High outflow channels on Mars indicate Hesperian recharge at low latitudes and the presence of Canyon Lakes

Neil M. Coleman a,*, Cynthia L. Dinwiddie b, Kay Casteel c

a US Nuclear Regulatory Commission, Mail Stop T2E26, Washington, DC 20555, USA
b Department of Earth, Materials, and Planetary Sciences, Southwest Research Institute®, 6220 Culebra Road, San Antonio, TX 78238-5166, USA
c Mercersburg, PA 17236, USA

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Abstract
A number of martian outflow channels were carved by discharges from large dilational fault zones. These channels were sourced by ground-water, not surface water, and when observed on high-standing plateaus they provide indicators of elevated paleo-groundwater levels. We identify three outflow channels of Hesperian age that issued from a 750-km-long fault zone extending from Candor Chasma to Ganges Chasma. Two of these channels, Allegheny Vallis and Walla Walla Vallis, have sources > 2500 m above the topographic datum, too high to be explained by discharge from a global aquifer that was recharged solely in the south polar region. The indicated groundwater levels likely required regional sources of recharge at low latitudes. The floodwaters that erupted from Ophir Cavus to form Allegheny Vallis encountered two ridges that restricted the flow, forming temporary lakes. The flow probably breached or overtopped these obstructions quickly, catastrophically draining the lakes and carving several scablands. After the last obstacle had been breached, a single main channel formed that captured all subsequent flow. We performed hydrologic analyses of this intermediate phase of the flooding, prior to incision of the channel to its present depth. Using floodwater depths of 30–60 m, we calculated flow velocities of 6–15 m s\(^{-1}\) and discharges in the range of 0.7–3 \(\times\) \(10^6\) m\(^3\) s\(^{-1}\). Locally higher flow velocities and discharges likely occurred when the transient lakes were drained. Variable erosion at the channel and scabland crossing of MOLA pass 10644 suggests that the upper 25–30 m may consist of poorly consolidated surface materials underlain by more cohesive bedrock. We infer that an ice-covered lake with a surface elevation > 2500 m probably existed in eastern Candor Chasma because this canyon is intersected by the Ophir Catena fault system from which Allegheny Vallis and Walla Walla Vallis originated. We introduce a new hydrology concept for Mars in which the groundwater system was augmented by recharge from canyon lakes that were formed when water was released by catastrophic melting of former ice sheets in Tharsis by effusions of flood basalts. This model could help to reconcile the expected presence of a thick cryosphere during the Hesperian with the abundant evidence for groundwater as a source for some of the circum-Chryse outflow channels.

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1. Introduction
The transition from the Noachian to the Hesperian era represents a significant shift in surface conditions on Mars. The crust formed during the Noachian and the planet was heavily bombarded by planetesimals. It is likely that the atmosphere was thickest and the water inventory highest during this time, because isotopic evidence shows that the planet has lost large fractions of water and other volatile inventories over geologic time (Jakosky and Phillips, 2001). Extensive valley networks developed in the cratered highlands (Carr and Chuang, 1997), and lakes possibly formed in many craters (Cabrol and Grin, 1999). Noachian erosion rates are estimated to have been ~1000 times greater than during later eras (Carr, 1996). By Hesperian time, which began ~3.5–3.7 Gyr ago (Hartmann and Neukum, 2001), the atmosphere had thinned and a thick, planet-wide cryosphere may have evolved (Clifford, 1993). Immense shield volcanoes began forming in Tharsis, and flood basalts inundated large
areas. Outflow channel activity peaked during the Hesperian (Tanaka, 1986) when enormous channels were carved and water bodies may have collected in the northern lowlands (Baker, 2001; Clifford and Parker, 2001). The largest channels emanated from the Valles Marineris canyons and adjacent regions and discharged their floodwaters to Chryse Planitia and the northern plains. In Xanthe Terra, the fluvial erosion that carved Ravi Vallis also triggered the formation of secondary chaos zones deep within the channel system. Analysis of those events provides evidence that the cryosphere was less than one km thick at the time of the Ravi Vallis flood at that location near the equator (Coleman, 2005).

This paper examines how and where water may have accumulated to provide regional sources for the floodwaters that formed three channels located south and west of Ganges Chasma (Fig. 1). Two of these channels, Allegheny Vallis and Walla Walla Vallis, have source areas at elevations greater than 2500 m, too high to be explained by flow in a global aquifer from distant recharge areas in the south polar region (Dinwiddie et al., 2004). We previously concluded (Coleman et al., 2003) that the outflow source for Allegheny Vallis in Ophir Planum would seemingly have required regional sources of recharge on and east of the Tharsis plateau. Here we examine these channel systems in more detail, along with the related Elaver Vallis. By inference, our results provide important clues for the sources of other Hesperian-aged floodwaters that emerged from the Valles Marineris.

2. Polar recharge model

Extensive groundwater recharge is inconsistent with the idea of a thick planet-wide cryosphere on Mars, which would inhibit the migration of water from the undergrounder aquifers. To resolve this dilemma, Clifford (1987) and Clifford and Parker (2001) suggested that Hesperian recharge occurred at the base of the polar layered deposits, and that this water migrated to low latitudes through a globally connected aquifer system. Carr (2002) tested this polar recharge model using the MOLA database of surface elevations obtained by Mars Global Surveyor. He identified 1500 m as an approximate upper limit for the elevation at which recharge could efficiently occur by melting beneath the south polar layered terrains. Carr (2002) concluded that major discharges of groundwater onto the surface at elevations above 1500 m are unlikely to have been derived from recharge sources in the south polar region. Examples of high-standing (> 1500 m) water-carved features include valleys on the flanks of Alba Patera and Ceraunius Tholus. These valleys could readily be explained by local processes, such as erosion by discharges from hydrothermal springs (Carr and Chuang, 1997). Harrison and Grimm (2004) used groundwater modeling results to suggest that recharge occurring on the Tharsis plateau would have been several times more efficient than south polar recharge as a source of groundwater for the circum-Chryse outflow channels.

Carr (2002) presented several alternative explanations that could reconcile high altitude water-carved features with the model of Clifford and Parker (2001). He suggested that volcanic melting of ice in the cryosphere could provide the water needed to erode the elevated channels. We agree that this process could melt large volumes of ice, but suggest that the interaction of flood basalt lavas with surface ice would far more efficiently release large volumes of water than the melting of ground ice by intrusive or extrusive volcanism. Carr (2002) suggested that the melting of ground ice in central Tharsis could have led to the movement of groundwater along radial fractures into the Valles Marineris to form canyon lakes. He also suggested another mechanism to account for elevated channels, in which the crust is ruptured by enhanced hydrostatic pressures created as a consequence of downward cryosphere expansion. This process would depend on the degree to which the crust was saturated by ice and water. Carr (2002) concluded that a coincidence between the ground inventory of water and ice and the holding capacity seemed unlikely.
For Allegheny Vallis and Walla Walla Vallis, we find the cryosphere expansion mechanism an unlikely way to rupture the crust and release floodwaters. Deep canyons of the Valles Marineris almost surround the source areas for Allegheny Vallis and Walla Walla Vallis. If these deep canyons existed when the channels formed, it is more likely that regionally enhanced hydrostatic pressures would have been released where overburden was minimal and fluid pressures were greatest (i.e., in low elevation areas of the proximal canyons). The chasmata present ideal conditions for the release of regional overpressures because the present-day canyon floors lie 4–5 km below the martian datum. Analysis by Coleman (2006) showed how the formation of lakes in the ancestral Valles Marineris would have been likely, perhaps inevitable, as a consequence of the rising potentiometric levels of groundwater systems.

