Hellas Planitia, Mars: Site of net dust erosion and implications for the nature of basin floor deposits

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Abstract. Hellas Planitia, located within an enclosed basin which includes the lowest topography on Mars, appears to be undergoing net erosion. Dust is removed from the basin. It probably contributes to global dust storms and should leave behind a coarse lag. The particle size distributions and particularly the rock or boulder populations in this lag might be useful for distinguishing between processes which formed the lithologic units that comprise Hellas Planitia. This report concludes that the abundance of rock particles larger than coarse sand is very low. Although this hypothesis awaits confirmation from forthcoming spacecraft data, the origins for Hellas floor deposits favored by this study are indurated volcanic airfall or ancient loess, lacustrine deposits, and some types of volcanic mud flows. The conclusions of this study tend to disfavor such geologic processes as blocky lava flows, glacial deposits (e.g., moraines), or boulder-laden catastrophic flood outwash.

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Introduction

Hellas Planitia, topographically the lowest region on Mars, is contained within an ancient ~2,000 km-wide impact basin. Hellas is relatively dynamic in the current climate: in winter, it is covered by bright frost; in summer it is a site of dust storm activity. Hellas Planitia has few impact craters, perhaps implying an extended history of depositional and/or erosional processes (e.g., Greeley and Guest 1987).

Greeley and Guest (1987) identified seven geomorphic "materials" units on the floor of Hellas basin, collectively termed the "Hellas Assemblage." These units were interpreted as sedimentary deposits of possible fluvial or aeolian origin, perhaps interbedded and/or partly covered by lavas. These units have subsequently undergone considerable aeolian erosion, leaving behind landforms such as yardangs, knobs, mesas, and inverted relic craters. Subsequent geomorphic studies of the Hellas region have focused upon the volcanic and channeled regions to the east of the basin, where many landforms appear to result from volcano-ground ice interactions, pyroclastic and effusive volcanism, and fluvial (some catastrophic) events (Squyres et al. 1987, Crown et al. 1992). To what extent these processes have contributed to the formation of deposits in central and western Hellas Planitia is yet to be fully mapped and quantified. In eastern Hellas Planitia there are deposits formed of sediments derived from channels that flowed into the basin (Crown et al. 1992). A different model for the origin of geomorphic features in Hellas was proposed by Kargel et al. (1991) and Kargel and Strom (1992). They identified a suite of landforms, widely distributed about the basin floor and slopes which together might be explained by continental-style glaciation and related proglacial lake formation.

The mapping results described above collectively indicate that Hellas Planitia is probably composed of depositional units (sedimentary and volcanic) perhaps laid down over a long period of time that began when the basin was formed by impact more than 4 Gya. The divergence of hypotheses and considerable uncertainties concerning Martian climatic history implies that an understanding of the physical and sedimentological properties of the materials on the floor of Hellas could be significant tools for addressing fundamental questions about Mars. The widespread occurrence of erosional landforms on Hellas Planitia indicate that the original deposits have been modified (Greeley and Guest 1987). Erosion within a deep, enclosed basin would likely result in redistribution of sediments from one location to another. This study considers a significant alternative: that aeolian erosion has resulted in removal of sediment from the basin interior. We discuss the properties of Hellas Planitia surface materials using Viking orbiter Infrared Thermal Mapper (IRTM) data, visual observations, and surface-wind computer modeling. We propose that dust has been winnowed out of the basin floor and removed from the region via suspension. This process may leave behind a lag of coarser material, allowing the possibility of testing hypotheses of Hellas Planitia deposit origins by considering the size distributions of the coarser sediments.

Observations

Historical telescope observations show that Hellas is the most active dust storm site on Mars (Martin and Zurek 1993). During the Martian year, the clearest period occurs between L_s 340° and L_s ~100° (late southern summer and autumn; Martin and Richardson 1993). IRTM-derived 9 µm opacities over Hellas were $\tau < 0.3$ during the clear periods, with $\tau < 0.2$ between L_s 20° to 25° (Martin and Richardson 1993). However, throughout most of the Viking mission (1976–1980), Hellas had a dusty atmosphere. Especially severe dust storm activity preceded the onset of the first global dust storm of 1977 (Martin and Richardson 1993).

