GEOLOGIC MAP OF THE TEMPE-MAREOTIS REGION OF MARS

By Henry J. Moore

INTRODUCTION

A principal motivation for mapping the Tempe-Mareotis region is to examine the low volcanic shields and lava plains that represent a style of volcanism unlike the styles of volcanism of the vast ridged plains and the large Tharsis Montes volcanoes. The goal is to place the volcanism in a spatial and temporal context with other volcanism on Mars, especially the volcanism that produced the Tharsis Montes volcanic chain (Arsia, Pavonis, and Ascraeus Montes). The Tempe-Mareotis region lies along a great circle that passes through the Tharsis shield volcanoes and Uranius Patera (fig. 1, see map sheet for all figures). Also of interest is the predominantly extensional tectonism that preceded, accompanied, and followed the volcanism. The geologic record of the region is complicated by the interplay of volcanic, tectonic, and fluvial processes.

The Tempe-Mareotis region has been the subject of volcanic (Plescia, 1980, 1981; Hodges, 1980) and tectonic (Scott and Dohm, 1990; Davis and others, 1995; Golombek and others, 1996) studies. The geology of this region has been mapped at 1:25,000,000 (Scott and Carr, 1978), 1:5,000,000 (Wise, 1979), and 1:15,000,000 (Scott and Tanaka, 1986) scales.

Six 1:500,000-scale controlled photomosaics (U.S. Geological Survey, 1991a–f) were combined and then reduced to 1:1,000,000 scale to form the base for this map. The 1:500,000-scale photomosaics and Viking Orbiter images were used extensively to prepare this map. Image quality varies because of resolution and, perhaps, atmospheric conditions (in parts of the northwest corner of the region). Two sets of images (448B and 299A) that are not included in the photomosaic have resolutions of 8–12 m/pixel. The 448B images provide useful, but limited, coverage from about lat 34.5° N., long 82.2° W. to about lat 36.9° N., long 81.8° W., but the 299A images are not useful because of hazy atmosphere. Problems with the photomosaic base that do not seriously affect the results of this study include: (1) mismatches of images from different orbits, which result in offsets or even missing topographic features, and (2) adjoining images with differing illumination directions.

GEOGRAPHIC SETTING

The Tempe-Mareotis region lies along the northeast trend of the Tharsis Montes volcanic chain, which includes four large, superposed volcanoes: Arsia, Pavonis, and Ascraeus Montes and Uranius Patera (fig. 1). The region also includes the elevated western margin of Tempe Terra and adjacent plains to the west (fig. 2; U.S. Geological Survey, 1985, 1989). In the north-central part of the region, terraced highlands are transected by the deep, north-south canyons of Tanais Fossae along the western margin of Tempe Terra. Farther west, isolated, irregular- to northeast-trending mountains (Tanaica Montes and Gonnus Mons) rise above the western plains, and Enipeus Vallis, a channel system, stretches 480 km from south to north across the western plains. Near the east-central part of the region, Baphyras Catena comprises three contiguous elongate abysses aligned northeast; the center depression, about 38 km long and 14 km wide, is nearly 1.7 km deep. In the southeast part of the region, the elevated, flat-topped ridges and intervening valleys of Tempe Fossae trend northeast and attain lengths that exceed 200 km. The plains of Ascuris Planum separate the terraced highlands to the north from the Tempe Fossae ridge-valley system to the south. Small, conical hills and low, broad ridges, having elliptical bases and, in places, summit depressions, are common features on southwestern Ascuris Planum, on patches of plains within the Tempe Fossae ridge-valley system, and on the western plains. These small hills and low ridges are interpreted to be shield volcanoes. The shield volcanoes are typically 2–10 km across, but Issedon Tholus (near the west-central edge of the region at lat 36.3° N., long 94.7° W.) is 50–60 km across. Several large volcanoes surrounding the map area include (1) an unnamed volcano described by Hodges and Moore (1994) about 190 km to the east in central Tempe Terra (near lat 37.6° N., long 75.9° W.); (2) Alba Patera, a large shield, about 600 km to the west (at lat 40.5° N., long 110.0° W.); (3) Uranius Patera, a large shield, about 400 km to the south (at lat 26.2° N., long 92.9° W.); and (4) older volcanoes to the east (Wise, 1979; Plescia and Saunders, 1979; Scott, 1982; Hodges and Moore, 1994). The volcano mapped by Scott and

Tanaka (1986) near lat 44° N., long 79° W. is questionable.

Topographic information (U.S. Geological Survey, 1989) indicates elevations along the south edge of the map area rise from 2,000 m in the east to 4,000 m in the west and elevations along the north edge are near 3,000 m. Photoclinometric estimates indicate the west edge of Tempe Terra rises up to 1,780 m above the floor of Tanais Fossae to the west (Davis and others, 1995). Relief of shield volcanoes about 5 km across have been estimated using photoclinometry (table 1; Davis and Soderblom, 1984; Tanaka and Davis, 1988) to average about 90 meters and to range from a few tens of meters to about 200 meters. The relief of larger shield volcanoes, such as Issedon Tholus, may be near 1,000 meters.

STRATIGRAPHY

The three rock-stratigraphic systems (Scott and Carr, 1978; Tanaka, 1986; Tanaka and others, 1992a) are represented in the Tempe-Mareotis map area, but the age assignments of some units are uncertain because resurfacing processes have altered crater size-frequency distributions and because the limited areal extents of the units counted result in large statistical uncertainties. Relative ages from crater size-frequency distributions are in general agreement with those from superposition relations and some previous assignments to systems and series. Ejecta from some impact craters are intercalated with volcanic deposits. Ages of significant units within the map area range from Late Noachian to Early Amazonian.

NOACHIAN SYSTEM

The Noachian System comprises the oldest map units. Five material units are mapped: (1) terraced plateau material (unit Nplt), (2) Reykholt crater material (unit Ncr), (3) lineated plateau material (unit Npll), (4) undivided material (unit Nu), and (5) degraded crater material (unit Nc). On this map, terraced plateau material (unit Nplt) includes the basement complex (unit Nb), hilly and cratered units of the plateau sequence (units Nplh and Npl_1), and part of the lower member of the Tempe Terra Formation (unit Htl) of Scott and Tanaka (1986) because the four units cannot be separated using stratigraphic evidence. For example, surface morphologies are the same on either side of the contact between hilly plateau material and cratered plateau material (units Nplh and Npl_1) as mapped by Scott and Tanaka (1986) between lat 40.0° N., long 85.5° W. and lat 37.5° N., long 85.5° W. Assignment of terraced plateau material (unit Nplt) to the Noachian is based on cumulative frequencies of craters with fresh to degraded morphologies within this region (fig. 3*A*). However, fluvial resurfacing of terraced plateau material (unit Nplt) is demonstrated by a fluvial channel that cuts into the southeast flank of crater Reykholt at lat 40.4° N., long 85.3° W. Furthermore, cumulative frequencies of craters greater than 2 km and 4 km that are superposed on the fluvial erosion surface of the terraced plateau (unit Nplt) and lineated plateau (Npll) materials indicate resurfacing near middle Hesperian (fig. 3*B*). Reykholt crater material (unit Ncr) is probably Noachian because Reykholt is a highly degraded impact crater having two or more superposed degraded craters, its flanks were eroded during the Hesperian, and the large diameter of Reykholt favors (but does not prove) an older age. Lineated plateau material (Npll) is interpreted to be ejecta from Reykholt mixed with local materials.

Undivided material (unit Nu) includes outliers that rise above the flat surfaces of the upper member of the Tempe Terra Formation (unit Htu) and mesas, hills, and ridges that rise above fractured plateau material (unit HNplf). Degraded crater material (unit Nc) appears degraded and (or) fractured and deeply embayed by younger material; smooth, flat floors are common and extensive.

HESPERIAN–NOACHIAN SYSTEMS

Fractured plateau material (unit HNplf) is essentially the areal equivalent of older fractured material (unit Nf) of Scott and Tanaka (1986). In contrast with their description of the unit, impact crater outlines are largely well preserved rather than "largely destroyed." Fractured plateau material (unit HNplf) is considered to be Hesperian-Noachian, but its stratigraphic position is uncertain. Previous workers have (1) suggested it may be the fractured equivalent of the material of Lunae Planum (Wise, 1979), which is Hesperian, (2) included it with Hesperian ridged plains material (unit Hprg of Scott and Carr, 1978), (3) described it as younger than the material of Lunae Planum (Plescia, 1980, 1981), and (4) placed it in the Noachian (Scott and Tanaka, 1986). Cumulative frequencies of craters greater than 2 and 4 km place fractured plateau material near the middle of the Hesperian (fig. 3*C*, *D*), but craters greater than 8 km (fig. 3*D*) suggest that Upper Noachian is possible. Adjacent to the east edge of the map area, faulted lava flows and tube ridges are part of fractured plateau material (unit HNplf; see Viking frames 627A17 and 18). These faulted lava flows and tube ridges extend 170 km to the northwest of an unnamed volcano caldera across units mapped as Noachian and Hesperian by Scott and Tanaka (their units Nf and Htm, 1986; see Viking frame 704B50). This unnamed volcano is Hesperian according to crater counts (Hodges and Moore, 1994); Scott and Dohm (1990) suggested a Noachian to Hesperian age on the basis of stratigraphic relations. Inspection of the contact

Table 1. *Locations, estimated dimensions, and ratios of selected shields and domes in Tempe-Mareotis region and dimensions of a shield in Syria Planum.*

[Relief and dimensions estimated from photoclinometry profiles; ratios assume symmetrical shields. Syria Planum is shallow and depth was not determined. Values listed without latitudes and longitudes are separate measurements of same preceding shield or dome, but on a different Viking Orbiter frame. Typical values for terrestrial shields from Pike (1978) and Pike and Clow (1981). -, not determined or not applicable]