3. Outflow channels near Ganges Chasma

We describe three specific channels that reveal new insights about Hesperian paleohydrology. These channels provide evidence that regional groundwater recharge occurred at low latitudes during one or more intervals of Hesperian time. In the following subsections we discuss the outflow nature of these channels and infer their relative age of formation.

3.1. Elaver Vallis

Elaver Vallis begins at the eastern rim of Morella, an 80-km-wide crater located south of Ganges Chasma (Fig. 2A). The flooding occurred during the mid- to late-Hesperian era because the channel eroded lower Hesperian strata (unit Hpl3) of the Plateau Sequence (Scott and Tanaka, 1986; Witbeck et al., 1991). The fact that Elaver Vallis begins at a gap in the wall of Morella Crater indicates that a lake formerly existed in the crater basin. The surface of this lake would likely have been frozen under the cold prevailing surface conditions during the Hesperian. The wall of Morella crater was eventually breached and lake waters catastrophically drained through the gap, carving the Elaver Vallis channel complex. A deep pit known as Ganges Cavus resides in the southern part of Morella Crater (Figs. 2A, 3A, and 3B). This feature is indeed a cavus and not a crater. It has no rim or ejecta blanket and is roughly ovate with a major axis length of 42 km and a minor axis length of 33 km. We interpret Ganges Cavus as a subsidence feature produced mainly by dilational faulting that was also partly undermined and excavated by the eruption of confined groundwater. The northern part of the cavus floor forms a 15-km-long linear depression that is aligned west to east. The deeper parts of Ganges Cavus have approximately the same elevation (~4000 m) as the deepest parts of western Ganges Chasma, located 100 km to the north. Groundwater is the only plausible source of the water that filled Morella Crater because Elaver Vallis exits Morella and no inflow channels exist. Thus, Morella Crater represents a Hesperian lake basin that was fed by groundwater, unlike numerous other paleolakes in impact structures that appear to have been dominantly fed by inflows of surface water (Cabrol and Grin, 1999). The existence of a Hesperian lake in Morella is

![Fig. 2.](image-url)
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Consistent with the finding of Cabrol and Grin (1999) that paleolake activity in martian impact craters began in the Noachian and continued through the Hesperian and into the Amazonian era.

It may not be a coincidence that Ganges Cavus formed within a sizable crater. The planetesimal impact that formed Morella Crater extensively fractured the country rock, weakening it and enhancing its permeability, creating favorable conditions for the eventual breakout of confined groundwater when the cryosphere was ruptured by faulting. The outflows from Ganges Cavus produced a lake in Morella that rose so high that the eastern crater wall was eventually breached, leading to catastrophic release of ponded waters and the carving of the Elaver Vallis channel complex. A small crossover channel preserved on the eastern rim of Morella Crater (Fig. 3C) reveals a high-water mark of 1780 m for the overflow elevation of the paleolake. Flow in this crossover channel ceased when the overflow gap to the north captured all the flow. The lake continued to drain via Elaver Vallis as long as the base of the downward-eroding gap in the crater wall was less than the stage elevation of the lake. Eventually, flow rates were insufficient to further deepen the gap and the lake level stabilized. THEMIS images indicate that the channel floor in the outflow gap (Fig. 3C) has a minimum width of ~2.3 km. Gridded MOLA data suggest that the valley floor in the center of the gap rises to an elevation of 1485 m. However, the presence of a precise curve in the MOLA topography (centered at 9.81° S, 50.63° W) reveals the high ground to be a false artifact caused by smoothing of elevations in an area without MOLA tracks. MOLA data exist within 4 km to the east and west, indicating a channel floor elevation of 1230 m on the western side of the gap and 1250 m on the eastern side. We interpret 1250 m as the final outflow elevation for the lake in Morella Crater. We subtract this elevation from the initial crossover elevation (1780 m) and thereby determine that a lake water column ~530 m deep was catastrophically drained by the flood. Most of the floor of Morella Crater lies at elevations <1250 m. A >50-m-deep lake would have remained in Morella after flows in Elaver Vallis ceased. The remaining...
water column eventually would have frozen to its total depth [e.g., Kreslavsky and Head (2002) calculated that a 500-m-deep ocean in the northern plains would have frozen within 2000–7500 years] before undergoing long-term sublimation. Morphological evidence suggests the lake depth in Morella Crater would have been much greater above Ganges Cavus. The cavus is currently ~3 km deeper than the crater floor, but we cannot be certain how deep it was when Elaver Vallis formed.

The floor of Morella Crater is relatively flat. Using gridded MOLA data, a 40-km-long profile from west to east across the center of the crater shows that elevation varies less than 90 m. We checked this result using the higher resolution of an individual MOLA track. MOLA pass ap10452L crossed eastern Morella Crater and indicates a north-to-south elevation change of ~100 m over a 45-km-long distance. This smooth floor morphology is consistent with lacustrine deposition or prior resurfacing by lava flows (or both). Using Eq. (1) (Garvin et al., 2000; Barlow and Perez, 2003),

\[ d = 0.19D^{0.55}, \]

where \( d \) = depth (km) of a fresh, non-polar crater of >7 km diameter (measured from the top of the rim to the lowest point within the crater), and \( D \) = present-day rim diameter of crater (km), we estimate a former depth of 2.1 km for Morella Crater. The present-day depth is ~90 m (mean rim height [~2050 m] minus mean floor elevation [~1175 m]). Therefore, we estimate the crater depth has been reduced 1.2 km as a result of wall collapse, infilling, and crustal relaxation.

Two craters of similar size occur on Morella’s floor north-east of Ganges Cavus (see Figs. 3A and 3B). Somerset Crater has a continuous rim, an east-to-west diameter of 4.8 km, a floor elevation of 720 m, and is >400 m deeper than the surrounding floor of Morella. East of Somerset Crater is Johnstown Crater, which has a similar diameter but a very different appearance (Fig. 3B). The rim of Johnstown (Fig. 3D) is highly degraded with numerous arcuate embayments where the walls were eroded. Several of the cone-like openings fully incise the crater rim to the floor elevation of Morella Crater. The overall appearance of Johnstown Crater is that of a highly eroded crater with scalloped edges. The lowest part of Johnstown Crater has an elevation of approximately 1090 m, less than 60 m deeper than the surrounding floor of Morella Crater. This suggests that fluvially transported and lacustrine sediments mostly filled Johnstown crater based on its greater depth and pristine appearance, Somerset Crater post-dates the lacustrine event in Morella Crater, whereas the shallow depth and degraded rim of Johnstown Crater provide evidence that it existed before the paleolake filled Morella Crater.

Various processes may have contributed to the erosion and infilling of Johnstown Crater. The initial outflows from Ganges Cavus likely had some erosive effects on the base of Johnstown. This early erosion could have ceased quickly, however, in the low-energy environment of lake formation and deepening. Eventually the lake level overtopped the rim of Johnstown Crater, resulting in rapid erosion of the crater walls at inflow points and deposition of eroded materials inside the crater. MOLA pass ap10452L crossed the eastern rim of Johnstown Crater, showing that the residual wall of Johnstown reaches an elevation of ~1210 m, which is below the final stage elevation of the paleolake (i.e., 1250 m). The formation, breakup, and movement of lake ice during the filling and draining of the paleolake could have further eroded any high-standing features within Morella Crater, including the walls of interior craters. After the paleolake had partially drained and frozen in place, the subsequent deterioration of surface ice and degradation of ground ice may also have further eroded the crater walls. Other workers have catalogued the characteristics of thousands of craters on Mars that are larger than Johnstown Crater, i.e., >5 km in diameter (Barlow, 1988, 2003; Barlow and Bradley, 1990). The morphology of Johnstown Crater is unusual among martian craters. Other craters may eventually be found that were eroded and modified by similar processes. We suggest that the scalloped rim of Johnstown Crater is a possible indicator of the erosive processes described above.

