

Tharsis dome, Mars: New evidence for Noachian-Hesperian thick-skin and Amazonian thin-skin tectonics

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Abstract. A photogeological reconnaissance of Viking mosaics and images of the Tharsis dome has been carried out. Fifteen new areas of transcurrent faulting have been located which, together with other structures previously detected, support a model in which the Thaumasia Plateau, the southeastern part of the Tharsis dome, is proposed to be an independent lithospheric block that experienced buckling and thrust faulting in Late Noachian or Early Hesperian times as a result of an E-W directed compression. Evidence is presented that this stress field, rather than the Tharsis uplift, was decisive in the inception of Valles Marineris, which we consider a transtensive, dextral accident. The buckling spacing permits us, moreover, to tentatively reconstruct a Martian Hesperian lithosphere similar in elastic thickness to the mean present terrestrial oceanic lithosphere, thus supporting the possibility of a restricted lithospheric mobility in that period. Tharsis lithosphere was again subjected to shear stresses in Amazonian times, a period in which important accidents, such as strike-slip faults, wrinkle ridges, and straight and sigmoidal graben, were formed under a thin-skin tectonic regime, while the lithosphere as a mechanical unit had become too thick and strong to buckle. The possible causes of those stresses, and especially their relationships to a putative period of plate tectonics, are discussed.

1. Introduction

Two of the paradigms of classic Martian geology are the nonexistence of plate tectonics and the consequent absence of transcurrent tectonics: In a planet without lithospheric mobility the only large horizontal stresses would be those related to density heterogeneities [e.g., Banerdt *et al.*, 1982, 1992]. The very old age of a significant part of the Martian surface and the outstanding isostatic anomalies shown by the giant volcanic constructs are the main facts supporting those assumptions. Forsythe and Zimbelman [1988] questioned this conclusion, since they found some support for a period of lithospheric mobility in Gordii Dorsum, a strike-slip fault system SW of Olympus Mons. Sleep [1994] was also against the orthodoxy when proposing a short (~100 m.y.) period of limited plate consumption in Hesperian times as an explanation for Mars' dichotomy.

The efforts to understand Mars tectonics have been in good part focused on the analysis of the Tharsis dome. Carr [1974], Wise *et al.* [1979], Melosh [1980], Banerdt *et al.* [1982], Plescia and Saunders [1982], Solomon and Head [1982], Watters and Maxwell [1983, 1986], Sleep and Phillips [1985], Tanaka and Davis [1988], Schultz and Lutz [1988], Scott and Dohm [1990], Tanaka *et al.* [1991], Watters [1991, 1992, 1993], Banerdt *et al.* [1992], Thomas and Allemand [1993], Schultz and Tanaka [1994], Anderson *et al.* [1997], Anderson and Grimm [1998], Anderson *et al.* [1999], Mangold *et al.* [1999], and Turcotte [1999], among others, have put forward hypotheses on the origin of the dome and its associated tectonic structures. In these studies, only two strike-slip fault systems were located in Tharsis: the aforementioned Gordii Dorsum and a set of faults cutting the ridged plains near Nectaris Fossae, south of Valles Marineris [Schultz, 1989]. Notwithstanding this, and after carrying out a thorough overhaul of the Viking mosaics and images covering the Tharsis dome, we have located (Figure 1) a number of new systems of transcurrent faults. It is important to emphasize that transcurrent systems were proposed in several of the previous theoretical models, which are discussed next.

1.1. Rotation Axis Reorientation Models

The coincidence with the present Martian equator of the Tharsis dome center led Melosh [1980] to propose a reorientation of the planet rotational axis by as much as 30°. This would produce on the lithosphere membrane stresses amounting to several kilobars, high enough to initiate faulting. The calculation for reorientations of 15°-30° [Melosh, 1980, Figures 1 and 2] predicts the presence in the low to medium latitudes of strike-slip faults, though this author makes an exception with Valles Marineris, which he

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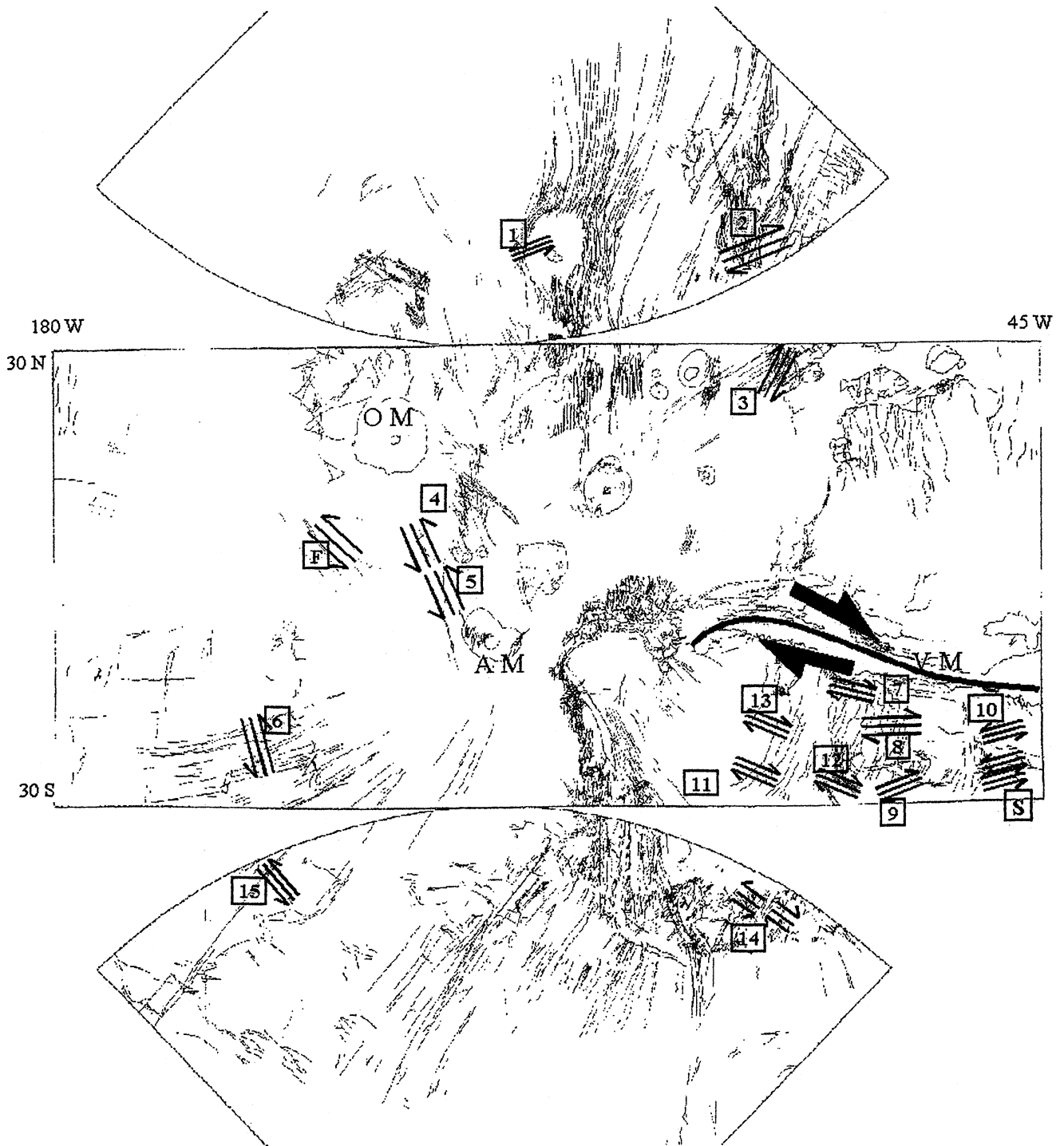


Figure 1. Location of the transcurrent faults identified on the Tharsis dome area. Numbers of the faults are referred to in Table 1. F and S indicate the strike-slip faults detected by *Forsythe and Zimelman* [1988] and *Schultz* [1989], respectively. Tectonic base after *Plescia*, as cited by *Carr* [1981]. OM, Olympus Mons; AM, Arsia Mons; VM, Valles Marineris.

defines as a pure extensional form. In 1988, *Schultz and Lutz* [1988] put forward an axial reorientation model that included the polar wandering of the whole Martian lithosphere but that predicted normal or reverse faults only, not strike-slip ones.

