GEOLOGIC MAP OF THE HELLAS REGION OF MARS

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INTRODUCTION

This geologic map of the Hellas region focuses on the stratigraphic, structural, and erosional histories associated with the largest well-preserved impact basin on Mars. Along with the uplifted rim and huge, partly infilled inner basin (Hellas Planitia) of the Hellas basin impact structure, the map region includes areas of ancient highland terrain, broad volcanic edifices and deposits, and extensive channels. Geologic activity recorded in the region spans all major epochs of martian chronology, from the early formation of the impact basin to ongoing resurfacing caused by eolian activity.

The Hellas region, whose name refers to the classical term for Greece, has been known from telescopic observations as a prominent bright feature on the surface of Mars for more than a century (see Blunck, 1982). More recently, spacecraft imaging has greatly improved our visual perception of Mars and made possible its geologic interpretation. Here, our mapping at 1:5,000,000 scale is based on images obtained by the Viking Orbiters, which produced higher quality images than their predecessor, Mariner 9. Previous geologic maps of the region include those of the 1:5,000,000-scale global series based on Mariner 9 images (Potter, 1976; Peterson, 1977; King, 1978); the 1:15,000,000-scale global series based on Viking images (Greeley and Guest, 1987; Tanaka and Scott, 1987); and detailed 1:500,000-scale maps of Tyrrhena Patera (Gregg and others, 1998), Dao, Harmakhis, and Reull Valles (Price, 1998; Mest and Crown, in press), Hadriaca Patera (D.A. Crown and R. Greeley, map in preparation), and western Hellas Planitia (J.M. Moore and D.E. Wilhelms, map in preparation).

We incorporated some of the previous work, but our map differs markedly in the identification and organization of map units. For example, we divide the Hellas assemblage of Greeley and Guest (1987) into the Hellas Planitia and Hellas rim assemblages and change the way units within these groupings are identified and mapped (table 1). The new classification scheme includes broad, geographically related categories and local, geologically and geomorphically related subgroups. Because of our mapping at larger scale, many of our map units were incorporated within larger units of the global-scale mapping (see table 1).

Available Viking images of the Hellas region vary greatly in several aspects, which has complicated the task of producing a consistent photogeologic map. Best available image resolution ranges from about 30 to 300 m/pixel from place to place. Many images contain haze caused by dust clouds, and contrast and shading vary among images because of dramatic seasonal changes in surface albedo, opposing sun azimuths, and solar inclination. Enhancement of selected images on a computer-display system has greatly improved our ability to observe key geologic relations in several areas.

Determination of the geologic history of the region includes reconstruction of the origin and sequence of formation, deformation, and modification of geologic units constituting (1) the impact-basin rim and surrounding highlands, (2) volcanic and channel assemblages on the northeast and south sides of the basin, (3) interior basin deposits, and (4) slope and surficial materials throughout the map area. Various surface modifications are attributed to volcanic, fluvial, eolian, mass-wasting, and possibly glacial and periglacial processes. Structures include basin faults (mostly inferred), wrinkle ridges occurring mainly in volcanic terrains and interior plains, volcanic collapse craters, and impact craters. Our interpretations in some cases rely on previous work, but in many significant cases we have offered new interpretations that we believe are more consistent with the observations documented by our mapping. Our primary intent for this mapping has been to elucidate the history of emplacement and modification of Hellas Planitia materials, which form the basis for analysis of their relevance to the global climate history of Mars (for example, Tanaka and Leonard, 1995).

HISTORICAL OBSERVATIONS

The immense size of the sharp albedo changes observed for Hellas have mystified observers for well over a hundred years (Blunck, 1982). Nineteenth-century and earlier observers believed this region (designated on older maps as “Lockyer Land”) to be a raised landmass or an island adjacent to the “Hourglass Sea” (Syrtis Major). Giovanni Schiaparelli mapped two large canals, Peneus and Alpheus, in the form of an “X” pattern within Hellas Planitia (Schiaparelli, 1877), yet today we find no features that correlate with this observation. Percival Lowell also mapped numerous canals in the region (Lowell, 1895). During some oppositions the region appeared brilliantly white when viewed near the edge of the martian disc, fooling some into believing it was some kind of polar cap. We now know this effect to be caused by high albedo CO_2 fog that seasonally condenses into lower Hellas Planitia.

Sending spacecraft to the red planet significantly increased our understanding of Mars and the Hellas region. The Mariner 4, 6, and 7 flybys combined to image about 20 percent of the planet's surface, including the Hellas region at fairly low resolutions (Moore, 1977; Snyder and Moroz, 1992). However, surface features of the Hellas region were concealed by thick dust and haze. Thus, throughout the early years of the spacecraft era, the Hellas region continued to be a mystery of the martian surface. Researchers speculated about large deserts and vigorous, perhaps ongoing, resurfacing, a view that is not entirely untrue. In fact, the Hellas region became one of three type areas (featureless plains) first differentiated by the Mariner data (Snyder and Moroz, 1992). The Mariner 7 spacecraft encounter in 1969 finally determined Hellas to be a large basin and not a raised plateau as previously believed.

The Mariner 9 orbiter in 1971 imaged the region in somewhat greater detail than previous spacecraft. However, dust and haze continued to obscure the basin floor several months after the great global dust storm of 1971 had subsided, and only a few frames from the Mariner 9 extended mission show part of the floor in some detail. The two Viking Orbiters, which visited just a few years later, showed that Hellas was not a featureless plain as previously thought; rather it is a broad, deep basin having a rugged, complex floor—now called Hellas Planitia—and large, ancient volcanoes and channels along the rim. The Viking Orbiter images finally allowed detailed geologic studies of the region to come to fruition.

One final distinction held by the Hellas region is that it is the site of the first manmade object to reach the surface of Mars. In late November 1971, the Soviet spacecraft Mars 2 jettisoned a small capsule that hit the red planet at about lat 44.2° S., long 313.2° W. (Moore, 1977), which is along the western rim of the basin near Hellespontus Montes. Apparently the capsule was intended to be a hard lander and carried only a pennant depicting the coat of arms of the Soviet Union; no scientific data were obtained (Snyder and Moroz, 1992).