Latitude (° N.)	Longitude (° W.)	Diameter (m) D	Height (m) h	Flank width (m) w_f	Crater diameter (m) w_c	Crater depth (m) d_c	Height/ diameter ratio h/D	Height/ flank ratio $2h/D-w_c$	Crater/ flank ratio w_c/w_f	Viking Orbiter frame number	Flank pattern
						Type 1					
38.69	86.81	9,746	44	3,993	1,760	80	0.005	0.011	0.441	627A04	Radial
37.05	85.25	5,154	25	2,211	732	17	0.005	0.011	0.331	627A27	Radial, hummocky
36.85	85.07	9,748	53	4,202	1,344	70	0.005	0.013	0.320	627A27	Radial
35.74	85.61	4,460	29	1,728	1,004	44	0.007	0.017	0.581	627A28	Radial
33.82	82.54	8,804	100	3,908	988	66	0.011	0.026	0.253	627A54	Radial
36.02	87.90	3,710	39	1,611	488	14	0.011	0.024	0.303	627A41	Smooth
		3,542	43	1,428	686	12	0.012	0.030	0.480	627A22	Smooth
35.75	87.87	6,835	90	2,660	1,514	100	0.013	0.034	0.569	627A41	Radial
36.79	88.40	8,808	143	3,661	1,486	44	0.016	0.039	0.606	627A22	Radial
Average ----------		7,137	66	2,974	1,189	54	0.009	0.023	0.448		
						Type 2					
38.56	88.01	4,620	112	1,766	1,088	30	0.024	0.063	0.616	627A02	Radial, hummocky
36.66	86.09	4,628	138	2,088	452	7	0.030	0.066	0.218	627A25	Smooth
		4,779	134	2,108	562	8	0.028	0.064	0.267	627A26	Smooth
		4,692	137	2,122	448	8	0.029	0.065	0.211	627A27	Smooth
33.76	82.69	4,949	129	2,024	900	61	0.026	0.064	0.445	627A54	Smooth
35.91	88.03	4,006	158	1,710	586	29	0.039	0.092	0.343	627A41	Smooth
36.15	84.99	5,130	205	2,119	892	105	0.040	0.097	0.421	627A28	Smooth
Average ----------		4,679	148	1,948	781	47	0.032	0.076	0.417		
						Type 3					
35.76	86.26	10,678	83	-	-	-	0.008	0.016	-	627A26	Hummocky
35.59	86.93	3,370	32	-	-	-	0.009	0.019	-	627A43	Hummocky
37.03	85.44	5,266	50	1,793	1,680	64	0.009	0.026	0.931	627A27	Hummocky
35.82	87.68	2,436	52	-	-	-	0.021	0.043	-	627A41	Hummocky
Average ----------		5,438	54	-	-	-	0.012	0.026	-		
						Domes					
33.78	81.95	1,664	84	-	-	-	0.050	-	-	627A54	Dome
		1,139	67	-	-	-	0.059	-	-	627A55	Dome
33.82	81.91	1,549	90	-	-	-	0.058	-	-	627A57	Dome
						Mars Shields					
Low --------------------		6,014	87	2,523	1,078	52	0.016	0.039	0.468		Average, this study
Syria Planum---------		43,900	600	20,500	2,900	-	0.014	0.029	0.141		Hodges and Moore (1994)
						Earth Shields					
Low --------------------		4,770	75	2,200	370	24	0.016	0.034	0.168		Pike (1978)
Steep -------------------		1,590	70	700	190	23	0.044	0.100	0.271		Pike (1978)
Iceland -----------------		8,595	480	4,050	475	33	0.056	0.118	0.117		Pike (1978)
Erta Ale----------------		49,150	625	24,000	1,150	50	0.013	0.026	0.048		Pike and Clow (1981)

between older fractured material and Hesperian ridged plains material near lat 34° N., long 77° W. (units Nf and Hr of Scott and Tanaka, 1986; see Viking frame 519A13) suggests that ridged plains material is buried by lava flows from the north and that the upper surfaces of horsts of fractured plateau material are composed of the same material that buries Hesperian ridged plains material. Finally, boundaries between fractured plateau material and facies 1a of the middle member of the Tempe Terra Formation (unit Htm_{1a}) are gradational.

In agreement with superposition relations, crater size-frequency distributions indicate fractured plateau material (unit HNplf) is older than the lower three facies of the upper member of the Tempe Terra Formation (units Htu_{1-3}; see fig. 3*C*, *D*, *F*). Thus, the upper bound stratigraphic position of this unit is well constrained, but the lower bound is not.

HESPERIAN SYSTEM

Hesperian map units record strong evidence for fluvial resurfacing, tectonism, and volcanism in the region. Channels, levees, terraces, eroded craters, and crater statistics represent fluvial resurfacing. The extent of resurfacing is difficult to establish, but it could have been great. Tectonism, which is discussed later, is represented by displacements along faults in fractured plateau material (unit HNplf) caused by northwest-southeast extensional stresses (see fig. 3*E*). Volcanism is represented by small volcanic shields, plains, rimae, and lava flows. Most of the volcanic materials came from local vents and sources, but some may have come from distant sources such as the flanks of Alba Patera. Size-frequency distributions of craters superposed on various surfaces and units, superposition of units, and geomorphic considerations show that the fluvial resurfacing, tectonism, and volcanism are Hesperian.

Fluvial resurfacing included both erosion and deposition. As noted earlier, statistics of superposed craters show that resurfacing occurred on terraced plateau material (unit Nplt) south of Reykholt during the Hesperian (fig. 3*B*). Channel material (unit Hch) of Enipeus Vallis is, at least in part, the result of fluvial deposition, because the channel has levees from lat 34.0° N., long 93.4° W. to lat 36.9° N., long 92.9° W. and because depositional bedforms are possible to the north, especially near lat 39.7° N., long 92.7° W. The stratigraphic position of channel material (unit Hch) has not been established with crater statistics. It is probably Hesperian because the channel is embayed by the lower member of the Alba Patera Formation and by facies 1 of the upper member of the Tempe Terra Formation (units Hal and Htu_1). In addition, unlike the nearby terraced plateau material (unit Nplt), fractures (which may be Hesperian) are not present. Channel material may be coeval with the fluvial resurfacing south of Reykholt.

Valley floor material (unit Hvf) occurs only at lat 42.5° N., long 85.4° W. Grooves and ridges that parallel the valley walls suggest this material flowed much like debris flows elsewhere (see, for example, Carr and Schaber, 1977; Squyres, 1978). The unit is probably Hesperian because it embays terraced plateau material (unit Nplt), grooves and ridges within it are truncated by facies 1 of the upper Tempe Terra Formation (unit Htu_1), and fractures in nearby terraced plateau material (unit Nplt; which may be Hesperian) are not present.

Hesperian volcanism is embodied in the Tempe Terra Formation. As originally mapped, the Tempe Terra Formation comprises lower, middle, and upper members (Scott and Tanaka, 1986). In this map area, the lower member is not recognized because the eastern part overlies fluvially resurfaced terraced plateau material (unit Nplt) from about lat 40.5° N., long 84.5° W. to lat 39.5° N., long 84.5° W., as well as the textured flanks of the crater Charlieu. Also, no evidence exists for different materials on either side of the contact between cratered plateau material and a lower member of the Tempe Terra Formation (units Npl_1 and Htl of Scott and Tanaka, 1986) from lat 40.5° N., long 85.0° W. to lat 38.8° N., long 85.0° W. In the western part, the lower member, as mapped by Scott and Tanaka (1986) near lat 38° N., long 87° W., includes a variety of materials and landforms, which show no consistent evidence for their contacts. Here, the middle member of the Tempe Terra Formation (Scott and Tanaka, 1986) is adopted in principal and divided into three facies (units Htm_{1a}, Htm_{1b}, and Htm_2). Facies 1a occurs in the northeast corner of the region where image resolution is poor. No constructional volcanic landforms are evident, but large collapse depressions are present. Facies 1a is roughly equivalent to and includes the type locality of the middle member of Scott and Tanaka (their unit Htm, 1986). Facies1b occurs in the west and includes volcanic vents and rimae, and collapse depressions; two rimae are truncated by facies 1 of the upper member (unit Htu_1). Facies 1a and 1b are middle or Upper Hesperian because they are superposed on terraced plateau material (unit Nplt) that was resurfaced in the middle Hesperian and because they are overlain by the lowermost facies of the upper member of the Tempe Terra Formation (units Htu_{1-3}). Facies 2 (unit Htm_2) is found in the west near lat 38° N., long 91° W. and in the east near lat 39.5° N., long 80.5° W. Both occurrences have well-developed lava flow fronts that overlie facies 1a and 1b (units Htm_{1a}, Htm_{1b}). In the west, facies 2 is overlain by facies 1 of the upper member of the Tempe Terra Formation (unit Htu_1); in the east, only crater materials overlie facies 2. In the western part of the map area, small, isolated, featureless patches of plains that are superposed on terraced plateau material (unit Nplt) and buried by facies of the upper member of the Tempe Terra Formation are assigned to facies 2.

The upper member of the Tempe Terra Formation comprises the volcanic shields, lava plains, and flows that represent the unique style of volcanism of this region. This upper member has been divided into six facies on the basis of dominant landforms and superposition; three of these facies may be Amazonian and are discussed later. The shields, plains, and flows are found in western Tempe Terra, as patches on the Tempe Fossae ridge-valley system, and on the western plains. In general, the local stratigraphic positions of each facies correspond to their numerical sequence, but considerable overlap may exist in stratigraphic positions of noncontiguous facies. Facies 1 (unit Htu_1) includes small shields, lava plains, and some noticeable lava flows that embay and bury faults and fissures in plateau materials (units Nplt and HNplf; fig. 4*A*) and in facies of the middle member of the Tempe Terra Formation (units Htm_{1a}, Htm_{1b}, and Htm_2). Facies 2 includes lava flow fields having conspicuous flow lobes and fronts that are superposed on facies 1 (fig. 4*A*); one or two shields may be present. Facies 3 includes noticeable shields or groups of shields that are superposed on facies 1, facies 2, or both (fig. 4*A*). Several isolated occurrences, such as Issedon Tholus and its neighbor, are mapped as facies 3 (unit Htu_3) on the basis of their morphologies.

They are partly buried by plains materials (units Hal and Ap_2) or younger facies of the upper member of the Tempe Terra Formation (unit $AHtu_4$) that completely encircle them. Frequencies of craters greater than 2 km on all three facies indicate they are Upper Hesperian (fig. 3*F*), in agreement with previous mapping (Scott and Tanaka, 1986).

Characteristic landforms of the upper member of the Tempe Terra Formation (unit Htu) as mapped by Scott and Tanaka (1986) are not the same everywhere. Low shields and lava plains are the characteristic landforms in Ascuris Planum, Tempe Fossae, and the western plains. In contrast, characteristic landforms from lat 34° N., long 73° W. to lat 39° N., long 68° W. (see Scott and Tanaka, 1986) are southeast-trending lava-tube ridges and lava plains that appear to be associated with Tempe volcano (fig. 1); low shields are absent. Parts of plains in the northeast part of the map area, near lat 45° N., long 82° W. and near lat 40° N., long 71° W., have few or no striking landforms.

Rima flank material (unit Hfl) of the two northeastern links of Baphyras Catena may be volcanic or impact in origin. Flow lobes and association with northeast-trending collapse structures argue for a volcanic origin. However, the raised ridges at the outer contacts are similar to those of many martian impact craters and suggest that craters produced by oblique projectile impacts could have preceded the collapses. This material is Hesperian because it was deposited after deposition of facies 1a of the middle Tempe Terra Formation (unit Htm_{1a}) and before slightly degraded crater material from Charlieu (unit Hc).

Member 2 of the Tharsis Montes Formation (unit Ht_2) occurs along the south-central edge of the map area, and the stratigraphic position of Scott and Tanaka (1986) has been adopted. Some of this unit, as mapped by Scott and Tanaka (1986), has been replaced with younger plains materials on this map (units Ap_3, Ap_4, and Ap_5). To the south (at lat 31.7° N., long 90.0° W.), member 2 of the Tharsis Montes Formation occurs on a broad ridge having a summit depression that suggests member 2 has genetic affinities with the upper member of the Tempe Terra Formation or the younger plains materials (units Ap_{1-5}). Lava-flow lobes of facies 5 of younger plains materials (unit Ap_5) and facies 3 of the upper member of the Tempe Terra Formation (unit Htu_3) are superposed on this unit, but it may be coeval with facies 1 of the upper member of the Tempe Terra Formation (unit Htu_1).