1.2. Vertical Loading Models

Banerdt et al. [1982, 1992] constructed three vertical models (flexural uplift, isostatic response, and flexural loading), though they discarded the first one and postulated the other two to happen in sequence: Tharsis would initially be isostatically compensated and only later would respond by flexural loading. Both models predicted a string of transcurrent faults around the dome.

1.3. Limited Lithospheric Mobility Models

Although unusual, the presence of some transcurrent systems attested that strike-slip movements had taken place in Mars' history, thus lending some credibility to mobilist hypotheses. *Schultz and Tanaka* [1994] proposed that the Nectaris Fossae strike-slip faults and the nearby wrinkle ridges were part of a wider system of structures extending through south Tharsis. The most important of these are a series of topographic rises such as the Coprates Rise, an asymmetric arch almost 1000 km long and which rises up to 4 km over the adjacent plains south of Coprates Chasma, eastern Valles Marineris (Plate 1). Those authors attributed to a combination of buckling and thrust faulting these arches and swells which are 150 to 500 km from each other, spacing typical of the deformation of the entire lithosphere. This fact, coupled with the apparent thickening of the lithosphere (see section 3), supports a horizontal compression. The same can be said of the two transcurrent structures mentioned before: *Schultz* [1989] calculated a minimum accumulated lateral displacement of 2 km for the strike-slip faults south of Valles Marineris, while *Forsythe and Zimbelman* [1988] estimated that the Gordii Dorsum system evidences as much as 40 km of horizontal slip. These important shortenings are difficult to understand without allowing for a certain amount of lithospheric mobility.

2. An Inventory of Tharsis Strike-Slip Faults in the Framework of Previous Models

Thus the identification and mapping of transcurrent accidents would also serve the goal of confirming or rejecting some of the competing models. The main criterion used to identify strike-slip faults was to select fractures that would clearly offset other features, in most cases wrinkle ridges. Some of the results of this search are portrayed in Figure 2. In two instances (e.g., Fig. 2d) the strike-slip faults were identified on the basis of their sigmoidal geometry, supported in one of the cases by its association with drag folds. In all (Table 1), 15 new transcurrent faults or systems of faults were tentatively identified; most of these were concentrated on Thaumasia Plateau (Plate 1), the southeastern part of the Tharsis dome. Then we plotted (Figure 3) the newly mapped strike-slip systems on the two models put forward by *Banerdt et al.* [1992], the (initial) isostatic compensation and the (later) flexural loading models. In both cases the results are poor: for the isostatic model, only four of the 15 systems are

coincident with the predicted transcurrent structures; for the flexural loading model the coincidence amounts to five, thus never approaching significant levels. It could be concluded that these tectonic arrays are apparently unrelated to the Tharsis bulge inception. Though most Tharsis structures can still be explained by the successive radial fields with vertical σ_1 proposed, for instance, by *Plescia and Saunders* [1982], we contend that the Tharsis uplift is unable to explain an important number of regional structures, which could mean that the dome's own tectonics is probably superposed on another stress field.

As can be seen, we include Valles Marineris among Mars transcurrent structures. Contrary to the standard explanation [e.g., *Melosh*, 1980; *Plescia and Saunders*, 1982], according to which the canyon is a system of pure graben, we contend that this chasm is a right-lateral transtensive megashear, an assertion supported by (1) its curved geometry, which is not explained by the pure extension origin usually accepted (whereas the graben can be explained as a result of transtension); (2) the strong negative Valles Marineris free-air gravity anomaly [*Smith et al.*, 1999; *Zuber et al.*, 2000], which is seemingly unrelated to a significant mantle upwelling, especially when compared with Hellas and Utopia basins (*Zuber et al.* [2000] see it as a consequence of lithospheric stretching, but the same structure would result from the action of a megashear); (3) the significant difference in crustal and lithospheric thickness [*Schultz and Tanaka*, 1994; *Zuber et al.*, 2000] between the southern (Thaumasia) and northern (Lunae Planum) realms of Valles Marineris; (4) the structural differences between north and south Tharsis (first, the wrinkle ridges of the latter area are distributed on a much larger surface; second, they are also connected by numerous strike-slip faults; third, compressional structures north and south of Valles Marineris seem to result from different stress fields [*Anderson et al.*, 1999]; and fourth, the absence of buckling structures north of Valles Marineris [*Schultz and Tanaka*, 1994]); (5) the E-W stress field proposed by *Schultz and Tanaka* [1994] and confirmed by, for instance, the faults illustrated in Figure 2b fits with the genesis of a megashear with the strike of Valles Marineris; and (6) the stress field deduced for the origin of Lunae Planum wrinkle ridges by *Watters and Robinson* [1997], who proposed that lithospheric compression in the area must be accommodated primarily by strike-slip faulting.

All these observations agree with our interpretation of Valles Marineris as a lithospheric discontinuity which would serve as a stress guide for the inception of the chasm. Since Valles Marineris could be considered as the northern margin of the Thaumasia block, and hence the border of two different lithospheric domains, a high rheological contrast must have existed here, which in turn caused stress and strain concentrations. The initial formation of Valles Marineris is believed to have occurred at Late Noachian, contemporaneous to the lithospheric buckling in Thaumasia [*Dohm and Tanaka*, 1999]. In this case both structures could be due to the same stress field. Additionally, the lithospheric properties at the time of formation of both tectonic structures were very similar, as we will discuss in the next section.