PHYSIOGRAPHY

The broad physiography of the map area is defined by topographic mapping at 1-km contour intervals having errors of as much as ±1.5 km (U.S. Geological Survey, 1991). This mapping shows that virtually the entire map area slopes into Hellas basin, which forms the deepest and broadest enclosed depression on the planet (about 9 km of relief; 2,000 km across) in the southern cratered terrain. Heavily cratered terrain ranges from –1 km elevation within the west rim of the basin to as much as 5 km in highland areas surrounding the basin. Individual massifs of the basin rim stand as high as 5 km above surrounding terrain and are more common and extensive on the southeast side of the basin (Leonard and Tanaka, 1993). The west side of Hellas basin is ringed by Hellespontus Montes, a system of mountains marked by high scarps circumferential to the basin.

The basin floor, Hellas Planitia, can be divided into two parts: (1) an annular ring about 1,500 km in diameter and 200–500 km wide made up of plains marked by ridges (including Zea Dorsa) at elevations of –5 to –1 km, and (2) an inner, rugged zone about 1,000 km across consisting of mesas, ridges, scarps, depressions, and hills (including Coronae Scopulus and Alpheus Colles); elevation of this zone is mostly –5 to –4 km, but parts lie as low as –5 km.

Paterae interpreted to be volcanic shields and circular calderas form local topographic highs (having 1 to 2 km of relief) on the basin rim. On the northeast rim are Hadriaca and Tyrrhena Paterae, and on the south rim are Amphitrites and Peneus Paterae. Narrow and broad channels dissect the flanks of these features. In turn, the paterae are surrounded by high plains (Hesperia and Malea Plana) that also are locally marked by channels oriented downslope. These channels include Axius Valles, which dissect the entire south rim of Hellas basin. Extensive wrinkle ridges also cross the paterae and surrounding plains along varying trends.

The large Niger and Dao Valles originate as 1- to 2-km-deep broad depressions along the southeast flank of Hadriaca Patera. Apparently related to these valles are isolated depressions, Ausonia and Peraea Cavi, just northeast of Hadriaca (lat 30° S., long 264.5° W.). The large valles appear to head from these depressions. Southeast of Niger and Dao Valles is another similar channel system—Reull and Harmakhis Valles. Reull Vallis originates in southern Hesperia Planum (northeast of the map area) and traverses rugged Promethei Terra; Harmakhis begins at a 1.5-km-deep depression near the termination of Reull Vallis amid the peaks of Centauri Montes. Lower parts of Niger, Dao, and Harmakhis Valles form braided channel systems incised less than 1 km into the Hellas basin rim. One large fretted channel system, Mad Vallis, also occurs on the south rim of the basin, along the east flank of Axius Valles. Slopes adjacent to all of these large channels are carved by shallower, parallel channel systems.

STRATIGRAPHY

Definition and description of map units and their stratigraphic relations are based on photogeologic observations (including geomorphology, topography, and albedo) from Viking images (see Description of Map Units). Many of the images were manipulated on computer display systems to enhance visualization of the image data. In addition, crater statistics have been measured using digital techniques to provide further analysis of relative ages, including assignment of units into the martian time-stratigraphic classification system (Tanaka, 1986); a compilation of cumulative crater densities for most map units is shown in table 2. For areally extensive units within Hellas Planitia, we performed detailed crater counts down to 2 km diameter. These craters were measured using the Mars Digital Image Mosaic, which has a resolution of 231 m/pixel. For the rim units, we used the crater database of Barlow (1990) that includes all craters larger than 5 km diameter measured from the Viking 1:2,000,000-scale photomosaic series. Extensive obliteration of smaller craters on many units in Hellas Planitia poses problems for the interpretation of cumulative crater densities. Thus, we find that relative ages based on larger crater diameters are more likely to reflect material ages for extensively modified surfaces [see Crown and others, 1992; Tanaka and Leonard (1995), Gregg and others (1998), and Price (1998) for more detail examinations of crater distributions for many of the units]. Crater densities given herein are based on the cumulative number of superposed craters, in which N(*x*) = number of craters larger than *x* kilometers in diameter per million square kilometers.

In some cases, stratigraphic ages assigned by previous workers have been changed where the present data warrant. The entire range of martian series is represented in the map area, from Noachian basement and plateau rocks to Hesperian volcanic and basin-floor deposits to Amazonian surficial materials.

NOACHIAN SYSTEM

The oldest rocks in the map area consist of heavily cratered materials believed to have formed during about the first billion years of the planet's history. Of these, the impact materials of Hellas basin rim are the most ancient (Early Noachian); indeed they are among the most heavily cratered materials on the planet at very large crater diameters (N(32)~80; Tanaka, 1986). These materials include massif (unit Nm) and hilly (unit Nh) units of the Hellas rim assemblage formed by kilometers of uplift of crustal material caused by the Hellas impact. The massif unit consists of large, isolated, resistant blocks, whereas the hilly unit is relatively less rugged and includes uplifted material, basin ejecta, and younger, undifferentiated mantling deposits of uncertain origins. The rim materials are distributed unevenly around the basin; they extend for 1,500 km to the southeast, but are only 300 km across along the west side and mostly buried on the south side. This large-scale rim geometry may be explained by an obliquely impacting bolide (as discussed in a later section). Along the inner part of the east rim, rim material and overlying channel deposits are mapped together as the older dissected unit (unit HNd_1) of the Hellas rim assemblage, because at map scale these two materials are too closely mixed together to be delineated. The channeling may have resulted from the release of ground water near Hadriaca and Tyrrhena Patera during the Late Noachian and Early Hesperian Epochs. Somewhat younger Hesperian channel deposits (Crown and others, 1992) overlie the channel deposits of the older dissected unit; the unit also is combined with older material and forms the younger dissected unit (unit HNd_2).

