ELSEVIER

Geomorphology 37 (2001) 329–352

GEOMORPHOLOGY

www.elsevier.nl/locate/geomorph

Morphology, evolution and tectonics of Valles Marineris wallslopes (Mars)

Jean-Pierre Peulvast [a,*], Daniel Mège [b], Jan Chiciak [c], François Costard [a], Philippe L. Masson [a]

[a] UMR 8616-Orsay Terre, Université Paris-Sud, Bât. 509, 91405 Orsay Cedex, France
[b] Laboratoire de Tectonique, case 129 ESA CNRS 7072, Université Pierre et Marie Curie, 4 Place Jussieu, 75252 Paris Cedex 05, France
[c] Université de Paris-Sorbonne, UFR de Geógraphic, 191 Rue Saint-Jacques, 75005 Paris, France

Received 1 July 1994; received in revised form 1 August 1998; accepted 14 January 1999

Abstract

Hillslopes up to 11 km in height can be found along the walls of the Valles Marineris troughs. The widest and deepest troughs are grabens, in which tectonics probably exerted the primary control on the wall morphology. Geographical variations in the wall morphology and profiles show that they result from complex, persistent tectonic influences, and that significant changes in erosional processes occurred during this evolution, from late Hesperian to late Amazonian. Preliminary calculations suggest that about 85–95% of the fault-controlled wall relief probably formed in an "ancient" stage prior to this transitional period. A study of the volatile content of the wall rocks, based upon the morphology and distribution of impact craters on the surrounding plateaus, shows that extreme erosional widening of the Central Valles Marineris troughs occurred during the "ancient" stage of high ground ice content. During the subsequent "recent" stage of tectonic and morphological evolution, the wall materials were partly desiccated.

Keywords: Morphology; Valles Marineris; Mars

1. Introduction

For Mars and Earth, very steep hillslopes with high relief (several kilometers) are related to special conditions, including (1) powerful geodynamic processes, such as volcanism with induced gravitational tectonics (Cipa, 1994) or mountain tectonic uplift; (2) exceptionally strong slope materials; and (3) deep erosion in the uplifted areas. The longevity and preservation of steep, high hillslopes also depend on the age of the tectonics and on the efficiency of the erosional processes that lower high-relief areas. Moreover, the high potential energy related to great slope heights (Embleton and Whalley, 1979), the possible effects of unloading related to rapid erosion, and vibrations related to the seismic–tectonic activity result in the frequent occurrence of mass move-

* Corresponding author.
E-mail address: peulvast@geol.u-psud.fr (J.-P. Peulvast).

0169-555X/01/$ - see front matter
PII: S0169-555X(00)00085-4

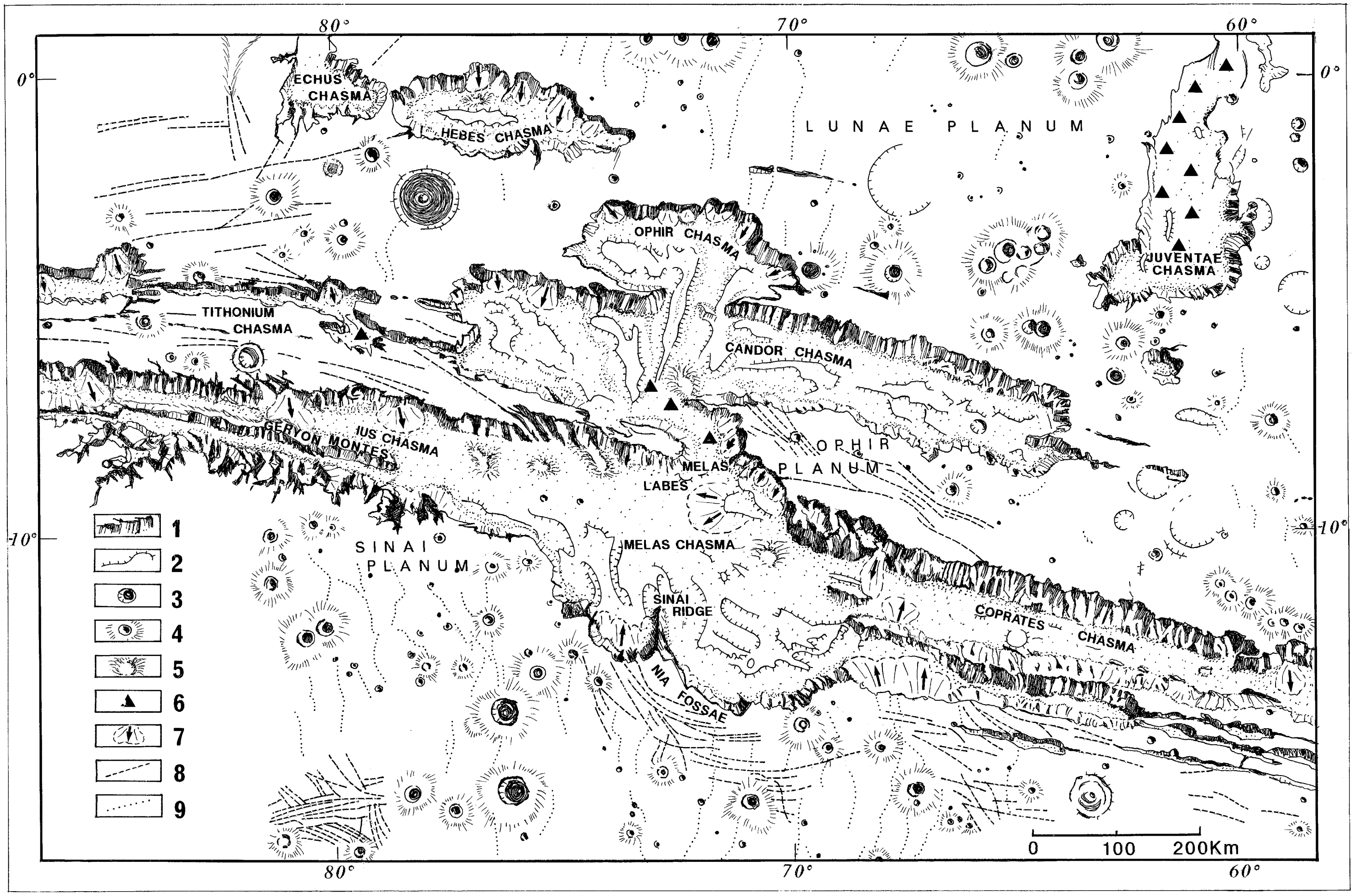


Fig. 1. Simplified geomorphological map of Valles Marineris (modified from Peulvast and Masson, 1993a). 1: Trough edge, ridge; 2: lower escarpment, bluff; 3: impact crater; 4: impact crater with lobate ejecta; 5: closed depression; 6: chaotic terrain; 7: landslide; 8: fault; 9: wrinkle ridge.

ments (Brunsden, 1979; Lucchitta, 1979; Fort, 1993; Fort and Peulvast, 1995).

The Valles Marineris walls in the Tharsis region of Mars have a relief up to 11 km in the central parts of a 4000-km-long system of troughs that lie just south of the martian equator (Fig. 1). The remarkable scale of this relief arose in part "because the absence of recent pluvial and fluvial activities on Mars prevented rapid infilling of the depressions" (Lucchitta et al., 1992, p. 478). The very high hillslopes of Valles Marineris form deep natural sections through the Tharsis Plateau. The morphology reflects properties of the wall materials, thereby constraining models of structure, composition, and evolution of the upper layers of the martian crust (Carr, 1981). Fault-related morphology is clearly evident along many wall segments (Carr, 1981; Mège and Masson, 1996; Schultz, 1991; Witbeck et al., 1991; Peulvast and Masson, 1993a). Relationships among these structures, and the erosional landforms of the walls and wallfoot deposits show the sequences of events that shaped the Valles Marineris troughs (Lucchitta, 1987; Lucchitta and Bertolini, 1989; Peulvast and Masson, 1993b). Thus, a geomorphological study of these landforms is critical to understanding the very high wallslopes of Valles Marineris in regard to regional tectonics.

2. Morphological relationships

This study will focus on the central part of Valles Marineris (Fig. 1), which consists of a series of parallel troughs or "chasmata". Most of these troughs are interconnected, except for Echus, Hebes, and Juventae Chasmata, which form a discontinuous parallel set a few hundred kilometers north of the main system; these chasmata form the deepest depressions within the whole Valles Marineris. The troughs are of two types (Lucchitta et al., 1990): (1) scalloped troughs and pit crater chains (Tithonium and south Coprates Chasmata) and (2) straight and well-aligned troughs (Ius and Coprates Chasmata), which widen between 64° and 77°30′ longitude to form a 700-km-wide system of isolated (Hebes Chasma) or interconnected troughs (Ophir, Candor and Melas Chasmata). The latter comprise the "Central Valles Marineris troughs", and are 120 to 160 km wide. Type (2) troughs are between 6 and 8 km deep, and up to 9 to 11 km deep in Melas Chasma (U.S. Geological Survey, 1986a,b). The very high trough walls overlook either: (1) wide and relatively flat parts of the trough floors, or (2) moats that separate them from interior benches or tablelands of layered deposits (e.g. Hebes Chasma: Peterson, 1981). In both cases, excepting local landslide areas, only minor volumes of waste deposits occur along the walls. In several places, the pristine morphology of downfaulted segments of upland plateau can be recognized on the trough floor, indicating the tectonic origin (Blasius et al., 1977; Schultz, 1991; Peulvast and Masson, 1993a,b).

The Valles Marineris structures comprise a rift system (Hartmann, 1973; Sharp, 1973; Frey, 1979; Masson, 1977, 1985; Witbeck et al., 1991), belonging to the Tharsis radial pattern of fractures (Carr, 1981), whose trend is expressed in many shallow grabens in the surrounding plateaus, and in structures inside the troughs. These fractures are perpendicular to the trend of wrinkle ridges on the plateaus, which are also related to Tharsis stress patterns (Chicarro et al., 1985; Tanaka et al., 1991; Watters, 1993). The boundary faults that control long and rectilinear sections of the chasmata are some of the most conspicuous among the regional structures (Schultz, 1991). Though rifting alone cannot explain the box-like terminations of Candor Chasma, or the transitions between pit crater chains, scalloped troughs, and grabens, good evidence exists for floor downfaulting and probably asymmetric rifting.

Lucchitta (1978) attributes the present configurations of the Valles Marineris walls to erosional scarp retreat, recognizing three major types of walls: (1) the spur-and-gully type, which is most common along straight and well-aligned troughs (e.g. Coprates and Ius Chasmata, interior ridges); (2) walls dissected by tributary canyons (south Ius Chasma); and (3) landslide scars forming broad curved or straight recesses in the chasma walls. This classification was later simplified (Bousquet et al., 1987; Lucchitta et al., 1992) to distinguish only spur-and-gully morphology, landslide scars, and small-scale talus slopes. Our classification (Fig. 2) emphasizes the systematic relationships among morphology, tectonics, and other structural elements in Central Valles Marineris (Peulvast and Masson, 1993a).