3. South Tharsis Mechanical Lithosphere Thickness

South Tharsis could retain the keys for understanding the evolution of the mechanical properties and dynamic history of

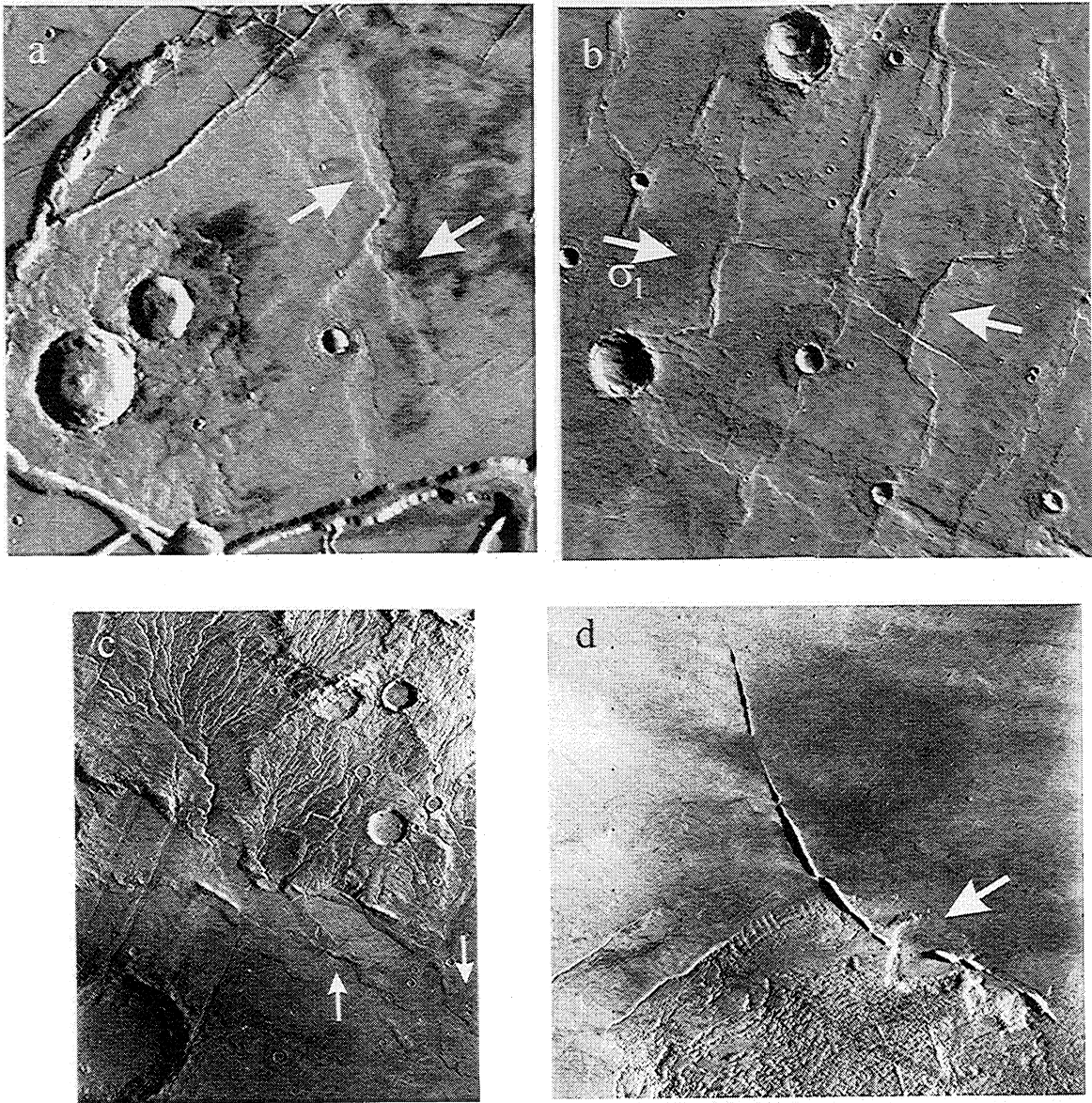


Figure 2. Examples of transcurrent faults in the Tharsis area. (a) Labeatis Fossae (fault 4 in Table 1; 32°N, 72°W): sinistral strike-slip faults which offset wrinkle ridges by 100 km. They are parallel to, or coincident with, graben systems. (b) Felis Dorsa area (fault 9 in Table 1; centered on 26°S, 65°W): Several pairs of Riedel shears with 80° and 130° strikes and offsets around 25 km are clearly seen, thus permitting the calculation of a σ_1 of N105E. (c) Warrego Valles area (fault 14 in Table 1; 43°S, 92°W): group of transcurrent accidents, both dextral and sinistral, making up another Riedel system, offset (by some 5 km) a hogback-type relief. (d) Sigmoid structure near Aganippe Fossa (fault 5 in Table 1; 3°S, 126°W). This 500-km-long, sinistral fault cuts the Arsia Mons aureole, a craterless terrain. A thick volcanic flow (arrow) where curvature changes has been interpreted [Anguita and Moreno, 1992] as evidence for transtension.

the Martian lithosphere: Was the lithosphere ever capable of plastic deformation? If so, when and at what scale? Were the main stresses vertical or horizontal? In the second case, what was their cause? Were they regional in scale, or are we witnessing in the Tharsis dome the traces of a plate tectonics period such as the one advocated by Sleep [1994] for the northern plains?

A key question for the understanding of the tectonic history of any planet is the mechanical evolution of its lithosphere. Mars' lithospheric thickness is poorly constrained. The analysis of the local load-induced flexures

[Banerdt *et al.*, 1992] points out that at the time when volcanic constructs were built, the Tharsis elastic lithosphere thickness h was, in general, higher than 150 km, excepting the central area, where it perhaps was thinner than 50 km. Mars Global Surveyor (MGS) gravity and topographic data [Zuber *et al.*, 2000] lead to h values from > 200 km below Tharsis volcanoes and Olympus Mons, to 50 km below Alba Patera, Valles Marineris (Melas Chasma), and Solis Planum.

The spacing of the arches identified by Schultz and Tanaka [1994] in south Tharsis permits an independent h estimation during buckling, through a comparison with the Earth. The