Hoffman (2000) proposed that martian outflow channels were eroded by gas-supported debris flows rather than by water floods. However, the morphology of Elaver Vallis demonstrates why gas-supported debris could not have formed this channel complex. If liquid CO₂ had erupted from Ganges Cavus, the resulting debris clouds would have rapidly depressurized, dropping their suspended loads within the rim of Morella Crater. Only a relatively stable and non-compressible fluid, like liquid water, could have accumulated to form a lake within the crater, which eventually breached the eastern crater wall and carved the channel complex. The erosive action of a liquid was required. Carbon dioxide-based models (Hoffman, 2000; O’Hanlon, 2001) face additional theoretical difficulties that now appear insurmountable (Stewart and Nimmo, 2002; Coleman, 2003; Urquhart and Gulick, 2003).

### 3.2. Allegheny Vallis

Allegheny Vallis, an outflow channel west of Elaver Vallis, emerges from an elongated pit known as Ophir Cavus (Figs. 2A, 2B, and 4). The main channel was nearly imperceptible in Viking imagery, but MOLA data and MOC images confirm that it is a continuous valley more than 190 km long. The plain incised by Allegheny Vallis is Noachian in age (units Npl₁ and Np₂ of Witbeck et al., 1991). We interpret that its flows may have occurred during the same epoch as the flows in Elaver Vallis (mid- to late-Hesperian) because both sources originated from the fault system of Ophir Catena. Erosional features at the northwest rim of Ophir Cavus show that the water that carved Allegheny Vallis erupted from the surface at an elevation of 2500 m (see Fig. 4), far above the 1500 m threshold described by Carr (2002) for contributions from polar basal recharge. From the source at Ophir Cavus to the terminus at the rim of Ganges Chasma the channel floor gradually drops in elevation by 435 m (Fig. 4, inset at lower right). Over this 190 km distance, the average slope or gradient of the channel floor is 0.0023. It is possible that elevations at the cavus source and along the channel have changed over time. However, extensional tectonics in the Valles Marineris seems more likely to have lowered the channel than to have raised it. Therefore,
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Fig. 4. Allegheny Vallis imagery and MOLA elevation, from its source at Ophir Cavus to the terminus at Ganges Chasma. Water overflowed the northwest rim of Ophir Cavus, carving erosional features at an elevation of \( \sim 2500 \text{ m} \). Two ridges [R1 and R2] temporarily dammed flow. Two scablands [S1 and S2] were created along the channel. Cultor Crater (inset at upper left) is centered at 8.85° S, 53.93° W. [Image credit: THEMIS Daytime IR Valles Marineris Mosaic (Christensen et al., http://themis-data.asu.edu/).]

if the channel source elevation has not changed appreciably, has been lowered since Hesperian time, or if relative elevation differences have been preserved, then the outflows must have derived from regional recharge at higher elevations.

Immediately north of Ophir Cavus, Allegheny Vallis crosses a large, subdued, unnamed crater (Fig. 5) that is probably Noachian in age. The outflows may have formed a temporary lake within the crater walls, but it is unclear whether the degraded rim of this crater was high enough at the time of the flooding to obstruct flow. Continuing outflows from Ophir Cavus eroded a deep channel across the crater floor (Fig. 5). Multiple channels and a flood-carved island occur northeast of the crater (Fig. 5). The main channel north of the island is \( \sim 100 \text{ m} \) deeper than the channel south of the island, which is “hanging” at both ends. The top of the island is fringed by a terrace that shows the floodwaters reached a minimum elevation of 2240 m at this location.

In its western reaches, Allegheny Vallis meanders in a series of sweeping, nearly right-angle turns (Fig. 4), which suggests that topographic lows in the pre-flood terrain may have been influenced by structural features in the upper crust. Northeast of Ophir Cavus the flows encountered the first of two ridges that temporarily dammed the flow (Fig. 4). The floodwaters eventually broke through or overtopped this barrier, eroded a mid-channel island at 9.08° S, 54.03° W, and continued flowing to the northeast where they encountered another topographic obstruction. A temporary lake was created behind this natural dam, with bypass flow occurring around the obstruction until it gave way (Fig. 6A). Based on the overflow elevation (2032 m) of a bypass channel at 8.745° S, 53.602° W, we used gridded MOLA data to estimate that the lake had a minimum surface area of 35 km². The lake drained catastrophically and floodwaters charged northeastward, carving a spectacular scabland with anastomosing channels and streamlined erosional features (Fig. 7). MOLA pass 10644 reveals 25–30 m of topographic relief in this broad scabland. Flow and erosion in the scabland ceased as an inner main channel deepened and eventually captured most or all of the waning floodwaters. This concentration of flow in one channel carved a narrow, well-defined reach that can be traced eastward to the western margin of Ganges Chasma. This main channel cross-cuts a wide, shallow channel that had exited the scabland to the east, producing a hanging valley (Figs. 4 and 6). The main channel formed a broad, branching, channeled scabland near its terminus that is truncated by the western rim of Ganges Chasma (Fig. 4). The abrupt termination of this scabland provides evidence that Ganges Chasma continued to expand westward after the fluvial episode had ended. The southernmost channel at the terminus of Allegheny Vallis disappears into a deep side canyon with a V-shaped cross-section that grew westward from the southwest corner of Ganges Chasma. The channel appears to descend 4 km through this side canyon to the floor of Ganges Chasma, but the elevation change is so abrupt that the side canyon more likely formed after the fluvial episode by faulting and mass wasting processes (Fig. 4, inset at lower right).

In addition to the geomorphologic evidence that groundwater carved the channel features, the ejecta pattern of two local craters provides evidence for a crust formerly rich in volatiles. Water ice was the likely volatile, given the difficulties of producing and storing CO₂ ice in the crust (Stewart and Nimmo, 2002). South of Ophir Cavus at 10.70° S, 55.26° W is Bronkhorst, a crater with a single-layer ejecta morphology and
Fig. 5. (A) Ophir Cavus is the source of Allegheny Vallis. The cavus formed along the southern rim of a large, eroded, unnamed crater. Bronkhorst Crater, centered at 10.7° S, 55.3° W, displays a layered (fluidized) ejecta blanket and central pit that suggest the presence of shallow volatiles in the crust at the time of impact. [Image credit: THEMIS Daytime IR Valles Marineris Mosaic (Christensen et al., http://themis-data.asu.edu/).] (B) Where Allegheny Vallis turns eastward, floodwater eroded two channels and carved a central island (Is). The layer that caps the island has an irregular periphery, suggesting it is softer than the underlying layer. Differential erosion of these layers formed a terrace all around the island. Additional terraces are identified on the northern bank of the deeper (main) channel. Two terraces are continuous over more than 25 km.

a central pit (see Fig. 5A). Craters with this layered ejecta pattern are thought to have formed by the vaporization of subsurface volatiles during crater formation (Barlow and Perez, 2003). Barlow and Bradley (1990) proposed that craters with single-layer ejecta excavated a zone consisting primarily of ice. The presence of a central pit in Bronkhorst Crater is also an indicator that volatiles were present in the upper crust at the time of impact (Barlow and Hillman, 2006). Another crater with layered ejecta, Culter Crater, is superimposed on Allegheny Vallis more than halfway along its course (Fig. 4) at 8.85° S, 53.93° W.