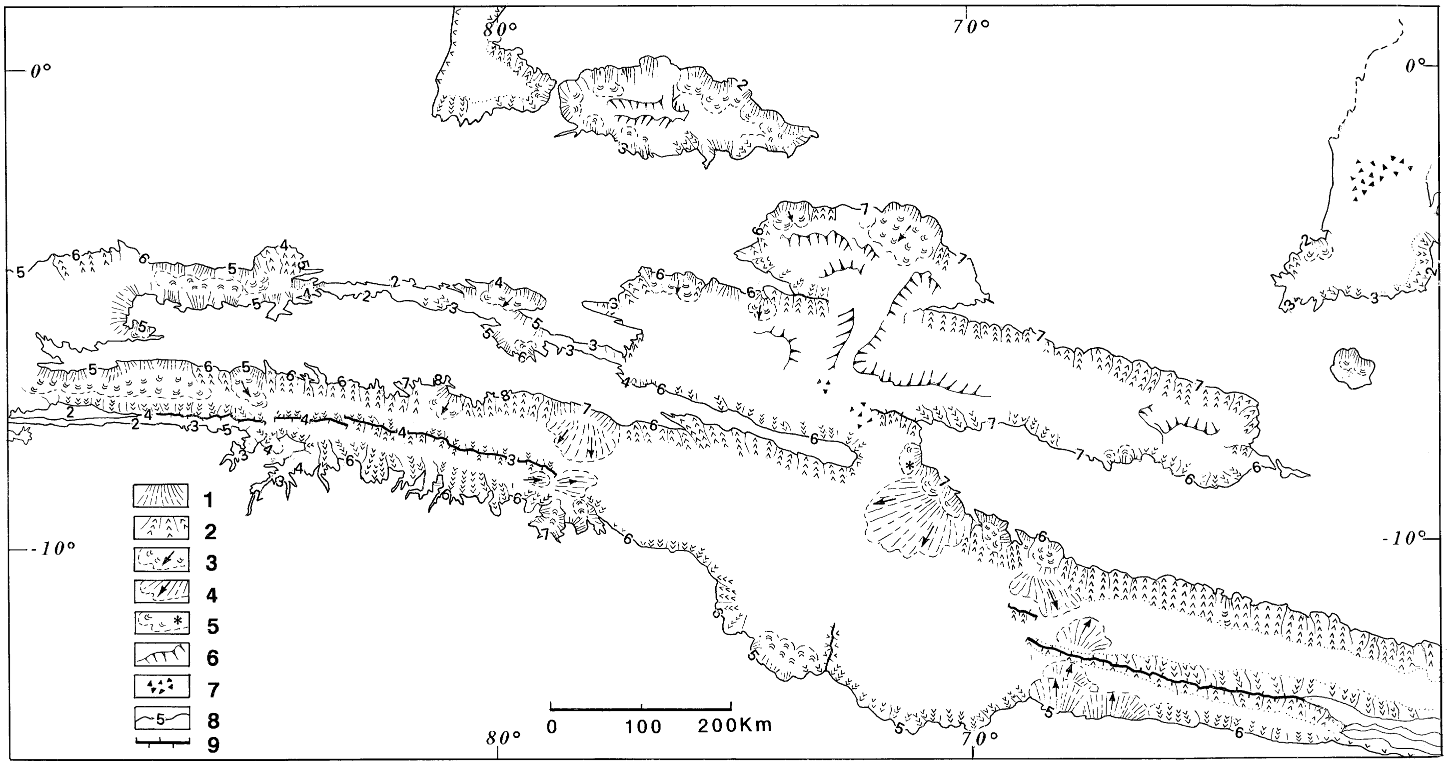


Fig. 2. Wall types of Valles Marineris. 1: Smooth talus slope; 2: spurs and gullies slope; 3: landslide deposits with hummocky material; 4: landslide deposits with hummocky and fan-shaped materials; 5: rotational slump; 6: aeolian flutes; 7: chaotic terrain; 8: slope gradient; 9: crest.

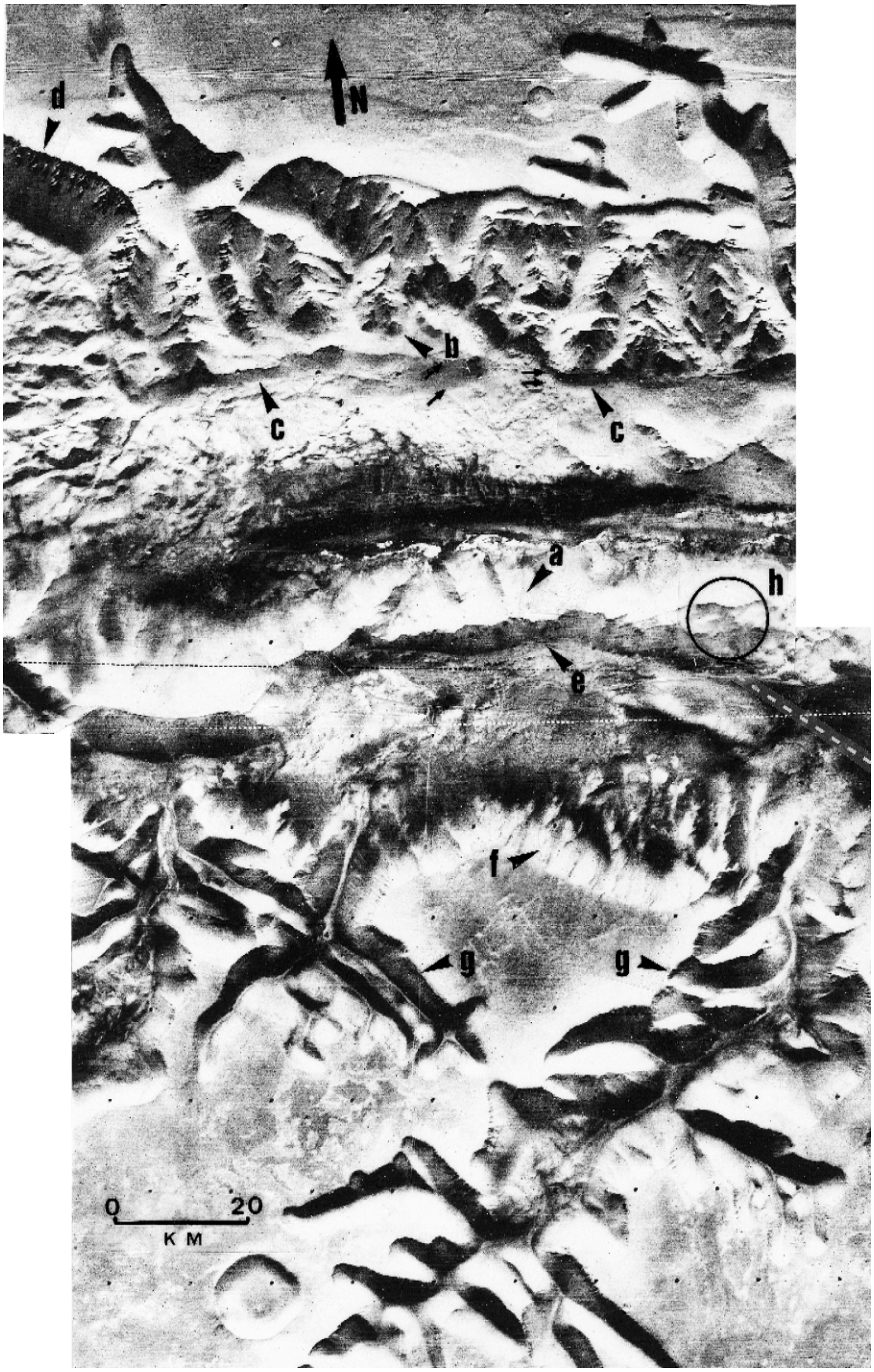
High, fault-controlled walls are well displayed along Ius and Coprates Chasmata (Figs. 1 and 3) and on interior ridges paralleling the chasma walls (for example, in the straight and well-aligned troughs). They are also found in north Melas Chasma and in the northern walls of Ophir and Hebes Chasmata (Lucchitta et al., 1992). Overall, rectilinear outlines and long rows of triangular facets cut into the lower spurs (northern wall of Coprates Chasma) allow a precise mapping of the main normal faults along the fault scarps (Schultz, 1991). These walls generally show spur-and-gully morphology with widths that range from 15 to 30 km and slopes in the head walls up to 30°. Dissection is locally controlled by minor oblique sets of synthetic and antithetic faults (Peulvast and Masson, 1993a). Though in many places the gullies merge with the troughs at the level of the floors, the presence of transverse fault steps and continuous rows of triangular facets and fault scarps at the base of the intervening spurs shows that faulting locally continued after the end of the gullying process. Narrow interior ridges, such as Geryon Montes in Ius Chasma, are probably horsts or half horsts (Mège and Masson, 1996; Schultz, 1991). The dissected ridge flanks intersect in sharp crests, where local splitting into two parallel crests suggests that slow mass movements contributed to the wasting process (Flageollet, 1988).

Tectonic control is also obvious along wall segments dissected by tributary canyons with blunt heads and V-shaped cross profiles (Lucchitta, 1978). For example, the 30- to 130-km-long Louros Valles, South of Ius Chasma, dissect a straight fault scarp, and appear to be passively controlled by intersecting oblique arrays of fractures. Northwest of Melas Chasma and in parts of Tithonium Chasma, solitary canyons are clearly incised along shallow oblique grabens intersected by the main troughs. These canyons have average gradients between 2° and 5°; the canyons merge with the trough floors and are considered to have been formed by structurally controlled sapping processes (Sharp, 1973; Lucchitta et al., 1992).

Most walls that lack obvious structural control were probably formed by erosional retreat along fault scarps (for example, the main fault lines are obliterated by erosion and/or deposition processes; Fig. 4). Moreover, the headwalls of the erosional landforms, which intersect the grabens and the wrinkle ridges of the surrounding plateaus, are only locally controlled by these structures, as in the case of some landslide scars. Most of these walls are located in Central Valles Marineris, where extreme widening and interconnection of parallel troughs (Ophir, Candor and Melas Chasmata) are at least partly related to erosional wall retreat (Peulvast and Masson, 1993a,b). Some of these walls display spur-and-gully morphology, and they generally overlook interior plateaus of layered deposits or the moats that separate these plateaus from the walls. Embayments of these deposits commonly occur inside the erosional re-entrants, and even inside the mouth of some gullies.

Landslide scars form broad curved or straight segments of chasma wall up to 100 km long and locally recessed 5 to 10 km from the adjoining walls (Lucchitta, 1978). The scars are the only erosional landforms whose deposits are recognizable on the chasma floors, with slump blocks at the head and vast aprons of longitudinally ridged or smooth material (Fig. 5). The scars occur mainly in central Valles Marineris, between segments of dissected walls without clear structural control; however, they are also found in rectilinear segments of Ius, Hebes, Ophir, and Melas Chasmata, and in Coprates Chasma, and in scalloped troughs of western Valles Marineris (Tithonium Chasma). Whereas some landslide deposits occur on the chasma floors, especially in the straight and well-aligned troughs, many partly embay or bury the interior layered deposits, in the moats, below the level of the tablelands or benches.

The basal fault scarps, which are restricted to rectilinear segments of the main troughs, and especially to the northern sides (Schultz, 1991), suggest that faulting continued after the end of the gullying process. Gullying probably implies some kind of vertical erosion and longitudinal waste transport by fluids or viscous interstitial material, probably ice (Lucchitta, 1978), related to the widening of the Central Valles Marineris troughs during the late Hesperian (Lucchitta et al., 1992), and to the emplacement of interior layered deposits. Embayment relations inside the mouth of some gullies shows that most of the erosional work was completed before the end of deposition. In all cases, gullying predates the landslides, whose deposits generally obliterate and destroy spur-and-gully topographies. The formation of these relatively young features (middle to late Amazonian) was followed in places by faulting (Mège and Masson, 1996). Because the morphology of some landslides seems to be related to a certain content of groundwater or ice (Lucchitta, 1987), the



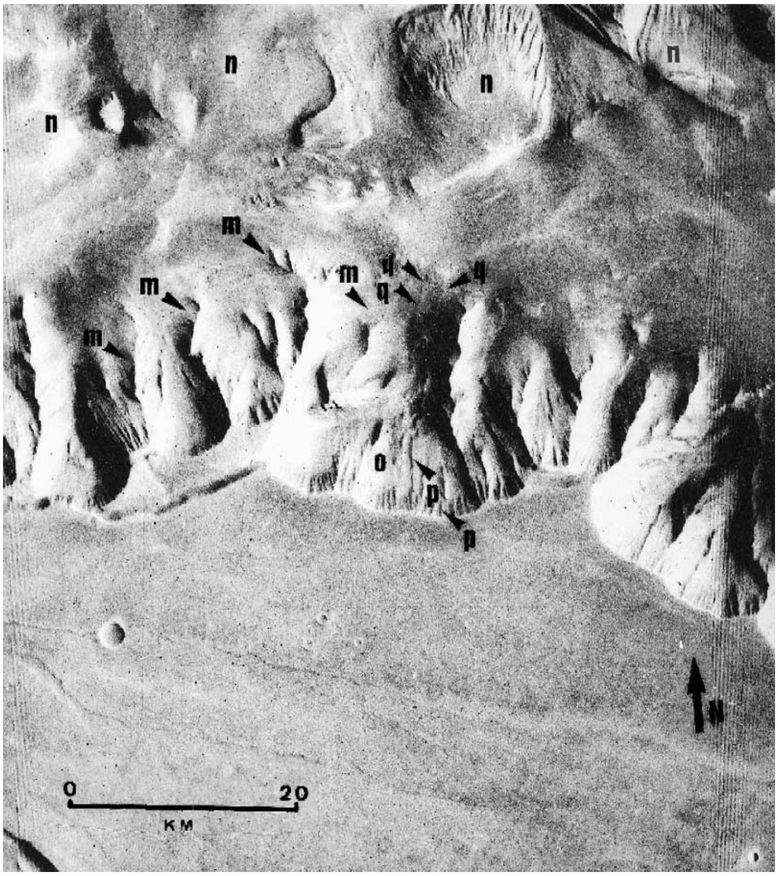


Fig. 4. Southern wall in eastern Candor Chasma. Note the morphological contrast of spur-and-gully topography between the northern (Fig. 6i) and southern walls (m), and the lack of basal scarps at the bottom of the southern wall. The spurs associated with a tectonic activity (Fig. 6) have more orderly branching and shorter crests than other walls. The debris from spur-and-gully erosion may have been incorporated into interior deposits (n) or is simply removed by unknown processes. Small, "recent" mass wasting processes, expressed on walls by talus slopes (o), and fine grooves on cornices and talus slopes (p), are responsible for debris tongues (q) at the base of the Valle Marineris walls. Viking picture 912 A 13 (49 m/pixel).