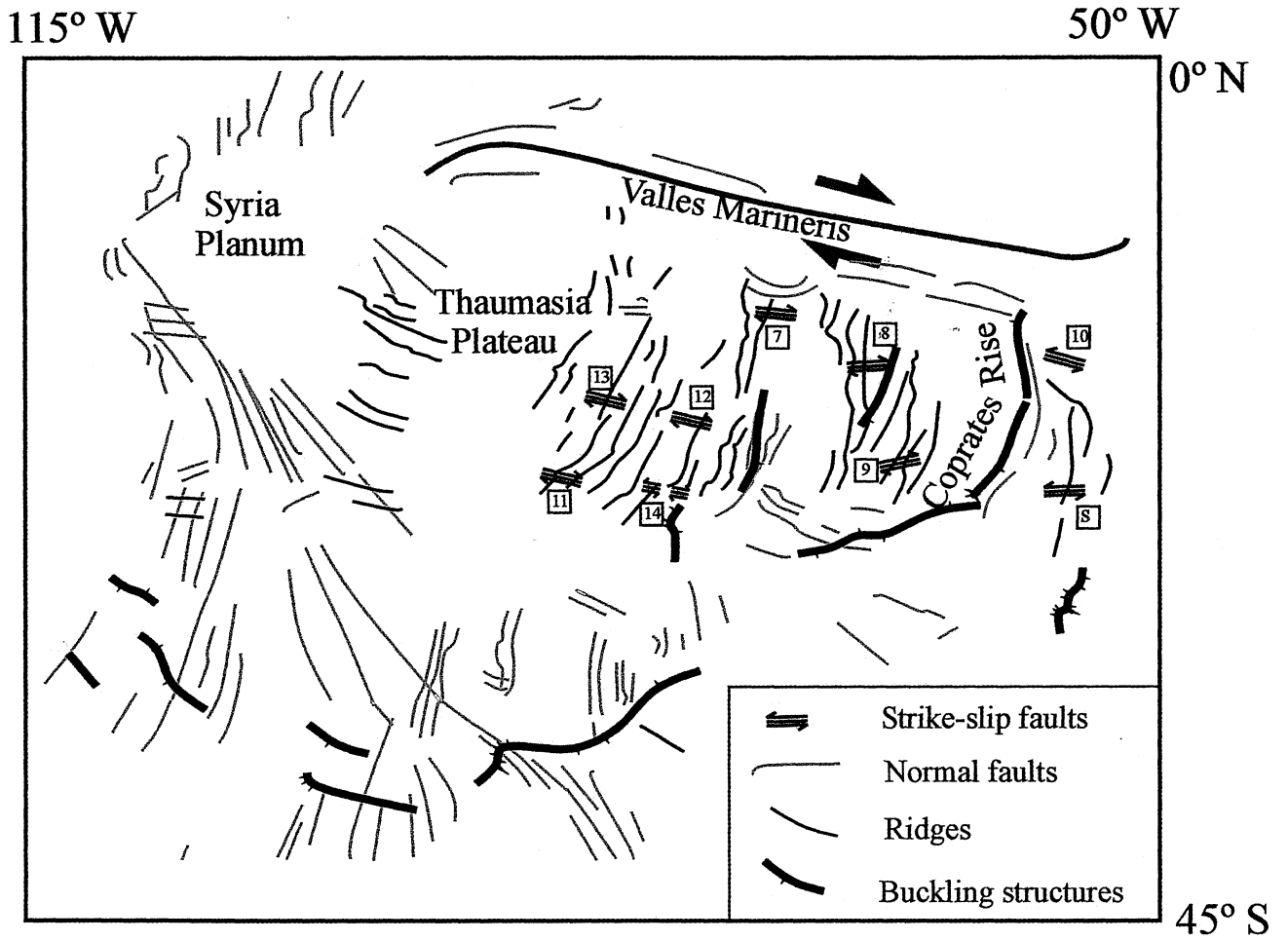


Plate 1. Tectonic features of the Thaumasia Plateau, the southeastern area of the Tharsis dome. Numbers of the faults are referred to in Table 1.

Table 1. Fault Systems Description

Fault	Latitude	Longitude	Slip (km) and Type (D or S)	Approximate Age	Identificative Criteria
1	41°N	113°W	10 (S)	Amazonian	displaces ridge
2	35°N	63°W	30 (D)	post-Early Hesperian	displaces ridges
3	32°N	72°W	9 (D)	post-Early Hesperian	displaces ridge
4	3°-5°N	130°W	? (S)	Late Amazonian	sigmoid shape
5	2°-4°S	125°-127°W	? (S)	Late Amazonian	sigmoid shape
6	21°-22°S	151°W	6 (S)	post-Noachian	displaces graben
7	16°S	72°-73°W	12 (D)	post-Early Hesperian	displaces ridges
8	20°S	65°-72°W	15 (D)	post-Early Hesperian	displaces ridges
9	26°S	64°-66°W	25 (Riedel shears)	post-Early Hesperian	displace ridges
10	22°S	56°W	15 (S)	post-Early Hesperian	displace ridges
11	25°S	81°-84°W	24 (D)	post-Early Hesperian	displace ridges
12	25°S	79°W	14 (D)	post-Early Hesperian	displace ridges
13	18°-21°S	80°-84°W	15 (D)	post-Early Hesperian	displaces ridges
14	43°S	90°-92°W	5 (Riedel shears)	post-Hesperian	displace ridges
15	29°-32°S	150°-154°W	40 (S)	post-Noachian	displace ridges
F	2°-8°N	142°-146°W	30-40 (S)	Noachian-Early Hesperian	Riedel shears; en echelon push-ups
S	20°-27°S	50°-57°W	> 2 (Riedel shears)	post-Early Hesperian	displace ridges

subrecent deformation of the oceanic Indo-Australian plate has also been explained through buckling [McAdoo and Sandwell, 1985]. In this case the arches' wavelength is ~200 km, and the estimated elastic lithospheric thickness is 40-50 km. Watters [1987] proposed the application of a wavelength/thickness (λ/h) ratio of ~4 to south Tharsis arches and swales. In this case, taking 300 km as the mean spacing, the Martian elastic lithosphere at the time (Late Noachian to Early Hesperian) of this thick-skin tectonics would have been ~75 km thick. However, in general, theoretical models that explain buckling do not propose a linear relation between λ and h . For elastic buckling models [e.g., Turcotte and Schubert, 1982] we have $h^5 \propto g \lambda^4$, where g is gravity. Although the parameters relative to rocks do not vary, and even ignoring the effects of the probable discontinuities in Mars and in the Earth, we should bear in mind the difference in g between the two planets. With this the elastic lithosphere thickness could be approximated to

$$h = h_0 \left(\frac{g}{g_0} \right)^{1/5} \left(\frac{\lambda}{\lambda_0} \right)^{4/5}, \quad (1)$$

where h_0 , g_0 and λ_0 are the reference (Earth) values for thickness, gravity, and buckling spacing. Taking $h_0 = 40$ -50 km, $g_0 = 9.8 \text{ m s}^{-2}$, and $\lambda_0 = 200$ km, and (for Mars) $g = 3.7 \text{ m s}^{-2}$ and $\lambda = 300$ km, we obtain a value of about 50-60 km for h . In viscous buckling models there is no simple expression relating h to λ , although the rate λ/h is generally larger than the one for elastic buckling models [Solomon and Head, 1984; Watters, 1991]. Therefore these values of h must be considered as an uppermost limit. It is interesting that the thickness of the layer which was buckled in south Tharsis, estimated from (1), is very similar not only to terrestrial areas whose lithosphere could be presently experiencing buckling [McAdoo and Sandwell, 1985], but also to those obtained independently by Zuber *et al.* [2000] for Solis Planum and Melas Chasma.

The depth of the base of the mechanical (or rheological) lithosphere can be defined as the depth for which the strength goes down to a low value and under which there are no

significant discontinuities in strength [McNutt, 1984; Ranalli, 1994, 1997; Anderson and Grimm, 1998]. This critical value has been estimated in ~10 MPa for Ranalli [1994]. The ductile strength, which describes the lower lithosphere and sublithosphere strengths, can be deduced by means of the flow law creep,

$$\sigma_1 - \sigma_3 = \left(\frac{\dot{\epsilon}}{A} \right)^{1/n} \exp \left(\frac{Q}{nRT_z} \right), \quad (2)$$

where $\dot{\epsilon}$ is the strain rate, A and n are constants, Q is the activation energy of deformation by creep, R is the gas constant ($8.3144 \text{ J mol}^{-1} \text{ K}^{-1}$), and T_z is the temperature at z depth. Therefore a useful approach to the problem of the evolution of the mechanical lithosphere thickness with a planet cooling is to assign a critical value to $\sigma_1 - \sigma_3$ at the lithosphere base. The temperature at the base of the lithosphere, T_{bl} , for which the critical value is reached for a definite lithology can be calculated from (2). The mechanical lithosphere thickness l can be estimated as a function of the heat flow, if one assumes a linear conductive gradient dT/dz :

$$l = \frac{T_{bl} - T_s}{dT/dz}, \quad (3)$$

where T_s is the surface mean temperature (220 K). Last, on the basis of MGS data [Zuber *et al.*, 2000] we take 60 km as the mean value for South Tharsis crust thickness. This value is consistent with former estimations [Bills and Ferrari, 1978; Esposito *et al.*, 1992; Anderson and Grimm, 1998] for South Tharsis. As for lithologies, we suppose that Martian crust consists of diabase and that the mantle is dry peridotite (Table 2). Strain rates of 10^{-15} s^{-1} and 10^{-19} s^{-1} are assumed.