We estimate the transient excavation depths of these craters, and thereby estimate the depth of ice reservoirs in the upper crust at the time of the impacts. Using Eq. (2) (Garvin et al., 2000; Barlow and Perez, 2003),

$$d_e = 0.131D_a^{0.85},$$  (2)

where $d_e$ = transient crater excavation depth (km), and $D_a$ = present-day rim diameter of crater (km), we estimate that the 18-km wide Bronkhorst Crater and the 6-km wide Culter Crater had transient excavation depths of 1.5 and 0.6 km, respectively (note that the diameter of Culter Crater is near the lower limit for application of Eq. (2)). The transient excavation depths suggest that water ice reservoirs were located within $\sim$1500 m (or less) of the surface at the time the craters formed. No channels originated at these craters. From this, we infer the depths of excavation were apparently insufficient to rupture a low-permeability confining layer and generate large-scale flow of water from a deep aquifer, or water flowing into the craters from below did not overtop the rim, or conditions for such flow did not exist at that time. Culter Crater is relatively fresh; it overlies Allegheny Vallis and therefore post-dates the flooding that formed this channel. We do not know whether the Bronkhorst impact occurred before or after Allegheny Vallis was formed. The terrain around this crater has been mapped as ridged plains material of lower Hesperian age (unit Hr of Witbeck et al., 1991). The degraded rim of Bronkhorst and the presence of sizable craters on its floor and ejecta blanket suggest the crater is Hesperian rather than Amazonian in age.

We interpret the following sequence of events in the formation of Allegheny Vallis: (1) gradual development of a groundwater potentiometric surface in Ophir Planum to an elevation >2500 m; (2) tectonic rupture of the cryosphere at the location of eastern Ophir Catenae, forming Ophir Cavus and dramatically releasing confined groundwater onto the surface as floodwater; (3) possible ponding of discharge just north of Ophir Cavus within the rim of a large, subdued crater; (4) floodwater incision of multiple channels at 9.43° S, 54.63° W; (5) ponding of water behind a ridge obstruction at 9.08° S, 54.20° W;
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Fig. 6. Location of channeled scabland that formed when a transient lake breached the northernmost topographic barrier. (A) MOLA topography contoured using GRIDVIEW (GSFC, 2004). Dotted blue line shows the minimum area of a transient lake that formed behind a low ridge. Water release carved a water gap and the scabland to the north. Arrows delineate a small channel where ponded water apparently bypassed the topographic barrier. (B) Flow calculations in Section 5 were performed along the straight reach of channel “A” shown in the dashed box. [Image credit: THEMIS Daytime IR Valles Marineris Mosaic (Christensen et al., http://themis-data.asu.edu/).] (C) Topography along profile A–A’ (located as shown in panel A), generated with GRIDVIEW. Profile A–A’ coincides with MOLA pass 14761. (D) Estimated pre-flood topography measured at 15 locations along the south bank of Allegheny Vallis between longitude 53.493° W and 53.126° W (located as shown in dashed box in panel B). We interpret locations where the slope of the bank changes abruptly as marking the lateral and vertical limit of fluvial erosion.

Fig. 7. Ridge R2 dammed flow, causing a temporary lake to form in the lower left corner of the image. Bypass flow eroded a shallow channel at lower right, but this flow ended when the main obstruction to the north was surpassed. Floodwaters rushed to the northeast, carving a water gap, the scabland, and the main channel. This channeled scabland formed before incision of the main channel, which ultimately captured the remaining discharge. An erosional scarp marks the western margin of the scabland. MOLA elevations along profile B–B’ (MOLA pass 10644) provide additional evidence for event sequence. Crest to trough relief in the scabland ranges 25–30 m. The topography suggests that the upper 25–30 m of material was more easily eroded, with an underlying layer of more resistant material (perhaps intact basalt). [Image credit: THEMIS visible image V06781022 (Christensen et al., http://themis-data.asu.edu/).]
Fig. 8. Walla Walla Vallis (A) Context image. (B) View of Walla Walla Vallis source area and passage through Wallula Crater. (C) A sketch of the river valley is shown offset to the right as a visual cue. The outflow channel emerges from a pit chain south of Wallula Crater; flow entered Wallula Crater, and eroded a channel to the north–northwest. Walla Walla Vallis, which appears to intersect Allegheny Vallis, likely pre-dated it, although concurrent flows are plausible. [Image credit A & C: THEMIS Daytime IR Valles Marineris Mosaic; B: THEMIS V10176001 (Christensen et al., http://themis-data.asu.edu/).] (D) MOLA topography crossing pit chain, Wallula Crater, and Allegheny Vallis.

3.3. Walla Walla Vallis

A pit chain (e.g., Wyrick et al., 2004) with margin elevation 2525 m is located east of Ophir Cavus. These collapse pits are the surface expression of the source area for the outflow channel Walla Walla Vallis (Fig. 8), which was not resolved in imagery until return of THEMIS data began in late 2002 (Dinwiddie et al., 2004). Walla Walla Vallis floodwaters followed the natural topographic relief down into nearby Wallula Crater, located 2 km to the north at a base elevation of approximately 2400 m (Fig. 8). Wallula Crater is a subdued 12.2-km-diameter feature located near 9.88° S, 54.66° W. The map by Witbeck et al. (1991) indicates the terrain upon which Walla Walla Vallis and Wallula Crater are superimposed consists of the cratered unit of the Plateau sequence, unit Npl1. The north–northwest-trending erosional channel across the center of Wallula Crater is visible in Fig. 8B. Ponded water may have accumulated in Wallula Crater until the water breached or overtopped the northwestern rim of the crater through a weak point or topographic low.
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Fig. 9. Parallel pit chains of Ophir Catenae extend ∼750 km from eastern Candor Chasma to near Ganges Chasma. Allegheny, Walla Walla, and Elaver Valles are shown in relation to this interpreted dilational fault system. Average land surface elevations decrease from the rim of Candor Chasma toward Ganges Chasma. This regional slope would have favored eastward migration of groundwater along structurally enhanced flowpaths. Viking image draped over MOLA megdr data.

Fig. 10. Schematic illustration of steep dilated faults and their potential relationship to pit chain development (after Fig. 2 of Ferrill et al., 2003). Compare pits in illustration to pits (cavus) and pit chains (cavi) shown in Figs. 2A, 2B, 3A, 3B, 5A, 8B, and 9.

Walla Walla Vallis split into two channels within a small semi-circular depression at the northwestern edge of Wallula Crater (Fig. 8C); one channel flowed north–northwestward, appearing to terminate as a hanging valley, while the second channel flowed west before turning to the north (Fig. 8C). THEMIS images suggest the second channel may end as a hanging valley truncated by Allegheny Vallis. Flows in Walla Walla Vallis thus likely predate late-stage flows in Allegheny Vallis, although concurrent flows during the early stage formation of Allegheny may be plausible, and Walla Walla floodwaters may have contributed to the total discharge through Allegheny Vallis. Higher resolution imagery is needed to confirm this possibility.

4. Structurally enhanced groundwater flow

The floodwaters that formed Elaver Vallis, Allegheny Vallis, and Walla Walla Vallis probably discharged during the Hesperian era from beneath a thick cryosphere and through a deep-seated dilational fault system. The Ophir Catenae structural complex is a long chain of pits that extends 750 km from eastern Candor Chasma to western Ganges Chasma (Fig. 9). We interpret the en echelon segments of this pit chain as the surface manifestation of a dilational fault zone (Wyrick et al., 2004; Ferrill et al., 2004), i.e., an incipient graben system that was part of the Valles Marineris structural complex. Collapse pits likely formed at the source during or after the fluvial episode as surface material dropped into subsurface cavities, was lowered by graben formation, or was carried away with the flow of water. Fig. 10 (after Ferrill et al., 2003) schematically illustrates steep dilated faults on Mars and their potential relationship to pit chain development. This kind of fault system would likely have created highly anisotropic conditions in the aquifer system, greatly enhancing the eastward flow of groundwater. We propose that groundwater breakouts coincided with the climactic formation of Ophir Catenae faults. Discharges from Ganges Cavus may indicate the easternmost extent of Ophir Catenae. At least two and possibly three channels were carved by aqueous outflows from the same dilational fault system, which suggests that the flows may have been concurrent. More discharge could have occurred in Elaver Vallis over a longer duration because its outflow zone is ∼1300 m lower in elevation than the sources of Allegheny and Walla Walla Valles.

