GEOLOGICAL PROCESSES AND EVOLUTION

J.W. HEAD[1], R. GREELEY[2], M.P. GOLOMBEK[3], W.K. HARTMANN[4], E. HAUBER[5], R. JAUMANN[5], P. MASSON[6], G. NEUKUM[5], L.E. NYQUIST[7] and M.H. CARR[8]

[1]*Department of Geological Sciences, Brown University, Providence, RI 02912 USA*
[2]*Department of Geology, Arizona State University, Tempe, AZ 85287 USA*
[3]*Jet Propulsion Laboratory, 4800 Oak Grove Drive, Pasadena, CA 91109 USA*
[4]*Planetary Science Institute, Tucson, AZ 85705 USA*
[5]*DLR Institute of Space Sensor Technology and Planetary Exploration, Rutherfordstrasse 2, 12484 Berlin-Aldershof, Germany*
[6]*University of Paris-Sud, 91405, Orsay Cedex France*
[7]*Johnson Space Center, Houston TX 77058 USA*
[8]*US Geological Survey, Branch of Astrogeological Studies, 345 Middlefield Road, Menlo Park, CA 94025 USA*

Received: 14 February 2001; accepted: 12 March 2001

Abstract. Geological mapping and establishment of stratigraphic relationships provides an overview of geological processes operating on Mars and how they have varied in time and space. Impact craters and basins shaped the crust in earliest history and as their importance declined, evidence of extensive regional volcanism emerged during the Late Noachian. Regional volcanism characterized the Early Hesperian and subsequent to that time, volcanism was largely centered at Tharsis and Elysium, continuing until the recent geological past. The Tharsis region appears to have been largely constructed by the Late Noachian, and represents a series of tectonic and volcanic centers. Globally distributed structural features representing contraction characterize the middle Hesperian. Water-related processes involve the formation of valley networks in the Late Noachian and into the Hesperian, an ice sheet at the south pole in the middle Hesperian, and outflow channels and possible standing bodies of water in the northern lowlands in the Late Hesperian and into the Amazonian. A significant part of the present water budget occurs in the present geologically young polar layered terrains. In order to establish more firmly rates of processes, we stress the need to improve the calibration of the absolute timescale, which today is based on crater count systems with substantial uncertainties, along with a sampling of rocks of unknown provenance. Sample return from carefully chosen stratigraphic units could calibrate the existing timescale and vastly improve our knowledge of Martian evolution.

1. Introduction and Constraints on Geological Evolution

Geological processes represent the major dynamic forces shaping the surfaces, crusts and lithospheres of planets. They may be linked to interaction with the atmosphere (e.g., eolian, polar), the hydrosphere (e.g., fluvial, lacustrine), the cryosphere (e.g., glacial and periglacial), or with the crust, lithosphere and interior (e.g., tectonism and volcanism) (Figure 1). Geological processes may also be linked to interaction with the external environment (e.g. impact cratering), and may vary in relative importance with time or at any point on the surface of a planet at a given

Space Science Reviews **96:** 263-292, 2001.

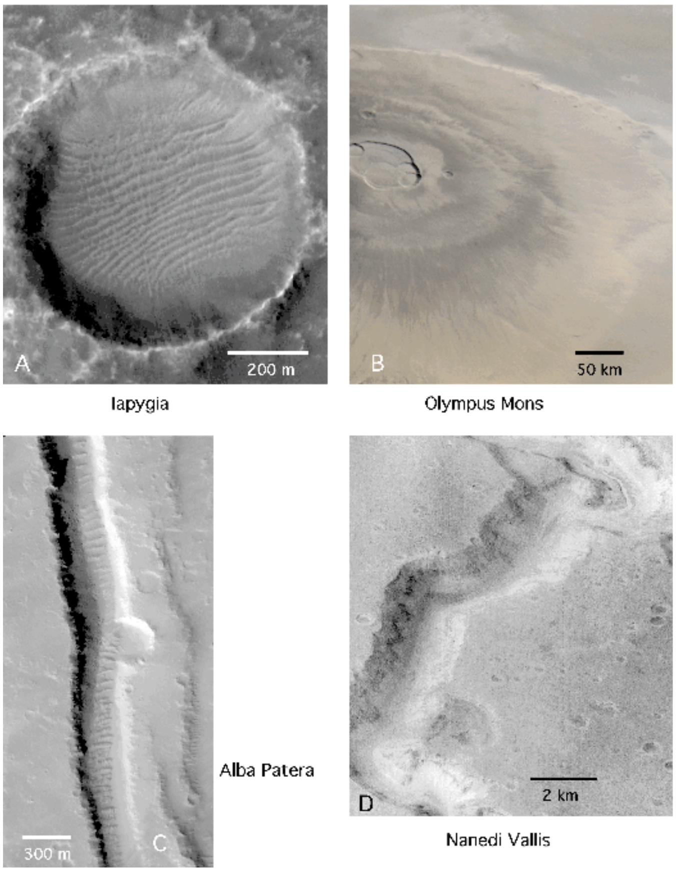

Figure 1. Examples of landforms produced by different geological processes and forming the units of the Martian crust, as seen in very high resolution Mars Orbiter Camera (MOC) images. a) Impact crater modified by dunes. MOC image M10-03652; 279.14W, 16.88S. b) Volcanic activity as exemplified by the 500-600 km diameter volcanic edifice Olympus Mons. MOC images 26301, 26302. c) Tectonic activity as exemplified by Alba Patera graben. MOC image M07-03562; 102.91W, 30.42N. d) Fluvial activity as exemplified by Nanedi Vallis channel in the Xanthe Terra region. MOC image 8704.

time. For example, impact cratering was a key process in forming and shaping planetary crusts in the first one-quarter of solar system history, but its global influence has waned considerably with time.

The geological history of a planet can be reconstructed from an understanding of the products or deposits of these geological processes and how they are arranged relative to one another. The geological history of Mars has been reconstructed through careful mapping using the global Viking image data set to delineate geological units (e.g., Tanaka, 1986; 1990; Tanaka *et al.*, 1992; Scott and Tanaka, 1986; Tanaka and Scott, 1987; Greeley and Guest, 1987), and superposition and cross-cutting relationships to establish their relative ages. Also important in estab-

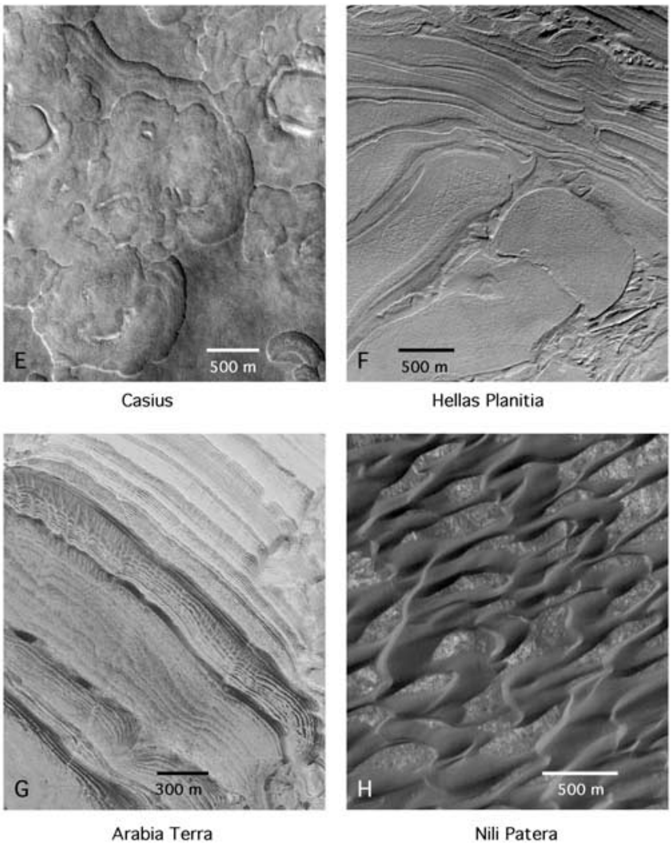

Figure 1. (continued) e) Periglacial features as exemplified by irregular depressions in the northern lowlands. MOC image M04-02077; 268.56W, 45.52N. f) Smooth deposits on the floor of the Hellas basin, interpreted as modified lake deposits. MOC image MOC2-277, 306.7W, 39.7S. g) Layering on a crater floor in western Arabia Terra. MOC image M09-01840; 7.33W, 8.34N. h) Eolian activity as exemplified by a major dune field in Nili Patera in Syrtis Major. MOC image FHA-00451; 292.93W, 8.83N.

lishing relative ages is the number of superposed impact craters on the surface of these units. Together these data have permitted the reconstruction of the relative geological history of Mars, establishment of the main themes in the evolution of Mars (Figure 2), and the relative importance of processes as a function of time. Three major time periods are defined during which the observed surface features and units have been formed: Noachian, Hesperian and Amazonian. The rocks emplaced during these periods form systems and each system is subdivided into series with corresponding time units (epochs).