Results are shown in Figure 4. When the thermal gradient was high, the lithosphere base was shallower than the crust base; with the waning of internal heat sources, and the base of the mechanical lithosphere becomes gradually deeper, provided that the strength at the top of the mantle stays low enough (for any given temperature, the mantle is stronger than the crust). For sufficiently high values of the strength at the

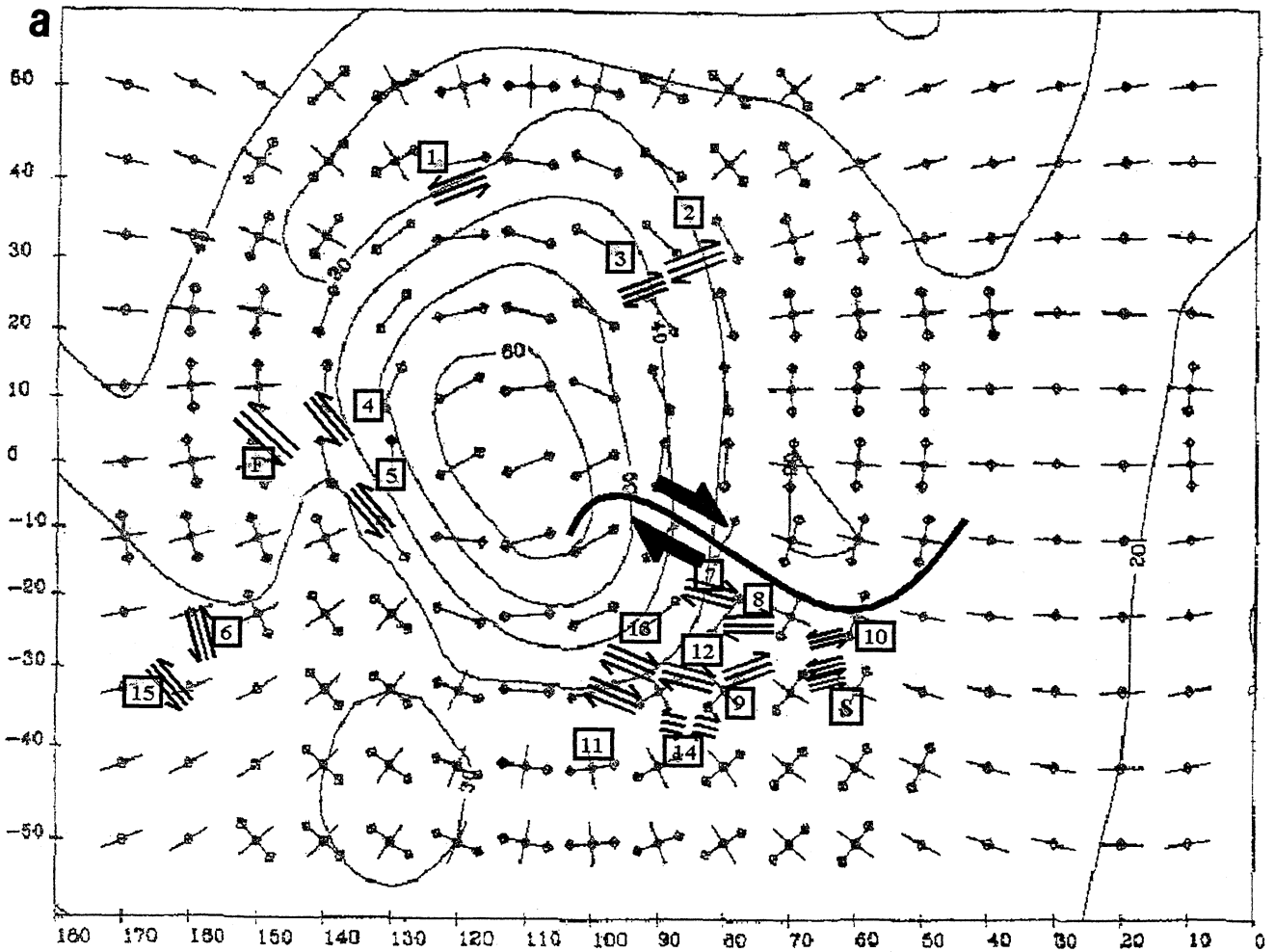


Figure 3. Plots of (a) the isostatic and (b) the flexural loading models of *Banerdt et al.* [1992] for Tharsis, with the situation of the transcurent faults identified. F and S have the same meaning as in Figure 1.

top of the mantle, the depth of the base of the mechanical lithosphere becomes controlled by mantle rheology. The jump in the curve (Figure 4), representing the evolution of l , results from the definition of this parameter strictly as a function of a critical strength in the base of the mechanical lithosphere. Though we concede that this is an extreme position, it serves our purpose of stating neatly that the transition from a control of the lithosphere thickness by crust rheology to a situation in which this parameter is controlled by mantle rheology must have been relatively sudden in Mars geologic history.

The evolution of l as a function of the thermal gradient can be used to calculate the evolution of the total lithospheric strength Σ in South Tharsis. Σ can be estimated from

$$\Sigma = \int_0^l (\sigma_1 - \sigma_3) (z) dz \quad (4)$$

where $(\sigma_1 - \sigma_3) (z)$ is the smallest strength (brittle or ductile) to every depth z . The brittle strength in a compressive tectonic regime (obvious for a buckling process) is

$$(\sigma_1 - \sigma_3) = 3 \rho g z (1 - \lambda_f) \quad (5)$$

[*Sibson*, 1974], where ρ is density (2900 kg m^{-3} for the crust and 3300 kg m^{-3} for the mantle) and λ_f is the ratio of pore

fluid to lithostatic pressure. Following *Anderson and Grimm* [1998], we use $\lambda_f = 0$, owing to the uncertainties in the amount of free water in the Martian lithosphere.

Figure 5 shows results for a strain rate of 10^{-15} s^{-1} (results for a strain rate of 10^{-19} s^{-1} are similar). The sharp increase in the total lithospheric strength occurs when the upper mantle becomes strong enough, owing to the drop of the thermal gradient at around $10\text{--}12 \text{ K km}^{-1}$. These values are similar to those of $7\text{--}12 \text{ K km}^{-1}$ obtained by *Zuber et al.* [2000] for an elastic lithosphere thickness of $\sim 50 \text{ km}$ when the topography of Solis Planum and Melas Chasma was formed. The sharp increase in the total lithospheric strength occurs at $\sim 10^{13} \text{ N m}^{-1}$ (Figure 5), which is similar to the maximum value estimated for tectonic forces in the Earth [*Ranalli*, 1997]. When South Tharsis was subjected to buckling, tectonic forces could deform the lithosphere because their magnitude was comparable to the total lithospheric strength. However, when the upper mantle became strong, the total lithospheric strength far exceeded the magnitude of tectonic forces, so that significant lithospheric deformation could not occur. Therefore it can be concluded that when the thermal gradient dropped below $\sim 10 \text{ K km}^{-1}$, the South Tharsis lithosphere became too strong to be deformed as a whole. The present Tharsis tectonic models do not quantify the stresses