"wet" conditions of erosion and trough widening associated with landsliding must have been followed by "dry" conditions. The present mean annual temperature exceeds the frost point by 20 K in the study area (Clifford, 1993), and other current conditions imply a lack of ground volatiles (Squyres et al., 1992).

The distribution of the various types of very high wallslopes (Fig. 2) shows that the highest slopes are mainly restricted to the aligned troughs, which are

Fig. 3. Western Ius Chasma, displaying two grabens, located on the both sides of Geryon Montes. *North of Geryon Montes*: The southern wall displays spurs and gullies only (a), and the northern wall displays spurs and gullies associated with "ancient" faceted spurs (b) and "recent" basal structural scarp (c). All these features are cut by a landslide (d) on the left of the image. *South of Geryon Montes*: The whole northern wall displays a continuous fault scarp 1 km high (e), suggesting a "recent" origin. The southern wall displays a spur-and-gully morphology (f) and tributary canyons (g). Geryon Montes crests are locally split into two parts (h), suggesting the occurrence of slow mass movements. The white broken line underlines a 1-km topographic step on the trough floor. The two sets of small parallel arrows indicate the ends of two "recent" scarps about 800 m (left) and 500 m (right) high (from U.S. Geological Survey, 1986a,b) corresponding to 11% and 7%, respectively, of the total wall relief. Viking pictures 65 A 09, 12 (75 m/pixel).

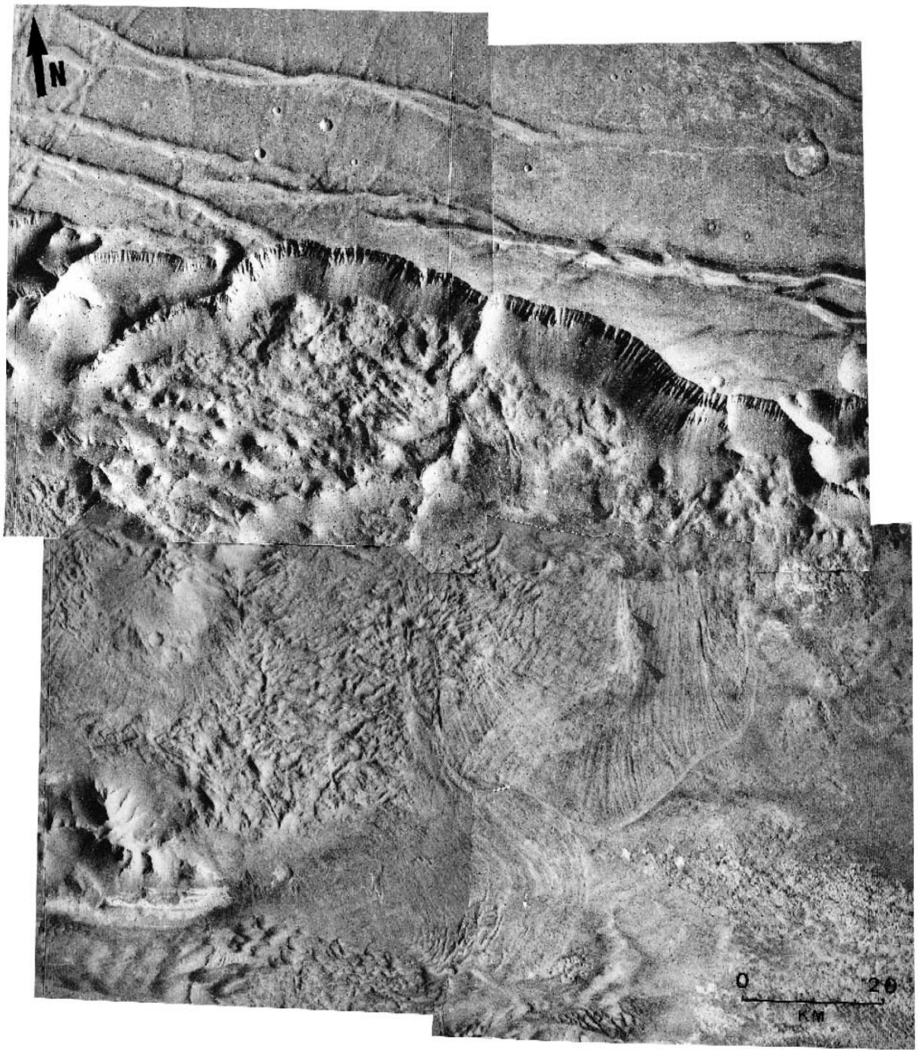


Fig. 5. Landslide in Ius Chasma described by Lucchitta (1979). Arrows: fissure-related volcanic flow. Viking pictures 919 A 15, 16, 17, 18 (46 m/pixel).

true grabens. The lesser depths of the highly scalloped troughs and pit crater chains are probably related to a different structural origin (Schultz, 1989; Lucchitta et al., 1990). Wall heights are also related to differences between the altitude of the Tharsis plateau and the altitude of the floors, which tend to decrease from west to east, from 8 km in parts of the Central Valles Marineris, to 3 km in East Coprates Chasma. In Central Valles Marineris and in Coprates Chasma, the northern walls are generally the highest. This is partly related to the location of the grabens on the southern side of the topographic crest that forms the eastern continuation of the Tharsis rise, and it may also be related to asymmetric faulting (Schultz, 1991; Peulvast and Masson, 1993b; Mège and Masson, 1996).

The morphological contrast between the widened Central Valles Marineris and the narrow troughs of the western and eastern parts of the system cannot be explained by structural and tectonic factors, because wall retreat without clear tectonic control appears to be the main mechanism of widening. No significant wall retreat and deposition is related to the deep dissection by tributary canyons of Ius and Tithonium Chasmata, whereas gullying of the walls and deposition of thick layered deposits are chronologically, if not genetically, linked in the widened troughs. Such relationships suggest complex spatial heterogeneity in the properties of the wall materials, including composition, structure, shear strength, volatile content, and time-related changes in the erosional/depositional processes and in the structural conditions of trough formation. The long-lived resurfacing history of Mars (Tanaka et al., 1992) suggests that the morphological diversity of the Valles Marineris walls involves the coexistence of landforms shaped during several stages of development, which may be revealed by geomorphological study.

3. Structural interpretation

3.1. Valles Marineris and the origin of Tharsis

The Ius, Melas, Coprates, Candor, Ophir, and Hebes Chasmata have a structural origin that is likely to be linked to the Tharsis tectonics. Using Valles Marineris as a constraint on Tharsis stress modeling, Sleep and Phillips (1979) showed that isostasy is consistent with the observed martian gravity field. The structural patterns expected from this mechanism, however, are not fully consistent with the location and orientation of Valles Marineris. The latter may be explained by modeling a succession of tectonic stages, first isostatic, followed by flexural (Banerdt et al., 1982; Willeman and Turcotte, 1982; Solomon and Head, 1982; Sleep and Phillips, 1985; see reviews in Banerdt et al., 1992; Mège and Masson, 1996). Alternatively, Tanaka et al. (1991) and Banerdt and Golombek (1992) hypothesize a synchronous mechanism based upon a different lithospheric structure beneath the internal and external parts of the Tharsis surface. Convective forces may have been important in the earliest history of Tharsis prior to later flexural relaxation (Kiefer and Hager, 1989; Schubert et al., 1990). Because the most appropriate model has to be consistent with structural data, inferred from study of morphological features, especially along the walls of Valles Marineris, the morphotectonic interpretation of these walls is critical to the geophysical understanding of Tharsis.

A body of arguments suggests a major change in morphogenetic processes on Mars at the Hesperian/Amazonian boundary (e.g. Lucchitta, 1984; Baker and Strom, 1992; Parker et al., 1993). This distinction between successive morphogenetic conditions will be referred to as "ancient" and "recent" in the discussion that follows.

One of two main extensional patterns for Valles Marineris consists of triangular faceted spurs, developed on fault scarps below the faceted spur-and-gully topography. Terrestrial faceted spurs indicate normal faulting (e.g. Hamblin, 1976; Wallace, 1978), and morphological comparisons between the terrestrial and martian features are striking (Figs. 3, 6, and 7). The Valles Marineris faceted spurs probably represent erosion of normal fault scarps under the "ancient" conditions. Faceted spurs in regions with especially clear tectonic control (e.g. eastern Candor Chasma) display spur-and-gully topography with more orderly branching than occurs on the walls, which are not fault-controlled. They also have shorter crests (Figs. 4 and 6), which can be compared to normal fault scarps subject to erosion on Earth (Fig. 7).

A second extensional pattern is expressed as normal fault scarps at the wall bases. These scarps evolved during conditions of reduced erosion, probably by wind, and they probably formed under current ("recent") conditions. For the prolonged and continuing tectonic activity of Valles Marineris, "ancient" features are expected to be observed on the upper parts of the walls, whereas "recent" features are expected to be restricted to the lowest parts.

3.2. The Ius Chasma example

A detailed description of Ius Chasma illustrates the morphological distinction of "ancient" patterns (Fig. 8) from "recent" patterns (Fig. 9). We studied about 150 pairs of Viking stereo images, with resolution mostly ranging from 40 to 75 m/pixel. East of Calydon Fossa and west of 83°W, the southern Geryon Montes wall is entirely composed of a continuous, 1 km high fault scarp (Figs. 3 and 9). Lacking faceted spurs, this scarp probably only expe-

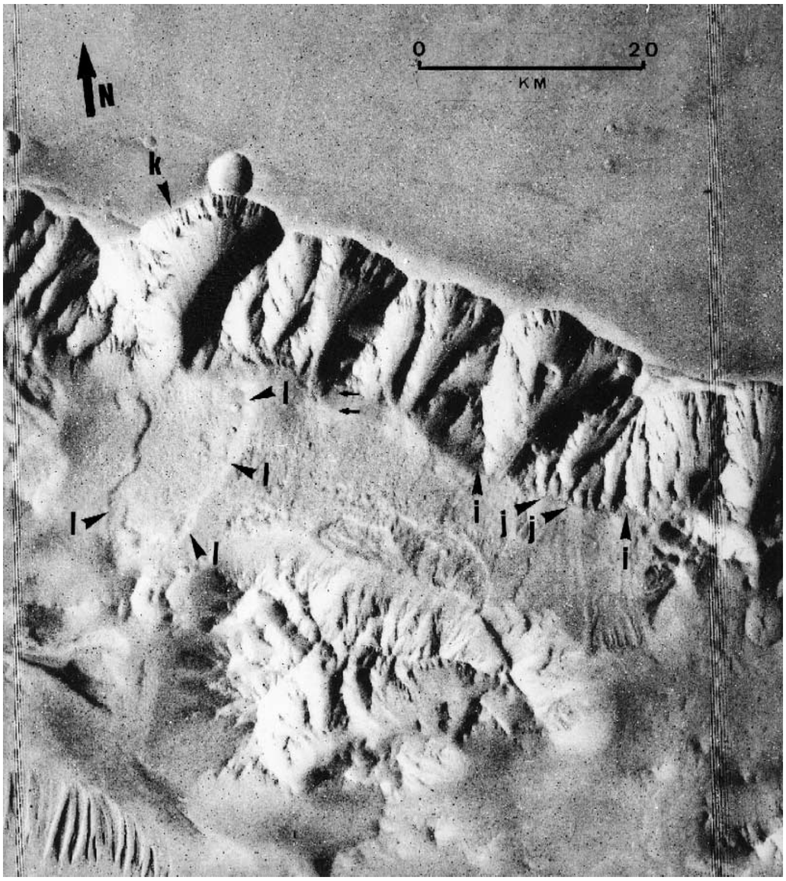


Fig. 6. Northern wall in eastern Candor Chasma, showing clear tectonic control, expressed by "ancient" faceted spurs beneath spurs and gullies (i), and a "recent" basal scarp (j). (k): Small, "recent" landslide cutting across the "ancient" spur-and-gully morphology, and probably across "recent" basal scarp. (1) Landslide material. Small parallel arrows indicate a "recent" scarp about 800 m high (U.S. Geological Survey, 1986a,b), corresponding to about 10% of the whole wall relief. The "recent" j scarp is about 400 m high and corresponds to about 5% of the wall relief. Viking picture 911 A 12.

rienced "recent" morphogenetic conditions. If the scarp were older, its height would likely have allowed spur-and-gully morphology to develop. Although this scarp has been well preserved, the wall to the south is intensely eroded down to its base, indicating an older age and the possible occurrence of a volatile-rich wall material that facilitated erosion during trough formation.

