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Stress models for Tharsis formation, Mars

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Abstract. A critical survey is presented of most stress models proposed for the formation of the tectonic structures in the Tharsis volcano–tectonic province on Mars and provides new constraints for further models. First papers, in the 1970s, attempted to relate the Tharsis formation to asthenospheric movements and lithosphere loading by magma bodies. These processes were then quantified in terms of stress trajectory and magnitude models in elastic lithosphere (e.g. Banerdt *et al.*, *J. Geophys. Res.* **87**(B12), 9723–9733, 1982). Stresses generated by dynamic lithosphere uplift were rapidly dismissed because of the poor agreement between the stress trajectories provided by the elastic models and the structural observations. The preferred stress models involved lithosphere loading, inducing isostatic compensation, and then lithosphere flexure. Some inconsistency with structural interpretation of Viking imagery has been found. In the early 1990s, an attempt to solve this problem resulted in a model involving the existence of a Tharsis-centred brittle crustal cap, detached from the strong mantle by a weak crustal layer (Tanaka *et al.*, *J. Geophys. Res.* **96**(E1), 15617–15633, 1991). Such a configuration should produce loading stresses akin to those predicted by some combination of the two loading modes. This model has not been quantified yet, however it is expected to reconcile stress trajectories and most structural patterns. Nevertheless, some inconsistencies with observed structures are also expected to remain. Parallel to this approach focused on loading mechanisms, the idea that volcanism and tectonic structures could be related to mantle circulation began to be considered again through numerical convection experiments, whose results have however not been clearly correlated with surface observations. Structural clues to early Tharsis dynamic uplift are reported. These structures have little to do with those predicted by elastic stress modelling of dynamic lithosphere uplift. They denote the existence of unsteady stress trajectories responsible for surface deformations that cannot be readily predicted by elastic models. These structures illustrate that improving current stress models for Tharsis formation shall come from deeper consideration of rock failure criterion and load growth in the lithosphere (e.g. Schultz and Zuber, *J. Geophys. Res.* **99**(E7), 14691–14702, 1994). Improvements should also arise from better understanding rheological layering in the lithosphere and its evolution with time, and from consideration of stress associated to magma emplacement in the crust, which may have produced many tectonic structures before loading stress resulting from magma freezing became significant (Mège and Masson, *Planet. Space Sci.* **44**, 1499–1546, 1996a).

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1. Introduction

Tharsis, the major magmato-tectonic province of Mars (Fig. 1), covers one-third of the planet's surface. Its bulged topography reaches 11 km in elevation in Syria Planum (U.S.G.S., 1991a, 1992). Tharsis is mantled by lava flows, and displays three giant shield volcanoes, the three Tharsis Montes, which are 22–27 km high above the reference topographic surface. The Tharsis western flank is bordered in part by Olympus Mons, another shield volcano of similar elevation. Tectonically speaking, Olympus Mons appears not to belong to the Tharsis unit. The latter was built south of the dichotomy boundary, on the intensely cratered uplands, whereas Olympus Mons was built on the more recent lowlands. Whether Alba Patera, a nearly flat volcano located north of the Tharsis Montes, belongs or not to the Tharsis province is conjectural, because the dichotomy boundary has been buried by the huge Alba Patera lava flows.

Very crudely stated, the Tharsis tectonics is characterized by five major features: (1) Valles Marineris, a