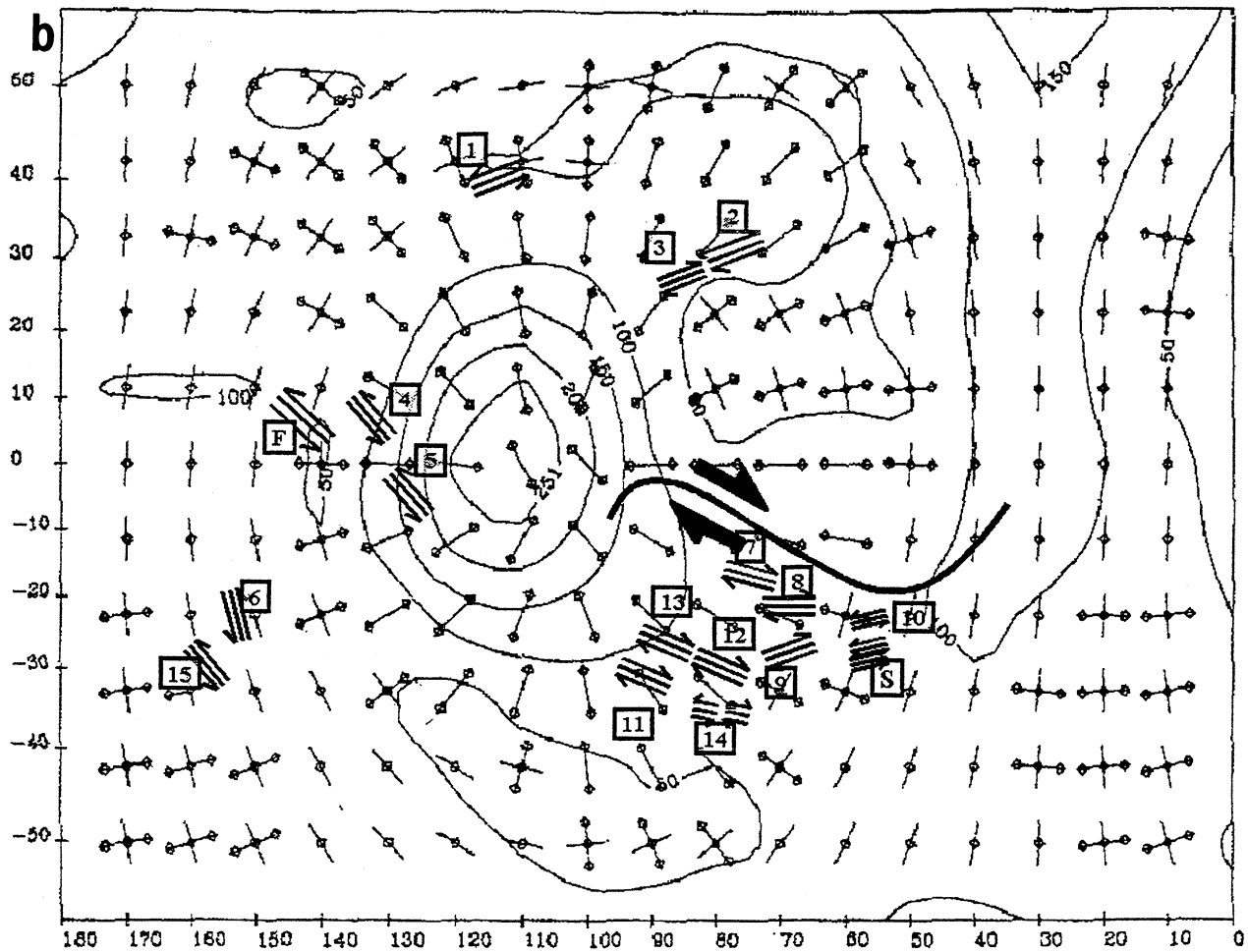


Figure 3. (continued)

needed for deformation [Schultz and Tanaka, 1994], so it is unclear whether those were sufficiently high to cause lithospheric buckling, or whether a different and larger stress source was necessary.

The evolution of l and Σ as a function of the thermal gradient can be compared with the published models of the Mars thermal evolution [Schubert and Spohn, 1990; Spohn, 1991; Schubert et al., 1992; Grasset and Parmentier, 1998; Choblet et al., 1999]. Using thermal conductivities k for the Martian lithosphere of $2.5\text{--}3.3\text{ W m}^{-1}\text{ K}^{-1}$, we obtain heat flow values [$q = k(dT/dz)$] at the surface of $\sim 25\text{--}36\text{ mW m}^{-2}$ at the moment of the sharp increase in the total lithospheric strength. These values are not very different from those supposed for present Mars ($\sim 20\text{--}40\text{ mW m}^{-2}$). However, caution is essential when comparing the evolution of l or Σ as a function of heat flow with Mars thermal evolution models. In these models we deal with the evolution of mean planetary heat flow, while the computing of thermal gradients (carried out on the basis of estimations of elastic lithosphere thickness

through flexure measurements [Solomon and Head, 1990]) and gravity and topography data [Zuber et al., 2000] does not regional control. Concurrently, it has been proposed [Zuber et al., 2000] that a high heat flow could have been dissipated through the northern plains in early Mars, compensating the considerably lesser heat flow of the southern highlands.

4. Discussion: The Age of the Structures and the Search for a Mechanism

A critical point for understanding Martian tectonics is the origin and age of the wrinkle ridges and strike-slip faults (including those found by the present authors) of Tharsis. As for the first aspect, the standard view [e.g., Schultz and Tanaka, 1994] explains all those structures as being caused by the Tharsis uplift; nevertheless, a clean pair of Riedel shears such as the one in Figure 2b is a perfect example for a stress field with large horizontal displacements, probably involving lithospheric mobility. The geometric and structural

Table 2. Flow Law Parameters for the Crust and the Mantle

Lithology	A , $\text{MPa}^{-n}\text{ s}^{-1}$	n	Q , kJ mol^{-1}
Diabase (crust)	2.0×10^{-2}	3.4	260
Peridotite dry (mantle)	2.5×10^4	3.5	532

Source: Ranalli [1997]