The most common case encountered is the occurrence of spur-and-gully topography, and the triangular facets on the higher portions of the trough walls. "Recent" fault scarps are located on the lower portions (Fig. 3). Sometimes, however, triangular facets are located at the basal part of the slope, and "recent" scarps are located inside the trough (for example, inside the "ancient" trough geometry; Fig. 9B). This "recent" deformation probably occurred on faults different than "ancient" boundary faults, and it locally defines narrower grabens.

The most important process to occur after the formation of basal scarps is landsliding. Gullying and sapping generally occurred before landsliding (during early Amazonian for sapping, according to Witbeck et al., 1991), perhaps just at the turning point of the erosional condition change. Although sapping is often difficult to interpret except for the dendritic network area in Ius Chasma (Louros Valles), some observations indicated that sapping

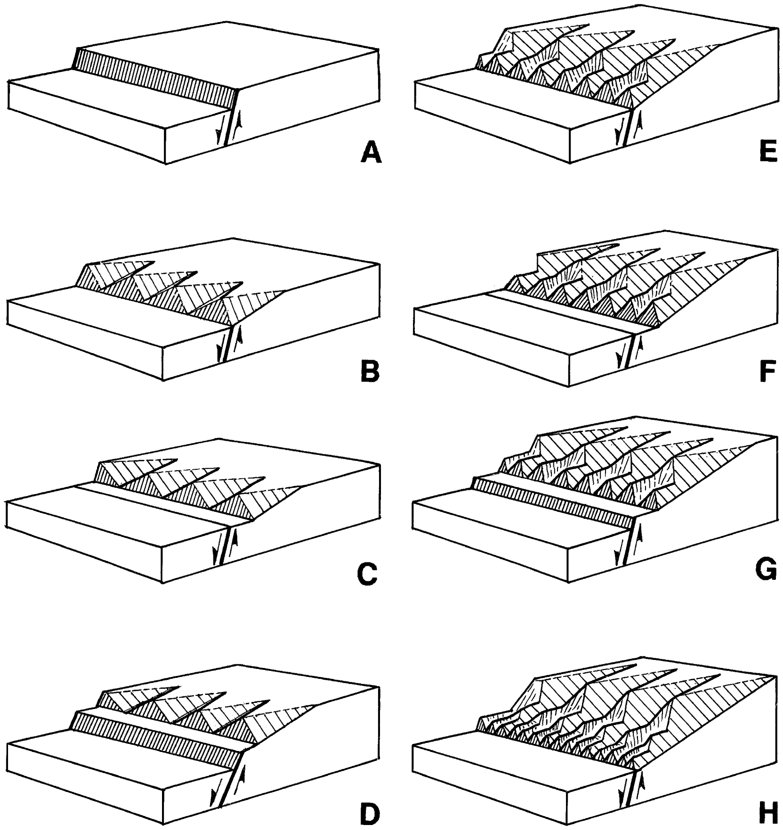


Fig. 7. Diagrams showing the evolution of faceted spurs produced by periods of movement separated by periods of stability on Earth. (A) Undissected fault scarp; (B) development of faceted spurs by streams cutting across scarp; (C) period of stability with slope retreat and development of a narrow pediment; (D) recurrent movement; (E) dissection of new segment of scarp by major streams and by those developed on the face of faceted spurs formed in B; (F) new period of stability with slope retreat and development of another narrow pediment at base of mountain front upthrown block; (G) recurrent movement; (H) dissection of scarp formed in G, resulting in a line of small faceted spurs at base of mountain front. Remnants of narrow pediments are preserved at apices of each set of faceted spurs (redrawn from Hamblin, 1976). Compare the final morphology shown at stage H with the i and j sites of Fig. 6. Moreover, note the similarities between stage G and the profile defined by b and c in Fig. 3. Whereas, stage G corresponds to fault reworking before fault scarp erosion on Earth, this stage corresponds in Valles Marineris to the "recent" conditions, with a reduced erosional activity. Stage H is not further expected as far as the current morphogenic conditions exist on Mars.

might have locally lasted until the "recent" period. For example, on the southern side of Geryon Montes, some "recent" sapping might have cut across both spur-and-gully topography and basal scarps. Locally, some tributary canyons wholly or partially cut across basal scarps, and others hang above them. The chronology between the last tectonic activity and mass-wasting processes is thus variable, perhaps because both are contemporaneous at the geological scale.

Although the "ancient" fracture set was partly destroyed by the later geomorphological events or covered by deposits, it is considered to be responsible for the major part of the extensional deformation. Vertical offsets from the "recent" period are usually limited to a few hundred meters and only exceptionally exceed 1 km. Two main orientations are recognized in the fault patterns (Figs. 8 and 9): (1) east-trending, following the main walls of Ius Chasma, and (2) oblique to the trend (1).

The east-trending fault orientation extends from the boundary between Noctis Labyrinthus and Ius Chasma to 83°W. Except for the "recent" walls, all of the walls of Ius Chasma have a structural and "ancient" origin. Deformations under the "recent" conditions are shared among boundary faults and other faults of various directions. East of 83°W, the Ius troughs follow the N75°W direction, up to Melas Chasma. All the trough walls north and south of Geryon Montes have an "ancient" genesis. East of Geryon Montes, the "recent" trough does not display a clear graben geometry, for example, boundary faults cannot be continuously followed.

The oblique fault set includes northeast, northwest, east, and north orientations. Both the northeast and northwest directions are strongly expressed from the boundary with Noctis Labyrinthus to 83°W. Farther east, the northwest trend becomes more and more predominant. West of 83°W, the east trend is the boundary fault orientation, whereas this trend links right-stepping N75°W trending faults at the base of the northern wall between 73°W and 76°W. Locally, several faults in the southernmost part of Melas Chasma follow the north trend.

North of Geryon Montes, widespread landslides complicate recognition of the "ancient" structural geometry. Nevertheless, left- and right-stepping faults developed "anciently" below the northern wall from 73° to 82°W. The "recent" tectonic patterns form a more complex fracture system. The landslides generate numerous northeast- and northwest-trending features, whereas inter-landslide areas preserve the dominantly trough-trending boundary faults. The landslide faults may not have any link with gravity-controlled movements, which would be oriented preferentially perpendicular to the sliding direction (e.g. Jibson and Keefer, 1993). Both northeast- and northwest-trending faults are aligned with tributary canyons to the north and south, which align along fault sets that predate Sinai Planum, (for example, Noachian in age; Scott and Tanaka, 1986). These may result from faulting, whose offset is concealed by the block morphology, or from the collapse of wet landslide material, preferentially accenting old, Noachian weakness zones. Some northeast- and northwest-trending faults were active during the "ancient" period, and these may also display inherited Noachian trends.

South of Geryon Montes, Ius Chasma may be divided into four major graben (G1 through G4). The

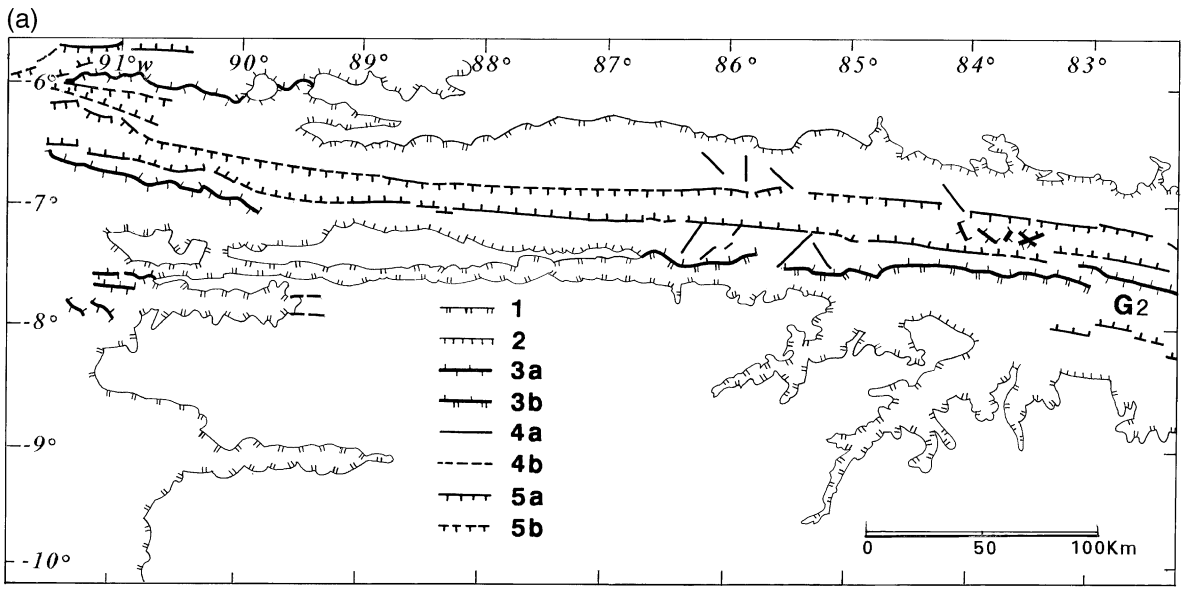


Fig. 8. Location of "ancient" structural features in eastern (a) and western (b) Ius Chasma from morphoclimatic arguments. *Morphology*: (1) Cornice with smooth talus slope; (2) cornice with spurs and gullies; (3a) crest with spurs and gullies on both sides; (3b) crest with spurs and gullies on one side and smooth talus slope on the other side. *Structure*: (4a) Fault; (4b) conjectured fault; (5a) normal fault; (5b) conjectured normal fault. G2–G4: Graben subdivisions south of Geryon Montes, discussed in the text.

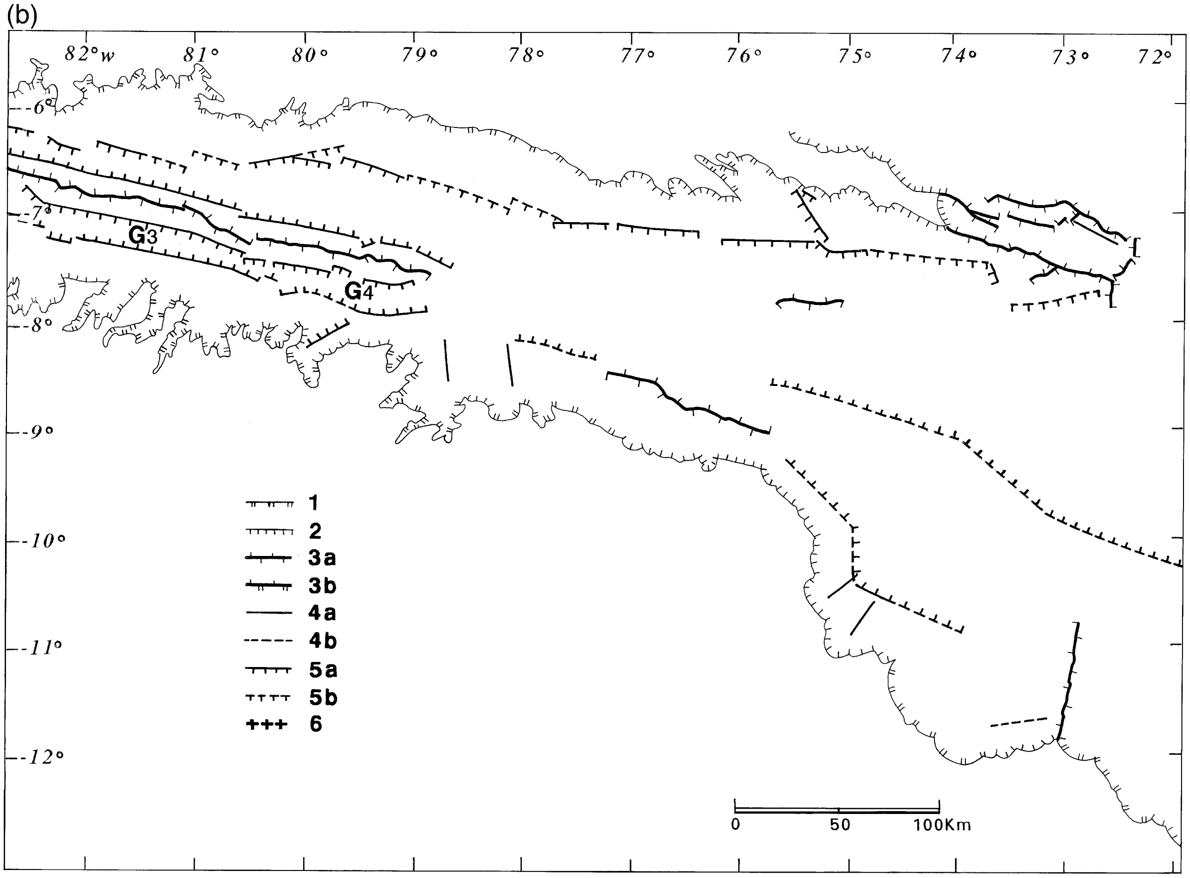


Fig. 8 (*continued*).

westernmost graben partly formed under "recent" conditions (Fig. 9: G1), and its floor is cut by a northwest-trending topographic step, (Fig. 3), 1000 m high (U.S. Geological Survey, 1986a,b). The second graben (Figs. 8 and 9: G2) is bordered by faults stepping southward, has a poorly constrained wall geometry, and is shifted northward compared to the third graben (Figs. 8 and 9: G3). Considering the "ancient" faults responsible for the main offsets, the transition zone between G2 and G3 displays northwest-trending extensional faults, with triangular faceted spurs. The third graben is connected with the fourth one (Figs. 8 and 9: G4) by two east–northeast-trending extensional faults. The graben orientation reverts back to a north trend, south of Geryon Montes.