Rim material of larger impact craters of Noachian age are mapped as older crater material (unit Nc). Noachian craters are heavily degraded, resulting in receded ejecta blankets and worn-down rims. This degradation may have resulted from intensified eolian and fluvial activities that were made possible by an early, relatively thick and warm atmosphere (for example, Craddock and Maxwell, 1993) or by vigorous hydrothermal activity (Brakenridge and others, 1985; Tanaka and others, 1998).

The outer rim of Hellas basin is embayed by the Middle Noachian cratered (unit Npl_1), Middle Noachian dissected (unit Npld), and Upper Noachian subdued cratered (unit Npl_2) units of the plateau sequence. These highly cratered materials are respectively characterized by (1) a dominance of large craters, (2) channel dissection of crater rims and other slopes, and (3) embayed or worn-down crater rims and resurfaced intercrater areas. The subdued cratered material may have resulted mainly from the deposition of alluvium on cratered material and possibly also from local eolian or volcanic resurfacing. Also, many Noachian craters appear rimless, particularly those in the northern map area outside of the Hellas rim units (southwestern Tyrrhena Terra). Rimless craters have very low to no discernible raised rim material at all. The highly degraded state of the highland materials suggests that they are more susceptible to degradation than Hellas basin rim materials, which are likely composed of resistant, crustal material uplifted by the Hellas impact. However, the high relief of the rim materials apparently has accelerated erosion of crater structures, accounting for their low crater densities relative to the cratered and dissected units of the plateau sequence (units Npl_1 and Npld; see table 2).

Centralized volcanism northeast of the basin occurred first at Tyrrhena Patera, which consists of the basal (unit Ntb) and summit (unit Nts) shield members of the Tyrrhena Patera Formation (see detailed geologic mapping and crater statistics of the summit region by Gregg and others, 1998). The summit shield member shows a few concentric ring faults, which in part define a central caldera structure at Tyrrhena. Both basal and summit members are deeply eroded by channels that radiate from the summit region. The channels and overall subdued morphology of the volcano (including an absence of lava-flow fronts) indicate a dominantly pyroclastic origin for the edifice (Greeley and Crown, 1990). Densities of larger craters indicate Late Noachian construction of the volcano (table 2). Shortly thereafter, the flank member (unit HNhf) of the Hadriaca Patera Formation was emplaced and then cut by shallow valleys radiating from the summit and extending about 400 km downslope toward Hellas Planitia. Crater density data (N(5)) indicate a Late Noachian or Early Hesperian age for the cessation of flank development (table 2).

Also during the same period, extensive volcanism resurfaced the southern rim of Hellas basin. Prominent vents include Peneus and Amphitrites Paterae. Peneus consists of a simple caldera structure and no substantial edifice. Amphitrites includes a discernible shield mapped as the shield member (unit HNas) of the Amphitrites Formation. The shield is dissected by radiating channels and is cut by a complex summit caldera. Amphitrites and Peneus Paterae each display a rugged, ring-faulted zone, composed of the rim member (unit HNar) of the Amphitrites Formation, and a partly collapsed floor made up of the floor member (unit HNaf) of the Amphitrites Formation. The rim member at Peneus shows several well-developed, arcuate normal faults and grabens, whereas at Amphitrites, it shows only a few, poorly developed faults. These structures may indicate multiple episodes of magma withdrawal and collapse at the paterae centers. Wrinkle ridges are common on both paterae floor units. The paterae, and perhaps other buried vents, likely were the main sources of the dissected member (unit HNad) of the Amphitrites Formation. This member appears layered in places and is marked by wrinkle ridges and by channels that debouch into Hellas Planitia. In some cases, the channels dissect the wrinkle ridges; in others, the wrinkle ridges postdate the channels. Given that extensive dissection also has affected Hadriaca and Tyrrhena Paterae, the dissected ridged material also may consist of pyroclastic material. Furthermore, the implied distance of runout of the dissected member, given its topographic slope, is amenable with the martian pyroclastic runout model of Crown and Greeley (1993). The channels have little tributary development and thus may have formed by shallow sapping of ground water.

Within Hellas basin, much, if not virtually all, of the floor (Hellas Planitia) was covered by the older ridged plains material (unit HNpr) beginning in the Late Noachian; this material is exposed on the basin floor wherever younger materials are absent. The unit may consist of thin, widespread lava flows that may have originated from vents both within the basin and along the northeastern and southern basin rim areas. The subsequent load of materials probably produced sufficient stress within the basin to generate ridge development. Older ridged plains material also was emplaced about the same time or afterward on Malea Planum, perhaps as lava extruded from Amphitrites and Peneus Paterae or other local vents. This occurrence also includes some pedestal craters south of the map area. The crater distribution indicates that many of the craters several kilometers in diameter and smaller have been obliterated. The Late Noachian/Early Hesperian age assigned to the older ridged plains material of Malea Planum reflects the crater densities of larger craters (table 2; see also Tanaka and Leonard, 1995, tables 3 and 4).

Etched material (unit HNe) occurs as partly degraded plains deposits of varying thickness scattered about on the Hellas basin rim. The unit fills larger craters that include Schaeberle, Terby, and Millochau, and covers local high plains on the rim. Significant occurrences of this material likely are not mapped, because their contacts are not traceable in areas imaged only in low resolution. For instance, a thin (probably a few tens of meters thick) veneer of etched material is evident north of Peneus Patera in high resolution images (Viking images 578B0106,19-23); it probably extends across large parts of northern Malea Planum for which only low-resolution imaging exists. The etched margins and deep depressions of this unit suggest a fine-grained, friable composition consistent with dust, loess, or tephra, possibly including interstitial ice. Some outcrops of the smooth plains units (units Hps and AHps) also fill craters and may be the uneroded equivalent of etched material.

HESPERIAN SYSTEM

The base of the Hesperian System is defined by younger ridged plains material (unit Hpr) in Hesperia Planum (Scott and Carr, 1978), which is partly contained in the northeastern part of the map area and includes ridges of Tyrrhena and Hesperia Dorsa. This unit consists of smooth, lobate, and wrinkle-ridged plains material that has been interpreted by some to be broad, thin lava flows (for example, Greeley and Spudis, 1981). The material embays shield members of Tyrrhena Patera and may have been erupted from local, presently buried fissures.