Martian general circulation models (GCM) have consistently calculated that the interior of Hellas is the site of very high winds (Pollack et al. 1981). The direction of wind circulation computed for Hellas when dust opacity is high (Skypeck 1989, Greeley et al. 1993, Fig. 2e) mimics the pattern of dust storms observed by Viking (see Martin and Richardson 1993). The surface wind shear stress in Hellas Planitia is greatest in the northern portion of the basin, suggesting that this area, if not all of Hellas, could be most susceptible to aeolian activity and/or erosion. Except for scattered patches of dark-hued material presumed to be sand (e.g., Thomas 1984), Hellas Planitia appears to have a relatively uniform, intermediate albedo (0.23 ± 0.02 in Christensen 1988).

The physical nature of the upper 2–10 cm of Hellas Planitia is estimated from IRTM-derived thermal inertias. Explained in detail elsewhere, thermal inertia (the square root of the product of bulk density, specific heat, and thermal conductivity) is most sensitive to thermal conductivity at Martian atmospheric pressures because it is related to the particle size of unconsolidated materials (e.g., Neugebauer et al. 1971, Edgett and Christensen 1991). Here, we express thermal inertias derived using the Kieffer et al. (1977) thermal model in units of 10^{-3} cal cm^{-2} s$^{-½}$ K^{-1} (S.I. units = 41.8 J m^{-2} s$^{-½}$ K^{-1}), and interpret effective particle sizes as discussed by Edgett and Christensen (1991). The results of Zimbelman and Greeley (1983) suggest that the thermal inertias presented here are good to within ~1.5 units with respect to elevation-pressure dependent corrections. Haberle and Jakosky (1991) showed that thermal inertias determined by the Kieffer et al. (1977) model are too high by 0.5 to 2.0, owing to infrared backradiation from atmospheric CO_2 and suspended dust. Therefore, thermal inertias and particle sizes presented here represent upper limits. Effective particle sizes are given in terms of the Wentworth (1922) sedimentological scheme; again, owing to the differences between the Kieffer and Haberle–Jakosky thermal models, these grain size classes are upper limits, the lower is the next-lower grain size class. It is important to note that there are no Viking IRTM data of sufficient temperature contrast to model the amount of rocks, boulders, or lithic outcroppings in Hellas (Christensen 1986a).

Three separate sets of Viking 1 IRTM predawn scans (~30 km resolution, obtained during the clear-atmosphere period of L_s 20°–50°) were mapped in 0.5° by 0.5° latitude-longitude bins. Set 1 consists of orbits 549-9, 551-2, 553-4, and 555-8 (in which 549-9 is orbit 549, sequence 9). Set 2 has data from orbits 586-3, 587-4, 590-8, and 592-9; and Set 3 has orbits 622-3, 625-5, 627-7, 629-4, and 632-4. Set 3 covers the greatest amount of Hellas Planitia, but is complicated by late autumn CO_2 frosts over much of the southern half of the basin. Set 1 has the next best spatial coverage but there are six relatively small (< 90 km diameter), thermally-distinct clouds in part of the northern region (Figure 1). Set 2 provides less coverage, but is sufficient to demonstrate that the clouds in Set 1 are ephemeral atmospheric features. Set 2 data also confirm the placements and magnitude of thermal inertias mapped in Set 1. Set 1 (Figure 1) and Set 2 data indicate that the Hellas floor surface units have effective particle sizes that are generally > 125 µm (fine sand) with most in the medium to very coarse sand range (250–2000 µm). The surface materials are coarsest in the northern portion of the basin, consistent with the region of highest wind shear.

Discussion

On the basis of the GCM computations and dust storm observations cited above, we propose that the surface of Hellas Planitia is being stripped of suspendable particles (grains << 100 µm). Particles ~100 to 1000 µm should saltate in Hellas (e.g., Iversen and White 1982). Saltating sand, probably found in the observed small, dark-hued patches, would provide a mechanism to lift the observed clouds of dust into suspension. Dust grains would then be completely removed from the basin via dust storm activity. This erosion should leave behind coarser materials: sand, granules, pebbles, and, if initially present, anything larger. The relative proportions of these coarser materials may be indicators of processes which formed the original lithologic units (Hellas Assemblage). For example, a preponderance of poorly sorted rocks and boulders might support a glacial hypothesis.