Scott and Wilson (2002) investigated the origin of pit chains near the martian volcano Alba Patera. They propose these pit chains formed as a result of volcanic activity above long regional dikes. They conclude that smaller pits formed when sur-
face rocks collapsed into spaces vacated by the leakage of gas that accumulated at the tops of dikes. They suggest that larger pits are produced after explosive Plinian eruptions ejected substantial fractions of magma from dike cores (i.e., the eruption rates exceeded the supply rate of magma feeding the eruption). Scott and Wilson (2002) suggest that the implosion and collapse of dike walls produced the larger pits. They explain that there are no visible products of the eruptions because areas near the vents are destroyed by the terrain collapse that formed the pits. Also, the thin martian atmosphere increases the vigor of the explosive eruptions, scattering debris over large areas rather than concentrating it near the vents. Wyrick et al. (2004), however, find no direct evidence that pits were formed by intrusive volcanism. We agree with their conclusion that dilational faulting is more consistent with Earth analogs and provides a simpler explanation than the volcanism model of Scott and Wilson (2002).

We can now add Allegheny, Walla Walla, and Elaver Valles to the group of martian channels carved by groundwater outflows from large fault zones. Such channels also include Athabasca Valis, which emanated from the Cerberus Fossae (Burr et al., 2002a, 2002b; Murray et al., 2005), and Mangala Valles, which discharged from the Memnonia Fossae west of Tharsis (Tanaka and Chapman, 1990; Wilson and Head, 2004). The observation that some large martian channels debouched from fault zones supports a groundwater origin, and contradicts other suggested mechanisms for outflow channel formation, including carving by wind (Whitney, 1979), erosion by gas-supported debris flows (Hoffman, 2000), or glacial scour (Lucchitta, 1982). Although Lucchitta (1982) proposed glacial scour as one mechanism to erode channels, she also suggested another more likely scenario in which ice jams enhanced the erosive power of catastrophic floods. This latter mechanism may explain why some channel landforms appear to have been sculpted by ice. The absence of distinct terminal moraines near the mouths of the large outflow channels further supports a fluvial rather than a glacial origin. The extreme smoothness of the northern plains of Mars (Aharonson et al., 1998; Head et al., 1998) can readily be explained by resurfacing of these lowlands by aqueous transport and sedimentation.

5. Discharge estimates for Allegheny Valis

Studies of catastrophic paleofloods on Earth and Mars have tended to be controversial. Over the last 40 years, however, evidence has grown that many megafloods were associated with continental ice sheets of Pleistocene age (Baker, 2002). The resulting massive discharges of fresh water are thought to have significantly influenced ocean circulation and climate. The study of martian megafloods is central to resolving questions about the global evolution and fate of water on Mars. It is useful to estimate the magnitude of flows that carved Allegheny Valis to compare with other channel studies on Mars and prehistoric megafloods on Earth. We focus on Allegheny because it is an elevated channel (>1500 m) indicative of recharge at low latitudes. Walla Walla Valis is also an elevated channel, but available data are not sufficient to define cross-sections for this small channel. While MOLA and image data are available to perform detailed analyses of Elaver Vallis paleo-flows, such analyses are beyond the scope of this study.

Hydrologic evaluation of the paleoflood in Allegheny Valis is challenging because the channel is highly sinuous and, as discussed earlier, there were at least three obstructions that could have caused temporary ponding and catastrophic release of floodwaters, resulting in erosion of several scablands. Such flows would have been highly non-uniform. We therefore chose to evaluate flows in the long straight reach of Allegheny Valis that occurs beyond the northernmost topographic obstruction (Figs. 6 and 7). The temporary lake that had accumulated behind this barrier eventually broke through and eroded the scabland shown in Fig. 7. Once the lake had drained, the southernmost channel (henceforth channel “A”) in the scabland captured all remaining flow in the channel system and continued eroding to its present depth until the discharge from Ophir Cavus ceased. We surmise that the discharge rate in channel “A” would have become quasi-uniform after the transient lake had drained and before the channel reached its present depth.

Dingman (1984) makes a fundamental point about open-channel flows: All the flow variables (i.e., slope, width, depth, velocity, and flow resistance) are interdependent and cannot be considered as truly independent variables. Four elements are needed to estimate the physical parameters for the Allegheny Vallis paleofloods: (1) energy slope; (2) width of eroded terrain; (3) hydraulic radius of the floodwater channel; and (4) an estimate of the flow resistance for the channel bed (i.e., the Chézy or Manning coefficients). The hydraulic radius refers to the cross-sectional area of flow divided by the wetted perimeter. The wetted perimeter is the distance across a channel measured along the channel bottom from the water’s edge on one side to the other (Dingman, 1984). For flows that are much wider than deep, the hydraulic radius can be estimated using the average flow depth. This assumption appears to be reasonable for the intermediate-stage flows we have analyzed here, prior to full incision of the channel.

The channel is ∼3800 m wide and ∼110 m deep at the location of MOLA pass 14761, with a floor elevation of 1794 m (MOLA 8.499° S, 53.256° W) (Fig. 6C). The width and estimated depth of flooding were used to calculate the cross-sectional area of the flood channel. Wilson et al. (2004) describe a 30 m floodwater depth as a realistic lower limit for channels on Mars, based on expected rates of evaporation and freezing that would prevent shallower floodwaters from traveling long distances. We apply this lower limit to Allegheny Vallis. Because we evaluated flow conditions for the period after channel “A” captured all the flow, we estimate the elevation of the flood stage using the base level elevation in the channeled scabland (i.e., 1895 m) located north of and parallel to channel “A.” At this intermediate stage of flooding, the scabland would have drained, but the main channel would not yet have eroded to its full depth. Therefore, we estimated flow velocities and discharge rates using a 30–60 m range of floodwater depths. A 60 m depth would represent a channel floor elevation of 1835 m, which is ∼40 m less than the present-day channel depth shown in Fig. 6C. We used Manning coefficients in the
range of 0.04 to 0.06 [see discussion of resistance factors by Wilson et al. (2004) and Coleman (2005)]. The pre-flood terrain slope was determined along a 23-km-length of the south bank of Allegheny Vallis using slope inflections that mark the lateral and vertical extent of erosion. The elevation change over 23 km is −100 m (see Fig. 6D) from which we obtain an energy slope of 0.0043. We used this value (rather than the present-day thalweg slope) to estimate the energy slope for intermediate-stage flooding prior to full channel incision. Results are presented in Table 1.

For intermediate stage flooding in Allegheny Vallis, our results indicate flow velocities of 6–15 m s$^{-1}$ and discharges in the range of $0.7–3 \times 10^6$ m$^3$ s$^{-1}$. We estimate that bottom shear stress was in the range of 400–1000 N m$^{-2}$, and that flow power was in the range of 3000–15,000 W m$^{-2}$.