The lower member of the Alba Patera Formation (unit Hal), along the west edge of the map area, is partly the same unit of Scott and Tanaka (1986). Flow lobes are common at the margins of this unit and within it. The combination of superposition and crater statistics places this unit in the Hesperian. Steep scarps of this unit encircle older volcanic shields, Issedon Tholus and its neighbor (unit Htu_3), which rise as steptoes above surrounding plains of the unit; it embays channel material (unit Hch). In turn, younger Amazonian lava flow lobes of facies 2 and 3 of younger plains materials (units Ap_2 and Ap_3) flowed in westerly directions onto this unit. Frequencies of craters greater than 2 km and 4 km indicate this unit is Upper Hesperian (fig. 3*G*), in agreement with Scott and Tanaka (1986).

Slightly degraded crater material (unit Hc) originated from bolide impacts with the rock units. Unit Hc is assigned stratigraphic position using superposition and morphologic criteria. Charlieu crater (lat 38.4° N., long 83.9° W.) is Hesperian because it is interleaved with facies 2 and 3 of the upper member of the Tempe Terra Formation (units Htu_2 and Htu_3). A small impact crater at lat 36.0° N., long 86.4° W. is also Hesperian because its north flank is overridden by flows (unit Htu_2) and the southwest flank is buried by a small shield (too small to map as facies 3; unit Htu_3), but the crater material rests upon facies 1 of the upper member of the Tempe Terra Formation (unit Htu_1) (fig. 4*B*). Morphological criteria include an appraisal of degree of degradation of the flanks and the morphology of the crater floor for areas without mantling material. Small depth-diameter ratios and smooth planar floors suggest they are older craters that have trapped eolian deposits.

AMAZONIAN–HESPERIAN SYSTEMS

Facies 4, 5, and 6 of the upper member of the Tempe Terra Formation (units $AHtu_4$, $AHtu_5$, and $AHtu_6$) are near the central part of the map area. Lava of facies 4 was deposited along the flanks of rimae at lat 40.4° N., long 88.0° W. and at lat 39.4° N., long 87.3° W. and issued from them as indicated by paleoflow directions and marginal lobate scarps. Facies 4 is assigned to the Amazonian-Hesperian because the collapse that produced the southern rima is Amazonian (see below), but the lava could have been deposited earlier. Facies 5 (unit $AHtu_5$) forms three small patches (near lat 88.5° N., long 39.0° W.) having well-developed flow lobes at their margins. A tongue of lava from one of these patches flowed onto facies 4. Volcanic shields, similar to those of facies 3 (unit Htu_3), are the principal landforms of facies 6 (unit $AHtu_6$). These shields are near lat 38.4° N., long 88.2° W. and form a northeast-trending chain; flow lobes of this facies cross facies 4 (unit $AHtu_4$).

Rima wall material (unit AHrw) has been mapped within the Baphyras Catena collapse depressions and another nearby large depression southwest of Baphyras Catena at lat 38.2° N., long 85.8° W. This unit is partly stratigraphic and partly morphologic because the material is mapped on steep slopes where talus deposits that formed subsequent to the initial collapse and earlier than older rima floor material (unit AHrf) dominate, but local exposures of Noachian rocks could be present. The material of the wall of the depression to the southwest of Baphyras is assigned to the Amazonian-Hesperian because of its proximity to similar material in Baphyras.

Older rima floor material (unit AHrf) occurs on floors of six northeast-trending rimae: three comprise Baphyras Catena (at lat 39.3° N., long 83.4° W., lat 39.2° N., long 83.9° W., and lat 38.9° N., long 84.5° W.), one is nearby to the southwest on the plateau (at lat 38.2° N., long 85.8° W.), another is still farther to the southwest in Tanais Fossae (at lat 39.3° N., long 83.4° W.), and the sixth lies to the southeast (at lat 33.7° N., long 83.4° W.) and is small. Five of the rimae involved collapse of Hesperian materials or erosion surfaces so that the floor material is Hesperian or younger. The rima in Tanais Fossae is Hesperian because it transects upper Tempe Terra Formation (unit Htu_1), and ejecta from a nearby Hesperian impact crater (unit Hc) was deposited on the rima floor. The material may be partly lava and partly eolian material.

AMAZONIAN SYSTEM

Materials of the Amazonian System are the result of volcanism, mass wasting, and eolian processes. Volcanism is represented by five facies of younger plains materials (units Ap_{1-5}) in the southwest corner of the map area (fig. 4*C*). The lava of facies 1, 3, and 5 (units Ap_1, Ap_3, and Ap_5) appear to have issued from vents having little or no relief that are aligned northeast-southwest, but three volcanic shields are part of facies 4 (unit Ap_4). Lava of each subsequent facies flowed onto preceding facies producing well-defined lobes, so their relative ages are known. The lava of facies 2 and 3 (units Ap_2 and Ap_3) flowed on top of the lower member of the Alba Patera Formation (unit Hal); lava of facies 3 (unit Ap_3) also flowed onto facies 1 (unit Ap_1). Although the base of facies 1 (unit Ap_1) is not exposed, it is placed with the younger plains materials because its northeast-southwest aligned vents are similar to those of facies 3 and 5 (units Ap_3, and Ap_5). These facies were included as part of the Hesperian units by Scott and Tanaka (1986; their units Htu, Hal, and Ht_2). Frequencies of craters greater than 1 and 2 km for all five of the facies are consistent with Lower Amazonian (fig. 3*H*).

Apron material (unit Aa) is a product of mass wasting; it is confined to the bases of steep slopes and steep-walled valleys. Earlier workers attributed similar material to flow of debris with interstitial ice (Carr and Schaber, 1977; Squyres, 1978). An occurrence was mapped as slide material (unit As, type 3) by Scott and Tanaka (1986) beyond this map area at lat 46° N., long 84° W. and was interpreted as debris flow. This material is Amazonian because it is uncratered and superposed on almost all adjacent materials, like similar material elsewhere on Mars (Carr and Schaber, 1977; Squyres, 1978; Tanaka, 1986).

Younger rima floor material (unit Arf) comprises the smooth floors of two rimae (at lat 40.4° N., long 88.0° W. and lat 39.4° N., long 87.3° W.), which may include lava deposited during and after the collapses that produced the rimae; some eolian material may also be present. The southern rima truncates some of the nearby apron material (unit Aa), but a small lobe of apron material has flowed onto the rima floor (fig. 4*D*). Thus, material of the southern rima is Amazonian; the northern rima is mapped the same as the southern floor because the morphologies of the rimae and their flanks are similar.

Crater floor material (unit Acf) forms smooth, level surfaces on the floors of nearly all impact craters assigned to the Hesperian and Noachian and in heavily mantled craters in the northwest. The extent of crater floor material (unit Acf) is an important part of the criteria for separating craters assigned to the Amazonian and Hesperian except in the northwest corner of the map area where there is extensive mantling by mottled material (unit Am).

Mottled material (unit Am) occurs as an eolian mantle in the northwest corner of the map area and was not recognized by previous mappers. The material partly to totally obscures subjacent topography at all elevations so that "ghosts" of channels, vents, craters, and aprons are visible in some places but absent in nearby mantled areas where their presence would be expected. The material is interpreted to be deposits of eolian dust. The eolian mantle covers adjacent units, including apron material (unit Aa).

Well-preserved crater material (unit Ac) is assigned stratigraphic position using superposition and morphologic criteria. Some well-preserved crater material (unit Ac) must be Amazonian because the craters overlie facies of the younger plains materials (units Ap_{1-5}). Depth-diameter ratios are large; smooth, flat floors, if present, are small.

STRUCTURE

The topography of the Tempe–Mareotis region is dominated by northeast-trending elements, except near the center where the west edge of Tempe Terra is dominated by the north-south canyons of Tanais Fossae. The dominant, northeast-trending elements are controlled by extensional faulting. Orientations of some plateau margins, plains margins, valleys, and mountains are subparallel to northeast-trending faults within or near them, but close inspection reveals the faults have a variety of orientations (fig. 5; see also Scott and Dohm, 1990). The deformation and faulting were intense in Hesperian–Noachian time and then declined (Scott and Dohm, 1990); stratigraphic uncertainties do not permit an accurate assessment of earlier deformation and faulting. Faulting and structures may be divided into (1) Tanais Fossae, (2) terraced plateau, (3) Tempe Fossae, (4) Mareotis Fossae–Ascuris Planum, and (5) rimae systems. These systems, along with other faults, are discussed below and compared with the Tharsis trend. Here, the Tharsis trend is represented by a great circle that passes through the centers of the calderas of Pavonis Mons (lat 0° N., long 112.8° W.) and Uranius Patera (lat 26.2° N., long 92.9° W.) and through Tempe Fossae in the southwest corner of the map area (from lat 32.5° N., long 86.65° W. to lat 38.06° N., long 80° W.). Many of the large valleys in Tempe Fossae are sub-parallel to the great circle. Caldera centers of Arsia and Ascraeus Montes are about 60 km west and east, respectively, of this circle. Azimuths of this great circle in this map area range from N. 42.4° E. (near lat 32.5° N., long 86.6° W.) to N. 50.5° E. (near lat 42.5° N., long 73.5° W.). These azimuths are used to compare fault trends with the Tharsis trend. The expectations for extension fault trends from some tectonic and geophysical models are similar (Plescia and Saunders, 1982; Banerdt and others, 1992), but they predict systematic changes in azimuths across the Tempe–Mareotis region. In the isostatic model presented by Banerdt and others (1992; see their fig. 9), predicted trends for extension faults (normal to extension directions) change from N. 22° E. in the northwest corner of the map area to N. 65° E. to N. 70° E. in the southeast corner. No such systematic changes were observed here nor were they reported by Scott and Dohm (1990).

TANAIS FOSSAE SYSTEM

Landforms resulting from this system include the north–south-trending canyons and valleys of Tanais Fossae at the west margin of the Tempe Terra plateau, the scalloped margins of adjoining plateaus to the west (fig. 4*D*), and the north- to northwest-trending canyons north of the crater Reykholt. The canyons, valleys, and adjoining plateaus have numerous attributes. (1) The canyons are incised into Reykholt and into lineated and terraced plateau materials (units Ncr, Npll, and Nplt, respectively). (2) Relief of the canyon walls is locally 1,200 m to 1,780 m (Davis and others, 1995). (3) Canyon walls of Tanais Fossae and the plateaus to the east, west, and north are terraced with risers measured in hundreds of meters and treads in kilometers. (4) The margins of terraced plateaus to the west are scalloped with septa and reentrants that trend northeast. (5) Canyons in the north, and some in the south, are box canyons that terminate in arcuate scarps. (6) Canyon floors are filled with volcanic and other materials.

Formation of the canyons and plateaus of this system probably precedes the middle Hesperian because they are filled with and embayed by facies 2 of the middle member and facies 1 of the upper member of the Tempe Terra Formation (units Htm_{1b} and Htu_1). Terraces may have formed on them during a fluvial resurfacing event in the Hesperian (fig. 3*B*). Onset of the formation of the canyons and plateaus could have begun at any time in the Upper or Middle Noachian (see fig. 3*A*, *B*) or even Lower Hesperian, because the stratigraphic positions of the older units involved (units Ncr, Npll, and Nplt) are poorly known. For example, parts of lineated plateau material (unit Npll) are adjacent to and in a shallow depression (at lat 39° N., long 85° W.) that may be part of the Tanais Fossae system because it is elongated in a north-south direction. This depression may have formed before (or after) the lineated plateau material (unit Npll) was deposited.