As a current source for dust storm particulates, we suspect that most of the fines being eroded from the Hellas floor were not deposited as fall-out from current-epoch global dust storms. We further speculate that regardless of Martian climate or obliquity, Hellas has never been a site of net aeolian dust deposition following global dust storms, because only high-elevation areas are currently seen to have thick (up to ~1 m) deposits of dust that settle out of global storms (e.g., Christensen 1986b). Thus, a significant amount of Hellas dust is probably derived from the erosion of in situ basin floor units.

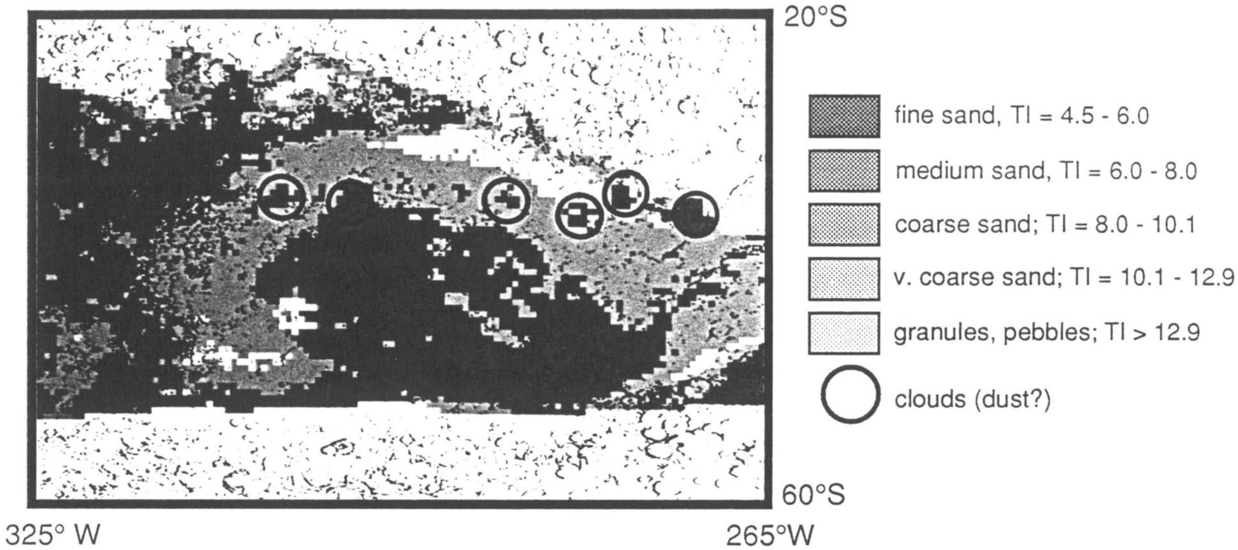

Figure 1. Thermal inertia map of Hellas region. Image constructed from predawn Viking IRTM observations binned by ½° latitude–longitude, superimposed on shaded relief map. The thermal inertias (TI) and corresponding effective particle sizes presented here are considered to be upper limits; the lower limit for each unit would be one size class lower than shown. For example, much of central Hellas has thermal inertias of 6.0 to 8.0 x 10^{-3} cal cm^{-2} $s^{-½}$ K^{-1}, or effective particle sizes of medium sand. The lower limit for this unit would be thermal inertias of about 4.5 to 6.0 x 10^{-3} cal cm^{-2} $s^{-½}$ K^{-1}, or fine sand. The circled areas indicate clouds; data within the circles should be ignored. **See last page of this self-archived report for a color figure similar to this one.**

One important question is that of rock abundance (Christensen 1986a). If Hellas' floor is being stripped of dust, then this region likely has a low rock abundance (< 5%), because non dust-covered surfaces with high rock abundance, like found in Acidalia Planitia, typically have a lower albedo and higher thermal inertia than Hellas Planitia (e.g., Christensen and Moore 1992). Moreover, rocky, windswept surfaces are rough and typically prevent erosion of finer grains between the stones (Chepil and Woodruff 1963).

