A QUANTITATIVE INVESTIGATION OF FLUVIAL ACTIVITY IN THE HESPERIA PLANUM REGION, MARS.

T.K. Jones[1] ,T.K.P. Gregg[1], and D.A. Crown[2] [1]State University of New York at Buffalo, Dept. of Geology, 876 Natural Sciences Complex Buffalo 14260, tessaj@gmail.com, [2]Planetary Science Institute, 1706 E. Fort Lowell Rd., Suite 106, Tucson, AZ 85719

Introduction: Hesperia Planum is located NE of the Hellas basin and covers ~2 x 10^6 km^2. It is surrounded by highlands to the west and east and is characterized as plains material containing mare-type wrinkle ridges and ridge rings [1,2]. This area has been interpreted to be composed of lava flows (possibly flood basalts) that filled topographically lower regions within the Martian highlands [3,4]. These observations and interpretations were made using Viking Orbiter images, which typically have resolutions of >150 m/pixel. The acquisition of new high-resolution data sets has enabled new stratigraphic and geologic interpretations of Hesperia Planum deposits [see Crown et al. and Gregg and Crown, this volume]. Recently it has been suggested [5-7] that Hesperia Planum includes significant sedimentary deposits, some of which may be lacustrine. Here, we investigate fluvial activity in and around Hesperia Planum to assess potential water supplies.

Detailed observations of Hesperia Planum using THEMIS and MOLA data have led to the identification of several small basins within the adjacent dissected highlands. These basins are filled with ridged plains that have the same morphologic characteristics as the Hesperia Planum ridged plains, although they are located at higher (>500 m) elevations. Channels and valley networks interpreted to have formed by precipitation and groundwater sapping are identified in the highlands adjacent to these basins [6,8]. We have employed a quantitative study of the channels and valleys around these smaller basins to constrain their potential fluvial input. Quantifying the discharge rate of channels and groundwater flux around the highland basins will determine if reasonable recharge rates could have existed during Noachian/Hesperian times.

Methodology: For this analysis, we assume that the highland basins within the dissected highlands contain lacustrine deposits, and were therefore once regions of an exposed water table. We further assume that liquid water was present and available during the time of channel formation. The purpose of this study is to evaluate the potential volume of water available given the size of the basins and their associated watersheds.

Using MOLA and THEMIS daytime IR data, we have mapped channels in the highlands adjacent to Hesperia Planum. Theatre-heads, flat-floors, and distal narrowing of single channels are the main features used to identify groundwater sapping channels [9]. Based on channel morphology, we infer that groundwater sapping accounts for the most recent modification of the fluvial features located around Hesperia Planum. Channels that are v-shaped and part of a dendritic valley network are interpreted to have been most recently modified by surface runoff, suggestive of precipitation.

To date, we have examined three basins around Hesperia Planum: basin #1, centered at 5º N and 95º E; basin #2, centered at 9° N 88° E, and basin #3, centered at 21º N 128º E. We have considered 2 sources of water for each basin: 1) dendritic valley networks formed from precipitation; and 2) groundwater flow. THEMIS daytime IR data indicate that highland basin #2 has no visible channels extending from the highlands toward its interior. Basins #2 and #3 have numerous channels interpreted to have been modified by ground water sapping as well as dendritic valley networks formed from surface runoff.

Potential groundwater contributions were calculated for each basin. Assuming that an available, full aquifer supplied groundwater flow around each basin that was topographically driven, Darcy's Law can be used to estimate the rate of discharge (Q):

$$Q=-k\,(dh/dr)\,(LH)$$

where k = hydraulic conductivity (km/s), dh = change in head (km), dr = distance between head measurements (km), Q= discharge (km^3/s), L = representative length perpendicular to profile (km) and H = depth of proposed lake (km). Between 52 and 63 detailed topographic profiles drawn perpendicular to the basin margin were extracted using Gridview (128 pixels/degree) for each basin. A parabolic best-fit curve was obtained for each profile, the apex of which marked the watershed boundary for the basin (Fig. 1). From each topographic profile, a value for dh/dr was acquired; dh = the relief between basin under investigation and an adjacent basin. Where an adjacent basin is not present, dh = the vertical distance between the parabola apex and H. Because hydraulic conductivity (k) is a difficult parameter to constrain, we used a likely range from

4×10^{-7} to 2×10^{-2} m/s [9]. Length (L) was defined as the total perimeter of the basin divided by the number of topographic profiles. Several values for height (H) were used to represent different depths of the proposed lake. The results presented here use 0.08 km. Discharge (Q) of each profile was then calculated. The total discharge of each topographic profile is added together to calculate the total groundwater input for each basin.