East of Geryon Montes, the N75°W boundary fault direction continues into West Melas Chasma up to 73.5°W, but this trend is underlined by few fault traces. Geryon's gently east-dipping topography close to Melas Chasma may indicate that the horst approximately ends at its current topographic terminus, implying lesser extension eastward than westward in Ius Chasma. Conversely, this horst may have undergone increasing extension eastward, leading to complete burial under more recent sediments at its current topographic terminus. The latter hypothesis is supported by observations that the Geryon Montes boundary fault set may well prolongate further East, with fragmented segments illustrated in Fig. 9.

Both the "ancient" and "recent" fault sets show some northwest control, which becomes especially strong between 72°W and 75°W. This trend also occurs in Melas Chasma (Peulvast and Masson, 1993a,b). The "recent" tectonic activity appears to be responsible for very small offsets. The north-trending faults in southwest Melas Chasma may be a structural expression of sediment packing, together with undulations observed by Peulvast and Masson (1993a,b).

3.3. "Ancient" and "recent" tectonics in Valles Marineris

A striking morphological feature is the clear prevalence of "recent" deformation at the base of the northern walls rather than southern walls. This is especially true for Candor and Ophir Chasmata, and in parts of Ius, Coprates, and Tithonium Chasmata. Spurs and gullies are often located on the upper slopes and lie above continuous basal scarps (Figs. 3, 6, and 10). Landslides have locally destroyed the previous spur-and-gully morphology (Fig. 5), and they postdate basal scarps. Triangular faceted spurs seem to be widespread on the northern slopes above basal scarps. In contrast, the southern walls display few traces of either "ancient" or "recent" tectonic activity. The lack of scarplets suggests that "recent" tectonics is reduced in comparison to the northern wall tectonics. "Ancient" faulting may have occurred, however, and been obliterated by further erosion. For instance, a part of the western Candor south wall has sapping morphological characteristics (Sharp, 1973; Lucchitta, 1978; Kochel et al., 1982) or collapse features (Baskerville, 1982; Tanaka and Golombek, 1989). This morphology is further developed at the western border of Candor Chasma and linked to sapping channels (Fig. 11). In these regions, the current morphology is that of a fault scarp which underwent significant retreat, probably during "ancient" as well as "recent" stages of wall development. The original boundary fault may coincide with the pit chain joining Candor and Tithonium Chasmata.

The current rectilinear morphology of some interior deposit boundaries within East Candor Chasma suggests that the "ancient" Candor development might have been influenced by transverse structures (Fig. 10). The borders of the western branch of Candor Chasma are not exactly aligned with the borders of the eastern branch. The central Candor Chasma area might mask a buried transition zone, perhaps responsible for the shifting of the Candor

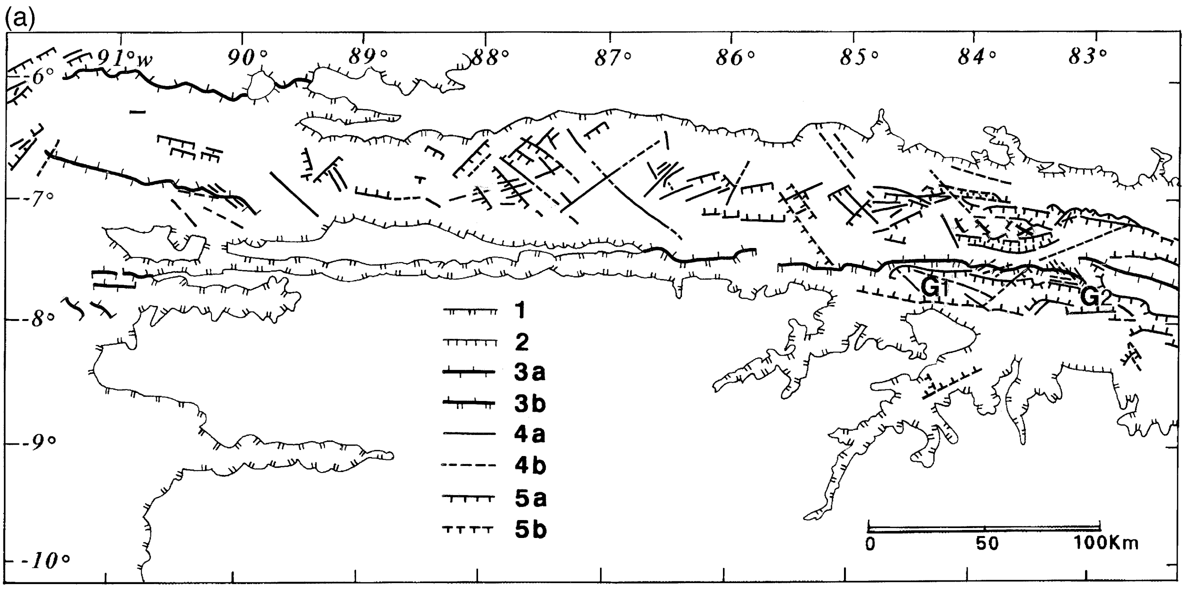


Fig. 9. Location of "recent" structural features in eastern (a) and western (b) Ius Chasma from morphoclimatic arguments. Symbols as in Fig. 8; (6) second-order drag fault (Mège and Masson, in preparation). G1–G4: Graben subdivisions south of Geryon Montes, discussed in the text. (A) Faults possibly connecting Geryon Montes boundary faults and Melas Chasma internal structures. (B) Examples of "recent" extensional movements in grabens on faults, which are different than "ancient" boundary faults. The two easternmost A faults may be correlated with a 2- to 3-km high topographic scarp (U.S. Geological Survey, 1986a,b).

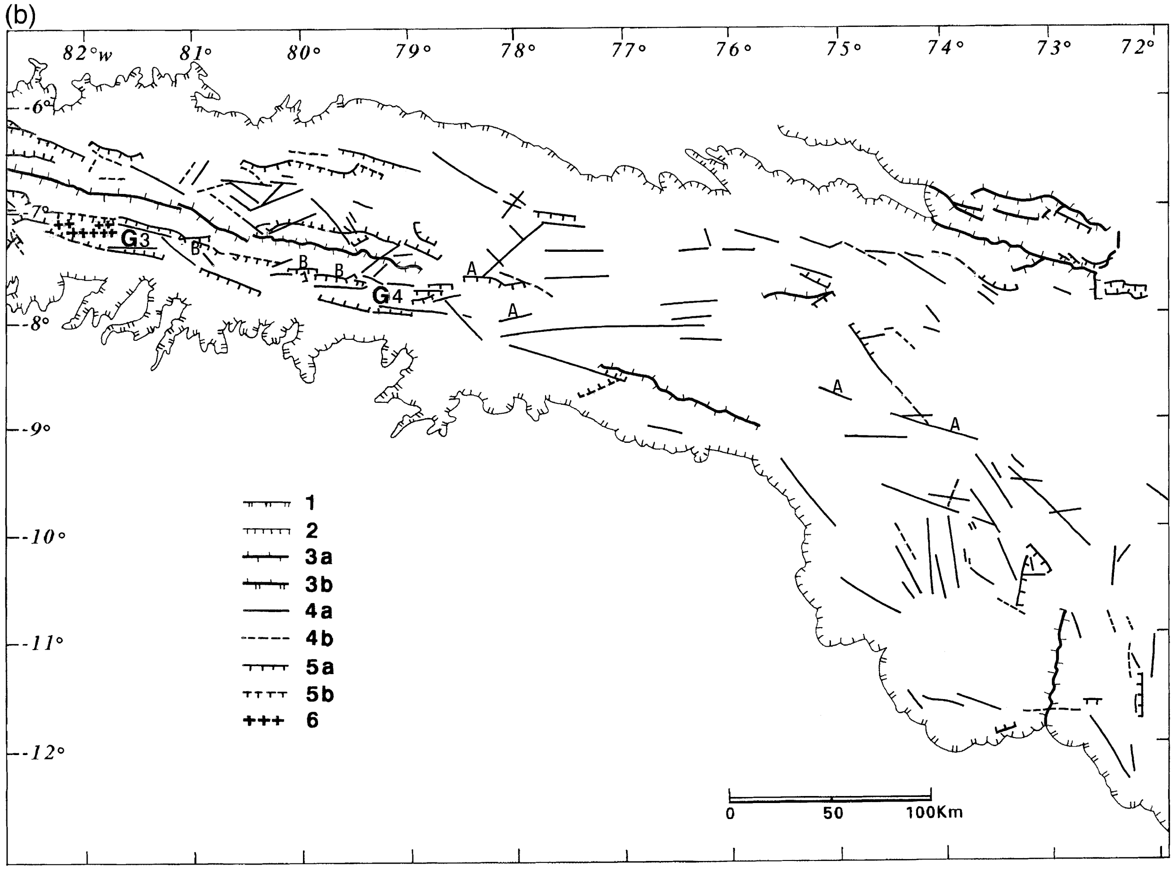


Fig. 9 (*continued*).

west and east grabens, as in Melas Chasma (Peulvast and Masson, 1993b).

Coprates Chasma has a complex structure of alternating horsts and grabens. Several horsts occur close to Melas Chasma and close to Eos and Capri Chasmata. Five major grabens may be observed (Fig. 10b: G5–G9), including one (G6) whose original geometry was destroyed by landsliding on the southern wall. In contrast to Ius, the oblique trends between these grabens cannot be correlated to tributary canyon directions (for example, ancient buried structures). "Ancient" and "recent" faulting is widespread, except in the pit chains. "Recent" movements are reduced at their proximity.

Despite the local occurrence of spur-and-gully topography, little clear evidence exists for faceted spurs in Tithonium Chasma. Close to Noctis Labyrinthus, Tithonium looks like Ius Chasma, a trough whose morphology is primarily controlled by tectonics and further enlarged by erosion (Figs. 1, 3, 10, and 12). The southern wall of this part of Tithonium Chasma displays spurs and gullies and some possible faceted spurs, whereas the northern wall displays basal scarps under a very degraded upper slope with spur-and-gully remnants. This part of Tithonium is clearly a graben, in which "recent" tectonic activity occurred on the northern wall only. The morphology of Tithonium in the other parts of the trough seems to be primarily related to erosional processes and secondarily to tectonics. Some other "recent" faulting is reflected in two discontinuous trends following the trough direction at the bottom of both walls (Fig. 10b).