Post-Noachian crater materials are distinguished from Noachian crater materials in that they generally have preserved ejecta blankets. However, Amazonian versus Hesperian craters cannot be distinguished solely by morphology and therefore are mapped collectively as younger crater material (unit AHc). Most floors of these younger craters appear mostly devoid of infilling and thus are mapped together with their surrounding rim and ejecta as younger crater material. However, the floors of most older and a few younger craters are infilled with material that is generally smooth, with few superposed craters. Crater-fill material locally may comprise fluvial, eolian, volcanic, and (or) mass-wasted sediments. Some craters are infilled with material of an adjacent unit, particularly where the unit embays the crater or breaches the crater rim.

Local, relatively featureless highland areas resurfaced during the Hesperian Period form the smooth unit (unit Hpl_3) of the plateau sequence. The smooth material probably includes volcanic, eolian, and fluvial deposits, which partly bury low areas and crater floors on the Hellas basin rim.

Superposed on the older ridged plains material in Hellas Planitia is a relatively friable and heavily degraded interior deposit, which is divided up into plateau (unit Hipl), knobby (unit Hik), hummocky (unit Hih), and smooth (unit His) interior units of the Hellas Planitia assemblage. The plateau interior unit represents the somewhat smooth and unmodified upper part of the interior deposit and mostly occurs in the eastern half of the deposit. For most outcrops, the ejecta of large craters on this unit appears to have armored the surface from erosion. More deeply eroded parts of the interior deposit are divided into hummocky and knobby interior units based on their morphology. The hummocky interior unit consists of large, somewhat subdued hills superposed on broad rolling topography and occurs more in the southern part of the interior deposit. The knobby interior unit includes a high density of sharply defined knobs as large as a few kilometers across. This unit occurs on either planar or sloping surfaces and within generally more deeply eroded areas in the central and northwestern parts of the interior deposit; Alpheus Colles partly include large areas of the knobby interior unit. Locally, the knobs are elongate and oriented along common trends that suggest either structural control or sculpturing by prevailing winds. Several groups of knobs have been interpreted as yardangs with their long axes correlating well with the strongest winds predicted by a recent Mars global circulation model (Tanaka and Leonard, 1995). The crater density of the mostly uneroded plateau unit probably reflects more closely the emplacement age of the interior deposit, whereas the more highly eroded hummocky and knobby units have lower crater densities commensurate with the timing of erosion (table 2). However, the hummocky and knobby units are generally stratigraphically lower than the uppermost part of the plateau unit.

The smooth interior unit comprises the eastern margin and other parts of the interior deposit. This unit appears somewhat uneroded and has higher crater retention and lower relief than the other interior units. Hence, it likely includes material more resistant to erosion, perhaps coarser grained sediments deposited by channels in the adjacent dissected units (units HNd_1 and HNd_2) on the inner basin slope.

The interior units overlie older ridged plains material (unit HNpr) and are cut by the vallis unit (unit AHv) of the Hellas rim assemblage, indicating an Early Hesperian age. Viking albedo and infrared thermal measurements and atmospheric-circulation models led Moore and Edgett (1993) to conclude that the interior deposits consist of sand and smaller particles and are presently undergoing erosion. Also, dust storms have been observed to originate from Hellas Planitia (Martin and Zurek, 1993), possibly involving dust from the interior deposits or from other surficial materials. The apparently poor lithification of the interior deposits indicates a weak cementing agent such as ice. Types of origins consistent with these surficial observations of the interior units include dust, glacial loess, volcanic airfall, lahars, and lacustrine deposits (Kargel and Strom, 1992; Tanaka and Leonard, 1995). Furthermore, the interior units collectively form a broad, thick plateau, having wind-sculptured features such as hollows and curvilinear scarps, that is similar to the gross morphology of the martian polar layered deposits (Cutts, 1973; Howard, 1978; Cutts and others, 1979; Squyres, 1979). This similarity suggests that the interior units may have formed by a polarlike mechanism of dust and ice deposition within the Hellas topographic low, perhaps because of atmospheric cooling at the end of the Noachian Period.

Hadriaca Patera apparently became active again during the Hesperian Period, producing the summit member (unit Hhs) of the Hadriaca Patera Formation, which fills the summit caldera. This smooth material is less densely cratered than the flank member of the Hadriaca Patera Formation (unit HNhf) and may be Late Hesperian or younger (Crown and others, 1992).

Tyrrhena Patera also became active during the Late Hesperian to Early Amazonian Epochs (Crown and others, 1992; Gregg and others, 1998), at which time large amounts of lava and perhaps pyroclastic materials were deposited to the west and southwest of the summit materials. The plains member (unit AHtp) of the Tyrrhena Patera Formation occurs just north of Ausonia Montes and has few flow lobes and no levees. The margins of this unit are locally etched, suggesting that the member may comprise pyroclastic flows (Greeley and Crown, 1990). Also at this time, a large rille formed on the summit and southwestern flank of Tyrrhena Patera. The rille may have been a conduit for an extensive lava-flow field, mapped as the flank member (unit Htf) of the Tyrrhena Patera Formation. The flank member extends about 1,000 km southwest from Tyrrhena Patera, embays Ausonia Mensa and Ausonia Montes, and terminates near the southeast flank of Hadriaca Patera and near the heads of Dao, Niger, and Harmakhis Valles; theses valles were active at about the same time.

The floor materials of Dao, Niger, Harmakhis, and Reull Valles are mapped as the vallis unit (unit AHv) of the Hellas rim assemblage. Shadow measurements indicate that parts of the channels attain depths of as much as 1 to 2 km. Reull Vallis heads a single channel at the south end of Hesperia Planum (northeast of the map area). From there it winds through Promethei Terra (perhaps along a Hellas basin-ring fault) and appears to debouch near the head of Harmakhis Vallis onto plains of the younger dissected unit (unit HNd_2). The mouth of Reull is buried by apron material from a nearby massif of Hellas Montes. Dao and Niger Valles join into a single channel and, along with Harmakhis Vallis, debouch into Hellas Planitia. Some anastomosing channel development at the mouths of Dao and Harmakhis Valles is evident in these plains. Other, smaller valles cut the southern and western rim of Hellas basin.