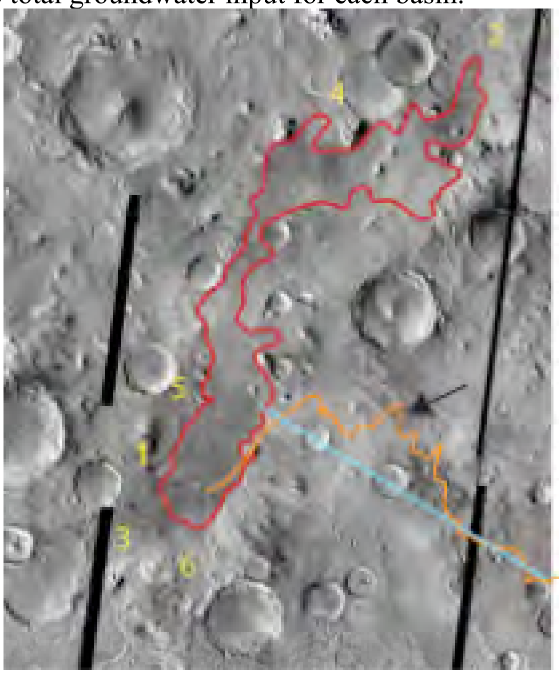

Figure 1. THEMIS daytime IR mosaic of basin #1, which is ~150 km across. Red line outlines plains material. V-shaped, dendritic network (1); flat-floored channels (3-6). Blue line marks location of topographic profile (orange line). Arrow marks the watershed boundary for this profile.

The Manning equation was used to calculate the discharge from trunk valleys of dendritic networks around each basin.

$$Q=0.51/n(AR^{2/3}\,\Theta^{1/2})$$

in which Q = discharge (km^3/s), n = roughness coefficient (s $m^{-1/3}$), A = cross-sectional area (km^2), R = hydraulic radius (km), and Θ = slope. A roughness coefficient (n) of 0.0545 s $m^{-1/3}$, characterized by Wilson [10], was employed. The width, height, and slope of the channels were measured using Gridview and the MOLA 128 px/degree DEM, and then used to estimate the cross-sectional area of the channel (A). Maximum discharge was estimated using the channel depth, which assumes bank-full flow. A more conservative calculation was done using 1/10 the channel height. Basin # 1 has one dendritic valley network; basin #3 has two.

Preliminary Results: A dendritic, v-shaped valley located on the western edge of highland basin #1 (Fig. 1) is at most 4.2 km wide and 0.165 km deep. The calculated discharge rate for this valley ranges from 0.01 to 0.07 km^3/s, requiring between 0.7 and 6.5 Earth years to fill the basin 100 meters deep. The potential contribution from groundwater flow is less, with a total influx of 0.37 m^3/s. A channel draining this basin is located on the northern border of basin #1 (Fig. 1). It is inferred that water entered the basin through a channel located on the southwest border, traveled north through the outflow channel and from there, into another adjacent highland basin.

Highlands basin #2 has no visible channels associated with it, suggesting that groundwater flow may have been the most significant contributor. Calculations indicate that if the hydraulic conductivity is $\geq 2 \times 10^{-3}$ m/s, basin #2 could be filled 80 m deep within 6 years. These numbers do not take into account evaporation, freezing, sublimation, or draining of the lake by a channel. Results for basin #3 will be presented in the future.

Conclusions: A quantitative investigation of the discharge values around the basins indicates that if sufficient water were available, it may have been possible to fill these basins, supporting the hypothesis that they are paleo lakes. Ground water sapping may have contributed a large volume of water to the basins; future work will further explore the fluvial history of the highlands and Hesperia Planum proper, integrating geologic observations with quantitative constraints on fluvial processes.

References: [1] Scott D.A. and M.H. Carr, (1978) *Misc. Series I-1083*. [2] Goudy C. et al. (2005) *JGR, 110*, E10005. [3] Greeley R. and Guest J.E. (1987) *U.S. Geol. Surv. Misc. Series* I-1802B [4]Greeley R. and Spudis P.D. (1981) *Rev. Geophys. 19*:13-41 [5] Crown D.A. et al. (2005) *JGR* 110, E12S22 [6] Ivanov et al., (2005) *JGR* 110, E12S21. [7] Gregg, T.K.P. and D.A. Crown (2005) LPSC XXXVI, abstract #1962. [8] Luo W. and Stepinksi, T.F (2006) *GRL*, 33, L18202. [9] Domenico P.A. and Schwartz F.W. (1998) John Wiley and Sons Inc. [10]Wilson L. et al. (2004) *JGR* 109, E09003. [11] Carr M.H. (1979) *JGR* 84, 2995-3007. [12] Crown, D.A and Mest S.C. (2004) USGS OFR 2004-1289 [13] Crown, D.A. et al, this issue [14] Gregg, T.K.P. and D.A.Crown, this issue [15] Kleinhans M.G. (2005) *JGR* 110, E12003. [16] Komar P.D. (1980) *Icarus* 42, 317-329. [17] Scott D.H. and K.L. Tanaka (1986) *U.S. Geol. Surv. Misc. Series I-1802A*. [18] Tanaka K.L. (1986) *JGR suppl*., 91, E139-E158