The Tanais Fossae canyons resemble box canyons of some fretted terrain having large relief and arcuate scarps at their heads (Sharp, 1973; Baker and others, 1992). However, terracing of canyon walls and adjoining plateaus in fretted terrain is rare, if not absent. Orientations of the canyons in fretted terrain are controlled structurally by joints and faults (Sharp, 1973). The floors of some canyons in fretted terrain have parallel to concentric ridges and grooves that have been interpreted to result from the flow of ice-laden debris (Squyres and others, 1992). The valley floor materials that fill the floor of the canyon at lat 42.5° N., long 85.4° W. may be similar to these ridged and grooved materials. The 1.2 km to 1.8 km relief along Tanais Fossae (and in Baphyras Catena) may represent the base of ice-laden rock, whereas fault-intersection depths may represent the top of the cryosphere (Davis and others, 1995). The Tanais Fossae canyons also resemble the troughs that may be sources for the lahar of the Elysium region (Christiansen, 1989), but again terraces are not evident in the Elysium troughs. Explanations for both the fretted terrain and Elysium troughs involve volcanic heating and sublimation or melting of ice in frozen ground (Sharp, 1973; Christiansen, 1989). A similar explanation is plausible for the formation of the Tanais Fossae canyons, but collapse and foundering of materials due to magma withdrawal at depth also are possible. A nominal relief (1.6 km), length (near 150 km), and width (10 to 25 km) of the canyons suggest that 1,000 to 6,000 km^3 of material could have been removed.

Schultz and Tanaka (1994) suggested that the Tanais Fossae structure might have formed by buckling and thrusting. East and northeast of Reykholt, some surfaces of terraced plateau material slope westward, suggesting possible rotation, but no evidence exists for buckling and thrusting. Rotation and, perhaps, buckling, has occurred immediately to the northwest of Tempe volcano (fig. 1) at lat 38.5° N., long 77.5° W., where faulted surfaces of lava flows slope toward the southeast.

TERRACED PLATEAU SYSTEM

The most obvious landforms resulting from this system are the northeast-trending plateau margins, mountains, valleys, and ridges associated with terraced plateau material (unit Nplt). These trends appear to be produced by displacements along extensional faults and grabens parallel to the Tharsis trend. Northwest–southeast elongation of craters, such as Gandzani, and analyses of fault-scarp widths suggest northwest–southeast extension and strain near 2% to 3% (Golombek and others, 1996). However, close inspection reveals that faults with northwest and northeast trends are present west of Tanais Fossae in the isolated plateaus and mountains of terraced plateau material (unit Nplt). Assessing faults of this system is difficult because they occur in isolated plateaus and mountains and because fluvial resurfacing has modified plateau surfaces. The faults may have formed over a long interval. For example, a fault east of Reykholt is buried by Reykholt crater material (unit Ncr), but nearby grabens to the south offset Hesperian terraces.

TEMPE FOSSAE SYSTEM

Landforms of this system are the flat-topped ridges and intervening valleys in the southeastern part of the region. Most valleys trend northeast and are several tens of kilometers long, but the largest valleys, 5 to 15 km wide, extend northeastward for 100 km or so. The ridges and valleys are interpreted to be horsts and grabens, and their relief ranges from about 10–30 m to 550–630 m based on photoclinometry and shadow measurements. The larger relief is consistent with the results of Davis and others (1995) for valleys in the older fractured material of Scott and Tanaka (unit Nf; 1986). Detailed inspection reveals small northeast- and north-trending grabens and faults intersecting one another to form diamond-shaped blocks. Large valleys have zigzag segments with edges that parallel the edges of the diamond-shaped blocks (fig. 6*A*, *B*). Overall trends of many large valleys are parallel to the Tharsis trend, but few of them form straight segments more than 60 to 100 km long.

The principal attributes of the Tempe Fossae system are as follows: (1) Faults and grabens cut fractured plateau material (unit HNplf). (2) Faults and fractures form mosaics of diamond-shaped blocks. (3) Many large valleys parallel the Tharsis trend, but they are made of zigzag segments having more northerly and northeasterly trends. (4) The frequency distribution of zigzag segments and faults is broad and bimodal (fig. 5). (5) The upper surfaces of horsts show little rotation. (6) Faults that displace the materials of craters superposed on fractured plateau material (unit HNplf) have similar orientations as those that do not displace crater materials. (7) Only part of the population of superposed craters has been faulted.

Several lines of evidence show that the horsts and grabens probably formed in the Hesperian (but a very Late Noachian age is possible). (1) The gradational contacts between facies 1a of the middle member of the Tempe Terra Formation (unit $\mathrm{Htm_{1a}}$) and fractured plateau material (unit HNplf) imply that the two units may be coeval. (2) Size-frequency distributions of all craters larger than 2 and 4 km that are superposed on the fractured plateau material (unit HNplf) permit Upper to Lower Hesperian (fig. 3*C*, *D*) and that very Late Noachian craters were fractured. The distribution of unfractured craters suggests that faulting nearly stopped in the Upper Hesperian (fig. 3*E*). (3) In agreement with crater frequencies, an upper bound of Upper Hesperian is established by superposition of the Hesperian facies of the upper member of the Tempe Terra Formation (units $\mathrm{Htu_{1-3}}$) on nearly all of the fractures and grabens of this system (fig. 6*A*). If the lava flows and tube ridges that issued from Tempe volcano (fig. 1) are Hesperian, as indicated by Hodges and Moore (1994) and by the structural study of Scott and Dohm (1990), then they are transected by faults of this system just east of the map area (near lat 37° N., long 78° W.; frames 627A17 and 18). (4) The fresh, noneroded appearances of faults, horsts, grabens, and superposed craters imply that they were unmodified by fluvial resurfacing. According to crater statistics, faulting did not stop until after the Hesperian fluvial resurfacing. Thus, the displacement that produced the Tempe Fossae horsts and grabens appear to have formed within a short interval near the middle of the Hesperian (fig. 3*B–E*), but some Late Noachian faulting is possible.

The rhomboidal fault patterns or diamond-shaped blocks of Tempe Fossae are similar in form and scale to those of the Basin and Range Province in south-central and southeastern Oregon and northeastern California (see for examples, Lydon and others, 1960; Donath, 1962; Walker and MacLeod, 1991). The diamond-shaped blocks of Tempe Fossae, which have little or no rotation, and modal fault orientations near N. 7° E. and N. 40° E. suggest vertical conjugate shear fractures could have formed in response to horizontal maxima and minima principal stresses and subsequent displacements were vertical (Donath, 1962). Adjacent to the northern and eastern boundaries of this map area and at the northeast corner, stress orientations compatible with vertical conjugate fractures are predicted by the isostatic model in Banerdt and others (1992), but the predicted azimuths would be more easterly than those observed. Clearly, similar patterns can be formed experimentally (Withjack and Jamison, 1986). Locally, horizontal offsets of diamond-shaped blocks and grabens are observed, but a consistent pattern of horizontal offsets is not evident nor can vertical (or inclined) fault planes be demonstrated with confidence (see also Davis and others, 1995). Lava injected along vertical conjugate fractures would form dikes that intersect at large angles, and subsequent erosion could produce ridges that intersect at large angles (see fig. 6*A*). However, rhomboidal fault patterns are common in regions of extensional faulting (Thompson and Burke, 1974). Elongation of faulted craters, such as at lat 37.0° N., long 80.1° W., is consistent with inclined fault planes (see Golombek and others, 1996), but not all craters are elongate in the same direction. Possible causes for complex fault patterns in extensional systems include (1) variations in the stress orientations with time, (2) variations in mechanical properties of the crust, and (3) the influence of preexisting structures (Thompson and Burke, 1974). Complex patterns also are produced in one episode during laboratory extension experiments (see for examples, Oertel, 1965; Stewart, 1971). In any event, the large valleys (with zigzag segments) that are sub-parallel to the Tharsis trend argue for extensional stresses in the northwest–southeast directions at a large angle to the Tharsis trend.

An analysis of scarp widths of faults and grabens along a traverse from lat 40.5° N., long 89.5° W. to lat 30.8° N., long 76.0° W. (and through lat 34.0° N., long 80.0° W.) yields northwest–southeast extension and a total strain of 22±16 km, or about 2% to 3% (Golombek and others, 1996). Modal values of fault-intersection depths near 1.0 km to 1.2 km for Tempe Terra have been attributed to a discontinuity, such as the top of the cryosphere (Davis and others, 1995).

MAREOTIS FOSSAE–ASCURIS PLANUM SYSTEM

Landforms of this system include sparse grabens that trend N. 44° E. to N. 64° E. In Mareotis Fossae and Ascuris Planum, most of the faults transect the lower three facies of the upper member of the Tempe Terra Formation (units Htu_{1-3}), but at lat 35.4° N., long 86.9° W., a lobe of lava from facies 2 of the upper member of the Tempe Terra Formation (unit Htu_1) flows into a graben, about 100 m deep and 1,400 m across, in facies 1 of the upper member of the Tempe Terra Formation (unit Htu_1). Elsewhere, northeast-trending faults and fissures transect the lower facies of the upper member of the Tempe Terra Formation (unit Htu_1). Also included in this group are the faults in the northeast corner of the map area. These faults cross facies 1a of the middle member of the Tempe Terra Formation (unit Htm_{1a}), but not facies 1 of the upper member of the Tempe Terra Formation (unit Htu_1). Again, extension at large angles to the Tharsis trend is implied by the orientations of the valleys.

RIMAE SYSTEM

Rimae are interpreted to be collapse depressions caused by the removal of magma and other fluids, such as water, from depth. They are among the most awesome landforms in the area. Baphyras Catena has three links (at lat 39.3° N., long 83.4° W.; lat 39.2° N., long 83.9° W.; and lat 38.9° N., long 84.5° W.) that attain lengths from 25–38 km, widths from 5–14 km, and depths to 1.7 km (or 1.62 km according to Davis and others, 1995). Most of the depressions trend in northeasterly directions, as represented by Baphyras Catena (N. 51° E., N. 61° E., and N. 70° E.). Most collapse depressions are probably Hesperian, such as the two near lat 39.65° N., long 84.6° W. and lat 39.9° N., long 84.9° W. They are incised into resurfaced terraced plateau material (unit Nplt) and nearly filled by facies 1a of the middle member of the Tempe Terra Formation (unit Htm_{1a}). However, some are Amazonian. At Baphyras Catena, Hesperian crater material from Charlieu (unit Hc) rests upon rima flank material (unit Hfl) but is truncated by the rimae; the nearby large rima at lat 38.2° N., long 85.8° W. is hosted by terraced plateau material (unit Nplt) that was resurfaced in the Hesperian. Facies 1a of the middle member of the Tempe Terra Formation (unit Htm_{1a}) in the northeast corner of the map area hosts many collapse depressions. At lat 37.1° N., long 91.8° W., the southeastern end of the collapse depression in facies 2 of the middle member of the Tempe Terra Formation (unit Htm_{1b}) contains lava of facies 1 of the upper member of the Tempe Terra Formation (unit Htu_1). Three small depressions to the northeast are associated with sinuous rimae that are superposed by facies 1 of the upper member of the Tempe Terra Formation (unit Htu_1). Among the youngest collapse depressions are the two at lat 40.4° N., long 88.0° W. and lat 39.4° N., long 87.3° W. that trend N. 60° E. and N. 67° E., respectively. Younger rimae floor material (unit Arf) covers their floors. Again, the southern rima transects apron material (unit Aa).