The structural control of Echus Chasma is not clear. "Ancient" east-trending tectonic structures may have existed, like those in the Hebes trough. Some

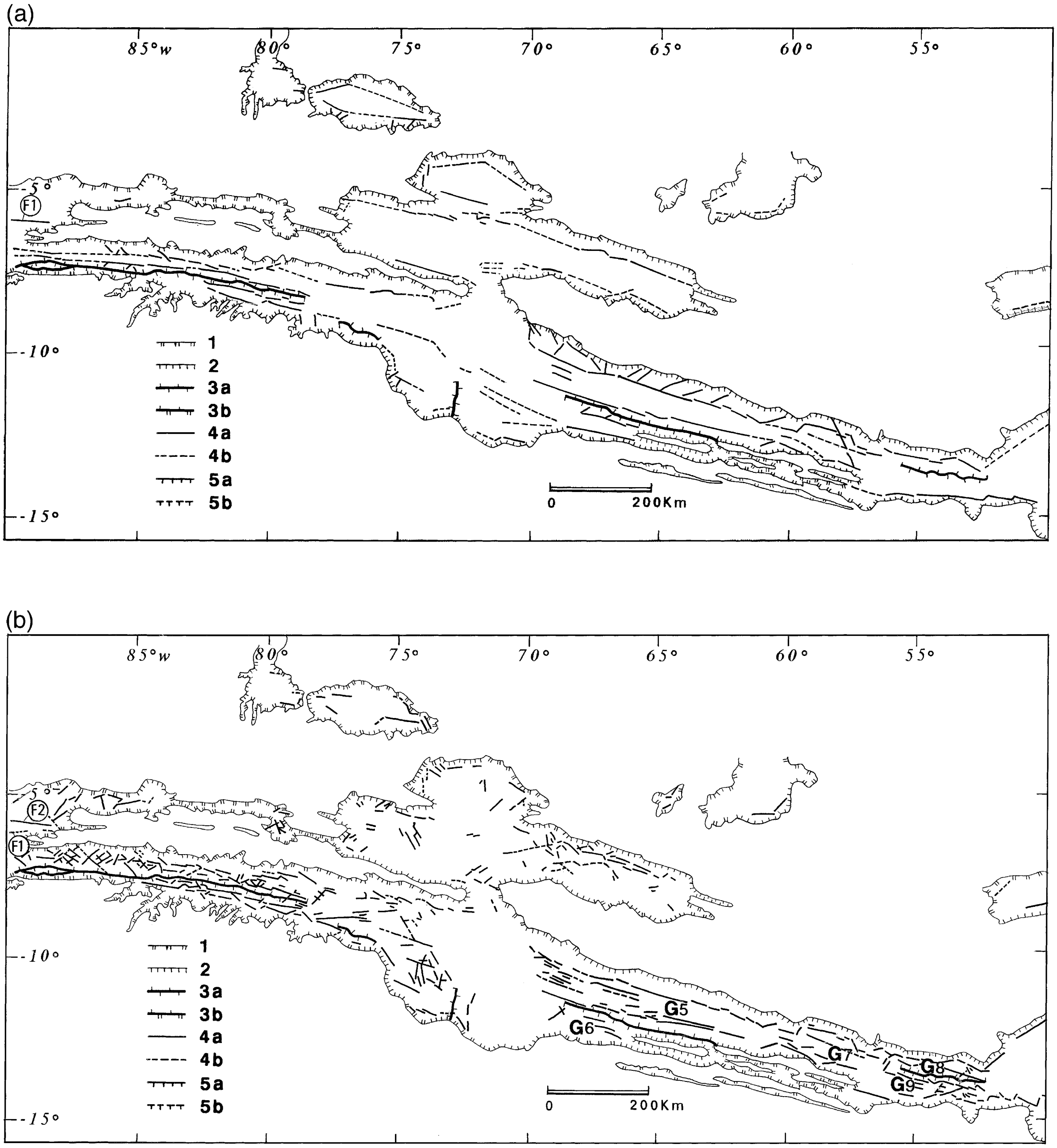


Fig. 10. (a) Generalized Valles Marineris tectonic patterns formed under "ancient" morphogenic conditions. b: Generalized Valles Marineris tectonic patterns formed later, under "recent" morphogenic conditions. Symbols as in Fig. 8. G5–G9: Graben subdivisions in Coprates Chasma mentioned in the text. F1, F2: Faceted spurs on the Tithonium Chasma southern wall (F1), and northern scarp on (F2) in Tithonium Chasma, discussed in the text.

evidence of slight "recent" evolution of Echus Chasma occurs on the southern wall. The further evolution of Echus Chasma may have included the same events as in Eos and Capri Chasmata, such as

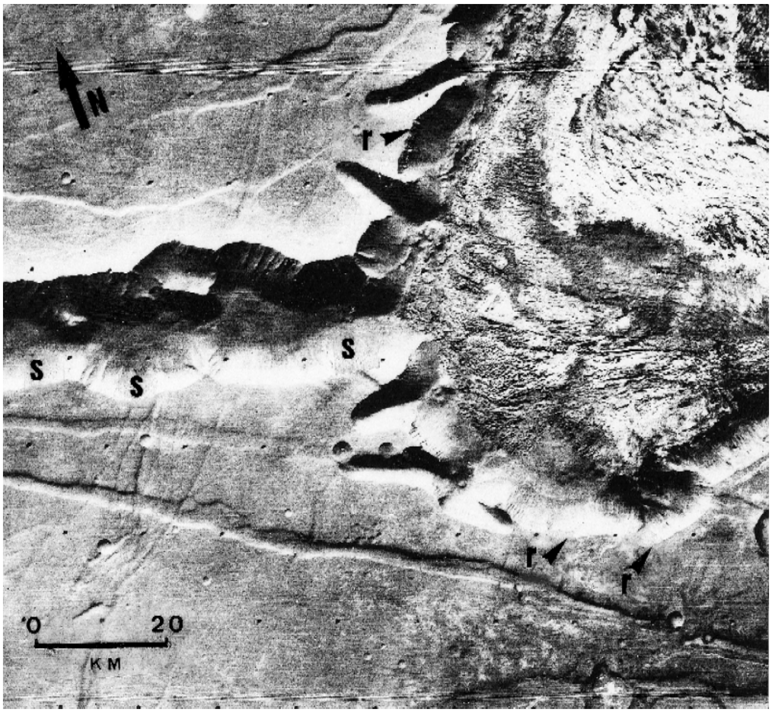


Fig. 11. Southwestern and western walls of Candor Chasma, displaying a collapse-like geomorphology (r). The southern wall might have originally followed the pit-crater chain (s) trend at the West, and developed a spur-and-gully morphology, further destroyed by another process, such as collapse. Viking picture 66 A 18 (75 m/pixel).

catastrophic flood discharge, which formed Kasei Valles and modified Chryse Planitia.

3.4. Relative magnitudes of "recent" and "ancient" tectonism

Because central Valles Marineris troughs are likely to have formed originally as grabens, it is possible to estimate the relative relief generated by "recent" tectonic movements, in contrast to that attributed to "ancient" tectonics. For instance, in the western Ius area (Fig. 3), "recent" scarp height measurements (U.S. Geological Survey, 1986a,b) give 800 and 500 m at two different sites, corresponding to 11% and 7% of the entire wall relief, respectively, assuming that deposits are thin near these walls (which is likely because of the layered deposits, e.g. Schultz, 1991). The "ancient" movements are, thus, expected to be responsible for 89% to 93% of the tectonic relief at these sites. The following values for the "recent" period are approximate maximum values in other central Valles Marineris sites: (a) 300 m (6%), (b) 400 m (5%), (c) 800 m (10%), (d) 1000 m (14%), and (e) 800 m (15%), respectively, for (a) Tithonium (Fig. 12), (b) and (c) Candor (Fig. 6), (d) Ophir, and (e) Hebes Chasmata northern walls (The Hebes southern wall might have undergone movements comparable to those on the Hebes northern wall). The remaining percentages correspond to what results from "ancient" movements. The amounts of "recent" movement tend to increase northward, from Ius to Hebes Chasmata, but this trend needs to be confirmed by further study.

3.5. Volatile distribution and wall development

The following features have been used to infer the present or past occurrence of volatiles in Valles Marineris region: (1) rampart craters in the surroundings of the troughs (Costard, 1990), (2) the tributary canyons, most probably formed by sapping (Kochel and Piper, 1986), (3) landslide deposits in Ophir–

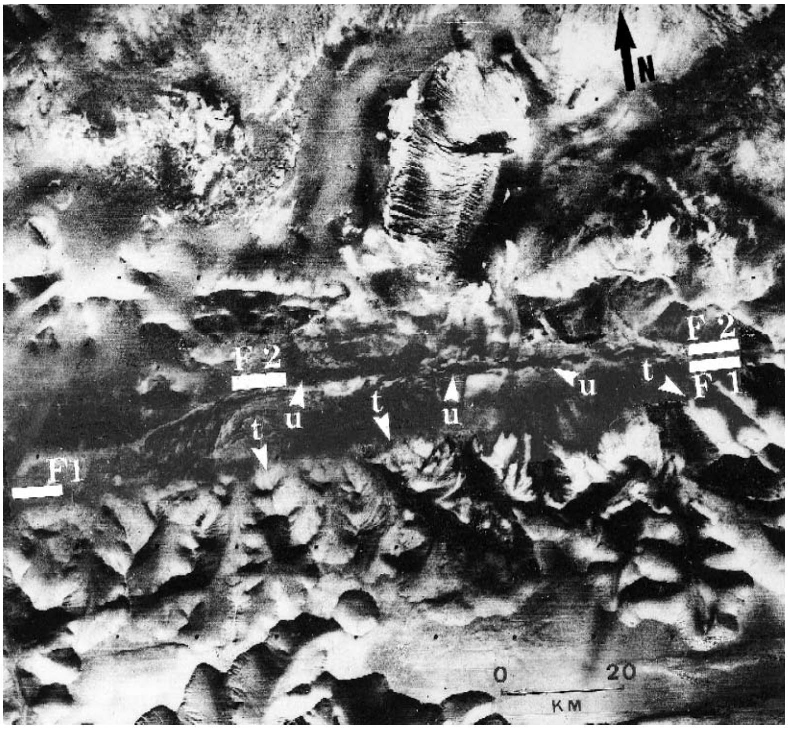


Fig. 12. Eroded faceted spurs (t) and basal scarps (u) in Tithonium Chasma near Noctis Labyrinthus. F1, F2: See Fig. 10 and the text. The u scarp is about 300 m high (U.S. Geological Survey, 1986a,b) and corresponds to about 6% of the whole slope gradient, assuming that the northern wall originally reached the same height as the southern wall. Viking picture 63 A 63 (75 m/pixel).

Candor Chasmata may have been emplaced as "gigantic wet debris flows" (Lucchitta, 1987), (4) the layered deposits, which may have been emplaced in ice-covered lakes (Nedell et al., 1987), and (5) the morphology of certain gully-mouth lobate deposits, which may be related to the presence of interstitial ice (Peulvast and Masson, 1993b). All of these features are most abundant in the central Valles Marineris troughs. Although "dry" processes have also been imposed for explaining some of these features (McEwen, 1989), the general view is that volatiles or ground ice were involved in the formation and erosional widening of the canyons (McCauley, 1978). Fanale (1976) estimates that the 3- to 7-km wall heights coincide with the depth of the permafrost, such that landslides would involve volatile-rich layers present at depth or would correspond to the emergence of underground aquifers (Carr, 1979). Battistini (1985) proposed a strong association between the ground ice and the morphology of Valles Marineris, linking collapse depressions, aligned along grabens, to the presence of volatile rich materials.

An objective sampling of the volatile distribution around the Valles Marineris may be accomplished by studying the distribution and morphological characteristics of rampart craters, which have lobate ejecta terminating distally in ramparts (Mouginis-Mark, 1979; Horner and Greeley, 1987). This morphology implies emplacement by flows around craters over the surface just after the impact event through the melting of volatiles (Carr et al., 1977; Mouginis-Mark, 1979). Thus, rampart craters are considered to be excellent ground-ice indicators (Squyres et al., 1992). In our study, all rampart craters were surveyed within $\pm 5°$ of the Valles Marineris. Data were collected for 62 rampart craters in the size range 5–40 km located on the plateau surfaces above the troughs. We did not survey the small number of impacts on the trough floor. As a first hypothesis, we infer that lobate ejecta located "near" the troughs (less than 50 km) reflect the nature of terrains that form the Tharsis plateau around Valles Marineris and the wall rocks.

We assume that the volatile content of the excavated material directly influences the ejecta mobility at the time of the impact event, as shown experimentally (Gault and Greeley, 1978; Wohletz and Sheridan, 1983). Because the extension of ejecta lobes is theoretically proportional to the volatile content (for appropriate material properties, porosity, etc.), ejecta mobility (EM) can be expressed by the ratio of the maximum diameter of ejecta deposits normalized by the diameter of the parent crater (Mouginis-Mark, 1979; Battistini, 1984; Kuzmin et al., 1988). The EM ratio indicates the volatile content of ground-ice at the time of the impact event. EM ratios of 2 to 3 indicate low-mobility ejecta, while EM ratios of 5 to 7 indicate high-mobility ejecta.

Fig. 13 shows the characteristics of individual rampart craters around Valles Marineris. Fig. 14 shows variations of the EM ratio for 62 rampart craters, according to their location in longitude along the troughs. The two regression lines in Fig. 14, showing the EM variation with longitude, stop between 70°W and 75°W because of insufficient data in that region. The scatter in the data may result from target heterogeneity, variations in the total ice content of the excavated material, the relatively sparse data set, and/or related variations in the content of ground ice. Fig. 13 shows a clear relationship between ejecta mobility (located at 5° on each side of the troughs), and the distribution of the widened parts of the troughs with the relative abundance of volatile-related erosional landforms. Rampart craters with high ejecta mobility deposits (EM values close to 4.5) predominate, especially south of Melas Chasma (the widest trough), but the number of mobile-ejecta craters decreases both to the west and east of the central Valles Marineris, except in the Louros Valles area. This same trend is reflected in the regression lines of the EM ratios in Fig. 14.