During the Late Hesperian and Amazonian, local areas of the Hellas basin rim and surrounding highlands were covered by smooth plains material that may have been emplaced by volcanic, eolian, fluvial, or mass-wasting processes. These deposits, which form the older and younger smooth plains materials (units Hps and AHps, respectively), locally fill impact craters and embay and partly bury older highland and Hellas rim units. The older unit has a higher crater density than the younger unit (table 2). Some margins of the older smooth plains unit appear etched and abrupt whereas others are gradational with adjacent units. Within the interior units of Hellas Planitia, the younger smooth plains unit forms some of the smoothest surfaces in the map area. Locally, the unit includes lobate scarps, indicating that the material may include debris or volcanic flows.

AMAZONIAN SYSTEM

Geologic activity in the map area during the Amazonian Period was greatly reduced and apparently dominated by eolian processes and local mass wasting. Apron material (unit Aa) forms relatively smooth slope deposits and slump features at the base of steep scarps of the massif unit (unit Nm) and the plateau interior unit (unit Hipl) of the Hellas rim assemblage, and crater Barnard (unit AHc); craters are rare (Crown and others, 1992, table 1). Apron material occurs around large massifs that flank Reull Vallis, coinciding with indications that this area may have an anomalously volatile-rich, near-surface layer (Costard, 1989). The surfaces buried by the apron material are mostly Hesperian to Amazonian in age. Within Hellas Planitia, two occurrences of lobate material (unit Al) may be young lava or debris flows.

Surficial deposits in the northern part of Hellas Planitia are made up of patterned sets of small ridges. The arcuate ridge unit (unit Ara) of the Hellas Planitia assemblage occurs in lobate patches and is characterized by low, arcuate ridges forming fingerprintlike patterns. These patterns may be formed by glacierlike creep or moraine deposition of unconsolidated, ice-rich material. The reticulate ridge unit (unit Arr) of the Hellas Planitia assemblage appears to be made up of dunes (Tanaka and Leonard, 1995) that show marked similarity to terrestrial systems of compound crescentic dunes (Breed and Grow, 1979; Walker, 1986). Dune networks of this sort are formed by overlapping systems of dunes, usually consisting of transverse dunes oriented to different (perhaps seasonal) wind regimes (Cooke and others, 1993). Their occurrence is associated with the lowest parts of Hellas Planitia (U.S. Geological Survey, 1991), directions of highest calculated seasonal wind velocities (Greeley and others, 1993), and sites of dust-storm origin (Martin and Zurek, 1993).

BASIN FORMATION AND STRUCTURAL HISTORY

The Early Noachian Hellas impact produced the broad-scale topography and the earliest preserved structures in the map area (N(32)~80, Tanaka, 1986; N(5) > 1,000, Wichmann and Schultz, 1989). Other early impact basins may control the location of Hadriaca and Tyrrhena Paterae and other features in the map region (Schultz, 1984; Schultz and Frey, 1990).

Although as many as seven ring structures have been proposed surrounding the inner basin of Hellas (Pike and Spudis, 1987), the impact's transient cavity probably has been buried by volcanic infilling and other modifications (Spudis, 1993). The main ring is best defined by a 2,200-km-diameter annulus of high-standing rugged massifs and mountain ranges (Pike and Spudis, 1987). Most remaining rings consist of concentric ranges of arcuate massifs (Wilhelms, 1973; Potter, 1976; Peterson, 1977; Pike and Spudis, 1987).

The extensive elevated terrain of massifs southeast of the basin led Wichmann and Schultz (1989) to propose that this area was a zone of enhanced disruption at the site of an earlier impact basin. By contrast, the northwestern rim forms only a narrow, discontinuous band that averages about 300 km wide.

Alternatively, this asymmetry could be due to an oblique impact. Laboratory studies and observations of impact craters on planetary bodies show that obliquely impacting projectiles produce features that are both symmetric about (normal to) their trajectory path and asymmetric along their range path (Gault and Wedekind, 1978; Moore, 1969; Schultz, 1994). The Hellas impact structure displays both types of features along a trend of S. 60° E. (Leonard and Tanaka, 1993; Tanaka and Leonard, 1995). Features symmetric across the proposed trajectory path include (1) broad concentric troughs at distances 1,800 to 2,500 km north and west of the basin (outside of study area); (2) prominent scarps that bound the southeastern region of uplift and are fronted by areas that appear to be structurally lower; (3) large areas between the narrow and broad parts of the rim, which show a marked absence of massifs; and (4) large volcanic paterae, surrounded by vast volcanic plains, which also lie just outside the basin rim where the rim narrows. Asymmetric features along the range path include (1) taller (generally by 1–2 km) and more densely concentrated massifs in the large southeast rim lobe (inferred downrange direction) than in the narrow northwest rim margin; (2) deeper excavation and steeper walls on the uprange profile; and (3) an uprange zone of subarcuate fractures and rim collapse, particularly the extensional faulting that formed the scarps along the margins of Hellespontus Montes along the western rim of Hellas basin. The oblique-impact hypothesis remains speculative, however, because deformation of the lithosphere by oblique impacts at the scale of multiring basins has not been simulated by laboratory experiments or predicted by theory.

Late Noachian basin deformation produced, or perhaps enhanced, the concentric fractures (interpreted to be normal faults) at Hellespontes Montes (N(5)~350; Wichmann and Schultz, 1989). These features may represent reactivated faults created by the Hellas impact event and have been interpreted as results of elastic response of the lithosphere to basin fill equivalent to at least 1.2 km of basalt (Wichmann and Schultz, 1989). The older ridged plains material (unit HNpr) within Hellas Planitia may form such a deposit.