But the evolution of a planet, the absolute ages of its events, the duration of the periods, and the phases of history cannot really be known without the pinning of the relative chronology to an absolute chronology. This step can be accomplished in several ways: first, if the flux of impact craters is known with certainty, as is generally the case in the lunar environment, then the number of superposed craters can be equated with an absolute age. Secondly, units can be directly dated by returning samples from broad geological units which act as marker horizons. The absolute ages of several of these units constrain the ages of other units, and the crater flux can be calibrated to date other similarly cratered units more confidently. Due to the lack of samples from Mars whose context and provenance are known, the assignment of absolute ages to the epochs based on crater densities depends on models to estimate cratering rates. Different earlier models have placed the Amazonian-Hesperian boundary between $1.8 - 3.5$ Gyr and the Hesperian-Noachian boundary between $3.5 - 3.8$ Gyr (Tanaka *et al.*, 1992; Neukum and Wise, 1976; Hartmann, 1978; Hartmann *et al.*, 1981; Neukum and Hiller, 1981). Ages associated with Hartmann tended to be younger, and those associated with Neukum, older. As discussed below, Hartmann and Neukum (2001; Figure 14) have revisited this problem, placing the divisions between the Periods and Epochs generally within the ranges suggested by the earlier literature, but tightening the age constraints.

2. Geological Processes and their Importance with Time

2.1. Impact Cratering

Impact cratering dominates the character of Mars, especially the heavily cratered, old uplands (Figure 1; Strom *et al.*, 1992; Smith *et al.*, 1999a; Kreslavsky and Head, 1999, 2000; Aharonson *et al.*, 2001), and several large basins (Hellas and Argyre) dominate the topography of that part of the planet. Impact cratering causes the vertical excavation and lateral transport of crustal material, and the ejecta deposits in younger craters provide important information on the nature of the substrate and of the cratering process. Impact craters have also provided geothermal sites due to heating and impact melt emplacement, penetrated the cryosphere to release groundwater, and served as sinks for ponded surface water (e.g., Carr, 1996; Cabrol and Grin, 2001).

Impact cratering is also important in terms of chronometric information. Papers in Part I of this volume review the methodology. Using this system, Hartmann and Neukum (2001) show that appreciable areas of late Amazonian young lavas and other units have ages in the range of a few 100 Myr, agreeing with martian meteorite ages, while, at the other end of the time scale, most of the Noachian dates before 3.5 Gyr ago.

Hartmann and Neukum (2001) use the Tanaka *et al.* (1987) tabulation of areas (km^2) resurfaced by different geological processes in different epochs, to graph the *rate* of resurfacing by those processes as a function of time. A consistent result from those diagrams is that many endogenic processes, including volcanic and fluvial resurfacing, show much stronger activity before roughly 3 Gyr ago, and decline, perhaps sharply, to a lower level after that time. Those results support

Mars: Major events in geological history

	Volcanism	Tectonism	Fluvial events	Cratering	Erosion and surficial processes
AMAZONIAN	• Late flows in southern Elysium Planitia. • Decreased volcanism in northeren plains. • Most recent flows from Olympus Mons. • Emplacement of massive materials at S. edge of Elysium Planitia. • Waning volcanism in Tharsis region. • Waning volcanism in Elysium region. • Widespread flows around Elysium Mons.	• Tharsis tectonism continued through the Amazonian, mostly associated with the large shield volcanoes • Formation of Elysium Fossae. • Initial formation of Olympus Mons aureoles.	• Channeling in southern Elysium Planitia. • Late period of channel formation. • Formation of channels NW of Elysium Mons.		• Emplacement of polar dunes and mantle. • Development of polar deposits? • Formation of ridged lobate deposits on large shield volcanoes. • Emplacement of massive materials at S. edge of Elysium Planitia. • Local degradation and resurfacing of northern plains. • Erosion in northern plains. • Deep erosion of layered deposits in Valles Marinaris. • Developmant of ridges, grooves, and knobs on northern plains.
HESPERIAN	• Volcanism at Syrtis Major. • Formation of highland paterae. • Volcanism at TempeTerra. • Major volcanism in Elysium and Tharsis regions. • Emplacement of ridged plains (Hr).	• Formation of Noctis Labyrinthus. • Formation of Valles Marinaris. • Formation of wrinkle ridge systems. • Memnonia and Sirenium Fossae, fractures around Isidis.	• Development of large outflow channels. • Infilling of northern plains. • Deposition of layered materials in Valles Marinaris.		• Degradation of northern plains materials. • Dorsa Argentea formation at South Pole. • Resurfacing of northern plains.
NOACHIAN	• Formation of intercrater plains. • Decreasing highland volcanism. • Beginning of widespread highland volcanism.	• Ceraunius, Tempe, and Noctis Fossae. • Tectonism south of Hellas. • Archeron Fossae. • Claritas Fossae.	• Formation of extensive valley networks.	• Waning impact flux. • Intense bombardment. • Argyre impact. • Hellas and Isidis impacts. • Formation of oldest exposed rocks.	• Extensive dessication and etching of highland rocks. • Formation and erosion of heavily cratered plateau surface. • Deep erosion of basement rocks.

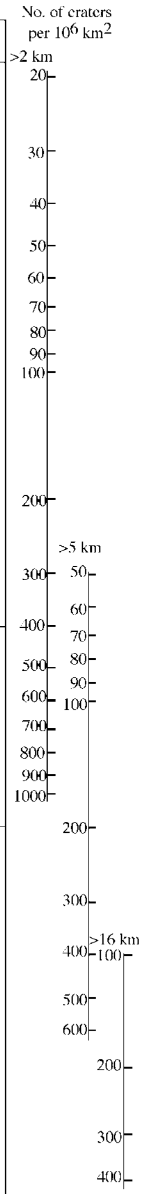

Figure 2. Geologic features formed and processes operating during the major periods of Martian history, the Noachian, Hesperian, and Amazonian. Derived from Scott and Tanaka (1986), Tanaka and Scott (1987) and Greeley and Guest (1987).

many suggestions (Jones, 1974; Soderblom *et al.*, 1974; Chapman, 1974; Hartmann, 1971, 1973; Neukum and Hiller, 1981) that Martian geological processes were more active in the early part of Martian history, and declined, perhaps rapidly, at the end of that period. The results also show the great advance to be made if the dates of the stratigraphic epochs could be measured with lower uncertainties, since the detailed history of various processes (Figure 2) could then be deciphered.

2.2. VOLCANISM

Mariner 9 provided the first clear evidence for the importance of volcanic processes in the history of Mars (McCauley *et al.*, 1972; Carr, 1973). Images revealed shield and dome volcanoes of the Tharsis and Elysium regions, extensive lava plains, and numerous other features of volcanic origin, including low-profile contructs, called patera, characterized by central craters and radial channels. Subsequently, data from the Viking Orbiters (Carr *et al.*, 1977b) allowed mapping and characterization of the extent, timing, and styles of volcanism on Mars (Figure 1; Greeley and Spudis, 1978, 1981; Mouginis-Mark *et al.*, 1992; Greeley *et al.*, 2000). High resolution images (Malin *et al.*, 1998) (Figure 1), information on surface compositions (McSween *et al.*, 1998), and topographic data (Smith *et al.*, 1998, 1999a, b) from the Mars Global Surveyor (MGS) permit comparison of Martian volcanism with theoretical analysis of the ascent and eruption of magma on Mars (e.g., Wilson and Head, 1994).