If the Hellas floor is generally free of "rocks," then the thermal inertias (Figure 1) indicate fine-or medium- to very coarse- sand. However, Martian sands usually have low albedos (e.g., Thomas 1984), yet most of Hellas Planitia has an intermediate albedo. Relatively rock- and sand-free surfaces that have thermal inertias in the range of 6.0 to 9.0 (lower limit ~4.0) might consist of indurated fines (mostly dust with some sand-sized material) (Jakosky and Christensen 1986). If indurated deposits form the bulk of the material being eroded to create dust storms in Hellas, then the proposed low rock abundance might be explained in one of two ways: either (a) there simply are no rocks, boulders, etc. present in the uppermost Hellas Planitia deposits or (b) the indurated layer is thick, buries rocks, and has not yet been eroded deeply enough to expose them. We speculate that the great depth of the modem basin may be the result of the long term (> 10^9 yr) removal of thick (several km) Hellas Assemblage materials (which were themselves deposited within a deep impact structure). It is worth noting that the highest model wind stresses and the most extensively wind modified landforms occur at the lowest elevations within Hellas.

If the surfaces of the Hellas Assemblage units consist of indurated fines with relatively few rocks (> 10 cm), then our interpretation of the data disfavors blocky lava flows, glacial deposits (e.g., moraines), or boulder-laden catastrophic flood outwash. Geologic processes favored by our interpretation are: indurated volcanic airfall or ancient loess, lacustrine deposits, and some types of lahars (volcanic mud flows). Deposits emplaced by catastrophic floods in association with, or formed entirely within, rock-poor deposits will have a low rock abundance. It should be noted that terrestrial lahars can incorporate large rocks picked up during emplacement, but if there is no rock source, then these deposits will not contain them. The deposits favored by our interpretation are types commonly composed of cemented or indurated fine material. Strong winds in conjunction with a local supply of sand can mechanically disaggregate clods and outcrops of indurated fines, thus providing a source for new dust storm particles.

Specific observations can be made by instruments aboard Mars Observer and other forth-coming Mars missions that will constrain further the nature of Hellas floor deposits. The question of rock abundance is the key to evaluate fully the proposals for Hellas floor lithologic units. Based upon our analysis, we suggest that rock abundance on the lower floor is very low and that the deposits there are undergoing net erosion. However, in order to settle the issue new data are needed. The Mars Observer Camera (MOC) and the Thermal Emission Spectrometer (TES) may provide much-needed high spatial resolution observations of the surface characteristics and block abundances in Hellas. Deposits formed by different geologic processes have different large-particle size-frequency distributions (Malin 1989). If present, the MOC will be able to observe the distribution of any meter-scale boulders present in Hellas. We predict that boulders will be scarce or nonexistent within the lower portions of Hellas. The TES will complement MOC observations of particle size distributions by providing thermophysical and spectral information. The higher spatial resolution of TES (3 to 9 km) compared to IRTM (~30 km) should allow for improved definition of the spatial variations in thermal properties. We predict that TES-derived "rock" abundances at the lowest elevations of Hellas will be nearly zero. It may also be possible to identify which geologic process deposited fine grained material within Hellas. Images (some in stereo) of ~25 m/pixel resolution from the Mars '94/'96 orbiters would permit a vast improvement over Viking images for characterization of landforms, which would permit the recognition of geomorphology unique to specific geologic processes. Compositional information from TES might also constrain the origin of Hellas floor materials; for example, the presence of abundant carbonate might be an indicator of lacustrine deposition. Gravitational potential plays a major role in the distribution of various facies during sediment deposition for a number of geologic processes, and the Mars Observer Laser Altimeter (MOLA) will also contribute to the evaluation of Hellas floor deposit origins.

Owing to frequent obscuration by airborne dust, the opacity results of Martin and Richardson (1993) suggest that the best time to observe Hellas is from L_s 340° to L_s 100°, when atmospheric opacities are likely to be low. The best TES and MOLA data will be likewise limited to this period; though the presence of frost after L_s 40° may limit thermal observations from L_s 340° to L_s 40°. For the Mars Observer primary mission, this will occur in September–December 1994.

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References

Chepil, W. S., and N. P. Woodruff (1963) The physics of wind erosion and its control, *Advances in Agronomy 15*, 211–302, doi:10.1016/S0065-2113(08)60400-9.