Changes in ejecta mobility appear to be best explained by a progressive enrichment in volatile materials within the Tharsis region around the central Valles Marineris (Costard, 1990). The widening of the chasmata in this region supports this interpretation. The inferred high volatile content would pro-

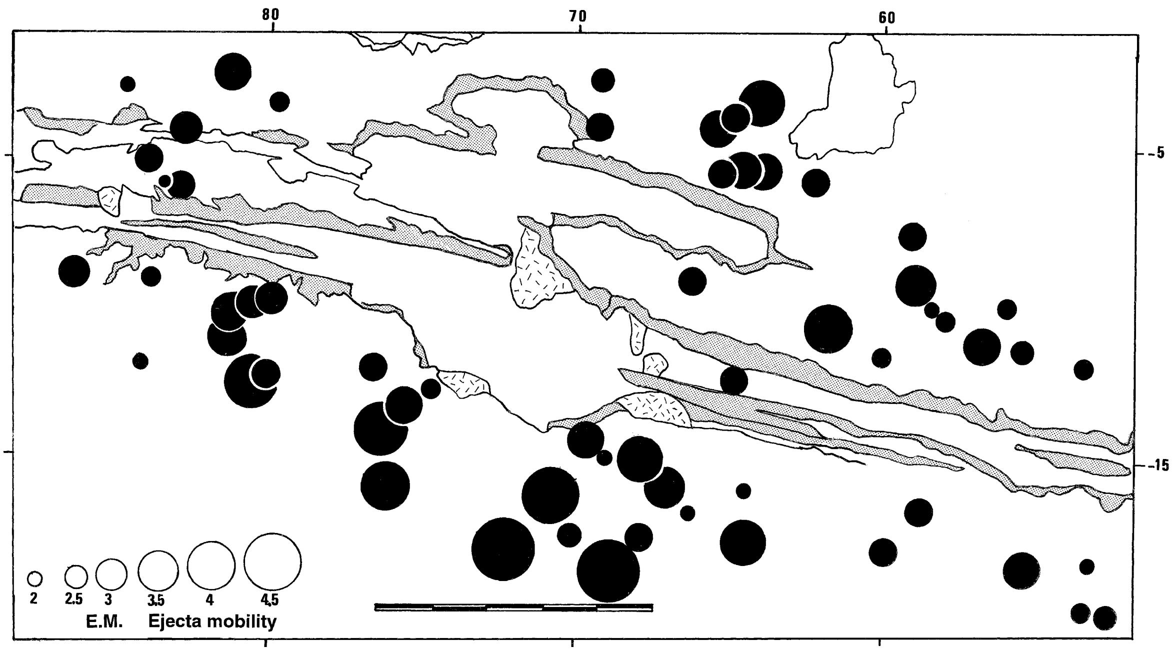


Fig. 13. Distribution of rampart craters within ±5° of the Valles Marineris. Ejecta mobility is calculated using EM ratio of the maximum range of ejecta lobes normalized to the diameter of the parent crater.

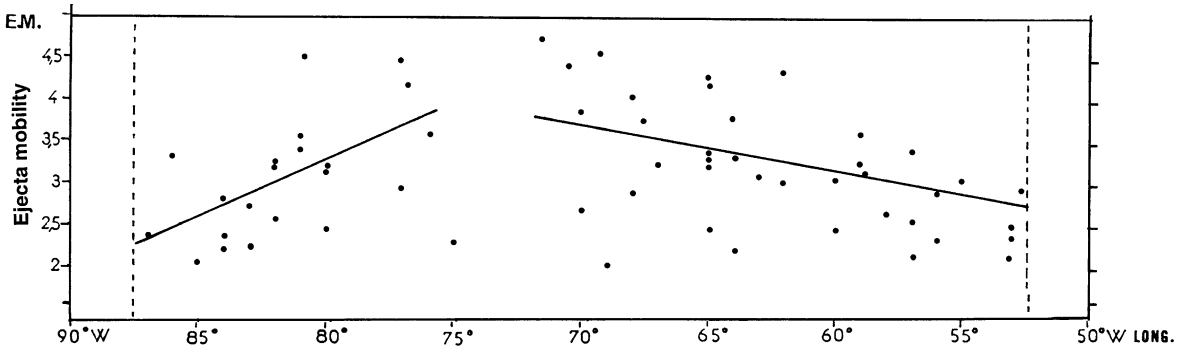


Fig. 14. Variations of the EM ratio (y-axis) for 62 rampart craters according to their location on the canyon (x-axis). Both regression lines exhibit a clear general rise in the EM ratio (ejecta mobility) towards the central part of the canyon. Note the enrichment in volatile materials from the margins to the Central Valles Marineris. This concentration of volatiles may have contributed to the widening of the chasma.

vide a plausible mechanism for several erosional processes including the very large wet debris flows that seemed to occur until middle to late Amazonian (Lucchitta and Bertolini, 1989). The reduction of wall retreat outside the central Valles Marineris region may be explained by the presence of a volatile poor permafrost and/or be the presence of more coherent and less porous wall materials.

4. Discussion and conclusions

Valles Marineris exposes some of the deepest vertical sections in the crust of any planetary body, providing a unique morphotectonic record for understanding the structural development of the upper martian crust. Although the range of slope processes is not yet fully understood because of uncertainties concerning the nature and evolution of the ground volatiles, gullying and widening processes, and the role of water or ice in mass movements, the regional evolution fundamentally underwent different stages from Hesperian to the present. A rather stable opening strain regime persisted in Valles Marineris for a long time (at least from Hesperian to present), which has major implications for the 5000-km Tharsis bulge. Moreover, some important differences occur between terrestrial rifts and Valles Marineris, including the lack of triple junctions (a major rifting process on Earth), transfer faults, and shifting major boundary faults. Like terrestrial rifts, however, prominent asymmetric wall topography occurs in Valles Marineris (Schultz, 1991). Finally, the major change from "ancient" to "recent" wall development, together with the change from possibly "wet" Noachian climatic conditions (e.g. Pollack et al., 1987) to the drier Hesperian and Amazonian conditions, are necessarily linked with other processes of major importance for the history of Mars.

The detailed study of a particular graben, Ius Chasma, reveals a long-lasting history and a complex geometry, with several imbricated grabens and horsts. Previous oblique structural trends, possibly Noachian, are included in this geometry.

The northern slopes of some tectonically controlled parts of Valles Marineris (Candor, Ophir, and Hebes Chasmata) underwent intense faulting during the Hesperian and Amazonian. In contrast, the southern slopes mostly have a poorly constrained "ancient" tectonic history, and the "recent" history is only characterized by rather weak erosion. The southern wall of Coprates Chasma and part of Hebes Chasma were most deformed, whereas the southern wall of Candor Chasma was less deformed. Similar to the conclusions of Schultz (1991) for Coprates Chasma, the morphotectonic study of the central Valles Marineris and Ius Chasma indicates that the former geometry of Valles Marineris may have been that of an asymmetric parallel graben set. This asymmetric rift evolved with varying active geomorphological processes, possibly related to a thermal anomaly centered under the Melas and central Candor Chasmata (Peulvast and Masson, 1993b).

Preliminary calculations suggest that, assuming thin floor deposits near the walls, the maximum "recent" extensional movements are about 5% to 15% of the wall relief, leaving 95% to 98% of the relief separated by tectonic movement. Although these values may not be representative of the whole trough system, the lowest values generally correspond to the southern troughs, and the highest, to the northern ones.

Ius and Coprates Chasmata, the simplest and most-aligned grabens, appear to have recorded slightly different deformational histories than the above-mentioned troughs. The southern walls clearly underwent "ancient" and "recent" normal faulting, and asymmetry is less apparent than in the other troughs. The western part of Tithonium Chasma also displays graben-like structures, but the evolution of its eastern part was dominated by geomorphological processes similar to those responsible for the scalloped troughs (Tanaka and Golombek, 1989).

The distribution of mobile ejecta in relation to the erosional landforms of wall retreat or dissection strongly suggests that water contributed to the widening of Central Valles Marineris troughs by reducing the shear strength of the wall rock. This implies localized enrichment of ice at depth, or the presence of groundwater below the permafrost (Battistini, 1985). Water saturating the pore space of the rock may have produced regolith mass movements at the base of some high walls (for example, in western Ophir Chasma), though the porosity is reduced at depth by the effect of lithostatic stresses (McKinnon and Tanaka, 1989). Groundwater may also have induced sapping along discontinuities in more coherent rocks (Louros Valles). Data on EM suggest that the central chasmata area may be relatively more porous, and formerly ice-rich than in peripheral areas, though the existence of very high walls implies a cohesion of the wall material that would derive from limited amounts of interstitial ice (Spencer and Croft, 1986; Croft, 1989). Lucchitta et al. (1992) consider the wall material to be either porous regolith (impact breccia) or volcanic materials such as tuffs (with volatiles trapped in their pores), probably capped with more coherent rocks less than 1-km thick. These coherent rocks, which may partly correspond to a Noachian basement, seem to be thicker in the western and eastern parts of the trough system.

The regional concentration of volatiles appearing at central Valles Marineris may result from accumulation of underground water from neighboring areas or underlying magma in porous rocks or regoliths (Battistini, 1985). The widening of the chasmata by sapping and poorly characterized mass movements, which was largely completed before the occurrence of most visible landslides, may have involved release of water from a confined aquifer (Carr, 1979; Higgins, 1982; Kochel and Piper, 1986). This process would also be genetically and chronologically consistent with a lacustrine origin for the layered deposits that were emplaced in central Valles Marineris during the same period (Nedell et al., 1987). Despite the probable dessication of the wall materials that followed these events, the effects of an active groundwater system on the reduction of shear strength may be one of the causes of the later stage of landsliding that occurred during middle- to late-Amazonian (Lucchitta et al., 1992). This study indicates, however, that a late tectonic stage, accompanied by wall heightening and oversteepening and by seismic activity, probably triggered these mass movements, and this stage was not followed by any erosional process which could be compared with the gullying of the previous stage.

Multi-stage faulting along wide grabens is the primary cause of the formation of high walls (Peulvast and Masson, 1993b), though the subsidence of the basins that prefigured the troughs during the first stage was locally accompanied by the emplacement of thick sedimentary and volcanic deposits. During the "ancient" stage, tectonic and probably volcanic activity was associated with a great efficiency of erosion, especially in central Valles Marineris, which resulted in wide re-entrants, and an unknown amount of wall retreat from fault lines or other down-warping structures in Melas Chasma. Gullying and even dissected landforms (Louros Valles) probably involved the action of water or ice released by wall rocks, but no large-scale mass movements developed. The end of this stage seems to be related to the end of active subsidence of the basin floors.

The last stages of faulting probably occurred when the wall materials were desiccated. This "recent" tectonic stage, which resulted in interconnection of basins along the main or annex grabens and in deepening to 7 or 9 km in the main troughs, seems to be related to regional up-doming and renewed magmatic activity of Tharsis (Geissler et al., 1990). Spectacular erosional processes, mainly landslides and talus formation, occurred, but the eroded volumes remained moderate, and the localized distribution of the corresponding landforms allowed the preservation of older wall topographies, even along segments of renewed tectonic activity. This second stage also gave rise to a local asymmetry, with mass movements mostly related to the reactivated northern boundary faults.

Acknowledgements

This work was supported by the Programme National de Planétologie from Institut National des Sciences de l' Univers (CNRS, France). The data used in this study were made available by the Regional Planetary Image Facility at University of Paris XI (Orsay, France) under the auspices of the NASA Planetary Geology and Geophysics Program. The figures were redrawn by Mrs. Geneviéve Roche (Unité C.N.R.S. de Géophysique et Géodynamique Interne, Université Paris-Sud).

References

Baker, V.R., Strom, R.G., 1992. Ancient Martian oceans. In: Zuber, M.T., James, O.B., Lunine, J.I., McPherson, G.J., Phillips, R.J. (Eds.), Planetary Geosciences 1989–1990. NASA SP, vol. 508, pp. 1–3.

Banerdt, W.B., Golombek, M.P., 1992. Constraints on the structure and evolution of the Tharsis region of Mars. In: Zuber, M.T., James, O.B., Lunine, J.I., McPherson, G.J., Phillips, R.J. (Eds.), Planetary Geosciences 1989–1990. NASA SP, vol. 508, pp. 4–6.

Banerdt, W.B., Phillips, R.J., Sleep, N.H., Saunders, R.S., 1982. Thick shell tectonics on one-plate planets: application to Mars. J. Geophys. Res., 87, 9723–9733.

Banerdt, W.B., Golombek, M.P., Tanaka, K.L., 1992. Stress and tectonics on Mars. In: Kiefer, H.H., Jakowsky, B.M., Snyder, C.W., Matthews, M.S. (Eds.), Mars. Univ. of Arizona Press, Tucson, pp. 249–297.

Baskerville, C.A., 1982. Collapse: a mechanism for Martian scarp retreat. NASA Tech. Memo., 85127, 244–252.

Battistini, R., 1984. Morphology and origin of ridges in low-latitude areas of Mars. Earth, Moon, Planets, 31, 49–61.

Battistini, R., 1985. Hydrolithosphere and problems of subsurface ice in the equatorial zone of Mars. In: Klinger, J., Bebest, D., Dollfus, A., Smoluchowski, R. (Eds.), Ices in the Solar System. Reidel, Boston, pp. 607–617.