Although much of the ancient crust of Mars is likely to be of volcanic origin, obvious morphological features, such as flows and structures which might be vents, are not seen in images of the oldest martian terrains. Nonetheless, by analogy with the early history of the Earth and the Moon (Stöffler and Ryder, 2001) and from models for Mars (Spohn *et al.*, 2001), magmatic activity was probably involved in the formation of the crust. Moreover, Thermal Emission Spectrometer (TES) data suggest primarily basaltic compositions for the martian highlands where most of the ancient crust is found (Christensen *et al.*, 2000). Subsequent geological processes, including those associated with wind and water, have modified the highlands so extensively that morphological traces related to early putative volcanism are not readily found with currently available data.

The first recognized volcanic features are the paterae in the Hellas region, e.g. Tyrhenna, Hadriaca, Amphitrites, and Peneus Paterae (Greeley and Spudis, 1978). The first two have been observed in sufficient detail to derive their volcanic history. Early eruptions seem to have involved magmas rising through water-rich megaregolith (Crown and Greeley, 1993), leading to extensive phreatic-magmatic activity and the emplacement of ash shields, the presence of which is suggested by the style of erosion on the flanks of the patera. These eruptions were apparently followed by effusive activity, emplacing complex sequences of flows which radiate from central calderas.

Alba Patera is a central vent structure covering more than 4.4×10^6 km^2, making it one of the largest volcanoes in the Solar System. It is composed of tube-and-channel fed flows and flows emplaced as massive sheets. It contains a caldera some 100 km across, the floor of which includes small cones of probable spatter and pyroclastic origin (Cattermole, 1987).

The most impressive volcanoes on Mars are in the Tharsis and Elysium regions, where more than a dozen major constructs exist, including classic shield volcanoes. High-resolution images show that these volcanoes were built from countless individual flows, many of which were emplaced through channels and lava tubes, signaling a style of volcanism analogous to Hawaiian eruptions (Greeley, 1973). Complex summit calderas suggest multiple stages of magma ascent and withdrawal (Crumpler *et al.*, 1996). The flanks of many of the volcanoes are marked by terrain which appears to have failed by gravitational collapse, leaving mass wasted deposits covering hundreds of square kilometers in some cases. Some of the deposits in the Elysium region have a morphology suggestive of lahars, i.e., to have been emplaced as water-rich slurries (Christensen, 1989). The subdued appearance of some of the Elysium volcanic summits has been proposed as pyroclastic material, suggesting Plinian styles of eruption (Mouginis-Mark *et al.*, 1988).

By far the greatest areal extent and inferred volume of volcanic materials on Mars are found in the various plains, the youngest of which show flow fronts and embayment into older terrains (Greeley and Schneid, 1991). These materials likely represent high-volume flood eruptions. The extensive ridged plains, typified by Hesperia Planitia, are more difficult to interpret. Characterized by "wrinkle" ridges, these units resemble mare basalts on the Moon. TES data suggest basaltic compositions in contrast to the younger northern plains which have more siliceous composition (Christensen *et al.*, 2000), suggesting magma evolution with time.

In addition to these major volcanoes and plains units, a wide variety of smaller volcanoes and volcanic features are recognized on Mars, including possible composite cones found in the highlands, small shield volcanoes (many of which have associated flowlike channels), and fields of small cones with summit craters, similar to pseudocraters in Iceland, which result from local phreatic eruptions as lavas flow over water-logged ground (Frey *et al.*, 1979).

In summary, volcanic processes appear to have operated throughout the history of Mars and could possibly be active today (Hartmann and Berman, 2000; Keszthelyi *et al.*, 2000). Although from the morphologies of the flows and most of the volcanoes basaltic compositions are inferred to be dominant, remote sensing data from orbit and measurements made on the surface by Mars Pathfinder suggest that slightly more evolved magmas might also have been present in the evolution of the surface (see discussion in Bibring and Erard, 2001).

The generation of magma and its extrusion onto the surface through volcanism is a direct indicator of interior activity. Age, location, and type of volcanic materials give insight into the evolution of the interior. Volcanism has dominated much of the history of Mars. Nearly half of its surface has materials inferred to be of volcanic origin (Greeley and Spudis, 1978; Mouginis-Mark *et al.*, 1992; Greeley *et al.*, 2000). These materials form either central volcanoes, such as the shields in the Tharsis region, or vast plateaus formed by the eruption of flood lava flows. From geological mapping of the areal extent of various volcanic deposits through time, coupled with estimates of the thicknesses of the deposits, the volumes of volcanic materials produced on Mars. Greeley and Schneid (1991) assumed a ratio of intrusive to extrusive materials of 8.5 to 1 (based on values for the Earth) and estimated a total magma volume of 6.54×10^8 km^3. For the 3.8 Gyr age-span of the volcanic materials analyzed, this gives a magma production rate of $\sim$0.17 km^3 yr^{-1}. Note that uncertainties in the derivation of these estimates include poorly constrained values for thicknesses and the fact that the intrusive to extrusive ratio might be different on Mars. The values based on photogeologic mapping can be refined through the application of MGS data, including topographic information.

The recent identification in Mars Observer Camera (MOC) images of extensive thin layering in the walls of Valles Marineris suggests that volcanism in the Noachian has been much more extensive than previously recognized (McEwen *et al.*, 1999). If these layers are flood lavas, as suggested by their morphology, the volume of material increases the magma volume discussed above by an order of magnitude. Furthermore, if most of this volcanism is Noachian in age as suggested by the mapping (McEwen *et al.*, 1999), a peak in the volcanic output occurred in the Noachian, with a general decrease with time.

2.3. Tectonism

The morphology of the martian surface as observed in imaging data provides ample evidence for tectonic processes (e.g., Carr, 1981; Banerdt *et al.*, 1992; Figure 1). Brittle failure of the crust and the lithosphere is indicated by a variety of structural features, both extensional (simple graben, complex graben, rifts, tension cracks, troughs, and polygonal troughs) and compressional (wrinkle ridges, lobate scarps). While the relative ages of the features and, therefore, the processes responsible for their formation may be dated by structural mapping and crater counts, additional information on topography and gravity is required to model loads and to derive stresses in the lithosphere (Golombek and Banerdt, 1999).

Global-scale processes like thermal contraction cooling, despinning, or polar wandering cause stresses in the lithosphere of a planet. While such processes may have operated on Mars, tectonic evidence for them is weak or ambiguous (Banerdt *et al.*, 1992). The global dichotomy is the most fundamental physiographic feature on the planet and formed early in Martian history. Exogenic processes (i.e. one or several mega-impacts, Wilhelms and Squyres, 1984; Frey and Schultz, 1988) have been invoked to account for it, but recent Mars Orbital Laser Altimeter (MOLA) investigations (Smith *et al.*, 1999a, b) did not find any single or several large circular topographic depressions to confirm this hypothesis (except for the Utopia basin, which had been speculated to be an impact basin even in the pre-MGS era; McGill, 1989). Endogenic processes seem to offer attractive alternatives, and a variety of convection or subduction mechanisms has been proposed. A plate- tectonics scenario has been suggested to explain many of the surface features in and around the northern lowlands (Sleep, 1994). Unfortunately, little photogeologic evidence supports the suite of features that would be expected to be observed at the various plate tectonic boundaries defined in Sleep's model (Pruis and Tanaka, 1995; Tanaka, 1995). Detailed structural mapping of key locations (e.g. the dichotomy boundary) required to further test the hypothesis is underway with new MGS data (Turcotte, 1999). An ancient phase of plate tectonics has also been proposed to explain a group of magnetic anomalies in a portion of the cratered highlands (Connerney *et al.*, 1999), although many alternate hypotheses are being considered to explain these features.