As noted earlier, relief of some rimae has been attributed to a discontinuity, the base of ice-rich rocks (Davis and others, 1995), but the relief of some rimae may represent other discontinuities. For example, photoclinometry indicates that the floor of the rima at lat 38.2° N., long 85.8° W. is at least 0.9 km higher than the floor of Tanais Fossae to the west.

OTHER FAULTS

The western plains in the map area are nearly devoid of faults and grabens. Two conspicuous sets of northeast-trending grabens transect the northwest corner of the map area with trends from N. 40° E. to N. 65° E. Part of one of the sets is superposed on the lower member of the Alba Patera Formation (unit Hal), and the other is partly obscured by mantle materials (unit Am). These grabens are extensions of Acheron and Tractus Catenae of Middle to Late Amazonian age (Tanaka, 1990). Two sets of grabens with trends about N. 41° E. transect facies 3 and 5 of younger plains materials (units Ap_3 and Ap_5) in the southwest corner of the map area. Fault-controlled vents and fissures in facies 3 of younger plains materials (unit Ap_3) are aligned N. 26° E.

VOLCANISM AND STRUCTURE

The interaction between volcanism and structure appears to have resulted in (1) preferred orientations of vents and eruption fissures and (2) inhibited faulting. One might expect that azimuths of extension faults would, more or less, parallel those of the great circle through Tharsis (about N. 47° E.), but actual orientations for faults reveal a more complicated situation (fig. 5). The systematic eastward increase of 30° in fault azimuths from N. 31° E. in the southwest corner of the map area to N. 61° E. in the southeast corner that is predicted by the isostatic model for Tharsis (Banerdt and others, 1992) is not evident in the data. A sample of fault orientations that predate volcanism (see figs. 3*D*, *E*, 5, and 6) range from about N. 162° E. to about N. 72° E. and show modal orientations near N. 7° E. and N. 40° E. Magma reached the surface along faults and fissures oriented chiefly between N. 32° E. and N. 72° E., but those oriented near N. 42° E. and N. 62° E. were preferred conduits as is shown by the two modes in the frequency distribution of vent and fissure orientations (fig. 5). These relations are embodied in figure 8, which shows the vents atop each of the two volcanic shields are aligned near N. 74° E., but surrounding faults buried by lava have a broad range of orientations. The differences between the two fault azimuth distributions (fig. 5) might have been caused by changes in orientations of stresses acting on preexisting fault structures. The disappearance of the mode of the pre-volcanism fault orientations near N. 7° E. and appearance of the mode of the vent and fissure orientations near N. 62° E. suggest that the stress field changed and extension directions shifted.

The onset of volcanism at the surface may have inhibited extensional faulting. This possibility is suggested by the paucity of faults in and the apparently consistent relative ages of the lower facies of the upper member of the Tempe Terra Formation (units Htu_{1-3}) and the unfaulted craters of fractured plateau material determined from crater statistics (compare fig. 3*D* to fig. 3*E*). A plausible model is that overpressure of intruding magma reduced the difference between the maximum (vertical) and minimum (horizontal) principal stresses to a value lower than that required for normal faulting (see Parsons and Thompson, 1991). The relief of the small shields and domes would seem to require modest magma overpressure at the surface. Thus, extensional displacement ordinarily accommodated by normal fault displacement could be accounted for by the intrusion of magma into vertical fractures, similar to the mechanism proposed for the eastern Snake River Plain in Idaho (Hackett and Smith, 1992) and Inyo Craters in California (Mastin and Pollard, 1988). If the fault-intersection depths near 1.0 to 1.2 km are the result of non-eruptive dikes (Davis and others, 1995), a change in stress magnitudes, as well as orientations, are implied because magma did reach the surface with, at least, modest overpressure. Other possibilities that might reduce or alter the differences between the maximum and minimum principle stresses include (1) deflation or removal of magma at depth, (2) changes in loading caused by the Tharsis volcanoes (see for example, Banerdt and others, 1992), and (3) planetary contraction (Solomon and Chaiken, 1976).

FLUVIAL EROSION

At least part of the Tempe–Mareotis region was substantially eroded by flowing water. Convincing evidence for fluvial erosion includes channels, terracing, unusual landforms, and crater statistics. A channel just southeast of the rim of crater Reykholt (fig. 7*A*) is incised into a deep channel that has eroded through the east and northeast flanks of Reykholt. In places, more than 300–400 m of ejecta (McGetchin and others, 1973; Pike and Davis, 1984) have been removed. Just north of this region near lat 43.7° N., long 86.1° W. (fig. 7*B*), channels emerge from beneath mottled material (unit Am) and become lost in the structural complex farther north. These northern channels have nearby terraced islands and bedforms that are consistent with catastrophic floods (Baker and Milton, 1974; Baker and Nummedal, 1978). The valleys near lat 42.5° N., long 83.0° W. and near lat 42.5° N., long 84.5° W. may be channels eroded by flood waters and subsequently covered with volcanic deposits. Fluvial erosion offers an explanation for the origin of at least part of the terraces (fig. 7*D*) of the terraced plateau materials (unit Nplt), terraces and missing crater materials of Reykholt (unit Ncr) and a somewhat larger crater to the north (near lat 43.4° N., long 84.0° W.), and the distribution of the lineated materials related to Reykholt (unit Npll). A variety of landforms can be explained by fluvial erosion, but among them is an unusual circular, concentric-ring structure at lat 39.8° N., long 84.6° W. (fig. 7*C*) that is interpreted to be an eroded complex impact crater (Pike and Davis, 1984), 19 km in diameter, for which more than 300 m of rim (Pike and Davis, 1984) have been removed-along with its central peak. Unusual northern landforms are possible scours and plucking depressions at the south base of a mesa near lat 43.5° N., long 84.0° W. (frame 255S32). Finally, the frequency distribution of superposed craters near Reykholt is consistent with Hesperian fluvial or other resurfacing (fig. 3*B*).

Fluvial resurfacing in areas mapped as fractured plateau material (unit HNplf) is problematic because the area is so complex that deciphering all the geologic relations is not possible. The erosion surfaces may have been (1) completely buried and then fractured, (2) simply fractured, or (3) partly buried and then fractured. Importantly, fluvial channels are not obvious in areas where fractured plateau material (unit HNplf) has been mapped, but erosional remnants are present. A number of ridges mapped as Noachian undivided material (unit Nu) that rise above the general surface of the plateau appear to have been scoured (for example, near lat 37° N., long 80.7° W.). Isolated mesas, some of which are terraced, rise above the general surface of the plateau; these mesas could be analogous to those on terraced plateau material (unit Nplt) near Reykholt. In other places, particularly in the southeastern part of the map area, narrow parallel and intersecting ridges may be dikes that were exposed by erosion (fig. 6*A*). If Hesperian fluvial resurfacing was pervasive, as the erosional remnants suggest, it preceded the faulting and displacement along fractures because the grabens and horsts appear too fresh and uneroded to have been extensively resurfaced. Crater size-frequency distributions, despite the uncertainties, indicate that fracturing of the fractured plateau material (unit HNplf) ceased after fluvial resurfacing. Erosion of terraced plateau material (unit Nplt) and erosion or emplacement of fractured plateau material (unit HNplf) could have preceded, postdated, or been coeval with one another (fig. 3*B–E*). Thus, parts of the areas mapped as fractured plateau material (unit HNplf) and Noachian undivided material (unit Nu) may actually be terraced plateau material (unit Nplt) that has been resurfaced and fractured.

Enipeus Vallis also appears to be fluvial in origin because the northern exposures of channel material (unit Hch) resemble the channeled scablands of eastern Washington that were produced by catastrophic floods. Both the scablands and the northern reaches of Enipeus have butte and basin topography and streamlined residual hills (see for examples, Baker and Milton, 1974; Baker and Nummedal, 1978). The lava flows that cover much of the southern part of the channel system probably have buried the source of the water. However, the central part of Enipeus resembles the plexus of channels near the source troughs of the Elysium lahar channel systems (Christiansen, 1989). Regardless of origin, both catastrophic floods and lahars involve large amounts of water that would create butte and basin topography and large streamlined islands.

VOLCANIC LANDFORMS

A broad spectrum of volcanic landforms is found in the map area. Among the landforms are small volcanic shields, domes, and lava tube ridges and flows. These landforms are discussed below.

SHIELDS

Numerous small, conical hills and low, broad ridges in the region have been interpreted to be volcanic shields, because they resemble the volcanic low shields of the Snake River Plain in size and shape (Plescia, 1980, 1981; Hodges, 1980), but several are larger. Three types of shields are present in the map area. Earlier workers (Russell, 1902; Stearns and others, 1938) called low shields of the Snake River Plain lava cones and lava domes. Russell (1902) also noted the broad similarity in shapes of the shields of the Snake River Plain and the much larger Hawaiian volcanic shields. Low shields of the Snake River Plain are composed mostly of pahoehoe lava flows (Greeley and King, 1977, p 13). Shapes of the martian shields include (1) conical rises with summit craters, (2) ridges with elliptical bases and elongate summit craters, and (3) irregular rises with elongate summit craters or knobs. Some martian shields have no summit craters similar to those of the Snake River Plain (Stearns and others, 1938), whereas others have summit knobs. On Mars, the shields may occur separately or shoulder to shoulder in clusters, similar to those of the Snake River Plain (Russell, 1902; Stearns and others, 1938; Jones, 1969; Greeley, 1982).

In order to compare the Tempe–Mareotis shields with those on Earth, the relief or heights of a number of shields with sufficiently uniform albedos were estimated using photoclinometry and high-resolution images; other dimensions were obtained from the photoclinometry profiles (table 1). The heights of the shields range from 25 m to 205 m and diameters range from 2,400 m to 10,700 m, but even smaller shields may be present. The independent estimates of the heights of some shields used here (table 1) and the revised estimates for the same shields (Davis and Tanaka, 1993; P.A. Davis, personal communication, 1994) are typically within 15–20%. Choice of flank widths is an important source of any differences. Despite the uncertainties, the results here appear reasonably consistent with geologic experience. Height to flank-width ratios (h/w_f) of the shields range from 0.011 to 0.097 and crater diameter and flank-width ratios (w_c/w_f) range from 0.211 to 0.931 (table 1).