Blasius, K.R., Cutts, J.A., Guest, J.E., Masursky, H., 1977. Geology of the Valles Marineris: first analysis of imaging from the Viking I Orbiter Primary Mission. J. Geophys. Res., 82, 4067–4091.

Bousquet, B., Jourdain-Bodart, J., Touzeau, J.F., 1987. Relief et modelé des versants de Valles Marineris: leur signification pour Mars. Bull. Soc. Geol. Fr., 3, 43–48.

Brunsden, D., 1979. Mass movements. In: Embleton, C., Phornes, J. (Eds.), Processes in Geomorphology. Arnold, London, pp. 131–186.

Carr, M.H., 1979. Formation of Martian flood features by release of water from confined aquifer. J. Geophys. Res., 84, 2995–3007.

Carr, M.H., 1981. The Surface of Mars. Yale Univ. Press, New Haven, 232 pp.

Carr, M.H., Crumpler, L.S., Cutts, J.A., Greeley, R., Guest, J.E., Mazursky, H., 1977. Martian impact craters and emplacement of ejecta by surface flow. J. Geophys. Res., 82, 4055–4065.

Chicarro, A.F., Schultz, P.H., Masson, P.H., 1985. Global and regional ridge patterns on Mars. Icarus, 63, 153–174.

Cipa, A., 1994. Le volcan Olympus Mons et ses environs, Mars. Thése de doctorat, Université Paris Sud, Orsay, 243 pp.

Clifford, S.M., 1993. A model for the hydrologic and climatic behavior of water on Mars. J. Geophys. Res., 98, 10973–11016.

Costard, F., 1990. Distribution et caractéristique du pergélisol martien: Son influence sur certains traits de la géomorphologie. Thése de doctorat, Université Paris IV, 327 pp.

Croft, S.K., 1989. Mars canyon systems: problems of origin. Abstr. 4th Int. Conf. Mars, Tucson, pp. 88–89.

Embleton, C., Whalley, B., 1979. Energy forces resistance and responses. In: Embleton, C., Phornes, J. (Eds.), Processes in Geomorphology. Arnold, London, pp. 11–38.

Fanale, F.R., 1976. Martian volatiles: the degassing history and geochemical fate. Icarus, 28, 179–202.

Flageollet, J.C., 1988. Les Mouvements de Terrain et Leur Prévention. Masson, Paris, 224 pp.

Fort, M., 1993. Etude d'une chaîne de collision intracontinentale (Himalaya du Népal, transversale des Annapurnas). Thése de doctorat d'Etat, Université Paris VII, 702 pp.

Fort, M., Peulvast, J.P., 1995. Catastrophic mass-movements and morphogenesis in the peri-Tibetan ranges: examples from West Kunlan, Pamir, and Ladakh. In: Slaymaker, O. (Ed.), Steepland Geomorphology. Wiley, Chichester, pp. 171–198.

Frey, H., 1979. Martian canyons and African rifts: structural comparisons and implications. Icarus, 37, 142–155.

Gault, D.E., Greeley, R., 1978. Exploratory experiments of impact craters formed in viscous-liquid targets, analog for rampart craters. Icarus, 34, 486–495.

Geissler, P.E., Singer, R.B., Lucchitta, B.K., 1990. Dark materials in Valles Marineris: indications of the style of volcanism and magmatism on Mars. J. Geophys. Res., 95, 14399–14413.

Hamblin, W.K., 1976. Patterns of displacement along the Wasatch fault. Geology, 4, 619–622.

Hartmann, W.K., 1973. Martian surface and crust: review and synthesis. Icarus, 19, 550–575.

Higgins, C.G., 1982. Drainage systems developed by sapping on Earth and Mars. Geology, 10, 147–152.

Horner, V.M., Greeley, R., 1987. Effects of elevation and ridged plains thickness on Martian crater ejecta morphology. J. Geophys. Res., 92, E561–E569.

Jibson, R.W., Keefer, D.K., 1993. Analysis of the seismic origin of landslides: examples from the New Madrid seismic zone. Geol. Soc. Am. Bull., 105, 521–536.

Kiefer, W.S., Hager, B.H., 1989. The role of mantle convection in the origin of the Tharsis and Elysium provinces of Mars. In: Frey, H.V. (Ed.), MEVTV Workshop on Early Tectonic Features on Mars. LPI Technical Rept. 89-04. Lunar and Planetary Institute, Houston, TX, pp. 48–50.

Kochel, R.C., Piper, J., 1986. Morphology of large valleys on Hawaii: evidence for ground water sapping and comparisons with Martian valleys. J. Geophys. Res., 91, E171–E192.

Kochel, R.C., Baker, V.R., Simmons, D.W., Lis, C.J., 1982. Surface channel networks developed by groundwater sapping in a laboratory model: analogy to sapping channels on Mars. Prog., NASA Tech. Memo., 85127, 227–229.

Kuzmin, R.O., Bobina, N.N., Zabalueva, E.V., Shashkina, V.P., 1988. Mars: estimation of the relative ice content in upper layers of the permafrost. Lunar Planet. Sci., vol. XIX, Lunar and Planetary Institute, Houston, TX, pp. 657–658.

Lucchitta, B.K., 1978. Morphology of chasma walls, Mars. J. Res. U.S. Geol. Surv., 6, 651–662.

Lucchitta, B.K., 1979. Landslides in Valles Marineris, Mars. J. Geophys. Res., 84, 8097–8113.

Lucchitta, B.K., 1984. A late climatic change on Mars. Lunar Planet. Sci., vol. XV, Lunar and Planetary Institute, Houston, TX, pp. 493–494.

Lucchitta, B.K., 1987. Valles Marineris: wet debris flows and ground ice. Icarus, 72, 411–429.

Lucchitta, B.K., Bertolini, L.M., 1989. Mars: interior structure of Valles Marineris, Mars. Lunar Planet. Sci., vol. XX, Lunar and Planetary Institute, Houston, TX, pp. 590–591.

Lucchitta, B.K., Balser, R.A., Bertolini, L.M., 1990. Valles Marineris, Mars: are pit chains formed by erosion and troughs by tectonism? Lunar Planet. Sci., vol. XXI, Lunar and Planetary Institute, Houston, TX, pp. 722–723.

Lucchitta, B.K., McEwen, A.S., Clow, G.D., Geissler, P.E., Singer, R.B., Schultz, R.A., Squyres, S.W., 1992. The canyon system on Mars. In: Kiefer, H.H., Jakosky, B.M., Snyder, C.W., Matthews, M.S. (Eds.), Mars. Univ. of Arizona Press, Tucson, pp. 453–492.

Masson, Ph.L., 1977. Structure pattern analysis of the Noctis Labyrinthus, Valles Marineris region of Mars. Icarus, 30, 49–62.

Masson, Ph.L., 1985. Origin and evolution of the Valles Marineris region of Mars. Adv. Space Res., 5, 83–92.

McCauley, J.F., 1978. Geologic map of the Coprates Quadrangle of Mars, scale 1:5,000,000. U.S. Geol. Survey Misc. Inv. Series Map I-897.

McEwen, A.S., 1989. Mobility of large rock avalanches: evidence from Valles Marineris, Mars. Geology, 17, 1111–1114.

McKinnon, D.J., Tanaka, K.L., 1989. The impacted Martian crust: structure, hydrology, and some geologic implications. J. Geophys. Res., 94, 17359–17370.

Mège, D., Masson, Ph.L., 1996. Stress models for Tharsis formation. Mars. Planet. Space Sci., 44, (12), 1471–1497.

Mouginis-mark, P.J., 1979. Martian fluidized crater morphology: variations with crater size, latitude, altitude, and target material. J. Geophys. Res., 84, 8011–8022.

Nedell, S.S., Squyres, S.W., Anderson, D.W., 1987. Origin and evolution of the layered deposits in the Valles Marineris, Mars. Icarus, 70, 409–441.

Parker, T.J., Gorsline, D.S., Saunders, R.S., Pieri, D.C., Schneeberger, D.M., 1993. Coastal geomorphology of the Martian northern plains. J. Geophys. Res., 98, 11061–11078.

Peterson, C., 1981. A secondary origin for the central plateau of Hebes Chasma. Proc. 12th Lunar Planet Sci. Conf. Geochim. Cosmochim. Acta, pp. 1459–1471.

Peulvast, J.P., Masson, Ph.L., 1993a. Melas Chasma: morphology and tectonic patterns in Central Valles Marineris (Mars). Earth, Moon, Planets, 61, 219–248.

Peulvast, J.P., Masson, Ph.L., 1993b. Erosion and tectonics in Central Valles Marineris (Mars): a new morphostructural model. Earth, Moon, Planets, 61, 191–217.

Pollack, J.B., Kastings, J.F., Richardson, S.M., Poliakoff, K., 1987. The case for a wet, warm climate on early Mars. Icarus, 71, 203–224.

Schubert, G., Bercovici, D., Glatzmaier, G.A., 1990. Mantle dynamics in Mars and Venus: influence on an immobile lithosphere on three-dimensional mantle convection. J. Geophys. Res., 95, 14105–14129.

Schultz, R.A., 1989. Do pit-crater chain grow up to be Valles Marineris canyons? NASA Tech. Memo., 4130, 539–540.

Schultz, R.A., 1991. Structural development of Coprates Chasma and western Ophir Planum, Valles Marineris Rift, Mars. J. Geophys. Res., 96, 22777–22792.

Scott, D.H., Tanaka, K.L., 1986. Geologic map of the western equatorial region of Mars, scale 1:15,000,000. U.S. Geological Survey Misc. Inv. Series Map I-1802-A.

Sharp, R.P., 1973. Mars: troughed terrain. J. Geophys. Res., 78, 4063–4072.

Sleep, N.H., Phillips, R.J., 1979. An isostatic model for the Tharsis province, Mars. Geophys. Res. Lett., 6, 803–806.

Sleep, N.H., Phillips, R.J., 1985. Gravity and lithospheric stress on the terrestrial planets with reference to the Tharsis region of Mars. J. Geophys. Res., 90, 4469–4489.

Solomon, S.C., Head, J.W., 1982. Evolution of the Tharsis province of Mars: the importance of heterogeneous lithospheric thickness and volcanic construction. Proc. 3rd Int. Colloquium on Mars. J. Geophys. Res., vol. 87, pp. 9755–9774.

Spencer, J.R., Croft, S.K., 1986. Valles Marineris as karst: feasibility and implications for Martian atmospheric evolution. NASA Tech. Memo., 88383, 193–195.

Squyres, S.W., Clifford, S.M., Kuzmin, R.O., Zimbelman, J.R., Costard, F.M., 1992. Ice in the Martian regolith. In: Kiefer, H.H., Jakosky, B.M., Snyder, C.W., Matthews, M.S. (Eds.), Mars. Univ. of Arizona Press, Tucson, pp. 523–554.

Tanaka, K.L., Golombek, M.P., 1989. Martian tension fractures and the formation of grabens and collapse features at Valles Marineris. Proc. 19th Lunar Planet. Sci. Conf., pp. 383–396.

Tanaka, K.L., Golombek, M.P., Banerdt, W.B., 1991. Reconciliation of stress and structural histories of the Tharsis region of Mars. J. Geophys. Res., 96, 15617–15633.

Tanaka, K.L., Scott, D.H., Greeley, R., 1992. Global stratigraphy. In: Kiefer, H.H., Jakosky, B.M., Snyder, C.W., Matthews, M.S. (Eds.), Mars. Univ. of Arizona Press, Tucson, pp. 345–382.

U.S. Geological Survey, 1986. Topographic orthophoto mosaic of the Tithonium Chasma region of Mars. U.S. Geol. Survey Misc. Inv. Series Map I-1294.

U.S. Geological Survey, 1986. Topographic map of the Coprates Northwest quadrangle of Mars. U.S. Geol. Survey Misc. Inv. Series Map I-1712.

Wallace, R.E., 1978. Geometry and rates of change of fault-generated range fronts, North-Central Nevada and some tectonic implications. J. Res. U.S. Geol. Surv., 6, 637–650.

Watters, T.R., 1993. Compressional tectonism on Mars. J. Geophys. Res., 98, 17049–17060.

Willeman, R.J., Turcotte, D.L., 1982. The role of lithospheric stress in the support of the Tharsis rise. Proc. 3rd Int. Colloquium on Mars. J. Geophys. Res., vol. 87, pp. 9793–9801.

Witbeck, N.E., Tanaka, K.L., Scott, D.H., 1991. Geologic map of the Valles Marineris region, Mars, scale 1:2,000,000. U.S. Geol. Survey Misc. Inv. Series Map I-2010.

Wohletz, K.M., Sheridan, M.F., 1983. Martian rampart crater ejecta: experiments and analysis of melt–water interaction. Icarus, 56, 15–37.