Most graben on Mars are narrow (few km wide) and long ($\sim 10-100$ km) structures bounded by inward dipping normal faults (Banerdt *et al.*, 1992). Wider (up to 100 km) and deeper structures (many km), more analogous to rifts on the Earth that rupture the entire lithosphere, are found in Tempe Terra, Valles Marineris, and Thaumasia. Linear to arcuate asymmetric topographic highs ("wrinkle ridges") are the most common compressional structures and form patterns of distributed compressional deformation (e.g., Chicarro *et al.*, 1985; Watters and Maxwell, 1986; Watters, 1988, 1993; Plescia and Golombek, 1986; Golombek *et al.*, 1991). Schultz and Tanaka (1994) have reported a system of large compressional ridges and buckles with greater strain in Noachian terrain to the south and southwest of Tharsis, and topographic data of the northern plains show a system of ridges generally concentric to Tharsis (Smith *et al.*, 1999a; Head *et al.*, 2001).

Regional-scale deformation seems to be responsible for most of the observed tectonic features. While minor faulting patterns are associated with the Elysium volcanic province and some impact structures, by far the dominant element in martian tectonics is the Tharsis bulge (e.g. Carr, 1981, Banerdt *et al.*, 1982, 1992; Golombek and Banerdt, 1999). In particular, a plethora of extensional structures (simple and complex graben, rifts, and troughs) radiate outwards from Tharsis. Additionally, Tharsis is the center of a concentric pattern of compressional structures (wrinkle ridges) (Lucchitta and Anderson, 1980; Maxwell, 1982; Chicarro *et al.*, 1985). Several processes have been proposed to explain the formation of the huge topographic bulge: Domal uplifting (e.g. Phillips *et al.*, 1973; Carr, 1973; Hartmann, 1973), magmatic intrusion (Sleep and Phillips, 1979, 1985; Willeman and Turcotte, 1982), and volcanic loading (Solomon and Head, 1982). For reviews of Tharsis stress models see Carr (1974b), Banerdt *et al.* (1982; 1992), Sleep and Phillips (1979), Solomon and Head (1982), Mége and Masson (1996a), Golombek and Banerdt (1999), and Banerdt and Golombek (2000). Prior to the MGS mission the formation of the extremely large system of graben (radial graben occur both on the elevated flanks of Tharsis *and* in regions outside Tharsis; Tanaka *et al.*, 1991) was explained with models that require two distinct stress states. The then available topographic and gravity data implied isostatic conditions to generate the

TABLE I

Major stages and locations of magmatic and tectonic activity in Tharsis (after Anderson and Dohm, 2000).

	Stratigraphy[1]	Age[2] (Gyr before present)	Centers of local and regional activity
Stage 1	Noachian	4.65–3.7	**Claritas**, Tempe, Ascraeus
Stage 2	Late Noachian - Early Hesperian	3.8–3.6	**Valles Marineris, Thaumasia**, Warrego Valles
Stage 3	Early Hesperian	3.7–3.6	**Syria NW**, Tempe, Ulysses, Valles Marineris
Stage 4	Late Hesperian - Early Amazonian	3.6–2.1	**Alba**, Tempe NW, Tempe SE
Stage 5	Middle - Late Amazonian	2.1–0	**Pavonis**, Tharsis Montes

(1) after Tanaka (1986)
(2) after Hartmann and Neukum (2001) (this volume)

graben on the flanks, while flexure was invoked to explain those farther out. However, the considerable improvement in the topography (Smith and Zuber, 1999) and gravity (Zuber and Smith, 1999) fields provided by MGS changed that view: Recent models of the lithospheric support of Tharsis based on topography, gravity, and structural mapping suggest that flexure alone can explain the structural pattern (Banerdt and Golombek, 2000). Many radial graben may also be the surface manifestation of dikes propagating from magmatic centers in central Tharsis (Wilson and Head, 2000; 2001).

The overall scheme of Tharsis as being the center of volcano-tectonic activity on Mars throughout most of its history is complicated by the fact that there are several local and regional sub-centers within Tharsis (Anderson and Dohm, 2000). A regional variation of Tharsis-related deformation is also in agreement with new modeling results (Banerdt and Golombek, 2000) which attribute the extensional faults *within* Tharsis (which previously seemed to be incompatible with pure loading models) to such variations unresolved in the pre-MGS data sets. Such centers of extensional as well as compressional features seem to have changed in space and time. Several studies tried to decipher the tectonic record in and around Tharsis as provided by surface faults (e.g. Wise *et al.*, 1979a, b; Plescia and Saunders, 1982; Watters and Maxwell, 1983, 1986; Tanaka and Davis, 1988; Golombek, 1989; Tanaka, 1990; Tanaka *et al.*, 1991; Schultz, 1991; Scott and Dohm, 1990; Mége and Masson, 1996b; Schultz, 1998; Dohm and Tanaka, 1999). Statistical analyses of recent hemispheric-scale structural mapping (Anderson *et al.*, 1998) indicate five measurable successive stages of volcano-tectonic activity within Tharsis (Table I) (Anderson and Dohm, 2000).

To summarize Martian tectonism, mapping of geologic and tectonic activity (e.g., Carr, 1974a, 1974b; Wise *et al.*, 1979a, 1979b; Plescia and Saunders, 1982; Scott and Dohm, 1990; Tanaka *et al.*, 1991; Banerdt *et al.*, 1992; Frey, 1979) within the stratigraphic framework of Mars (Tanaka, 1986, 1990; Scott and Tanaka, 1986, Tanaka and Davis, 1988, Tanaka, 1990; Dohm and Tanaka, 1999) has revealed a complex structural history involving five stages of tectonic activity with changes in the derived centers of activity through time (Anderson *et al.*, 2000). More than half of the structures mapped on Mars are Noachian in age (stage 1; $>3.8 - 4.3$ Gyr), concentrated in exposures of Noachian age crust exposed in Tempe Terra, Ceraunius Fossae, Syria Planum, Claritas Fossae, Thaumasia, and Sirenum. By Late Noachian-Early Hesperian (stage 2) activity was concentrated in Thaumasia and Valles Marineris. Middle Hesperian (stage 3) included the development of concentric wrinkle ridges concentrated along the edge of the topographic rise, although wrinkle ridge development may have continued later due to global compression (e.g., Tanaka *et al.*, 1991). Normal faulting during this time also occurred north of Alba, in Tempe Terra, in Ulysses Fossae, in Syria Planum and Valles Marineris, and in Claritas Fossae and Thaumasia. Stage 4 activity during the Late Hesperian-Early Amazonian was concentrated in and around Alba Patera and Middle to Late Amazonian activity (stage 5) was concentrated on and around the Tharsis Montes volcanoes with additional activity in Elysium Planitia along Cerberus Rupes. All of these events produced radial graben centered at slightly different locations (local centers of volcanic and tectonic activity) within the highest standing terrain of Tharsis, indicating that the basic loading pattern of Tharsis has changed little since the Middle Noachian (Anderson *et al.*, 2000).

Lithospheric deformation models based on MGS gravity and topography appear to have simplified the stress states required to explain most of the tectonic features around Tharsis (Banerdt and Golombek, 2000). Flexural loading stresses based on present day gravity and topography appear to explain the type, location, orientation and strain of most tectonic features around Tharsis (Banerdt and Golombek, 2000). These models require the load to be huge (of the scale of Tharsis and thus large relative to the radius of the planet) and the mapping requires the load that caused the flexure to have been in place by the Middle to Late Noachian ($>3.8 - 4.3$ Gyr), with minor changes since that time. Fine layers within Valles Marineris revealed by the Mars Orbiter Camera have been interpreted as being lava flows that are Late Noachian or older (McEwen *et al.*, 1999). These volcanics are therefore likely the load that caused the flexure around Tharsis. This enormous load appears to have produced a flexural moat, which shows up most dramatically as a negative gravity ring, and an antipodal dome that explains the first order topography and gravity of the planet (Phillips *et al.*, 2001). New altimetric data provide evidence that the circum-Tharsis mid-Hesperian wrinkle-ridge system extends into the northern lowlands, a radius of 7000 km from the center of Tharsis (Head *et al.*, 2001). Many ancient fluvial valley networks, which may have formed during an early warmer and wetter period on Mars, flowed down the present large-scale topographic gradient, further arguing that Tharsis loading was very early (Phillips *et al.*, 2001). It is in fact possible that the formation of Tharsis actually produced this early warmer and wetter environment. If the load is composed of volcanics as suggested by fine layers within Valles Marineris (McEwen *et al.*, 1999), water released with the magma would be equivalent to a global layer up to 100 m thick, which could have had a significant impact on the martian climate (Phillips *et al.*, 2001).