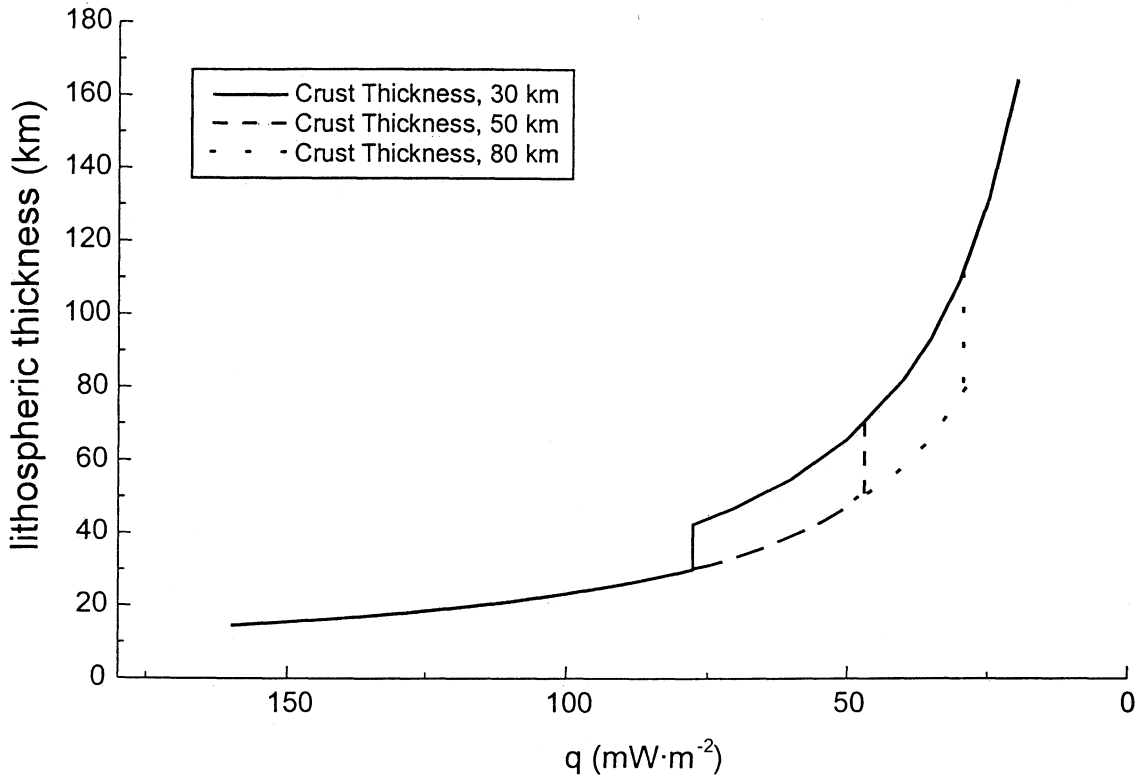


Figure 4. Evolution of the thickness of the Martian mechanical lithosphere with a diminishing thermal gradient. The abrupt step illustrates the change from a crust-controlled regime to a mantle-controlled regime.

relationship between wrinkle ridges and strike-slip faults that can be observed in Figure 2b is consistent with an E-W compression in the area. Therefore, here strike-slip faults can have both left-lateral (WNW-ESE trending faults) and right-lateral (ENE-WSW trending faults) movements. For horizontal strike-slip faults to develop in this stress field the direction of least compression must change from vertical to

horizontal. This is compatible with temporal permutations of the main stress axes, a common feature in terrestrial tectonics.

Concerning the age of faulting, we agree with *Schultz* [1989] that the strike-slip and the wrinkle ridges at Coprates could be approximately coeval, but we disagree when, in agreement with *Scott and Dohm* [1990], he proposes for them an Early to Middle Hesperian age. The attribution of those

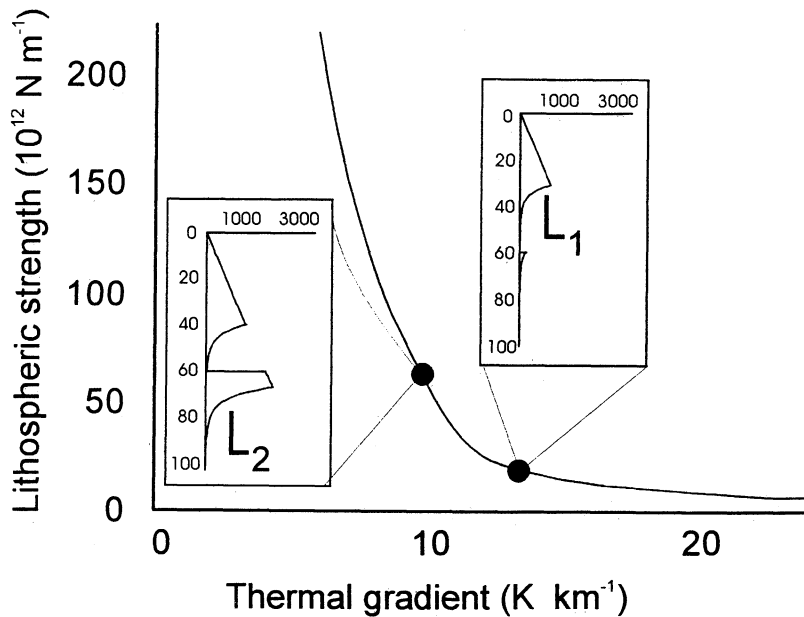


Figure 5. Evolution of the total lithospheric strength, in terms of thermal gradient, for a strain rate of 10^{-15} s^{-1} . Strength envelopes for thermal gradients of 10 and 13 K km^{-1} are shown.