To a first order, the martian shields are similar in size and shape to low shields on Earth. Their heights and diameters are compatible with those reported for terrestrial low shields that are 2–20 km across and 20–225 m high (Pike, 1978; Pike and Clow, 1981; table 1). Diameters of the martian shields are typically greater than those of terrestrial steep shields, and heights are typically less than Icelandic shields. Height to flank-width ratios reported for low shields on Earth are typically near 0.034 and range from 0.010 to 0.067 (Pike, 1978; Pike and Clow, 1981). All but two of the martian shields fall within this range, and these two are more or less compatible with one or two Icelandic shields. As a departure, crater diameter and flank-width ratios reported for low shields on Earth are typically 0.168 and range from 0.032 to 0.54 (Pike, 1978; Wood, 1979; Pike and Clow, 1981);

but, those of the martian shields are greater than 0.168 and five are greater than 0.54 (table 1). Wood (1979) suggested that crater diameters of low shields on Mars may be greater than those on Earth because of the difference in gravity of the two planets (Wood, 1979). Larger crater formation may be related to gravity and atmospheric pressure for many reasons. Crater diameter and flank width ratios for Hawaiian-type eruptions may be larger on Mars because lava fountains should be higher and broader on Mars (Wilson and Head, 1994). Also, bubble nucleation and lava fragmentation depths may be greater on Mars (Wilson and Head, 1994). Thus, mechanical failure of the upper parts of the vents in rocks having normal friction angles might result in wider crater diameters on Mars than on Earth.

Issedon Tholus is 50–60 km across and larger than other low shields in the region and those on Earth. The relief of Issedon Tholus could not be estimated because of albedo variations on it flanks, but the relief of a similar-size (approx 44 km across) and similar-appearing shield volcano in Syria Planum was estimated to be 600 m using photoclinometry (Hodges and Moore, 1994; table 1). This estimate suggests the relief of Issedon could be somewhat greater, perhaps 1,000 m, because of its larger diameter. Issedon has two summit pits, about 2 km across, and subdued summit depressions, about 3 km across, aligned along N. 58° E. Faint ridges and valleys on the flanks of Issedon radiate from a central region. Some continental shield volcanoes composed of nontholeiitic basaltic lava, such as Erta Ale in Ethiopia (Pike and Clow, 1981), may be analogous to Issedon Tholus and the shield in Syria Planum mentioned above because the diameters and heights are quite similar (table 1). Again, the diameters of the martian summit craters are larger than those of Erta Ale.

Three types of low shields exist in the map area (table 1; figs. 8, 9). The first type typically has radial ridges and valleys on its flanks, height and flank-width ratios near 0.011 to 0.039, and summit craters. The second type typically has smooth flanks, height and flank-width ratios near 0.064 to 0.097, and summit craters. The third type has hummocky flanks, height and flank-width ratios near 0.016 to 0.043, and summit knobs or summit depressions. Radial ridges and valleys on the flanks of shields are interpreted to be lava tubes, flows, and channels; the hummocks may be tumuli or other inflationary structures. Smooth flanks are difficult to interpret, but the possibility of pyroclastic deposits should be considered, as well as effusive pahoehoe lava flows. Summit knobs could be analogous to (1) the steeper summits of the Snake River Plain shields (Russell, 1902; Stearns and others, 1938; Greeley, 1982), (2) lava cumuli such as those of Erta Ale (Barberi and Varet, 1970), or (3) the large block on the south rim of Pillar Butte in the Snake River Plain (Greeley and King, 1977). In any event, the similarity in shape and size of shields on Mars and Earth suggest similar compositions, that is, basalt. The three types of shields may be the result of magmatic differentiation and (or) different eruption conditions such as effusion temperatures or gas contents (see for example, Wilson and Head, 1994).

DOMES

Eight domes are present in the southeastern part of the map area. Many of these are oval to circular in plan view and appear to be superposed on faults and adjacent materials; others are probably extrusions of lava, surfaces bowed by intrusions, and (or) erosional remnants. Two small extrusive domes, 1,700 and 1,600 m across, are 84 and 93 m high, respectively (table 1; figs. 8, 9). However, another dome at lat 34.0° N., long 82.3° W. (north of the dome shown in fig. 8) appears to be much lower than the others; it may be the surface expression of an intrusion or laccolith. Domes produced by intrusions of rhyolite in the Snake River Plain have uplifted basaltic rocks, and one of them, East Butte, is entirely uplifted basalt (Hackett and others, 1992); more subtle structures, such as Buckskin Dome and Ferry Butte, also may have resulted from intrusions (Kellog and Marvin, 1988).

For extrusive domes, Orowan (1949) calculated yield strengths (τ) using:

$$\tau = \rho * g * h^2/D$$

where ρ is the bulk density of the lava (2,400 g/m^3), g is the acceleration of gravity (3.7 m/s^2), h is the height of the dome, and D is the diameter of the dome. Yield strengths for the domes mentioned above are the same magnitude as those of lava domes on Earth (Blake, 1990) and between those used for rhyolite and basalt in an analyses of venusian domes (Bridges and Fink, 1992). However, no clear relation exists between compositions of domes and their yield strengths (Blake, 1990).

LAVA FLOWS

Lava flows are readily identified along most margins of lava fields (facies 2 of the upper member of the Tempe Terra Formation, unit Htu_2), the lower flanks of many shields having radial ridges and valleys (facies 3 of the upper member of the Tempe Terra Formation, unit Htu_3), and elsewhere in Tempe-Mareotis. As estimated using photoclinometry, lava flows range from 2 to 30 m thick and average 13 m thick; widths range from 500 to 5,000 m (table 2). According to nearby paleoflow directions and a topographic gradient from photoclinometry, the thickest measured flow issued from a fissure, flowed southeastward, ponded, and then cascaded into a graben; it is 26–30 m thick near the graben and 14–16 m thick upstream at a kipuka (fig. 10*A*). The thinnest measured flow, at 2 or 3 m, is some distance from a shield but aligned radial to it (fig. 10*B*).

Table 2. *Locations, estimated dimensions, estimated topographic gradients (Sin Θ), and adjusted thickness (ρ*g*H) of selected lava flows in Tempe-Mareotis region.*

[Relief estimated using photoclinometry; gradient estimated using various procedures (see text). Entries at lat 36.03°, long 85.77° and lat 36.03°, long 85.75° are same flow but measured at slightly different locations. -, no data]

Latitude (° N.)	Longitude (° W.)	Thickness (m)	Flow width (m)	Levee width (m)	Sin Θ	ρ*g*H (kPa)	Frame number	Notes
38.41	88.55	8-12	1,560	540	0.009	70	627A02	Flow
					0.006	110		
39.36	83.91	15-19	1,210	440	0.020	130	627A09	Flow
					0.014	170		
36.62	88.61	12-15	1,320	460	0.023	107	627A22	Flow
					0.010	134		
36.23	86.26	14-19	-	-	0.052	120	627A26	Kipuka
						170		
36.03	85.77	5-6.5	1,180	281	0.033	44	627A26	Flow 1
					0.008	58		
36.75	85.46	2-3	500	-	0.025	18	627A27	Flow 1
					0.005	27		
36.59	85.24	9-13	1,120	-	0.025	80	627A27	Flow 2
					0.010	120		
36.03	85.75	6-10	1,500	390	0.012	53	627A28	Flow 1
					0.007	89		
35.67	87.48	5-6	730	170	0.016	44	627A41	Flow 1
					0.008	53		
35.47	86.96	14-16	-	-	-	120	627A43	Kipuka
						140		
35.42	86.85	26-30	5,000	-	0.006	230	627A43	Near graben
					0.005	270		below kipuka

Comparisons of thickness, adjusted thickness, and yield strength for lava flows elsewhere on Mars and on Earth suggest that those in the Tempe-Mareotis region are basaltic in composition. The average thickness (13 m) of the Tempe-Mareotis flows is smaller than the average thickness of 23 m for flows measured on Mars, but comparable to the average (12 m) for flows on steeper slopes near the Ascraeus Mons summit. For comparisons with terrestrial flows, thickness is adjusted by multiplying them by the acceleration of gravity at the surface (g) and an estimate of the bulk density (ρ) of the flow (fig. 11). In preparing figure 11, gradients for the Tempe-Mareotis flows were estimated in three ways: (1) with photoclinometry, (2) by equating yield strengths (τ_1 and τ_2; τ_1 and τ_3), and (3) from the nearest shield. The equations for yield strength (Orowan, 1949; Hulme, 1974; Moore and others, 1978) are:

$$\tau_1 = \rho * g * H * \sin(\Theta)$$
$$\tau_2 = \rho * g * H^2/W_f$$
$$\tau_3 = \rho * g * W_l * \sin^2(\Theta)$$

where H is the thickness of the flow, Θ is the slope angle, W_f is the flow width, and W_l is the levee width. As shown in figure 11, adjusted thickness and gradients of the Tempe-Mareotis lava flows lie within and to the lower left of the cluster of data points formed by other martian flows; the adjusted thickness and gradients of some flows are about the same as some aa basalt on Earth. The adjusted thickness of the Tempe-Mareotis flows lies well below that of terrestrial rhyolite, trachyte, andesite, and basaltic andesite flows on comparable gradients. Thus, a basaltic composition seems likely. Lunar-like or slabby aa basalt may be present. Probable lava tube ridges and analogy with the basalt of the Snake River Plain suggest that thin pahoehoe flows would be the dominant flow type in the region.

Tempe-Mareotis flows have an average yield strength (1.5 (τ_1), 0.9 (τ_2), and 2.2 (τ_3) kPa) smaller than the average yield strength for other flows on Mars and yield strength somewhat consistent with terrestrial basalt (see for example, Moore and others, 1992). Average yield strength of rhyolite, trachyte, and andesite (between 123 and 287 kPa) is well above that of the Tempe-Mareotis flows. Yield strength (τ_1=0.6 kPa) for slabby aa basalt is smaller than that of the Tempe-Mareotis flows, and yield strength (τ_1=0.1 kPa and τ_2=0.2 kPa) of the lunar flow also is smaller than that of the Tempe-Mareotis flows; pahoehoe flows are not amenable to the analyses above.

IMPACT CRATERS

Impact craters within the map area are typical for Mars, but some aspects deserve comment, such as, clusters, large continuous ejecta-crater radii ratios, large bolide incidence angles, and the occurrence of lineated plateau material (unit Npll). At least three clusters of craters are present in the map area. In the first, seven craters larger than 3.5 km are within the area of the continuous ejecta (39 km^2) of Charlieu (lat 38.4° N., long 83.8° W.). Of the seven craters one is almost filled with ejecta, one is partly filled with ejecta, one has flanks partly buried by ejecta, three are superposed on the ejecta, and one is Charlieu. The second cluster consists of three craters larger than 3.5 km (near lat 40.0° N., long 83.3° W.) within an area of 42 km^2. The largest crater appears fresh and unfilled; the older two have flat floors and are deeply filled. The third cluster consists of three craters larger than 4.0 km (near lat 34.6° N., long 84.7° W.) within an area of 12 km^2. The smaller, younger one appears fresh, but the two older ones have flat floors and are deeply filled. The fill of one of these is superposed on the ejecta of the smaller, younger crater. Relative ages inferred from these clusters are unrealistically great, so they probably were produced by the impact of a cluster of bolides produced by atmospheric or tidal break-up of a larger bolide (see for example, Oberbeck and Aoyagi, 1972).

Large ratios of continuous ejecta to crater radii appear to be common within the region, but no detailed study was made of these ratios. For lunar craters, these ratios are near 2.348 (Moore and others, 1974) and ratios for most of the craters in this map area do not appear to differ significantly from the lunar craters, but some ratios are much larger than the lunar ratio. For example, the ratios for Charlieu and the crater to the north (near lat 40.0° N., long 83.3° W.) are near 4. A least squares fit to areas of the outer ejecta and diameters of craters in Acidalia Planitia imply a ratio near 4 (Mouginis-Mark and Carey, 1980), but those elsewhere are smaller (Mouginis-Mark and Cloutis, 1983). Mouginis-Mark (1981) attributed the differences in ratios to uniform targets containing volatiles (small ratios) and layered targets containing volatiles (large ratios).