2.4. Fluvial Features and Processes

Mariner 9 images in 1972 first showed giant channels and smaller branching valley networks on Mars, and Viking provided more detail indicating many flow-like features (Figure 1), which seemed to be formed by running water (e.g., Masursky *et al.*, 1977; Baker and Kochel, 1979; Pieri, 1980; Carr, 1981). Present-day conditions on Mars preclude liquid water flowing on the surface, however, liquid water may have existed on the surface in the past (e.g., Sharp and Malin, 1975; Carr, 1981, Mars Channel Working Group, 1983; Baker *et al.*, 1992). Alternatives to water for the outflow channels have been proposed (lava flows, winds, debris flows and liquid hydrocarbons; Carr, 1974a, b; Schonfeld, 1976; Cutts and Blasius, 1981; Nummedal, 1978; Nummedal and Prior, 1981; Yung and Pinto, 1978), but none of these theories explains the formation of the associated flow-like features as readily as water (e.g., Baker *et al.*, 1992; Carr, 1996; Lucchitta and Anderson, 1980). Most workers account for the observed channel features as due to the catastrophic release of groundwater under stable, metastable or non-stable surface conditions for liquid water. Masson *et al.* (2001) give a more thorough discussion of fluvial features such as outflow channels and valley networks.

The valley networks are almost entirely restricted to the old uplands and the simplest explanation is that the valleys are old themselves and the climatic requirements for valley formation were met early in the planet's history and rapidly declined during the subsequent evolution. A warmer, wet Mars with a dense atmosphere at the time after the heavy bombardment is supposed to provide the conditions for valley formation by running water (Malin and Edgett, 1999). Based on the evaluation of high resolution MOC images Malin and Edgett (1999; Carr and Malin, 2000) concluded that the valleys were formed by fluid erosion, and in most cases the source was ground water.

A detailed review on the current knowledge about the ages of the fluvial features is also given by Masson *et al.* (2001). We only refer to a few results here: The relatively small size of Martian valleys, the modification by aeolian processes, and restrictions by the resolution and coverage of existing images makes it difficult to use crater counting for age determination. Thus, only a small number of channels and almost no valley networks are dated by this method. For instance, Neukum and Hiller (1981) have estimated crater retention ages of the floors of particular channels and their significantly older surroundings. The number of craters in the surroundings and the number representing a younger resurfacing period translates to crater model ages of a few Gyr. Neukum and Hiller (1981) also show a sequence of the circum-Chryse valleys with most of the Kasei Vallis floors being oldest followed by the mouth of Maya Vallis, Ares Vallis, Tiu Vallis and the mouth of Kasei as the youngest unit in this area. Channels on volcanoes are too small for crater counting on their floors but they dissect volcanic units (for example at the flanks of Alba Patera) and are considered to be as young as a few 100 Myr (Gulick and Baker, 1990). In general, although valley networks are found on units that range in age from Noachian to Amazonian (Scott and Dohm, 1992; Carr, 1995), they are predominantly in the oldest units (Noachian). Formation of outflow channels appears to have peaked in the late Hesperian.

The distribution of channels in time can also be analyzed using relative superposition and intersection relationships (Ivanov and Head, 2001). For example, Nelson and Greeley (1999) mapped the circum-Chryse channels in detail and found the youngest units of different channels, ranging from Mawrth and Ares Vallis oldest, followed by Tiu and Simud Vallis to Shalbatana Vallis and the youngest channel units exposed at the mouth of Kasei Vallis, indicating that the last flood events of each channel were younger from east to west around Chryse.

Relatively young small-scale alcove-like gullies combined with small channels and aprons in the walls of impact craters, south polar pits and two of the larger valleys indicate even more recent groundwater seepage and probably short-term surface runoff under almost current climatic conditions (Malin and Edgett, 2000). These are discussed in more detail by Hartmann (2001).

2.5. Periglacial and Glacial Landforms and Processes

Freezing and thawing of the ground and the presence of permanently frozen ground or permafrost (Tricart, 1968; 1969) are the characteristics of periglacial surfaces on Earth. A number of Martian surface features seen in Mariner 9 and Viking Orbiter imagery have been attributed to periglacial or permafrost processes (Figure 1). Although there are analogies between terrestrial periglacial features and Martian landforms it has to be kept in mind that these landforms are the result of freeze-thaw cycles in the active layer above the permanently frozen ground which is not possible under the present climatic conditions on Mars (e.g. Carr, 1996). In addition, the dimensions of most of the periglacial-like features are one or more orders of magnitude larger than those on Earth. All types of terrain softening are mostly seen in a latitudinal belt between 30° and 60°. A detailed description of these landforms as geomorphologic evidence for liquid water on Mars and many references are given in the accompanying paper by Masson *et al.* (2001).

One of the periglacial landforms are the *lineated valley fills* which to some extent resemble terrestrial median moraines on glaciers. However it is not clear to what level transverse versus longitudinal flow contributed to the generation of the lineations. They are found at the upland/lowland boundary on Mars, where lobate debris aprons with well defined flow fronts and convex-upward surfaces extend from the highlands and from isolated mesas into the low-lying plains (Mangold and Allemand, 2001; Carr, 2001). In valleys, opposing scarp walls confine material flows and the flow fronts converge, resulting in ridges and grooves, commonly called lineated valley fill. In some places, terrain softening appears also in craters where material has obviously moved down the inner crater wall and probably forms *concentric crater fill*. Another periglacial landform on Mars is the viscous flow of ice-lubricated debris, also compared to *rock glaciers*. These features are flow-like but the rheology and composition of the ductile material is not clear (e.g. ice versus CO_2 clathrate hydrate; Hamlin *et al.*, 2000). The ejecta of so-called *rampart craters* on Mars are marked by distinct lobes and provide strong evidence for subsurface ice or water (Strom *et al.*, 1992; Mouginis-Mark, 1979; Barlow, 1988; Schultz and Gault, 1979). The lobes seem to flow around pre-existing obstacles and are outlined by a low ridge or rampart. Viking imagery showed fractured plains marked by giant *polygons* 30 km across, thought to originate from the cooling and fracturing of ice rich sediments and/or self-compaction and accommodation to the underlying terrain, or the desiccation of wet sediments deposited by water from the giant outflow channels (see review in Hiesinger and Head, 2000). Polygons of much smaller sizes (diameters of $10 - 100$ m) are also observed on Mars. They are interpreted to be the result of thermal contraction in ice-rich soils based on the similar scales as compared to terrestrial ice-wedge polygons (Mellon and Jakosky, 1995).

Other periglacial or glacial landforms on Mars are *thermokarsts, frost mounds, eskers*, and *moraines* (Masson *et al.*, 2001, and references therein). Among these, the most convincing morphological evidence for ancient glaciation on Mars are the eskers (e.g. Carr *et al.*, 1980), which are long sinuous ridges mainly found in mid and high latitudes. It seems most plausible that those features formed analogously to terrestrial eskers, as a result of infillings of ice-walled river channels by subglacial, englacial, or supraglacial drainage networks. Recent MOLA measurements of the heights and widths of the largest Martian features are consistent with the esker hypothesis (Head, 2000a; b). Kargel and Strom (1990) argued that only glaciation can account for features covering wide areas in Argyre, Hellas, and the south polar region in a simple and unifying way.

2.6. Lakes and Oceans

Numerous craters in the uplands of Mars show flat floors and channels entering the crater, and apparently sediment was deposited there (Figure 1). Numerous workers have proposed that these regions were the sites of standing bodies of water or lakes (e.g., Carr, 1996). Recent MOC data show evidence for layered deposits in many impact craters, suggesting that standing bodies of water occurred in these locations (e.g., Malin and Edgett, 2000). Such locations have been cataloged and are abundant and widespread (e.g., Cabrol and Grin, 2001). Lake sediments are also thought to characterize the floor of Valles Marineris (Lucchitta *et al.*, 1992).