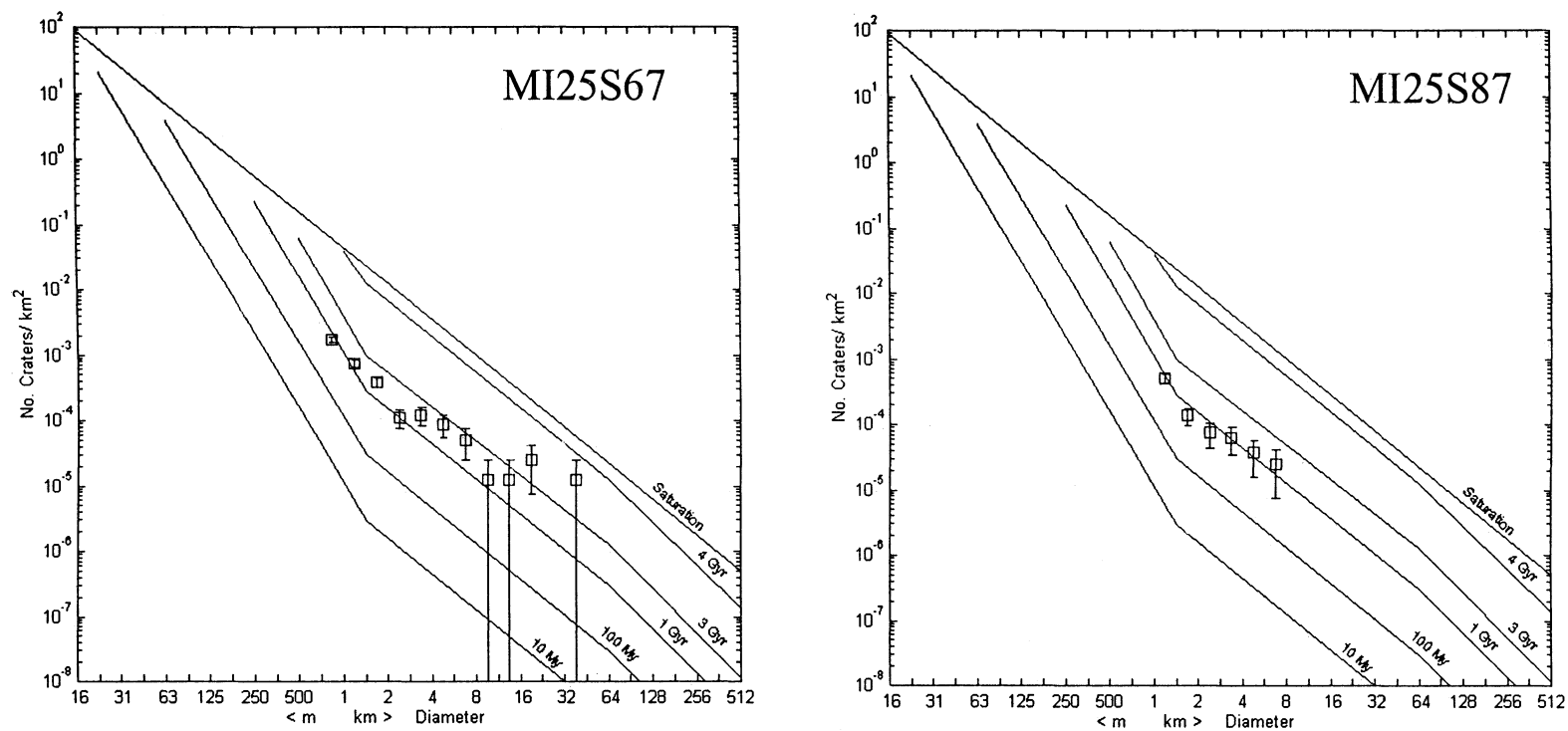


Figure 6. Crater countings for Viking mosaics centered on 25°S-67°W and 25°S-87°W. The counting characteristics are explained in Table 2. The isochrons are from *Hartmann* [1999].

Table 3. Crater Counting Statistics

Image	Scale	Minimum Latitude, °	Maximum Latitude, °	Minimum Longitude, °	Maximum Longitude, °	Surface (pixels)	$n > 2$ (per 10^6 km)	$n > 5$ (per 10^6 km)	$n > 16$ (per 10^6 km)
25S67	0.2313	-27.5	-22.5	64.994	70.00	1184x1280	320.53	61.64	12
25S87	0.2313	-27.5	-22.5	85.00	90.0	1184x1280	123.28	0	0

very old ages reflects, in our view, the prejudice that only nearly primordial tectonic activity took place on Mars. *Watters* [1993, p. 17049] states, for instance, that the wrinkle ridges are Early Hesperian “if the tectonic features are roughly the same age as the units in which they occur.” However, our observations are that many of these structures cut volcanic plains very sparsely cratered and that therefore neither the plains nor the structures can be 3.8 to 3.6 Gyr old. To substantiate our position, we carried out crater countings in areas cut by transcurrent faults. The results, shown in Figure 6, show ages between 1 and 3 Gyr for the area at 25°S-67°W and around 1 Gyr for the one at 25°S-87°W. The number of craters larger than 2, 5, and 16 km per million km² in the countings (Table 3) would date those areas as Early and Middle Amazonian respectively [*Scott and Tanaka*, 1986]. Thus some of the Thaumasia Plateau strike-slip faults are apparently incised on Amazonian volcanic plains.

In consequence, we have two spatially superposed tectonic systems, both apparently caused by horizontal compressive stresses but widely separated in time. These stresses, moreover, are coincident in orientation: as said, E-W for the buckling [*Schultz and Tanaka*, 1994]; E-W to N80W for the Nectaris Fossae strike-slip system [*Schultz*, 1989]; and with a maximum compressive stress of N75W as deduced by the present authors from the geometry of the Riedel shears at Felis Dorsa (26°S, 65°W). The dynamic evolution could be as follows: Once stabilized, the Martian lithosphere evolved through a period (Late Noachian to Hesperian) in which their mechanical features were similar to those of the modern Earth oceanic lithosphere; when subjected to compression, it buckled [*Schultz and Tanaka*, 1994] or even subducted [*Sleep*, 1994]. It probably showed also thin-skin tectonic features, most of which are now obliterated, at least in the Tharsis area. Much later, in Amazonian times, the lithosphere was compressed again, but this time it was too thick and strong to deform as a coherent unit, and thin-skin tectonics ensued. It is unclear whether this Amazonian deformation happened only once or several times.

The probably youngest structure detected, a 500-km-long sigmoid structure centered on 3°S, 126°W (Figure 2d) already studied by *Anguita and Moreno* [1992], cuts the Arsia Mons aureole, a deposit of uncertain origin but indisputably recent, since it is (at Viking typical resolutions) craterless. This accident is continued by another sigmoid structure (at 4°N-130°W and grossly parallel to the Gordii Dorsum fault) that is affecting an area mapped as Upper Amazonian in the U.S. Geological Survey map [*Scott and Tanaka*, 1986]. This set deserves further attention as a possible example of Martian neotectonics, an idea supported by the coincidence in length and strike between the Arsia Mons fault and the much older one at Gordii Dorsum [*Forsythe and Zimbelman*, 1988]: reactivated basement?

Every possible cause attempting to explain the deformations of the Martian lithosphere is loaded with

problems. The polar wandering models [*Schultz and Lutz*, 1988] have been disputed [*Grimm and Solomon*, 1986], and the palaeotectonics they propose is at odds with the evidence obtained from the images. The rapid and limited rotational reorientation proposed by *Melosh* [1980] generally agrees with the real tectonics; it has, moreover, the advantage of avoiding the necessity of attributing to pure chance the equatorial position of Tharsis. Nevertheless, as also pointed out by *Turcotte* [1999], the transient plate tectonics regime [*Sleep*, 1994] has definite advantages over the rest of the Martian tectonic hypotheses, since it can explain the dichotomy as well as the very flat topography of the northern lowlands [*Smith et al.*, 1998]. We must add that it is also remarkable that the epoch (Early Hesperian) and the maximum stress direction (for Tharsis, nearly E-W) proposed by *Sleep* [1994] for his “Martian plate tectonics” can account for the South Tharsis buckling structures. In turn, the lithosphere rheology we have deduced for this epoch is quite favorable for a period of plate mobility. This is also the approximate period when the “plate tectonics window” concept of *Condie* [1989] would apply to Mars.

A plate tectonics regime, while not being free of problems, is thus viewed by the present authors as a feasible hypothesis to justify the Early Hesperian tectonic phase. As for the Amazonian reactivation(s), the origin of the horizontal stresses remains a matter of speculation. Now that the existence of big-scale horizontal stress fields is sustained by the weight of observations, new schemes for Martian global tectonics are needed.

5. Conclusions

1. A revision of Viking imagery of the Tharsis dome has permitted the identification of a number of strike-slip faults. These had been predicted by theoretical models as systems radial to Tharsis. Their location and geometry, nevertheless, do not fit in a vertical σ_1 radial stress field, but in a horizontal, E-W-oriented σ_1 one.

2. This new evidence has been used in a quantitative reconstruction of the mechanical evolution of the Martian lithosphere through time. The buckling/thrust faulting of South Tharsis in Late Noachian/Early Hesperian times, which can be compared to the one shown by some areas of the present terrestrial lithosphere, could represent an epoch in which the mechanical Martian lithosphere could deform as a whole. It is proposed that Mars was then experiencing a transient stage of high lithospheric mobility, as previously proposed by *Condie* [1989] and *Sleep* [1994] on more theoretical grounds.

3. Geometrical, geophysical, geomorphological and structural observations support, or are compatible with, the hypothesis that Valles Marineris is a right-lateral transtensive megashear.

4. The crustal deformation recurred in Amazonian times, in

this occasion originating thin-skin tectonic structures such as the strike-slip faults found in South Tharsis. In some Tharsis areas observations point to a recent age for the deformations.

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