Wrinkle ridges occur throughout much of the map area, but particularly in the ridged plains materials (units HNpr and Hpr) in Hellas Planitia, Malea Planum, and Hesperia Planum. Many of the ridges in Hellas Planitia and Hesperia Planum have roughly basin-radial orientations; those in Hesperia locally form reticulate ridge patterns (Watters and Chadwick, 1989). Most ridges appear to have begun forming in the Late Noachian Epoch (in Hellas Planitia and Malea Planum) and continued forming into the Early Hesperian Epoch (particularly in Hesperia Planum). Wrinkle ridges probably resulted from horizontal contraction caused by subsidence due to loading by plains material or by crustal contraction caused by magma withdrawal and cooling (Solomon and Head, 1979; Raitala, 1988). In addition, global contraction during the Late Noachian and Early Hesperian may have been responsible for more ubiquitous wrinkle-ridge formation (Tanaka and others, 1991; Watters, 1993).

The youngest wrinkle ridges in the map area may be those that deform the flank member of the Tyrrhena Patera Formation. This unit is Late Hesperian. Two orientations of wrinkle ridges crosscut the unit; the northwest-trending ridges appear to be younger than the northeast-trending ridges (Porter and others, 1991).

EROSIONAL HISTORY

The Hellas region has undergone periods of local and regional erosion throughout geologic time. Following the Hellas basin impact in the Early Noachian, basin ejecta was eroded to the extent that ejecta deposits and possible grooves scoured by the impact are not recognized. Instead, resistant, high-standing rim materials that have withstood eons of exposure remain. Other ancient basin rims on Mars generally have had similar erosional histories (Schultz and others, 1982; Edgett, 1988). Later in the Noachian Period, local channeling on highland slopes produced dissected terrains, which may reflect a period of warmer, wetter climate and (or) enhanced hydrothermal and seismic activity relative to later periods (Craddock and Maxwell, 1993; Tanaka and others, 1998).

Erosion of the interior deposit of Hellas Planitia began soon after its deposition in the Early Hesperian Epoch, resulting in the hummocky interior unit. Continued erosion into the Early Amazonian wiped out smaller craters on the plateau and hummocky interior units and produced the knobby interior unit where erosion was deep. Knob orientations indicate that they have been sculptured by the wind because of their alignment with some of the highest wind stresses on the planet as predicted by a Mars general circulation model (Greeley and others, 1993). Also, many landforms within the interior deposit may have a thermokarst origin where interstitial ice and (or) lenses of ground ice are removed to form steep, curvilinear scarps (thermocirques); coalescing, steep-sided, flat-floored valleys (alases); lobate (apparently fluidized) debris aprons; collapse depressions; and chaotic terrain (see Czudek and Demek, 1970). Other local hills, ridges, and scarps may be glacial features, such as drumlins, eskers, and recessional moraines (Kargel and Strom, 1992).

The interior deposit may have once covered the entire floor of Hellas Planitia and was later partly excavated by wind erosion. Locally, the older ridged plains material (unit HNpr) in Hellas Planitia includes superposed patches of the interior units; these remnants sometimes fill exhumed craters. Long, inward-facing scarps forming the basin-side contact of the dissected member of the Amphitrites Formation (unit HNad) may delineate where that unit had been deposited against a former edge of the interior units. Many pedestal craters in the presumed area of retreat armor underlying material from erosion, thereby preserving evidence of a former, more widespread layer of sediments. The interior deposit is largely bound near the north and west edges by prominent escarpments that may have resulted from wind erosion or backwasting caused by volatile loss and disintegration; similar escarpments border parts of the layered deposits at the martian poles. Finally, etched deposits on the Hellas rim (unit HNe) also indicate local wind excavation of possibly similar materials mainly during Hesperian and Amazonian time.

During the Late Hesperian and Early Amazonian Epochs, and perhaps earlier, significant episodes of fluvial erosion carved small channels and larger canyons (valles) along the northeastern and southern rims of Hellas basin. This erosion may have been caused in part by volcanic activity, because many of the channels originate from volcanic materials of Hadriaca and Tyrrhena Paterae and of southern Hesperia Planum. Densities of smaller craters (see table 2) indicate that the Tyrrhena and Hadriaca Paterae both underwent detectable degradation until at least the Late Hesperian. Locally, sparse remnants of eroded layers occur on the paterae flanks. This erosion may have contributed considerable volumes of sediment derived from the northeastern rim of Hellas basin into eastern Hellas Planitia (Crown and others, 1992; Tanaka and Leonard, 1995).

Also, surrounding many isolated massifs, particularly along the eastern rim, are very large aprons of mass-wasted debris. The flowlike morphology and commonly pitted surfaces of these aprons suggest that the material may have been rich in volatiles (Crown and others, 1992).

GEOLOGIC SUMMARY

Early Noachian—Formation of Hellas basin by impact, resulting in uplifted crust that makes up rugged mountains on the basin rim.

Middle Noachian—End of high impact flux; basin rim and surrounding highlands partly resurfaced by impact breccia and other materials.

Late Noachian—Emplacement of lava within basin and on Malea Planum that become deformed by wrinkle ridges; pyroclastic volcanism forms Tyrrhena and Hadriaca Paterae on the northeastern Hellas basin rim and Amphitrites and Peneus Paterae on the southern rim; plateau materials locally modified by fluvial activity.

Early Hesperian—Lava flows marked by wrinkle ridges emplaced in Hesperia Planum; Hellas Planitia mostly filled by eolian and fluvial material; subsequent basin erosion produces steep scarps and hummocky and knobby terrain; small channels modify eastern and southern Hellas rim.

Late Hesperian—Extensive volcanism resulted in lava-flow emplacement southwest of Tyrrhena Patera, which likely triggered development of Niger, Dao, and Harmakhis Valles on northeast Hellas rim; wrinkle ridges on eastern basin floor partly subdued by influx of valles sediments; dust mantled parts of basin floor and rim.