Large outflow channels emptied into the northern lowlands and their floods (e.g. Lucchitta *et al.*, 1986) must have left extended deposits, large lakes, and possibly oceans. The water released by the outflow channels into the northern plains is estimated to amount to at least 6×10^6 km^3 but probably more (Carr *et al.*, 1987).

Baker *et al.* (1991) calculated that 6×10^7 km^3 of water was needed to fill up the northern plains. The northern plains cover an older rougher terrain that survived as hills and knobs commonly outlining old pre-plains impact craters. The plains are complex deposits probably formed by many processes, such as sedimentation from outflow channels, volcanism and mass wasting from the adjacent highlands modified by impact craters. Recent analyses using altimetry data suggest that the present surface deposits (mostly the Hesperian-aged Vastitas Borealis Formation) are underlain by early Hesperian ridged plains of volcanic origin, and the overlying Vastitas Borealis Formation is at least 100 meters thick (Head *et al.*, 2001). Many of the detailed surface features seen in MOC images seem to be recent and formed by action of ground ice and debris mantles (see Kreslavsky and Head, 2000).

Some features surrounding the northern plains form contacts interpreted as ancient shorelines (Parker *et al.*, 1989, 1993) marking the boundaries of former lakes or a northern ocean. Analysis of the elevations of the contacts and surface roughness using MOLA data show that the oldest contact interpreted to be a shoreline varies widely and is not a good candidate for an equipotential line. The younger contact also shows variations but is closer to an equipotential line (Head *et al.*, 1999). In addition, the plains inside these contacts are extremely smooth. Thus, some of these data are consistent with the presence of a large standing body of water, while others are not (Head *et al.*, 1999).

If there were standing bodies of water in the northern lowlands in the past, under current climate conditions such lakes or oceans would rapidly freeze and form an ice cover (e.g., Kargel *et al.*, 1996; Clifford and Parker, 1999). However they could stabilize if fed by meltwater or groundwater (McKay and Davis, 1991). Outflow channels, their source regions, and termination areas in the northern lowlands, provide the best evidence for surface water on Mars and a widespread groundwater system. The outflow channels could form by running water under current climatic conditions because the volume and flux of released water can change the atmosphere sufficiently to prevent freezing and sublimation at least during the relatively short periods of flooding (Carr, 1979, 1995; Gulick *et al.*, 1997). Again, for this topic we also refer to Masson *et al.* (2001).

2.7. Aeolian Processes

Wind processes dominate the current Martian environment in the absence of known or abundant active volcanism, tectonism, and running water on the surface (Greeley *et al.*, 1993). Despite the tenuous CO_2 atmosphere (average surface pressure is 6.5 mb), winds are capable of setting large quantities of dust into motion, at times obscuring the surface from view from orbit. A century of earthbased observations as well as surface monitoring from orbit show that aeolian processes, such as dust storms, occur predominantly in the southern hemisphere summer. However, data obtained from the Viking orbiters and MGS, and measurements made by the Pathfinder and Viking landers show that aeolian activity can also occur at other seasons.

Particles are transported by the wind in three modes: suspension (dust particles, which on Mars are a few μm in diameter), saltation (which involves sand, or grains $\sim 0.6 - 2$ mm in diameter), and creep (grains larger than a few mm), depending upon wind strength. According to Greeley *et al.* (2001), fine sand $\sim 80 - 100$ μm in diameter is most easily moved by the lowest strength winds.

A wide variety of wind related features is seen from orbit and from landers (Figure 1; Greeley *et al.*, 2001). These include wind depositional features, such as dunes, and wind erosional features, including sculpted hills termed yardangs. The most common aeolian feature seen from orbit are various albedo patterns (such as wind streaks) which change with time, while common features seen on the surface include deposits associated with rocks, called wind tails. Both wind tails and wind streaks are considered to represent the prevailing winds at the time of their formation and can be mapped to infer wind directions.

Wind streaks occur in several forms, including bright features which appear to be stable over periods of years and dark streaks, some observed to have changed in as little as 38 days (Greeley *et al.*, 2001). Many bright streaks are thought to be dust deposited in the waning stages of dust storms, while dark streaks appear to result from erosion of windblown particles, exposing darker substrate, or leaving a lag deposit of lower albedo material on the surface. General circulation models (GCMs) of the atmosphere show that bright streaks correlate with predicted regional wind directions (Greeley *et al.*, 1993). Dark streaks appear to be influenced more by local topography than by regional wind patterns.

Images from MGS (Edgett and Malin, 2000) show that mantles of windblown materials, inferred to be dust deposits settled from suspension, can form both bright and dark surfaces. Similarly, dunes and duneforms also occur as both bright and dark features. These observations suggest a variety of source material for windblown sand and dust, including weathered and unweathered mafic rocks.

As long as Mars has had an atmosphere and loose particles on its surface, it is likely that aeolian processes have operated. Deposition of windblown sediments to form mantling blankets provides a mechanism of "resurfacing", while aeolian deflation of sediments has the potential to exhume formerly buried surfaces. The resurfacing and exhumation involve depths up to at least a few km (Malin and Edgett, 2000). Greeley *et al.* (2001) emphasize that resurfacing and exhumation must be taken into account in estimating surface ages based on impact crater statistics.

On Earth, eolian features and deposits provide clues to past climates. A similar potential exists on Mars. For example, recent analysis of duneforms, windstreaks, and eroded craters seen from orbit and ventifacts studied on images from the Mars Pathfinder lander suggest the presence of a paleowind regime (Kuzmin *et al.*, 2001). Because eolian activity is dependent on the wind regime, potential periods of more active eolian processes in the past could signal the presence of a higher density atmosphere because the threshold winds for sand and dust transport would be lower. Thus, analyses of present and past eolian activity can give new insight into the evolution of the Martian surface and climate.

3. Integrated Geological History and Relation to the Evolution of Mars

Initial differentiation, core and crustal formation occurred very early in the history of Mars prior to the time of the visible record seen in presently preserved surface geological units as confirmed from geochemical/geophysical data by Halliday *et al.* (2001). The magnetic field persisted for a long enough time period to form the magnetic fabric of the southern uplands, apparently prior to the Hellas and Argyre basins (Connerney *et al.*, 1999). The Noachian Period (Figure 2) was the time of significant impact bombardment modifying this crust, the formation of large basins, and the emplacement of highland plains in lows and plateaus. The origin of these smooth plains is uncertain because of poor preservation and mantling, but it is likely that many of them are of volcanic origin because of the higher thermal fluxes typical of the earlier history of Mars.

The northern lowlands formed during the early part of the Noachian, but the nature of its origin (external impacts or internal processes) still remains to be determined. Any theory for formation of the northern plains must reconcile the major geological differences between the northern lowlands and the southern uplands as well as distinctive differences in average crustal thickness (e.g., Zuber *et al.*, 2000). One possibility is that the lowlands result from thinning associated with magma ocean cumulate overturn (Hess and Parmentier, 2001).

Recent analyses of the northern lowlands suggests the presence of early Hesperian ridged plains lying below the late Hesperian sedimentary veneer of the Vastitas Borealis Formation (Frey *et al.*, 2001; Head *et al.*, 2001). Detection of numerous large, shallow basins in the northern plains (Frey *et al.*, 2001) suggests that the early Hesperian rigded plains are no more than 1 km thick and underlain by an early Noachian heavily cratered surface.

Tectonic activity in the Noachian is seen in the form of tectonic structures south of Hellas and in Acheron and Claritas Fossae. There is also growing evidence that the Tharsis rise underwent considerable construction and tectonic activity during the Noachian and may have been almost fully developed in shape and scale by the end of the Noachian. Emplacement of ridged plains of apparent volcanic origin had begun by the late Noachian. Extensive erosion of the heavily cratered terrain occurred during this time and extensive blankets of debris were emplaced at high southern latitudes.