At lat 41.7° N., long 80.3° W., three elongate craters, up to 4 km wide, align along an azimuth near N. 6° W. Crater rims are missing or low along the axis of alignment, ejecta deposits are bilaterally symmetrical about the axis, and the ejecta blanket of the northernmost one appears to be partly buried by ejecta from its nearest neighbor. Contact relations between ejecta from the other two are not evident. These characteristics suggest that the craters were produced by the impact of one projectile traveling from south to north along an oblique trajectory with a very large or grazing incidence angle near 88°—somewhat like those produced in the laboratory (Gault and Wedekind, 1978). The two eastern links of Baphyras Catena, which contain rima flank material (unit Hfl), may have been the sites of oblique impact craters that were subsequently modified by collapse.

Finally, a plausible explanation for the occurrence of lineated plateau material (unit Npll) is that it is residua resulting from the dynamic deposition of Reykholt ejecta that survived fluvial resurfacing. Lineated plateau material (unit Npll) is related to Reykholt crater because various patches of the unit have ridges, grooves, and elongate pits that are aligned along radials from Reykholt; patches near the northeast flanks are topographically lower than the nearby crater material (unit Ncr) of the flank. Theory (Oberbeck, 1975) and observations of terrestrial craters (Hörz and others, 1983) show that the impact and ground-hugging outward motion of ejecta from large impact craters erode and mix with the substrate materials. Thus, the lineated plateau material (unit Npll) is interpreted to be a mixture of crater ejecta and substrate material on a pocked and grooved surface. These materials underlie crater materials on the flanks of Reykholt, and they have been entirely removed by fluvial erosion in many places.

Although not a preferred explanation, the lineated plateau material (unit Npll) could be the result of scouring and deposition during the fluvial resurfacing event. In this interpretation, the ridges, grooves, and elongate pits are erosional and depositional bedforms (Baker and Nummedahl, 1978; Baker and others, 1992).

MASS WASTING

Landforms created by mass wasting include talus slopes, landslides (geomorphic, unit ls), and lobate debris aprons. Talus slopes were not mapped as such, but talus is most assuredly present on the walls of grabens, rimae, and craters. Landslides are found throughout the region where suitable slopes are present. Three or four landslides occur along the channel east and southeast of crater Reykholt (fig. 7*A*). These landslides have hummocky runout masses below scars having slopes that face the runout. One landslide composed of slumped and apparently rotated blocks was mapped within a graben near lat 35.9° N., long 80.6° W. Other landslides are undoubtedly present in grabens, but identification of them is unclear. Possible landslide runout was mapped south of apron material (unit Aa, fig. 4*D*). Lobate debris aprons were assigned the nongenetic name apron material (unit Aa) for mapping purposes and are discussed below.

Lobate debris aprons may be the result of on-going or recent processes (Squyres, 1978). Lobate debris aprons near lat 39.6° N., long 86.6° W. are both younger and older than the younger rima floor material (unit Arf). Tanaka (1986) places the debris aprons in the Upper to Middle Amazonian. Within Tempe-Mareotis, they form aprons at the bases of steep slopes of terraced plateau and fractured plateau materials (units Nplt and HNplf, respectively). Unlike hummocky landslides, the aprons have smooth surfaces that are convex upward near their outer margins. The materials appear to have flowed into and around craters and onto adjacent volcanic units. Squyres (1978) proposed that they formed in a manner similar to terrestrial rock glaciers. Here, debris moves downslope under the influence of gravity and becomes admixed with ice that sublimates from the atmosphere; this accumulation of ice and debris then moves like a glacier.

The process of lobate debris apron formation and flow is probably not an on-going one. Moore and Davis (1987) analyzed the profile of a lobate debris apron in the southern hemisphere, which they assumed was a steady-state ice glacier, and concluded that it had an ablational form. They also suggested that moats at the bases of hills elsewhere were the former sites of ice that had completely ablated. A photoclinometric profile across one apron having relief of 124 m and length of 4,738 m (at lat 38.0°, long 86.6° W.; fig. 4*D*) revealed not only the convex-upward outer margin, but also a reversal of slope near the base of the scarp of about 1/1,000. For a steady-state glacier, this reversal of slope implies flow toward the scarp and (or) ablation (see Moore and Davis, 1987). Inspection of other aprons also suggests that gentle slopes that face the scarps are present. Moats, which may represent the locations of lobate aprons that have been completely removed by ablation, also are present in Tempe-Mareotis, such as the moat around the base of Pindus Mons (lat 39.8° N., long 88.5° W.) and elsewhere.

VOLCANISM IN TEMPE-MAREOTIS COMPARED TO EARTH

Volcanic landforms of the Tempe-Mareotis region are similar to landforms produced by basaltic plains volcanism on Earth. Exemplified by the Snake River Plain, this style of volcanism is intermediate between flood or plateau basalt volcanism and Hawaiian volcanism (Greeley, 1982). Tempe-Mareotis has abundant low shields (some of which are coalescing), lobate flows, lava plains, and domes—similar to those of the Snake River Plain (Greeley and King, 1977; Plescia, 1980, 1981; Hodges, 1980; Greeley, 1982; Hackett and Smith, 1992). However, the presence of significant amounts of rhyolitic volcanic rocks and sediments beneath the surface in Tempe-Mareotis, like those beneath the Snake River Plain (Hackett and Smith, 1992), seems unlikely. Additionally, pyroclastic cones and maar volcanoes (or their equivalents) are absent in Tempe-Mareotis. Abundant small volcanic shields, aligned vents and fissures, and flow fronts distinguish Tempe-Mareotis from Lunae and Hesperia Plana volcanism; these same features distinguish Snake River Plain and flood basalt volcanism. The extensive plana resemble lunar maria (Greeley and Spudis, 1981) and include few lobate lava flows or flow fronts and few small edifices, such as low shields. Small volcanic shields and lava plains distinguish the Tempe-Mareotis volcanism from the Tharsis volcanism that produced the large shields; these same features distinguish Snake River Plain and Hawaiian volcanism. Tempe-Mareotis shields have basal diameters typically less than 60 km, whereas montes, paterae, and tholi in Tharsis have basal diameters of 55 to 600 km. Similarly, the shields in Tempe-Mareotis typically have hundreds of meters or less relief; the great shields in Tharsis have relief between, at least, 1 and 26 km. On Earth, the 1969–1971 and 1972–1974 eruptions of Kilauea Volcano produced a small volcanic shield called Mauna Ulu (Swanson and others, 1979; Tilling and others, 1987). Likewise, shields similar to those of Tempe-Mareotis are present on and near the great shields of Tharsis, such as the north flank of Uranius Patera and south of Ascraeus Mons (fig. 1) (Hodges and Moore, 1994).

Volcanic styles in Ceraunius Fossae and Syria Planum (Hodges and Moore, 1994), at the base of Olympus Mons (Morris and others, 1991) (fig. 1), and in Cerberus (Plescia, 1990) are partly similar to the volcanic style of Tempe-Mareotis. The low shields of Ceraunius Fossae appear to be chiefly, if not entirely, characterized by summit depressions and flanks with radial ridges and valleys; those with steep smooth flanks appear to be absent. Similarity with Syria Planum volcanism is confined to the abundant shields of Syria at elevations near 9 km. The Syria Planum volcanic plains that extend south-southeast from the shields differ from the Tempe-Mareotis plains because of the large concentration of thick flows, large lava-tube ridges, and coalescing protuberances. In the two other volcanic plains, the base of Olympus Mons and Cerberus, vents do not attain the relief of some of the shields of Tempe-Mareotis.

Geographic association of the Tempe-Mareotis plains volcanism and Tharsis volcanism suggests a genetic relation between them. As with Syria Planum and Ceraunius Fossae, the Tempe-Mareotis volcanic plains occur along the margins of Tharsis; Alba Patera and Tempe volcano are to the west and east, respectively (fig. 1). Tempe-Mareotis also lies along the northeast extension of the Tharsis Montes volcanic chain, which hosts three great volcanic shields (Arsia, Pavonis, and Ascraeus Montes) (Carr, 1974) and which passes through Uranius Patera. Among the possible explanations for this geographic association are magma sources and tectonism. Mantle convection triggered by core separation is a suggested mechanism for the formation of the Tharsis Montes volcanic chain and related hotspots for magma supply (Carr, 1973, 1980). Subcrustal erosion and underplating by a single convection cell have been offered as an explanation for the crustal dichotomy, Tharsis Montes volcanic chain, and magma sources for Tharsis volcanism (Wise and others, 1979a, b). Extensional stresses related to the Tharsis Montes volcanic chain have been used to explain fault patterns that radiate from Tharsis, such as those of Claritus, Tempe, Ceraunius, and Memnonia and Sirenum Fossae (Carr, 1974; Wise and others, 1979a, b; Plescia and Saunders, 1982; Banerdt and others, 1992). This explanation is partly supported by the northeast-trending grabens of Tempe Fossae, but the zigzag segments, complex fault patterns, and mosaics of horsts in Tempe Fossae (fig. 6) and elsewhere show that the actual situation is more complex than the predictions of current models (see Scott and Dohm, 1990). Some vent azimuths are approximately radial to Tharsis, but azimuths near N. 62° E. are more easterly. The causes of the more easterly vent trends are unclear. Scott and Dohm (1990) suggest that their Upper Hesperian fault set #2, which is transverse to radial from Tharsis, may be related to stress fields associated with Alba Patera and Ceraunius Fossae, rather than Tharsis. Tanaka (1990) suggests that preexisting structures related to a major impact basin might have influenced later faulting.

Eruptions that produced the minor Tharsis, Tempe-Mareotis, and, perhaps, the major Tharsis shields began in the Hesperian (fig. 12) and possibly in the Noachian (Hodges and Moore, 1994). Minor Tharsis and Tempe-Mareotis shields were formed by the end of the Late Hesperian and Early Amazonian, but the major Tharsis shields continued to grow to elevations as high as 25–26 km and lava continued to form plains at elevations as low as 1–6 km; summit elevations of Tempe-Mareotis shields are near 3–4 km (fig. 12). Flank eruptions produced lava fans on the northeast and southwest flanks of the central shields of Arsia, Pavonis, and Ascraeus Montes and the apron on the east flank of Pavonis, and summit eruptions persisted within the caldera of Arsia Mons, and perhaps other calderas (Hodges and Moore, 1994). The increase of relief and elevations of the summits of the Tharsis and Elysium shields over time was recognized by Carr (1976a, b), who suggested that the relation may be related to the increase in lithosphere thickness over time. By analogy with Hawaii (Eaton and Murata, 1960), the maximum elevation or relief was attributed to the pressure head engendered by the density difference between the magma and surrounding rocks at depth (Carr 1974, 1976a, b, c, 1980). Carr estimated that the source of magma generation for the highest Tharsis volcanoes was about 200 km below the surface. Such a pressure difference is necessary for magma to rise above 27 km, but only very modest overpressures are required for the magma to rise to the summits of low shields. Magma need not rise to the same limit for a variety of reasons that are related to the evolution of the magma and reservoirs, the properties of the lithosphere, and regional and local stresses (see Head and Wilson, 1992). This is clearly illustrated by the caldera fill and lava fans of Arsia Mons and the low shields of Tempe-Mareotis. The amount of magma available in the reservoirs at Tempe-Mareotis was smaller than that which produced the great shields in Tharsis. For example, the volume of Olympus Mons, at 2 million km^3 (see for example, Morris, 1982), would cover two-thirds of the area of this map (about the same area as the base of Olympus Mons) with a uniform layer of lava nearly 7 km thick. Such a large quantity of lava in Tempe-Mareotis is not evident. Kuntz and others (1986) suggest that the cyclic eruptions of basalt along the Great Rift of the Snake River Plain are the result of increases and decreases in magma pressure within the reservoir. Eruptions begin and continue when the magma pressure exceeds a critical value related to the sum of the least principal stress and the tensile strength of the roof rock. Thus, the northeast-trending vents and grabens and relief or summit elevation of the low shields may be manifestations of reduced least-principal stresses (extension) and elevated magmatic pressure required for eruptions in Tempe-Mareotis. However, the pressure required to lift the magma to the summits of low shields was much less than at the summits of the great shields of Tharsis.