Christensen, P. R. (1986a) The spatial distribution of rocks on Mars, *Icarus 68*, 217–238, doi:10.1016/0019-1035(86)90020-5.

Christensen, P. R. (1986b) Regional dust deposits on Mars: Physical properties, age, and history, *Journal of Geophysical Research 91*(B3), 3533–3545, doi:10.1029/JB091iB03p03533.

Christensen, P. R. (1988) Global albedo variations on Mars: Implications for active aeolian transport, deposition, and erosion, *Journal of Geophysical Research 93*(B7), 7611–7624, doi:10.1029/JB093iB07p07611.

Christensen, P. R., and H. J. Moore (1992) The Martian surface layer, in *Mars*, edited by H. H. Kieffer, B. M. Jakosky, C. W. Snyder, and M. S. Matthews, p. 686–729, University of Arizona Press, Tucson.

Crown, D. A., K. H. Price, and R. Greeley (1992) Geologic evolution of the east rim of the Hellas basin, Mars, *Icarus 100*, 1–25, doi:10.1016/0019-1035(92)90014-X.

Edgett, K. S. and P. R. Christensen (1991) The particle size of martian aeolian dunes, *Journal of Geophysical Research 96*(E5), 22,765–22,776, doi:10.1029/91JE02412.

Greeley, R., and J. E. Guest (1987) Geologic map of the eastern equatorial region of Mars, *U. S. Geological Survey Miscellaneous Investigations Series Map 1-1802*-B.

Greeley, R., A. Skypeck, and J. B. Pollack (1993) Martian aeolian features and deposits: Comparisons with general circulation model results, *Journal of Geophysical Research 98*(E2), 3183–3196, doi:10.1029/92JE02580.

Haberle, R. M., and B. M. Jakosky (1991) Atmospheric effects on the remote determination of thermal inertia on Mars, *Icarus 90*, 187–204, doi:10.1016/0019-1035(91)90100-8.

Iversen, J. D.,and B. R. White (1982) Saltation threshold on Earth, Mars, and Venus, *Sedimentology 29*(1), 111–119, doi:10.1111/j.1365-3091.1982.tb01713.x.

Jakosky, B. M., and P. R. Christensen (1986) Global duricrust on Mars: Analysis of remote-sensing data, *Journal of Geophysical Research 91*(B3), 3547–3559, doi:10.1029/JB091iB03p03547

Kargel. J. S., and R. G. Strom (1992) Ancient glaciation on Mars, *Geology 20*(1), 3–7, doi:10.1130/0091-7613(1992)020<0003:AGOM>2.3.CO;2.

Kargel, J. S., R. G. Strom, and N. Johnson (1991) Glacial geology of the Hellas region on Mars, *Lunar Planetary Science XXII*, 687–689, Lunar and Planetary Institute, Houston, Texas.

Kieffer, H. H., T. Z. Martin, A. R. Peterfreund, B. M. Jakosky, E. D. Miner, and F. D. Palluconi (1977) Thermal and albedo mapping of Mars during the Viking primary mission, *Journal of Geophysical Research 82*(28), 4249–4292, doi:10.1029/JS082i028p04249.

Malin, M. C. (1989) Rock populations as indicators of geologic processes, *Reports of the Planetary Geology and Geophysics Program – 1988*, p. 363–365, *NASA Technical Memorandum 4130*.

Martin, L. J., and R. W. Zurek (1993) An analysis of the history of dust activity on Mars, *Journal of Geophysical Research 98*(E2), 3221–3246, doi:10.1029/92JE02937.

Martin, T. Z., and M. I. Richardson (1993) New dust opacity mapping from Viking IR thermal mapper data, *Journal of Geophysical Research*, in press. (later published in *Journal of Geophysical Research 98*(E6), 10,941–10,949, doi:10.1029/93JE01044).

Neugebauer, G., G. Münch, H. H. Kieffer, S. C. Chase Jr., and E. Miner (1971) Mariner 1969 infrared radiometer results: Temperatures and thermal properties of the Martian surface, *The Astronomical Journal 76*(8), 719–728, doi:10.1086/111189.

Pollack J. B., C. B, Leovy, P. W. Greiman, and Y. Mintz (1981) A Martian global circulation experiment with large topography, *Journal of the Atmospheric Sciences 38*(1), 3–29, doi: 10.1175/1520-0469(1981)038<0003:AMGCEW>2.0.CO;2.