A fundamental question about Mars is whether it was warm and wet in its early history. Evidence for water on the surface may be seen in the form of the extensive debris mantles and the presence of valley networks. Originally thought to represent the result of precipitation and surface runoff, valley networks are now considered by a number of researchers to have formed as a result of groundwater sapping processes. If this interpretation is correct, then it places the formation of the aquifer from which the sapping originates further back in the history of Mars. Presently, there is no compelling direct surface geological evidence to support a warm, wet early Mars.

The Hesperian Period (Figure 2) marks a critical transition from the Noachian early history characterized by high thermal fluxes and heavy cratering rates, to the later low volcanic and impact fluxes of the Amazonian. During this period some of the most important geological features and processes occurred, but ironically, we have little information about its duration, with estimates ranging from a few hundred million years to over a billion years.

Volcanic activity was extremely widespread in the Early Hesperian. Smooth plains, subsequently deformed by wrinkle ridges (mapped as Hr, Hesperian, ridged plains), form numerous regional patches such as Hesperia Planum and are testimony to global volcanism, followed by regional to global contractional deformation. Recent work suggests that the northern lowlands were also resurfaced by Hr (ridged plains) during the Early Hesperian (Head *et al.*, 2001), bringing the total resurfacing to over 40% of the planet. Individual volcanic edifices and flows are rare in these early deposits, but several centers of volcanism with unusual structure and morphology (e.g., Tyrrhena, Apollinaris, and Hadriaca Paterae) suggest that rising magma may have interacted with ground ice and ground water to produce explosive eruptions (e.g., Greeley and Spudis, 1981). Volcanic activity continued in the later Hesperian with the emplacement of Syrtis Major deposits, which differ from the Early Hesperian ridged plains in the preservation of flow fronts. A significant part of the volcanic activity during the later Hesperian shifted from broad plains to central-vent volcanism at Alba Patera, in Tharsis and in Elysium (Hecates Tholus, Albor Tholus).

Water is one of the hallmarks of the Hesperian Period. Midway through the Hesperian, the Dorsa Argentea Formation (Tanaka and Scott, 1987) was emplaced around the present south polar region. This fragmental and apparently ice-rich mantling material shows evidence of being an ice sheet that underwent meltback and drainage into surrounding low areas such as the Argyre and Hellas basins (Head and Pratt, 2001). The cause of the meltback, the fate of this water and the influence it had on the subsurface groundwater table (e.g., Clifford, 1993) are not presently known. There is evidence that volcanic eruptions interacted with this extensive ice sheet, causing melting and drainage (Ghatan and Head, 2001).

During the middle and later parts of the Hesperian, the valley networks characterizing the late Noachian period (and extending into the Hesperian, but in reduced number) transitioned to the major outflow channels. Numerous channels were emplaced along the southern uplands- northern lowlands boundary, emerging from the uplands and debouching into the lowlands. Accompanying this period of channel formation was the modification and retreat of the boundary scarp, particularly in the Deuteronilus Mensae area. Channels also entered the Hellas basin from the Hesperia Planum region. Evidence exists that the Hellas and Argyre basins were volcanically resurfaced in the Early Hesperian, and that sediments of fluvial, lacustrine and eolian origin were later emplaced on their floors.

A key question is the fate of the water involved in the formation of the outflow channels. Did the water pond in the northern lowlands to produce large bodies of water (e.g., the oceans of Parker *et al.*, 1989; 1993) or was there insufficient water to create large standing bodies? If there were large quantities of water, what was its fate? Did it persist for millennia in a liquid form, did it quickly soak back into the groundwater table, or did it rapidly freeze and sublime? Was there an ocean in the northern lowlands prior to the Hesperian? If so, when and for how long? The answers to these questions are not presently available and a key element is the duration of this period of channel formation, a number that is very poorly constrained (Figure 3). The Vastitas Borealis Formation (Tanaka and Scott, 1987), a Late Hesperian fragmental unit overlying the ridged plains (Head *et al.*, 2001), is temporally equivalent to the outflow channel deposits. A likely scenario is that the sediments of the Vastitas Borealis Formation are part of the deposits of the outflow channels, perhaps a residue remaining after the sublimation of a water-sediment mixture emplaced during one or more channel-forming events.

The Amazonian Period (Figure 2) saw the continuing emplacement of volcanic units such as the Elysium Formation, a veneer of volcanic material radial to Elysium Mons, late-stage flows from Tharsis Montes and other Tharsis volcanoes, and flow units emplaced on the upper margins of the northern lowlands (the Arcadia Formation). Volcanism apparently continued in the Elysium region until very recently, as very young flows in the Elysium-Marte Valles region have been detected (Keszthelyi *et al.*, 2000). Most volcanic activity in the Amazonian was associated with the major rises, Elysium and Tharsis, where it continued throughout this period. Outflow channel activity and even some valley networks continued into the Amazonian. Volcanic activity in the Elysium region was associated with significant quantities of groundwater. Lahar-like channels and deposits are seen debouching into the Utopia Basin, while to the east, Amazonian fluvial channels are closely related to the very recent volcanism with crater retention ages of the order 10 Myr or less (Hartmann and Berman, 2000; Hartmann and Neukum, 2001). Minor valley networks are seen on the upper flanks of Alba Patera.

In the Late Amazonian, the polar layered terrain was emplaced and modified. The crater retention ages for these deposits suggest a very young age (less than a few Myr; Herkenhoff and Plaut, 2000). There is a major hiatus between these very young Amazonian polar deposits and underlying Hesperian-aged materials at both poles (Dorsa Argentea Formation at the South Pole and the Vastitas Borealis Formation at the North Pole). Does the formation of these Late Amazonian deposits represent a major change in the atmospheric environment of Mars where it continued throughout this period? Were similar deposits formed and destroyed at several times in the Amazonian during the hiatus? Important to the understanding of these questions is a firm measurement of the duration of the Amazonian; current uncertainties in Martian cratering rates (Ivanov, 2001) mean that the duration could range from $\sim$1.8 Gyr to $\sim$3.5 Gyr. Hartmann and Neukum (2001), using the best estimates of crater rate, place the duration at $\sim$2.9 to 3.2 Gyr.

Late-stage Amazonian Period activity includes abundant eolian modification of the polar deposits (Tanaka and Scott, 1987; Fishbaugh and Head, 2000) and formation of a latitudinally distributed belt of mantling material (e.g., Squyres and Carr, 1986; Kreslavsky and Head, 2000), probably related to depositional variations accompanying obliquity cycles.

4. Outstanding Chronological Questions and Measurement Requirements

Two independent systems for dating Martian units complement each other, as discussed earlier in this volume by Nyquist *et al.* (2001), Neukum *et al.* (2001), and Hartmann and Neukum (2001). Dating of rock samples is believed to be relatively exact (barring any unknown systematic errors), but in the absence of sample return missions, we have samples only from unknown locations on Mars; thus, their provenance, context, and relation to geological units and history are unknown.

The cratering method has the advantage of being able to date any chosen terrain of sufficient area, but at present the uncertainties are large. As discussed by Ivanov (2001) the main uncertainty is in the ratio of Mars/moon cratering rates, used to calibrate the system. The formal uncertainties may be of the order 20% (in terms of the estimated asteroid impact rate, crater scaling, etc), but a more realistic uncertainty may be a factor 2 (taking into account unknown effects of comets, various main belt resonances, etc). This uncertainty propagates directly into uncertainty on age. This, in turn, means that features with crater retention ages less than, say, 200 Myr are probably formed in the last 10% of Martain time and have value in terms of geological and geophysical processes. But a crater retention age of, say 2 Gyr, is less useful in geophysical terms, because the true age might lie between 1 and 4 Gyr (Figure 3). At the oldest extreme, a crater retention age around 4.0 Gyr is relatively constrained because the high crater densities are associated only with high cratering rates before about 3.5 Gyr.

Currently, Martian meteorites provide the only basis for an absolute chronology of Martian processes. Accepting the Martian origin of the meteorites, methods must be developed to generalize observations about them to observations about Mars. Attempts to do so are hampered to variable degrees by lack of knowledge of the geologic settings from whence the meteorites came. Interpretations of some types of observations, such as the composition of Martian atmospheric gases trapped in the meteorites, are independent of the location on Mars from which the meteorites came. Interpretations of other types of observations, such as the mineralogy of the meteorites, benefit from and inform the knowledge of the general environment in which rocks parental to the meteorites crystallized, but do not require exact knowledge of those environments. Interpretations of still other types of observations, such as calibrating the Martian cratering rate, would benefit greatly from exact knowledge of the places of origin of the meteorites. Some aspects of the radiometric age data for Martian meteorites fall into all three of these categories.

