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Introduction: We are utilizing a variety of approaches to characterize the hydrologic regime characteristic of the late Noachian to early Hesperian transition. Earlier studies of Martian highland landscapes have demonstrated that extensive fluvial denudation occurred near the Noachian-Hesperian boundary [1-13]. In addition to incised fluvial networks, this epoch of intensive erosion is recorded in alluvial fans and deltas [14-19]. Channels dating to this time period are exposed locally on the highlands, permitting estimates of formative discharges from measurements of channel width, meander wavelength, and size of exposed boulders [12, 18, 20, 21]. These measurements suggest discharges of magnitude equivalent to terrestrial mean annual floods for basins of equivalent size, although the frequency of these discharges is not well constrained. We are employing a variety of techniques to further constrain the hydrological regime of early Mars, including computer simulation of landform evolution [4, 7, 22, 23], mapping of valley networks [12], and estimation of the hydrological balance between runoff, lake evaporation, and channel flow [24-27].

Simulations of Landform Evolution: A computer simulation model of Martian landform evolution has been developed that includes rock weathering, mass wasting, fluvial erosion/transport/deposition, hydrological routing through basins, impact cratering, groundwater flow, and eolian deposition [7, 22, 23, 28-30]. Among the conclusions that have resulted from this modeling are: 1) Crater degradation on the Noachian highlands occurred primarily through fluvial erosion and deposition [7]; 2) Groundwater sapping by direct fluvial erosion by emergent groundwater is quantitatively insufficient to erode the valley networks, although seepage weathering may have contributed to valley headwall erosion [30]; 3) The continuation of occasional impacts during (and after) the epoch of fluvial erosion has contributed to the apparently fragmentary nature of the valley networks [23]; 4) Fluvial erosion and sedimentation can integrate intersecting craters into large basins through a combination of rim erosion and burial, including formation of upland plateaus [23]; 5) Creation of integrated valley networks required high ratios of precipitation to lake evaporation [23]; The paucity of deltas in eroded Noachian crater basins reflects either dominantly arid conditions or high quantities of eroded fine sediment (silt and clays) delivered to basins [23]; and 6) A hydrologic scenario of episodes of multi-year, nearly continuous precipitation resulting from basin-scale impacts [31, 32] is inconsistent with highland landscape morphology because such conditions would result in erosional exit breaches of all or most craters, which is not observed [22]. We are continuing the use of simulation modeling by more fully investigating the role of regional slope and different hydrological settings upon the density and depth of valley incision and degree of network integration. We are also modeling the development of Martian alluvial fans.

Mapping of Valley Networks: Mapping of valley networks based upon Viking images suggested that they were fragmentary, so that runoff was not intense or widespread, and may have resulted from groundwater sapping, possibly from hydrothermal sources [33-40]. Recent high-resolution spacecraft images and MOLA topography have revealed that the drainage networks are strongly integrated and have high areal density on steep slopes, indicating a dominance of water sources from precipitation as rain or snow [1-4, 8, 9]. Methods for mapping valley networks include interpretation of visual images [41] and hand or automatic extraction from MOLA DEM’s [2, 42-44]. Mapping from automated techniques may differ depending upon criteria for recognition of valley heads, criteria for downstream continuation and termination of valleys, variable resolution of the original database, effects of craters on valley recognition, and post-valley landscape modification by a variety of processes, including cratering, eolian modification, mass wasting, lava inundation, and glaciation. A global mapping effort is nearing conclusion that has a different emphasis than earlier efforts. A combination of MOLA-derived contour maps plus overlays of THEMIS IR (day and night) and Viking image mosaics is used to map valley networks by hand. The emphasis is on the extent and connectivity of the valley network at the time of its formation, so that, where possible, valleys are extended through disruptions by later processes, such as cratering, and attempts are made to identify the continuation of valley systems through pre-existing depressions (presumably paleolakes). Not every
headwater tributary is mapped, so that the emphasis is not on identifying the original drainage density (which, in any case, is a fragile aspect of paleodrainage that is strongly affected by later modification). The digitized valley networks are analyzed by query of the MOLA PEDR database to reconstruct valley profiles, valley cross-sectional shapes, and incision depths. This database, when completed, will be made available together with search and extraction tools. Figure 1 shows a preliminary map of valley networks in the longitude belt of -30°E to 100°E and longitude range of 30°S to 45°N, and a comparison with the Viking-base map of [34]. This region includes the >4500 km valley system of Naktong Vallis, Scamander Vallis, and Mamers Vallis. For uncertain reasons there are few valley networks in the cratered terrain below about -1500 m in northwestern Arabia (west of 30°E and north of the equator).

Hydrological Modeling: The extent and intensity of valley network development on the cratered highlands depends crucially upon the degree to which runoff was capable of overflowing depressions and integrating a drainage network through fluvial erosion. We have developed a hydrological model that balances runoff from uplands with evaporation from ponded water to determine lake size and the degree of fluvial integration.

A Hydrologic Balance. Consider an enclosed drainage basin of total area \( A_T \) with an included lake of area \( A_L \). We consider a multi-year water balance with the average precipitation rate \( P \) over the watershed. On the uplands the fractional runoff yield is \( R_B \). Yearly evaporation rate on the lake is \( E \). With sufficient precipitation the lake may overflow at a yearly volumetric rate \( V_O \), and overflow from other basins may contribute to the present basin at a rate \( V_I \). A yearly water balance for the basin is thus:

\[
V_O = V_I + (A_L - A_T)P_R + A_L(P - E - A_L). \quad (1)
\]

If the lake does not overflow \((V_O=0)\), then (1) can be solved for the requisite size of the lake. Each basin has a maximum lake area \( A_{LM} \) at which overflow into an adjacent basin occurs, which depends upon the basin topography. If the solution assuming \( V_O=0 \) and \( V_I=0 \) indicates a lake area \( A_L > A_{LM} \) then \( V_O \) is determined by substitution of \( A_{LM} \) for \( A_L \) into (1).