Early Amazonian—Eolian processes continued to deposit and etch dust mantle on Hellas basin rim; smooth materials deposited in local catchments in central Hellas Planitia.

Middle Amazonian—Mass wasting and debris flows formed large aprons around rim massifs and along steep interior basin scarps.

Late Amazonian—Dunes, yardangs, and transient splotches active in Hellas region; interior basin fines continued to be removed by eolian transport.

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REFERENCES CITED

Barlow, N.G., 1990, Constraints on early events in martian history as derived from the cratering record: Journal of Geophysical Research, v. 95, no. B9, p. 14,191–14,201.

Blunck, Jürgen, 1982, Mars and its satellites—A detailed commentary on the nomenclature: Smithtown, New York, Exposition Press, 222 p.

Brakenridge, G.R., Newsom, H.E., and Baker, V.R., 1985, Ancient hot springs on Mars—Origins and paleoenvironmental significance of small martian valleys: Geology, v. 13, p. 859–862.

Breed, C.S., and Grow, T., 1979, Morphology and distribution of dunes in sand seas observed by remote sensing, *in* McKee, E.D. (ed.), A study of global sand seas: U.S. Geological Survey Professional Paper 1052, p. 253–302.

Cooke, R., Warren, A., and Goudie, A., 1993, Desert geomorphology: London, UCL Press Ltd., p. 393.

Costard, F.M., 1989, The spatial distribution of volatiles in the martian hydrolithosphere: Earth, Moon, and Planets, v. 45, p. 265–290.

Craddock, R.A., and Maxwell, T.A., 1993, Geomorphic evolution of the martian highlands through ancient fluvial processes: Journal of Geophysical Research, v. 98, no. E2, p. 3453–3468.

Crown, D.A., and Greeley, Ronald, 1993, Volcanic geology of Hadriaca Patera and the eastern Hellas region of Mars: Journal of Geophysical Research, v. 98, no. E2, p. 3431–3451.

Crown, D.A., Price, K.H., and Greeley, Ronald, 1992, Geologic evolution of the east rim of the Hellas basin, Mars: Icarus, v. 100, no. 1, p. 1–25.

Cutts, J.A., 1973, Wind erosion in the martian polar regions: Journal of Geophysical Research, v. 78, no. 20, p. 4211–4221.

Cutts, J.A., Blasius, K.R., and Roberts, W.J., 1979, Evolution of martian polar landscapes—Interplay of long-term variations in perennial ice cover and dust storm intensity: Journal of Geophysical Research, v. 84, no. B6, p. 2975–2994.

Czudek, T., and Demek, J., 1970, Thermokarst in Siberia and its influence on the development of lowland relief: Quaternary Research, v. 1, p. 103–120.

Edgett, K.S., 1988, Ejecta deposits of large martian impact basins—A useful geologic tool and window to early martian history?, *in* MEVTV Workshop on Early Tectonic and Volcanic Evolution of Mars, Easton, Maryland, 1988: Lunar and Planetary Institute Technical Report Number 89–04, p 32–34.

Gault, D.E., and Wedekind, J.A., 1978, Experimental studies of oblique impact, *in* Proceedings from the Ninth Lunar and Planetary Science Conference, Houston, March 13-17, 1978: Geochimica et Cosmochimica Acta, supplement 10, v. 3, p. 3843–3875.

Greeley, Ronald, and Crown, D.A., 1990, Volcanic geology of Tyrrhena Patera, Mars: Journal of Geophysical Research, v. 95, no. B5, p. 7133–7149.

Greeley, Ronald, and Guest, J.E., 1987, Geologic map of the eastern equatorial region of Mars: U.S. Geological Survey Miscellaneous Investigations Series I–1802–B, scale 1:15,000,000.

Greeley, Ronald, Skypeck, A., and Pollack, J.B., 1993, Martian eolian features and deposits—Comparisons with general circulation model results: Journal of Geophysical Research, v. 98, no. E2, p. 3183-3196.

Greeley, Ronald, and Spudis, P.D., 1981, Volcanism on Mars: Reviews of Geophysics and Space Physics, v. 19, no. 1, p. 13–41.

Gregg, T.K.P., Crown, D.A., and Greeley, Ronald, 1998, Geologic map of part of the Tyrrhena Patera region of Mars: U.S. Geological Survey Miscellaneous Investigations Series I–2556, scale 1:500,000.

Howard, A.D., 1978, Origin of the stepped topography of the martian poles: Icarus, v. 34, p. 581–599.

Kargel, J.S., and Strom, R.G., 1992, Ancient glaciation on Mars: Geology, v. 20, no. 1, p. 3–7.

King, E.A., 1978, Geologic map of the Mare Tyrrhenum quadrangle of Mars: U.S. Geological Survey Miscellaneous Investigations Series I–1073, scale 1:5,000,000.

Leonard, G.J., and Tanaka, K.L., 1993, Hellas basin, Mars—Formation by oblique impact [abs], *in* Abstracts of papers submitted to the Twenty-fourth Lunar and Planetary Science Conference, Houston, March 15–19, 1993: Houston, Lunar and Planetary Institute, p. 867–868.

Lowell, Percival, 1895, Mars: [originally published by] Boston and New York, Houghton Mifflin, 228 p.

Martin, L.J., and Zurek, R.W., 1993, An analysis of the history of dust activity on Mars: Journal of Geophysical Research, v. 98, no. E2, p. 3221–3246.

Mest, S.C., and Crown, D.A., 2002, Geologic map of MTM –40252 and –40257 quadrangles, Reull Vallis region of Mars: U.S. Geological Survey Geologic Investigations Series I–2730, 1:1,000,000 scale [in press].

Moore, H.J., 1969, Subsurface deformation resulting from missile impact, *in* Geological Survey Research, 1969: U.S. Geological Survey Professional Paper 650–B, p. B107–B112.

Moore, J.M., and Edgett, K.S., 1993, Hellas Planitia, Mars—Site of net dust erosion and implications for the nature of basin floor deposits: Geophysical Research Letters, v. 20, no. 15, p. 1599–1602.