### 5.1. Comparison of terrestrial and martian floods

Table 2 provides estimates of volumetric discharge published for various paleofloods on Mars and Earth. Also shown for comparison purposes are discharge estimates for several large Earth rivers. The discharges we estimate for Allegheny Vallis are similar in magnitude to those estimated for Athabasca Vallis (Burr, 2003), which also discharged from a fault system (i.e., Cerberus Fossae). The Allegheny discharges were similar in magnitude to the Missoula paleofloods on Earth. Non-uniform flows of much greater discharge rate likely occurred in Allegheny Vallis when obstacles to flow were breached by ponded floodwaters. The pre-MOLA extreme discharges calculated by Robinson and Tanaka (1990) for Northern Kasei Vallis need to be reconsidered based on the work of Williams et al. (2000). A pre-MOLA analysis by Komatsu and Baker (1997) reports a large peak discharge for Ares Vallis flooding. They presume deep, rim-high flows in a narrow reach of the channel, but do state that the discharge could well have been much lower than the estimated peak. MOLA pass 21 crossed their study area and verifies a channel depth of $\sim 1600$ m (Smith et al., 1998). This is significantly deeper than estimated by Komatsu and Baker (1997). We agree with Komatsu and Baker (1997) that the peak discharges in Ares Vallis were likely much lower than reported in Table 2. By the time the Ares Vallis channel was fully incised, the floodwater sources were likely to have been greatly depleted, resulting in flood stages well below the valley rims.

### 5.2. Theoretical floodwater hydrographs

Channels that were carved by groundwater erupting from large fault systems (i.e., Allegheny Vallis, Athabasca Vallis, Walla Walla Vallis, Mangala Valles) likely produced hydrographs with a signature that includes a rapid rise followed by a prolonged decline. Confined groundwater was the apparent source for the initial outflows that carved Allegheny and Walla Walla Valles. If this confined groundwater were the only source, the flow could not have been sustained because confined aquifers, once released, tend to depressurize rapidly. The de-watering of an unconfined aquifer is much more protracted. The hydrograph for flow released from a confined aquifer should display initial rapid release of water, resulting in broad scabland erosion. Subsequent drainage under unconfined conditions, augmented by distant recharge from an ice-covered lake in ancestral Candor Chasma, would produce a long recession of flood stage. These protracted flows of reduced discharge likely concentrated erosion, producing an inner channel in the fluvial system. In the case of Allegheny Vallis, channel “A” represents an inner channel that formed from slowly declining flood stages after all flow obstructions had given way. This type of hydrograph differs from those described by Baker (1978) for the terrestrial Missoula floods. He concluded that the extraordinary preservation of high-stage indicators throughout the Channeled Scabland suggests those floodwaters rose slowly, peaked, and then dropped abruptly, similar to the hydrographs commonly (but not always) seen for Icelandic jökullhlaups (Roberts, 2005). We expect that a similar hydrograph form may have represented martian channels carved by flows originating from catastrophic releases from surface impoundments located in the deep canyons. The Simud–Tiu Valles system may have been formed by such releases.

### 6. Hesperian recharge zones for high channels

We have identified the region of potential recharge for high channels like Allegheny Vallis and Walla Walla Vallis, and for most of the large outflow channels that emptied into Chryse Planitia. Large regions near the Valles Marineris stand higher than 2500 m (Fig. 11), including the elevated terrain of the Tharsis plateau, Olympus Mons, and Alba Patera. Other high-standing terrain includes the Hesperian plains of Syria Planum, Sinai Planum, and Solis Planum. Based on the geologic map-

<table>
<thead>
<tr>
<th>Hydrologic scenario</th>
<th>Flow width (m)</th>
<th>Slope</th>
<th>Mean flow depth (m)</th>
<th>Mean flow velocitya (m s$^{-1}$)</th>
<th>Discharge × 10$^6$ (m$^3$ s$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Flow in main channel “A”</td>
<td>3800$^b$</td>
<td>0.0043</td>
<td>30</td>
<td>6–10</td>
<td>0.7–1</td>
</tr>
<tr>
<td>after lake and scabland had drained (see Fig. 6)</td>
<td>45</td>
<td>8–13</td>
<td>1–2</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

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$^a$ Flow velocities calculated using the approach of Komar (1979) and a range of terrestrial Manning $n$ values from 0.04 to 0.06. Values of the Chézy friction coefficient ($C_f$) were obtained using these $n$ values and the estimated range of flow depths (Carr, 1996). For example, given a flow depth of 30 m and $n = 0.05$, $C_f = 0.008$.

$^b$ Determined where MOLA pass 14761 crosses Allegheny Vallis (profile A–A′ in Table 2.

### Table 1

<table>
<thead>
<tr>
<th>Hydrologic scenario</th>
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Table 2
Estimated volumetric discharges for martian and terrestrial paleofloods, and for several large present-day rivers

<table>
<thead>
<tr>
<th>Location</th>
<th>Discharge ($m^3 s^{-1} \times 10^6$)</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Martian floods</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mangala Vallis</td>
<td>0.01–40</td>
<td>Komar (1979)</td>
</tr>
<tr>
<td>Lunae Plenum scabland</td>
<td>0.1–10</td>
<td>Carr (1979)</td>
</tr>
<tr>
<td>Maja Vallis</td>
<td>~1</td>
<td>Chapman et al. (2003)</td>
</tr>
<tr>
<td>Ares Vallis</td>
<td>100–1000</td>
<td>Komatsu and Baker (1997)</td>
</tr>
<tr>
<td>Northern Kasei Vallis</td>
<td>900–1400</td>
<td>Robinson and Tanaka (1990)^a</td>
</tr>
<tr>
<td>Kasei Valles</td>
<td>3–20</td>
<td>Williams et al. (2000)^b</td>
</tr>
<tr>
<td>Vedra Vallis</td>
<td>1.3</td>
<td>De Hon et al. (2003)</td>
</tr>
<tr>
<td>Maumee Vallis</td>
<td>1.9</td>
<td>De Hon et al. (2003)</td>
</tr>
<tr>
<td>Cerberus Fossae (source of Athabasca Vallis)</td>
<td>1–2</td>
<td>Manga (2004)</td>
</tr>
<tr>
<td>Ma'adim Vallis inner channel</td>
<td>~1</td>
<td>Irwin et al. (2004)</td>
</tr>
<tr>
<td>Ravi Vallis Early overland flow</td>
<td>4–50</td>
<td>Leask et al. (2006)</td>
</tr>
<tr>
<td>Ravi Vallis Late stage flow (in an inner channel)</td>
<td>&lt;1</td>
<td></td>
</tr>
<tr>
<td>Allegheny Vallis Intermediate stage flow in main channel, &quot;A&quot;</td>
<td>0.7–3</td>
<td>This study</td>
</tr>
<tr>
<td><strong>Terrestrial paleofloods and present-day rivers</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Late Pleistocene Missoula</td>
<td></td>
<td></td>
</tr>
<tr>
<td>&gt;25 floods</td>
<td>&gt;1</td>
<td>Benito and O’Connor (2003)</td>
</tr>
<tr>
<td>&gt;15 floods</td>
<td>&gt;3</td>
<td></td>
</tr>
<tr>
<td>&gt;6 floods</td>
<td>&gt;6.5</td>
<td></td>
</tr>
<tr>
<td>1 flood</td>
<td>~10</td>
<td></td>
</tr>
<tr>
<td>Grand Coulee (Missoula)</td>
<td>5</td>
<td>O’Connor and Baker (1992)</td>
</tr>
<tr>
<td>Amazon River</td>
<td>0.3</td>
<td>Baker and Kale (1998)</td>
</tr>
<tr>
<td>Mississippi River</td>
<td>0.02–0.03</td>
<td>Komar (1979)</td>
</tr>
</tbody>
</table>

^a To reproduce the calculations of Robinson and Tanaka (1990), an energy slope of $1.3 \times 10^{-3}(2.5/1890)$ must be used (M. Robinson, personal communication, 1999).

^b Williams et al. (2000) obtained values two orders of magnitude lower than previous estimates for Northern Kasei Valles. They suggest that the high-water mark used by Robinson and Tanaka (1990) lies outside the banks of Northern Kasei Vallis.