The duration of volcanism in the Tempe-Mareotis region may have been long, but the duration of eruptions forming individual shields and lava fields probably was short. Duration indicated from ages inferred using models 1 and 2 (fig. 12) and the relative ages of the young plains materials (units $\mathsf{Ap}_{1\text{–}5}$; fig. 3*H*) and the three lowermost facies of the upper member of the Tempe Terra Formation (units $\mathsf{Htu}_{1\text{–}3}$; fig. 3*F*) are near 0.25 to 1.0 b.y. By analogy with small shield volcanoes on Earth, such as Mauna Ulu on Kilauea, Hawaii, individual martian shields and fields could form in a matter of years. The Mauna Ulu eruptions in 1969–1971 and 1972–1974 covered areas chiefly with pahoehoe at average rates near 19 km^2 per year; average volume rates were near 0.07 km^3 per year (Swanson and others, 1979; Tilling and others, 1987). At these rates, the lava field and shields near lat 36.5° N., long 88.5° W., which have an estimated area of 1,500 km^2 and volume of 32 km^3, would form in about 80 to 460 years—much shorter than 0.25 to 1.0 b.y. Effusion rates estimated for individual flows on the flanks of martian shields, using the methods of Hulme and Fielder (1977) and Pieri and Baloga (1986) and assuming that the lava flows issued from tubes, are 1–3 km^3 per year. Although these rates are poorly constrained (see Crisp and Baloga, 1990a, b), they imply that average eruption rates for Mauna Ulu are plausible rates for Mars because the estimated rates are well within the peak rates inferred from lava fountains at Mauna Ulu (Swanson and others, 1979). Eruptions from volcanoes on Earth are separated by periods of repose. These periods for recent eruptions on the Snake River Plain are near 500 to 3,200 years; the output rates for uncontaminated lava of the Snake River Plain are near 1.5 km^3 per 1,000 years (Kuntz and others, 1986). At this rate, the lava field and shields above would form in 21,000 years—again, shorter than 0.25 to 1.0 b.y. The contrast between the long duration of volcanism inferred from relative ages and by analogy with Earth is puzzling; nothing is known about repose periods for eruptions of martian low shields.

GEOLOGIC HISTORY

Rock units and landforms in the Tempe-Mareotis region are the result of geologic processes that began billions of years ago and continue to the present day. Very little is known about the oldest deposits because they are highly modified and partly buried; younger deposits are better understood because they are less modified than the older deposits; present-day deposition of dusts and surface changes due to winds have been observed by spacecraft. The oldest rocks in the Tempe-Mareotis region are Noachian in age. The Noachian Period ended about 3.5 to 3.8 billion years ago (Tanaka, 1986) or possibly about 3.4 to 3.8 billion years ago (see fig. 12). Noachian rocks include terraced plateau material (unit Nplt), which has been interpreted as eroded layered deposits. As with Noachian materials elsewhere (Tanaka, 1986), terraced plateau material may include some unknown combination of impact basin and crater ejecta, volcanic material, and fluvial and eolian sediments. Noachian undivided material (unit Nu) should consist of similar deposits. Degraded and Reykholt crater materials (units Nc and Ncr, respectively) are materials of impact craters. The oldest in-place crustal materials may be present in the walls of Tanais Fossae at the lowest elevations because relief, near 1.4–1.8 km, suggests a large section of rock is present.

Subsequent to deposition of terraced plateau material (unit Nplt), the valleys of Tanais Fossae may have begun to form by the melting of ground ice somewhere in the section by magmatic heating. The water released by this heating resulted in collapse to form the valleys and may have contributed water for the fluvial resurfacing of the plateau. The fluvial resurfacing etched terraces in the Noachian rocks, eroded impact craters, and incised channels. Isolated plateaus, ridges, and dikes that rise above the surface of fractured plateau material (unit HNplf) may suggest fluvial resurfacing was widespread in the map region. Enipeus Vallis and related channel deposits (unit Hc) may have formed during this fluvial resurfacing event.

The region has a complex history of faulting and fracturing, but it may have been subjected to horizontal maxima and minima stresses that produced a rhomboidal pattern of vertical conjugate joints. Subsequently, in the Hesperian, extensional stresses at right angles to the projected axis of the Tharsis Montes volcanic chain resulted in the formation of northeast-trending grabens and a mosaic of rhomboidal horsts and northeast-trending grabens having zigzag segments. Graben formation nearly ceased and was followed by the basaltic volcanism in the Late Hesperian. The basaltic magma reached the surface along vents and fissures chiefly oriented in two northeasterly directions. This Late Hesperian volcanism produced the lava fields and low shields of the lowermost facies of the upper member of the Tempe Terra Formation (units Htu_{1-3}). Lava continued to issue from local vents and fissures aligned in northeasterly directions from the Late Hesperian into the Early Amazonian.

Mass wasting and eolian processes during the Amazonian produced a variety of landforms and deposits. These landforms include landslides, talus slopes, lobate debris aprons, and an extensive mantle of eolian material in the northwest part of the region. Eolian processes, represented by windstreaks, bright red surface materials with high albedos, and dune forms in grabens, produced the most recent geologic deposits and landforms in the region. Eolian processes are still active.

CANDIDATE LANDING SITES

Selection of candidate landing sites requires an evaluation of the surface materials, as well as consideration of spacecraft capabilities and scientific goals. Below, surface materials are evaluated in a general sense using remote-sensing signatures, possible variations of materials and roughness are evaluated using high-resolution images, and several landing sites are suggested for several spacecraft capabilities.

Landed spacecraft likely will encounter extensive soils and rocks at the surface in the Tempe-Mareotis region. Remote-sensing signatures (table 3) indicate that the soils best fit unit 1 of Christensen and Moore (1992), chiefly loose dust. This fit is supported by fine component thermal inertia near 1.6 to 2.1 x 10^{-3} cal cm^{-2} $s^{-1/2}$ K^{-1} (Christensen 1982, 1986a), albedos near 0.30 to 0.32 (Kieffer and others, 1977; Palluconi and Kieffer, 1981), and red-violet ratios near 2.7 to 3.1 (U.S. Geological Survey, 1992a, b, c). The red-violet radiance factors also permit soil-like clods as well as bright-red dust at the surface (fig. 13; Arvidson and others, 1983, 1989a, b; Dale-Bannister and others, 1988). The prospects of sampling or acquiring rocks to accomplish the exploration goals appear to be reasonable because rock abundances are between 5 and 11 percent (Christensen, 1986b; table 3). If the rocks are similar to those at the Viking landing sites, dust-free surfaces on the rocks will be rare (Guinness and others, 1987) and small rocks at the surface will be scarce (Moore and others, 1987). The materials at the surface may be uniform because remote-sensing signatures are uniform and standard deviations are moderate (table 3).

The question of uniformity at the fine scale is difficult to answer because of the large footprints of the

Table 3. *Average albedo, red-violet ratio, average fine-component thermal inertia, and average rock abundance for six 1:500,000 quadrangles of Tempe-Mareotis region.*

[Albedo from Palluconi and Kieffer (1981), red-violet ratio from U.S. Geological Survey (1992a, b, c), fine-component thermal inertia and rock abundance from Christensen (1986a, b)]

Quadrangle	Albedo	Red-violet ratio	Fine-component thermal inertia (10^{-3} cgs units)	Rock abundance (percent)
40092	0.30-0.31	2.7-3.0	1.90±0.16	7.9±2.6
40087	0.30-0.32	2.8-3.0	1.82±0.22	9.1±1.6
40082	0.30-0.32	2.8-3.1	1.84±0.23	9.4±1.6
35092	0.30-0.31	2.8-3.0	1.87±0.10	7.1±1.1
35087	0.30-0.31	2.8-3.0	1.71±0.11	9.5±1.4
35082	0.30-0.31	2.8-3.1	1.77±0.13	8.8±1.4

remote-sensing observations (kilometer and larger; Kieffer and others, 1977; Palluconi and Kieffer, 1981; Christensen, 1982). Landforms in images having resolutions near 8.4 m/pixel suggest that considerable variations in surface properties are possible (fig. 14); these variations are not evident in images having resolutions of 55 m/pixel and could not be mapped (fig. 6*B*). In figure 14, the dark wind streaks are consistent with the removal and deposition of cohesionless dust (see for example, Moore and others, 1987). The duneforms could be composed of drift materials like those at the Viking landing sites (see Binder and others, 1977; Mutch and others, 1977; Moore and others, 1982, 1987) or sand. The sand could be cohesionless (if recently formed) or cemented. Textured impact crater flanks imply poorly sorted ejecta. Rocks would be expected in ejecta and in the talus of the scarps. By analogy with the Viking landing sites, cohesive soil-like materials are probably present (Moore and others, 1982, 1987).

The Tempe-Mareotis region has many interesting places to explore. The chief goals of geologic exploration there would be to understand (1) styles of volcanism, (2) fluvial processes, and (3) ancient materials. The first goal could be achieved with the pin-point landing of a rover between the two shields shown in figure 8 and, then, exploring and sampling along three or four traverses to the central shield, dome, eastern shield, and fractured plateau; traverses near 8 km would be needed. Although not as satisfying, a fixed-platform lander targeted to lat 35.5° N., long 87.5° W. (with a landing ellipse 80 km by 160 km) probably would acquire samples of volcanic material. Exploration of fluvial deposits and erosion surfaces for goal 2 might require small landing ellipses or long rover traverses, because areas having unequivocal landforms are small (see for examples, fig. 7). Achievement of goal 3 requires sampling of a complete section. Thus, sampling along east–west traverses (15–30 km long) across the terraces shown in figure 7*D* should represent 1.6 km of section. Even older materials from greater depths might be found on the rims of the largest impact craters. Samples from very large depths also could occur as inclusions in basaltic rocks from lava flows and shields (see, for example, Leeman, 1980).

EDITOR'S NOTE

Dr. Henry J. Moore died prior to the completion of some finishing touches to this map. The map had been improved by technical reviews from James Dohm and Jeff Plescia. Ken Tanaka and Derrick Hirsch performed minor editing on the map, while attempting to avoid changing any of the author's original meaning.

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