 2003), with a favoured timescale of at least $\sim 10^5$ yr (Bhattacharya et al., 2005) that is comparable to the time required to construct large alluvial fans in some Martian impact craters (Moore and Howard, 2005). A reasonable evaporation rate on the order of $\sim 1$ m/yr from the lake would suggest $\sim 0.1$ m/yr of runoff from the contributing basin (Irwin et al., 2005b).

Pieri (1980), Irwin et al. (2005a, 2005b), Jaumann et al. (2005) and Howard et al. (2005) have also recognized interior channels within valley networks (Figure 19.6(e)). Larger drainage basins typically yield wider channels, and their discharge from Equation (19.5) divided by a topographically defined contributing area suggests that runoff production rates locally exceeded 1 cm/day at times. Runoff production rates were smaller for larger drainage basins, which incorporate low-gradient plains and which may have been larger than the individual storm cells.
Summary

In many respects, the surface of Mars is intermediate between the cratered Moon and the deeply eroded Earth. Early telescopic observations suggested conditions favourable for life, but later observations and spacecraft imaging revealed a hyperarid planet with very low long-term erosion rates. The Martian highland landscape reflects the prolonged fluvial erosion of Noachian impact craters (Craddock et al., 1997), but this process may have been discontinuous and was not intense enough to fully regrade the landscape into a set of mature drainage basins. As a result, the length, order, network structure, junction angles, sinuosity, drainage density, cross-sectional and longitudinal profiles, and watershed topography of relict valley networks still reflect imposed crater topography. These generally sparse, immature valleys with poorly dissected interfluve areas were incised during one or more epochs of more intense fluvial activity around the Noachian/Hesperian transition (3.7 Ga). Dominant runoff comparable to terrestrial mean annual floods was associated with local ponding and the deposition of alluvial fans and deltas. The short longevity of Earth-like conditions, the inefficiency of fluvial abrasion under lower gravity and possible high infiltration capacities inhibited the development of headwater tributaries on a relatively cool and arid planet. Erosion rates declined suddenly and severely following this epoch, leaving undissected delta surfaces and well preserved channels and V-shaped valleys. Aeolian erosion and gardening by small impacts has degraded the valley networks somewhat, but their preserved morphometric characteristics are useful in deciphering the environmental conditions of early Mars.

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References

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