In order to determine the water balance for a large basin with multiple enclosed sub-basins an iterative approach must be used, because the output from several overflowing enclosed sub-basins can serve as inputs for the next sub-basin downstream. The solution for given input parameters \( P, R_B, \) and \( E \) starts by routing water through the channel network to the low point of the basin assuming no en-route evaporation losses. In addition, the basin area \( A_T \) and maximum lake area \( A_{LM} \) are determined. Then the lake area \( A_L \) is calculated for each basin from (1) assuming \( V_I=0 \). If \( A_L > A_{LM} \) then \( V_O \) is calculated and this is used as an input, \( V_I \) for the next sub-basin downstream during the next iterative calculation cycle. Iterations continue until there is no change of \( V_I \) into any sub-basin. Additional complications arise since two or more overflowing basins may mutually drain, and filling of a downstream basin may submerge the outlet for an upstream basin. In such occurrences the basins are combined into a new sub-basin for subsequent iterations. This model has been validated through predicting both the modern and the late Pleistocene lake distribution in the Great Basin region of the southwestern United States [24], with model calibration using spatially explicit data on yearly precipitation, and regression relationships predicting runoff and evaporation as functions of local precipitation, elevation, latitude, and mean annual temperature.

Application to Mars. Some constraints can be placed on runoff magnitudes in Martian channels from dimensions of preserved channels [18, 20, 45]. In addition, the magnitude of evaporation rates from ponded water can be estimated for Mars [46, 47]. However, the atmospheric conditions and precipitation magnitudes on early Mars are highly uncertain and controversial. Additional constraints about the hydrologic balance of early Mars can be obtained from determination of which crater and intercrater basins overflowed and which did not. The approach used here is to model the distribution of lakes and the flow magnitudes in valleys as a function of a range of assumed relative amounts of runoff and lake evaporation. These simulations can then be compared with the occurrence (or lack of) basin overflows to estimate the runoff/evaporation balance that pertained on Mars and how that balance depended upon latitude, elevation, and other possible controls. Absolute magnitudes of evaporation, precipitation, and runoff
yield are uncertain for Mars, but simulations can be made for a variety of assumptions about the ratio, $X$, of the net evaporation from lakes ($E-P$) to the runoff depth from contributing uplands, $PR$. If there are no inflows to a basin and the basin does not overflow, then from (1) the lake area in an enclosed basin is given by:

$$A_L = A_r / (X + 1)$$

Thus all lakes overflow as $X \to 0$ and lakes become indefinitely small as $X \to \infty$.

**Figure 2.** Hydrologic modeling of the region between 120°E to 230°E and 20°N to 80°S for an evaporation ratio, $X=1.5$. In the absence of information on the spatial distribution of runoff depth, we assume that it is areally uniform, and for convenience, we specify the runoff depth, $PR$, to be unity. Relative elevation is indicated by dark to light red, the location of ponded water by blue, and green indicates flowpaths, with thicker and brighter lines for greater discharge. Straight lines connect inflows to a lake to the lake overflow (if there is one) so that flowpaths can be readily traced. Post-Noachian craters have been cosmetically excised from the DEM using the “healing” tool of Photoshop®.

**Figure 3.** Simulation as in Figure 2 but with evaporation ratio of $X=3.5$. Note that the predicted drainage networks to the upper right and upper left are on post-Noachian volcanic terrain and are thus specious products of the routing procedure. Networks along the top portion of Figures 2 and 3 are on the post-Noachian northern lowlands and are also specious.

**Figure 4** shows the discharges for three valleys on the highlands as a function of $X$. Flows are normalized by assuming a unit runoff depth ($PR=1$ in Eq. (3)), so that for $X=0$ the flow equals the maximum upstream contributing area. If there were no upstream

depressions (craters, intercrater basins, etc.) the relative discharge would be independent of $X$ (no lake evaporation), and simulations for typical terrestrial drainage networks with no lakes or reservoirs show no dependence of relative discharge upon $X$. The Martian networks, however, all show strong decreases of discharge with increase in evaporation ratio (**Figure 4**). The Samara Valles system is a relatively maturely dissected drainage system, and exhibits a gradual decline in relative discharge with $X$. By contrast Uzboi is sampled just downstream from possible overflow into this valley from Argyre and Ma’adim is sampled near its head at the possible overflow into the valleys from the southern highlands. These show an abrupt drop in relative discharge when $X$ increases such that Argyre and the highlands, respectively, do not overflow into the basin.

Similarly, lake surface elevation within basins can be estimated as a function of $X$ (**Figure 5**). As $X$ decreases lake elevation increases, but then remains constant after the basin overflows.

More than 200 overflowing basins on the Martian highlands have been identified by [48]. Estimated
values of $X$ for these basins to have overflowed range down to about 2.0. Another constraint, which we are now investigating, is the minimum value that $X$ can reach before basins which appear not to have overflowed would have done so. An additional technique being applied is to compare the depth and volume of valley incision to predicted flow magnitudes for various values of $X$ to find the best regional fit.


Figure 4. Relative discharge through three valleys as a function of the evaporation ratio, $X$.  

Figure 5. Simulated levels of lakes in three basins as a function of evaporation ratio, $X$.  