Because the places of origin of the Martian meteorites are unknown, use of their ages for calibrating the cratering rate is distinctly limited. Nevertheless, the observation of young igneous crystallization ages, down to $\sim$165 Myr, among the meteorites shows that Martian volcanism continues essentially until the present day. The observation of a high proportion of young ages moreover suggests that Mars has been volcanically relatively active at recent times: 10 of 15 meteorites for which radiometric ages are summarized in Nyquist *et al.* (2001) have ages $<$500 Myr, 8 of those have ages in the range $\sim 165 - 185$ Myr. Only 1 in 15 has an age older than 1.3 Gyr. However, there are three outstanding questions: (a) Do the Martian meteorites give a statistically reliable sampling of Martian surface ages? (b) If not, how many Martian surface areas really have been sampled? (c) What are the potential bias factors in sampling them?

Because $\sim$40% of the Martian surface is highlands belonging to the oldest stratigraphic unit, the Noachian, whereas only 1 of 15 ($\sim$7%) of the meteorites has a correspondingly old radiometric age, the answer to the first question appears to be negative. This conclusion invites comparison to the lunar case. Of 20 lunar meteorites, 7 (35%) are from mare areas. The mare surfaces of the Moon represent about 17% of the lunar surface, thus the highlands again appear to be underrepresented by about a factor of two. In both cases there is a need for a better statistical sampling, but there is an implication that some bias factor may discriminate against older, brecciated, and therefore more friable samples, as has been suggested by a number of authors.

However, the grouping of Martian meteorite ages around certain preferred values emphasizes the need to correct for “launch-pairing” among them. If the crystallization ages of the meteorites are used to group the meteorites, the number of apparent ejection events is reduced to 4–5, and the number of Noachian age events (1) is the statistically expected number. If instead, the cosmic ray exposure ages are used, there are seven apparent ejection events, also statistically acceptable. The two approaches are combined by Nyquist *et al.* (2001). Better understanding will be facilitated via acquisition of new data and better statistics. If the lower number of events implied by the crystallization ages is confirmed, it may imply that secondary collisions in space contribute to the distribution of cosmic ray exposure ages, increasing their apparent number relative to the actual number.

The shock metamorphic histories of the meteorites can make a more important contribution to determining launch conditions and pre-launch sample locations than perhaps has been recognized. The currently known Martian meteorites appear to populate a “launch window” of peak shock pressures in the interval $\sim 15 - 45$ GPa. Qualitatively, this appears to be systematically higher than peak shock pressures experienced by the lunar meteorites, for example. This observation needs to be quantified, and its implications assessed for models of meteorite ejection, as well as possible strength-related biases in meteorite yields. Variations in peak shock pressures among launch-paired samples can give information on the relative depths at launch for the samples.

Finally, reliable launch-pairing of the meteorites will enable better interpretation of a variety of mineralogical, geochemical and isotope geochemical data obtained for the samples. Current data provide a strong suggestion that some mantle-derived basaltic magmas have assimilated Martian crustal materials (see Nyquist *et al.*, 2001, among others). Geochemical and isotopic variations among some of the meteorites appear to reflect variable degrees of this crustal contamination. If the assimilation processes can be adequately described, it may be possible to infer the geochemical processes of a trace-element-enriched Martian "crustal" component, which has not otherwise been sampled. This objective requires the correct grouping of samples for investigation of candidate assimilation processes and scenarios.

Finally, it must be noted that the degree of reliability currently achieved for radiometric dating of lunar samples has required the experience and improvement in laboratory techniques acquired over two decades. Not all of the problems encountered in dating these samples have been analytical. Martian rocks, as evidenced by Martian meteorites, bear the record of a complex series of processes, both primary igneous processes and secondary processes (weathering, deposition of secondary minerals, cf. Bridges *et al.*, 2001). The return of actual Martian samples to terrestrial laboratories is likely to be required to answer many of the outstanding questions of Martian chronology. Doubtless they, too, will hold surprises for unwary analysts.

For the above reasons, it is critical to determine absolute ages for several broad, homogeneous Martian surface units. This could be done by sample return, and even crude (20–50% uncertainty) in situ methods would be valuable, but the surface unit must be chosen carefully to reduce the error bar in ages of currently unknown units.

To calibrate the system in this way, Noachian units are less useful; their high crater densities already give some confidence that they formed before about 3.5 or 3.7 Gyr ago (see Hartmann and Neukum, 2001). The cratering rate is believed to have been changing rapidly before that time, so units with a wide range of crater densities may have about the same age, just as on the Moon. Sampling of regional and globally widespread units such as Hesperian-aged ridged plains (Hr) would provide absolute chronological information on the key transition between the Noachian and the Hesperian (Figure 2). The very widespread nature of this unit and the well preserved craters and crater ages would have the benefit of providing a global datum somewhat analogous to the Imbrium basin ejecta on the Moon. Also extremely useful would be dating of units near the Hesperian/Amazonian boundary (e.g. to address the question does this boundary fall at 1.8 Gyr, 3.5 Gyr, or somewhere in between) and broad units, such as lava flows, that are clearly in a specific Amazonian epoch. Widespread and accessible volcanic units of the Tharsis region offer such opportunities. Interestingly, a sample return from such a unit may seem less desirable on other scientific grounds, such as the search for water or for evidence of life. For example, there is much interest in testing environmental and biological conditions represented by Noachian samples. However, examination of Figure 2 shows that exploration of Mars requires a coherent scientific strategy which must include the establishment of a firm global absolute chronology. Too narrow a focus could result in no improvement over the present understanding of the absolute age scale for the geologic history of Mars, and thus no improved context for the rates and timing of fundamental geologic processes such as volcanism, river and lake formation and tectonism. Therefore, sample return missions and in situ dating methods, need to include assessment of these critical relative time markers.

In summary, the geologic history of Mars is characterized by rock sequences (geologic units) which are derived from the topographic, geomorphologic and spectral characteristics of remotely sensed surface features (Tanaka *et al.*, 1992). The stratigraphic position of geologic units is estimated by the means of superposition and intersection and by the concentration of impact craters superposed on geologic units. Due to the lack of samples from Mars of known provenance and context, the assignment of absolute ages to the epochs based on crater densities is dependent on cratering rates (Tanaka *et al.*, 1992; Neukum and Hiller, 1981; Hartmann, 1978; Neukum and Wise, 1976) and thus is model dependent. Different models define the Amazonian-Hesperian boundary between 1.8–3.5 Gyr. and the Hesperian-Noachian boundary between 3.5–3.8 Gyr. (Hartmann *et al.*, 1981; Neukum and Wise, 1976; Neukum and Hiller, 1981).

Hartmann and Neukum (2001) have reduced the formal uncertainty and disagreement between their two systems, but they and Ivanov (2001) emphasize that the ages are still proportional to 1/(Mars cratering rate) which is still uncertain by as much as a factor of two. These large uncertainties in the Martian absolute chronology prevent a correct interpretation of the initial stage and duration of geologic processes such as volcanism, tectonism, erosion, formation of channels and valley networks, glaciation and resurfacing by wind and their implications for the climate evolution. The next stage of the exploration of Mars must strive to resolve these fundamental uncertainties.

Acknowledgements

This paper is dedicated to the memory of David H. Scott, whose careful and diligent work and unassuming sharing of his knowledge has significantly helped to provide a first-order understanding of the geological evolution of Mars. We also thank Elizabeth R. Fuller, Mikhail Ivanov, Anne Cote, William Hartmann, Baerbel Lucchitta, Ken Tanaka, Reinald Kallenbach, and Ursula Pfander, for contributing directly and indirectly to the preparation and review of this contribution.

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Address for correspondence: James W. Head, Department of Geological Sciences, Brown University, Providence, RI 02912 USA; (james_head_iii@brown.edu)