Skypeck, A. P. (1989) *Comparison of a Mars General Circulation Model with Aeolian Features and Deposits*, M.S. thesis, Arizona State University, Tempe.

Squyres, S. W., D. E. Wilhelms, and A. C. Moosman (1987) Largescale volcano-ground ice interactions on Mars, *Icarus 70*, 385–408, doi:10.1016/0019-1035(87)90085-6.

Thomas, P. (1984) Martian intracrater splotches: Occurrence, morphology, and colors, *Icarus 57*, 205–227, doi:10.1016/0019-1035(84)90066-6.

Wentworth, C. K. (1922) A scale of grade and class terms for clastic sediments, *Journal of Geology 30*(5), 377–392, doi:10.1086/622910.

Zimbelman, J. R., and R. Greeley (1983) Elevation-dependent corrections for thermal inertias on Mars, *Lunar and Planetary Science XIV*, 879–880, Lunar and Planetary Institute, Houston, Texas

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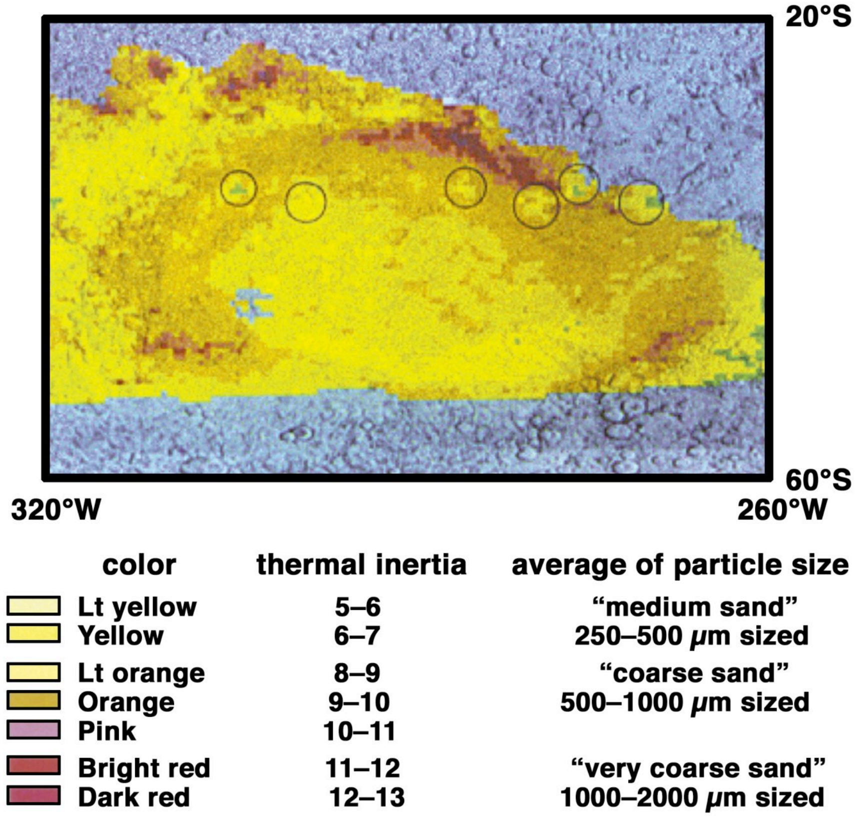

Supplemental Figure. Color version of Figure 1, the thermal inertia map of Hellas region. Image constructed from predawn Viking IRTM observations binned by ½° latitude–longitude, superimposed on shaded relief map. The thermal inertias and corresponding effective particle sizes presented here are considered to be upper limits; the lower limit for each unit would be one size class lower than shown. For example, much of central Hellas has thermal inertias of 6.0 to 8.0 x 10^{-3} cal cm^{-2} $s^{-½}$ K^{-1}, or effective particle sizes of medium sand. The lower limit for this unit would be thermal inertias of about 4.5 to 6.0 x 10^{-3} cal cm^{-2} $s^{-½}$ K^{-1}, or fine sand. The circled areas indicate dust clouds (night-time dust storms); data within the circles should be ignored.