...the occurrence of widespread Hesperian volcanism over most of the potential recharge areas for Allegheny, Walla Walla, and Elaver Valles suggests direct, causal relationships between long-lived volcanism, large-scale genesis of liquid water, and channel formation. Overall, our understanding of Hesperian regional hydrology would improve if we could identify plausible sources for the water that filled the canyon lakes. The key to this puzzle may be the accumulation of ice at low latitudes due to increased obliquity and the dramatic release of water caused by regional volcanism, particularly on the Tharsis plateau.

6.1. Water ice reservoirs and release mechanisms

The present martian climate is dominated by low obliquity. Under these conditions, water ice is unstable at low latitudes. Neutron data from Mars Odyssey demonstrate that rich deposits of ground ice are present poleward of $60^\circ$ latitude (Feldman et al., 2003). The areas with enriched hydrogen generally coincide with regions where ground ice was predicted to be stable under the present climate (Mellon and Jakosky, 1995). Jakosky and Carr (1985) explored how ice might accumulate at low latitudes during times of very high obliquity during the Noachian and later periods. They concluded this accumulation would have been efficient only at extreme obliquities that ensued before the Tharsis plateau formed, but may also have occurred during the highest obliquities over the rest of martian history.

Unlike the Earth, which has an obliquity stabilized by a large Moon (Laskar et al., 1993), Mars has experienced chaotic obliqu-
High channels indicate Hesperian recharge and Canyon Lakes

Fig. 11. Topography for the region south and west of Chryse Planitia. All terrain within the 2500 m contour line stands higher than the outflow locations of Allegheny and Walla Walla Valles. The terrain east and south of Tharsis Montes that exceeds 2500 m covers an area greater than $1 \times 10^7$ km$^2$. Map contoured using MOLA data grid at $\frac{1}{4}$ degree resolution.

Obliquity that varies widely. Laskar and Robutel (1993) performed a frequency analysis of the obliquity of Mars over tens of millions of years. They found the martian obliquity to be chaotic with possible variations from nearly 0° to $\sim 60^\circ$. Obliquity in the present epoch is estimated to oscillate between 13° and 42° (Ward, 1992; Carr, 1996). More recent results show that within the last 10 Myr the martian obliquity could have exceeded $\sim 45^\circ$ at times (Laskar et al., 2004). Beyond 100 Myr the evolution of the martian obliquity cannot be evaluated with precision. Nonetheless, Laskar et al. (2004) evaluated the density function of the obliquity over the age of the Solar System. Their statistical results suggest an average obliquity of $\sim 38^\circ$ over 5 Gyr (standard deviation of $\sim 14^\circ$) and a maximal value of $\sim 82^\circ$. Laskar et al. (2004) also note that the present obliquity of $\sim 25^\circ$ is far from the most probable value of $\sim 42^\circ$ evaluated over 4 Gyr. Obliquity variations may have more dramatic effects on the martian climate than was previously believed (Jakosky et al., 2005). Mischna et al. (2003) used a general circulation model for Mars to investigate the orbital forcing of martian water and CO$_2$ cycles. Their three-dimensional model of the water cycle significantly improves upon previous models (e.g., Jakosky et al., 1993, 1995). Mischna et al. (2003) found that, as obliquity increases, abundances of water vapor and water ice clouds increase substantially and the region of surface ice stability moves toward the equator. At $45^\circ$ obliquity, water ice is stable only in the tropics and would preferentially collect in regions of high topography or high thermal inertia. The highest elevations on Mars occur on the volcanic summits of Tharsis. Surfaces on Tharsis have very low thermal inertias, less than $\sim 180$ J m$^{-2}$ s$^{-1/2}$ K$^{-1}$ (Jakosky et al., 2000). However, once surface ice begins to form, the albedo would increase dramatically and positive feedback would induce the further condensation of atmospheric water onto the surface as ice.

Important clues about the possible range of Hesperian climatic conditions may be found in the mounting evidence for significant variability in Amazonian climates, including recent geologic time (Baker, 2005). Neukum et al. (2004) reported evidence that the Tharsis region was volcanically active over hundreds of millions to billions of years. The most recent summit caldera activity on the Tharsis volcanoes was clustered approximately 100-200 Myr ago. Neukum et al. (2004) describe glacial deposits at the base of Olympus Mons as evidence for repeated phases of ice deposition. Some of the ice deposits may be as young as 4 Myr. They present morphological evidence that snow and ice were deposited on Olympus Mons at elevations $> 7000$ m, leading to episodes of glacial activity. Neukum et al. (2004) suggest that buried water ice may be present today at high altitudes on the edge of the western scarp of Olympus Mons.

Head et al. (2005) reported evidence for geologically recent and recurring glacial activity at low and middle latitudes on Mars. They presented evidence for glaciers veneered with debris at the northwest edge of Olympus Mons, extending up to 70–120 km from the basal scarp of this volcano. These rock glaciers are interpreted to be just a few million years old, but currently inactive. Head and Marchant (2003) interpreted fan-shaped landforms that extend over 350 km from the base of Arsia Mons as the deposits and remnants of cold-based mountain glaciers. Shean et al. (2005) concluded that the northwestern flank of each of the Tharsis Montes volcanoes has a fan-shaped deposit of Amazonian age. These deposits were interpreted as remains of tropical mountain glaciers created by atmospheric deposition of water ice during periods of high obliquity. High-resolution climate simulations of Mars with an obliquity of 45° and present-day atmosphere predict ice accumulation by precipitation in regions where glacial remains have been found (Forget et al., 2006).
Head et al. (2003) concluded that Mars is now experiencing an interglacial epoch that has lasted $\sim 300$ kyr, during which obliquity has remained in the narrow range of $22^\circ$–$26^\circ$. They interpret dusty, ice-rich surface deposits at middle to high latitudes as forming during a geologically recent ice age from 2.1 to 0.4 Myr ago, when obliquity regularly exceeded $30^\circ$. Such insolation conditions likely favored transfer of water from the poles to mid-latitudes, where it was deposited as ice-rich mantles along with dust eroded from the polar layered terrains. Head et al. (2003) suggest that martian interglacial periods are much shorter than glacials, preventing the erosion of surface lag deposits and preserving much of the mantle material.

During the Hesperian when most of the large outflow channels formed, we suggest that large volumes of liquid water were periodically produced at or near the martian surface at low latitudes. A likely source for liquid water was volcanic–ice interaction (Carr, 1996). We previously suggested that atmospheric conditions in mid- to late-Hesperian, proximal to major volcanic eruptions, may have permitted precipitation of snow onto the elevated terrain of the Tharsis plateau and other high terrain east of Tharsis (Coleman et al., 2003). This process may have been greatly enhanced at times of high obliquity when surface ice became stable at low latitudes. Precipitation may also have been augmented by gaseous emissions from Tharsis volcanism. Ice could have nucleated on volcanic particulate emissions and other windblown dust, and ice may also have deposited directly as frost during the adiabatic cooling of air masses rising onto and traversing Tharsis. Even under the present tenuous atmosphere, orographic formation of water ice clouds is common over the Tharsis rise (Pearl et al., 2001). An increased atmospheric pressure during Hesperian time compared to the present-day (but still much lower than during the Noachian) could have further enhanced cloud formation and snow and ice precipitation, especially in association with volcanic eruptions (Hort and Weitz, 2001). Ice blankets layered with aeolian dust and volcanic ash could have accumulated at rates that outstripped sublimation, analogous to the deposits described by Head et al. (2003) and Shean et al. (2005). More extensive surface ice deposits that may have formed during the Hesperian could have provided substantial reservoirs of water ice that were susceptible to catastrophic melting. Eruption of flood basalts could have periodically melted these reservoirs and provided large volumes of water for surface runoff and infiltration. Given a somewhat higher global water inventory during the Hesperian, periods of high obliquity may have resulted in widespread accumulation of ice mantles at low latitudes, and not just in Tharsis. However, the locations where these mantles could have been rapidly converted to meltwaters should have been limited to regions of active volcanism, such as the Tharsis plateau.