Moore, Patrick, 1977, Guide to Mars: New York, W. W. Norton and Company, 214 p.

Peterson, J.E., 1977, Geologic map of the Noachis quadrangle of Mars: U.S. Geological Survey Miscellaneous Investigations Series I–910, scale 1:5,000,000.

Pike, R.J., and Spudis, P.D., 1987, Basin-ring spacing on the Moon, Mercury, and Mars: Earth, Moon, and Planets, v. 39, p. 129–194.

Porter, T.K., Crown, D.A., and Greeley, Ronald, 1991, Timing and formation of wrinkle ridges in the Tyrrhena Patera region of Mars, *in* Abstracts of papers submitted to the Twenty-second Lunar and Planetary Science Conference, Houston, March 18–22, 1991: Houston, Lunar and Planetary Institute, p. 1085–1086.

Potter, D.B., 1976, Geologic map of the Hellas quadrangle of Mars: U.S. Geological Survey Miscellaneous Investigations Series I–941, scale 1:5,000,000.

Price, K.H., 1998, Geologic map of the Dao, Harmakhis, and Reull Valles region of Mars: U.S. Geological Survey Miscellaneous Investigations Series I–2557, scale 1:1,000,000.

Raitala, Jouko, 1988, Superposed ridges of the Hesperia Planum area on Mars: Earth, Moon, and Planets, v. 40, no. 1, p. 71–99.

Schiaparelli, G.V., 1877, Osservazioni astronomiche e fisiche sull'asse di rotazioni e sulla topografia del planete Marte. *In* Atti della Royale Accademia dei Lincei, Memoria della cl. di scienze fisiche. Memoria 1, ser. 3, vol. 2, p. 308–439.

Schultz, P.H., 1984, Impact basin control of volcanic and tectonic provinces on Mars, *in* Abstracts of papers submitted to the Fifteenth Lunar and Planetary Science Conference, Houston, March 12–16, 1984: Journal of Geophysical Research, v. 89, supplement, p. 728–729.

———1994, Chicxulub as an oblique impact, *in* Abstracts of papers submitted to the Twenty-fifth Lunar and Planetary Science Conference, Houston, March 14–18, 1994: Houston, Lunar and Planetary Institute, p. 1211–1212.

Schultz, P.H., Schultz, R.A., and Rogers, John, 1982, The structure and evolution of ancient impact basins on Mars: Journal of Geophysical Research, v. 87, no. B12, p. 9803–9820.

Schultz, R.A., and Frey, H.V., 1990, A new survey of multiring impact basins on Mars: Journal of Geophysical Research, v. 95, no. B9, p. 14,175–14,189.

Scott, D.H., and Carr, M.H., 1978, Geologic map of Mars: U.S. Geological Survey Miscellaneous Investigations Series I–1083, scale 1:25,000,000.

Snyder, C.W., and Moroz, V.I., 1992, Spacecraft exploration of Mars, *in* Kieffer, H. H., Jakosky, B.M., Snyder, C.W., and Matthews, M.S., eds., Mars: Tucson, University of Arizona Press, p. 71–119.

Solomon, S.C., and Head, J.W., 1979, Vertical movement in mare basins—Relation to mare emplacement, basin tectonics, and lunar thermal history: Journal of Geophysical Research, v. 84, no. B4, p. 1667–1682.

Spudis, P.D., 1993, The geology of multi-ring impact basins—The Moon and other planets: New York, Cambridge University Press, 263 p.

Squyres, S.W., 1979, The evolution of dust deposits in the martian northern hemisphere: Icarus, v. 40, p. 244–261.

Tanaka, K.L., 1986, The stratigraphy of Mars, *in* Lunar and Planetary Science Conference, 17th, Houston, March 17–21, 1991, Proceedings: Journal of Geophysical Research, v. 91, no. B13, p. E139–E158.

Tanaka, K.L., Dohm, J.M., Lias, J.H., and Hare, T.M., 1998, Erosional valleys in the Thaumasia region of Mars—Hydrothermal and seismic origins: Journal of Geophysical Research, v. 103, no. E13, p. 31,407–31,419.

Tanaka, K.L., Golombek, M.P., and Banerdt, W.B., 1991, Reconciliation of stress and structural histories of the Tharsis region of Mars: Journal of Geophysical Research, v. 96, no. E6, p. 15, 617–15,633.

Tanaka, K.L., and Leonard, G.J., 1995, Geology and landscape evolution of the Hellas region of Mars: Journal of Geophysical Research, v. 25, no. E3, p. 5407–5432.

Tanaka, K.L., and Scott, D.H., 1987, Geologic map of the polar regions of Mars: U.S. Geological Survey Miscellaneous Investigations Series I–1802–C, scale 1:15,000,000.

U.S. Geological Survey, 1991, Topographic maps of the polar, western, and eastern regions of Mars: U.S. Geological Survey Miscellaneous Investigations Series I–2160, scale 1:15,000,000.

Walker, A.S., 1986, Eolian landforms, *in* Short, N.M., and Blair, R.W., Jr., Geomorphology from space—A global overview of regional landforms: National Aeronautics and Space Administration Special Paper 486, p. 510–515.

Watters, T.R., 1993, Compressional tectonism on Mars: Journal of Geophysical Research, v. 98, no. E9, p. 17,049–17,060.

Watters, T.R., and Chadwick, D.J., 1989, Crosscutting periodically spaced first-order ridges in the ridged plains of Hesperia Planum—Another case for a buckling model (abs.), *in* Watters, T., and Golombek, M.P., eds., MEVTV Workshop on Tectonic Features on Mars: Houston, Lunar and Planetary Institute, p. 36–38.

Wichmann, R.W., and Schultz, P.H., 1989, Sequence and mechanisms of deformation around the Hellas and Isidis impact basins on Mars: Journal of Geophysical Research, v. 94, no. B12, p. 17,333–17,357.

Wilhelms, D.E., 1973, Comparison of martian and lunar multiringed circular basins: Journal of Geophysical Research, v. 78, no. 20, p. 4084–4095.