Highly permeable fractured basalt aquifers could have efficiently transported meltwaters eastward from recharge areas in Tharsis to the ancestral Valles Marineris canyons. The Hesperian precursor of the structural complex of Noctis Labyrinthus ($10^\circ$ S, $100^\circ$ W) may have provided a network of surface and underground pathways that enhanced drainage from elevated terrain in central Tharsis. The underlying structural complex is probably larger than mapping indicates (Scott and Tanaka, 1986), because geologic units of Amazonian and upper Hesperian age appear to have overlapped and concealed part of its surface expression.

7. Lakes in the Canyons

Coleman (2006) examined the implications of high channels with respect to the formation of lakes in the ancestral Valles Marineris canyons. Analysis of potential fluid pressures beneath the canyon floors led to his conclusion that formation of lakes in the ancestral canyons would have been likely, perhaps inevitable, as a consequence of the rising potentiometric levels of groundwater systems. Conditions appear to have been so favorable for groundwater breakouts that ice-covered lakes likely existed in the canyons when the high channels formed.

The former presence of lakes in the Valles Marineris is supported by the discovery of spillover channels that formed when lakes overtopped canyon rims (Coleman and Baker, 2007; Coleman et al., 2007). There is also geochemical evidence of past aqueous activity in the canyons. The OMEGA hyperspectral imager has identified hydrated sulfates in the light-toned layered terrains on Mars (Gendrin et al., 2005). Sulfates have been identified in all the canyons of Valles Marineris, with the sulfate signatures being mainly found on the flanks of massive deposits and on several isolated interior layered deposits (Quintin et al., 2006). Sulfates have also been detected in Aeonium, Aram, and Iani Chaos (Gendrin et al., 2006). Gendrin et al. (2005) present several interpretations for the origin of these sulfates, including alteration of mafic minerals by acidic precipitation, evaporation of standing water bodies, or by the seepage of hydrothermal brines. The sulfates exist over a wide range of elevations, which suggests that sulfate sediments may have formed in shallow water bodies while the canyons continued to grow.

7.1. Alternate recharge model

We have identified plausible locations for Hesperian recharge sources on and east of the Tharsis plateau (Fig. 11). While flood basalts could have produced large-scale melting of surface ice deposits in this region, given the expected presence of a thick regional cryosphere, it is unclear how large volumes of meltwater could penetrate the surface to reach the aquifer system. An alternate model may be possible for Mars, in which the ancestral Valles Marineris canyons served as basins that collected the extensive meltwater runoff from volcano–ice interactions. The ancestral Noctis Labyrinthus canyon complex and its underlying fault systems could have effectively channeled groundwater eastward from central Tharsis to the main canyons of the Valles Marineris. This process could also have led to groundwater breakouts in the canyons (Coleman, 2006) and the gradual accumulation of lake waters beneath ice covers. The cryosphere beneath the canyon floors would eventually disappear under these conditions and the aquifer system would become directly linked to the lake water systems. In other words, much of the regional aquifer recharge would have occurred deep within ice-covered lakes rather than on the Tharsis plateau. This alternate
model could help to reconcile the expected presence of a thick cryosphere during the Hesperian with the abundant evidence for groundwater as a source for some of the circum-Chryse outflow channels.

8. Implications for life

Analysis of mid- to late-Hesperian thermal conditions suggests that ice covers on lakes at low latitudes would have been less than 3 km thick (Coleman, 2005). Therefore liquid water would have existed beneath the ice covers if lakes in the ancestral canyons were deeper than 3 km. Deep lakes and aquifers would have been ideal refuges for any extant life forms that could have evolved earlier in the Noachian. These environs would have protected organisms from the harsh conditions at the present-day surface. The evidence supporting Hesperian lakes in the canyons and in a large, flat-floored crater (Morella) at 10° S, 51° W provides potential low-latitude targets for landed science missions. Morella Crater represents a paleolake target analogous to the Noachian-aged Gusev Crater explored by the Spirit Mars Exploration Rover. The younger Hesperian-aged sediments deposited in Morella Crater (and in the canyons) may preserve chemical traces of life forms, microfossils, or even spores. Johnstown Crater pre-dates the fluvial event in Morella Crater, and appears to have been filled by lacustrine materials. This crater therefore represents a sediment “trap” that could be targeted for core recovery by a landed mission. As experiments on Earth have shown, spore-bearing bacteria 250 million years old can be isolated and grown (Vreeland et al., 2000). Therefore, the possibility of recovering evidence of life, or even rejuvenating ancient martian life, remains an intriguing prospect for future Mars missions.

9. Conclusions

The polar basal recharge model (Clifford, 1987; Clifford and Parker, 2001) may be a viable mechanism for the creation of Hesperian groundwater recharge on Mars. There are high-standing water-carved features at low latitudes, however, that cannot be explained by polar recharge and groundwater movement in a global aquifer system. Allegheny and Walla Walla Valles are high-standing outflow channels that discharged from terrain >2500 m above the martian datum. The indicated potentiometric surface elevations would have required regional sources of recharge on and east of the Tharsis Plateau. The floodwaters that erupted from Ophir Cavus to form Allegheny Vallis encountered topographic obstacles that restricted the flow, forming temporary lakes. These obstructions were probably breached or overtopped quickly, catastrophically draining the lakes and carving several scablands. After the last obstacle had been breached, a single channel formed that captured all subsequent flow. For this intermediate phase of flooding, we estimate flow velocities of 6–15 m s⁻¹ and discharges in the range of 0.7–3 × 10⁶ m³ s⁻¹. Higher flow velocities and discharges would have occurred when the transient lakes were drained.

We also examined the implications of high channels with respect to the potential formation of lakes in the ancestral Valles Marineris canyons. Formation of lakes in the ancestral canyons should have been likely, perhaps inevitable, as a consequence of rising groundwater potentiometric levels. Conditions appear to have been so favorable for groundwater breakouts that ice-covered lakes probably already existed in the canyons when the high channels formed. In particular, a high-standing, ice-covered lake likely existed in eastern Candor Chasma because this canyon is intersected by the Ophir Catena fault system from which Allegheny Vallis and Walla Walla Vallis originated. Mars appears to have experienced periods of hydrologic quiescence with intermittent and pronounced hydrologic activity that brought about dramatic evolution of landforms during post-Noachian time. The evidence for high-altitude recharge at low latitudes, coupled with inferences about ice-covered lakes in the ancestral canyons, can be used to calibrate models of a volcanic-hydrologic climax during Hesperian time. We also describe an alternate flow model in which groundwater in the Valles Marineris region periodically may have been enhanced by the presence of ephemeral ice-covered lakes in the canyons, created by eastward drainage of meltwaters from central Tharsis via the ancestral Noctis Labyrinthus canyon system. High lake level elevations could have led to the buildup of pressure in confined aquifers downhill from canyon lakes, creating conditions for incipient groundwater breakout. Tectonic activity could have contributed to the fault-controlled release of groundwater and the formation of chaotic terrain and outflow channels.

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References


GSFC, 2004. High-resolution MOLA grid in GMT format (data file 30s_0_330e_g0_330e.gd) is available at http://core2.gsfc.nasa.gov/gridview_gridview_grid.html from the Planetary Geodynamics Laboratory, Goddard Space Flight Center, Greenbelt, MD.