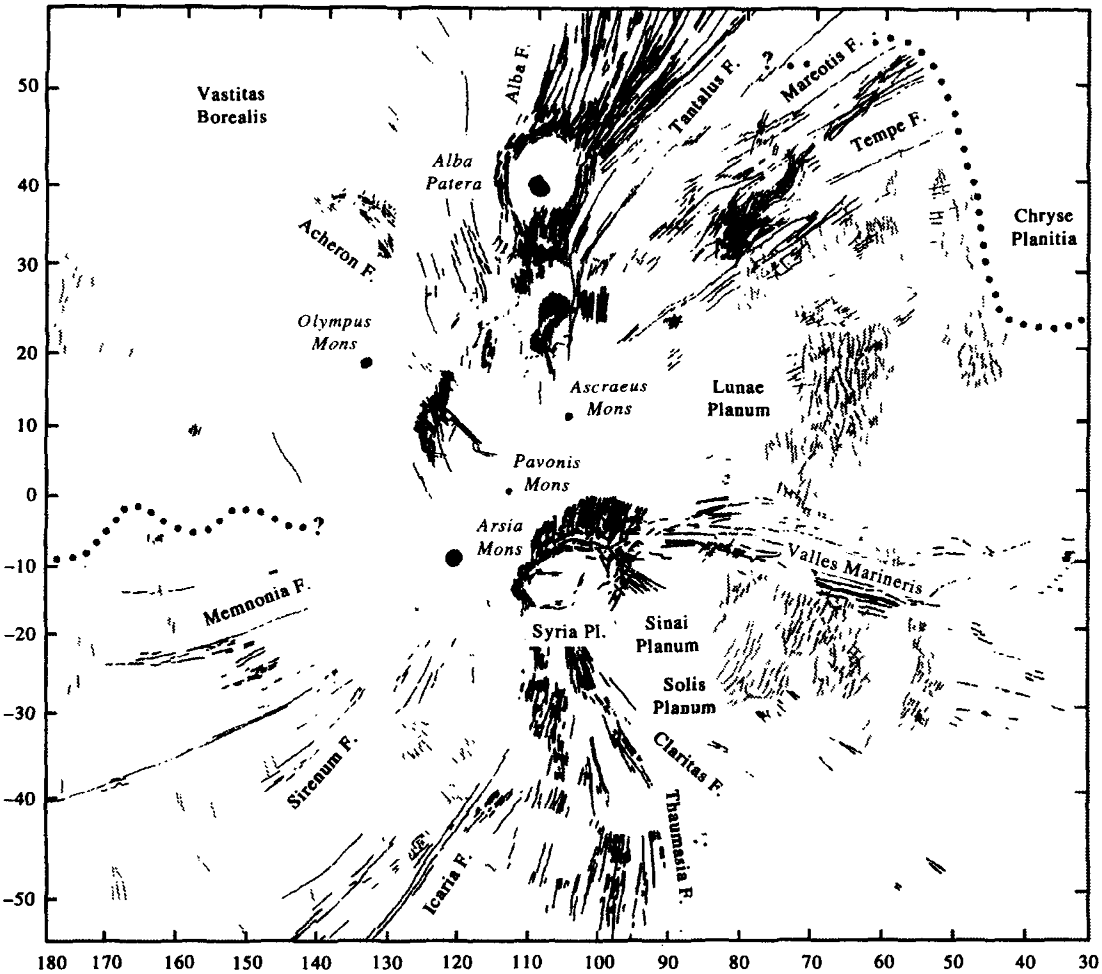

Fig. 1. Simplified structural map of the Tharsis province, displaying the location of Valles Marineris, the main graben sets, the wrinkle ridges (grey patterns, from Watters (1993)), and the caldeiras of Olympus Mons, Alba Patera, and the Tharsis Montes

graben system more than 200 km long and 600 km wide, located on the eastern Tharsis flank. The graben border faults have vertical throws up to 10 km. (2) Hundreds of narrow grabens radiating from the Tharsis central area. Their length is up to 2000–3000 km in some cases, their depth, 50–200 m, and their width, mostly less than 5 km. (3) The Alba Patera-related structural patterns, mostly concentric and N/S-trending grabens of characteristics similar to those of grabens in point 2. (4) Compressional ridges, formed in some of the lava fields. They are roughly concentric about the centre of the bulged area. (5) The South Tharsis Ridge Belt, recently found out by Schultz and Tanaka (1994), a series of broad-scale folds (?) up to 100 km wide, 300 km long, and 1 km high, located in the cratered uplands close to the boundary between the bulged and non-bulged areas. The belt is partly concentric about Syria Planum.

Our purpose is to focus on the nature and evolution of the Tharsis support and stress with time in a structural and historical perspective. This paper summarizes most of the works already done by stress workers and examines their relations to the structural patterns, reports new structural interpretations, and suggests research trends for further works.

Since the end of the 1970s, the available data on the Martian surface include (1) Mariner 9 and Viking Orbiters 1 and 2 imagery, (2) topography data, and (3) gravity data. All provide important constraints on Tharsis stress modelling. Structural observations are of primary interest for such a purpose; nevertheless, any geophysical model could successfully account for all of them up to now.

One difficulty is that the comparison between the Tharsis structures and models requires that the structural patterns entirely resulting from the Tharsis evolution and those that were also influenced by other phenomena can be distinguished. Relations between Tharsis and other major features such as Alba Patera and Valles Marineris are still unclear. A quite high level of autonomy has been suggested for Valles Marineris by Anderson and Grimm (1994) and Mège and Masson (1996b), and, for Alba Patera, by Raitala (1988), Tanaka (1990), and Janle and Erkul (1991).

Another major difficulty is the length of many grabens radial about Tharsis (up to 2000 km, sometimes 3000 km) compared with their width (a couple of kilometres), which suggest a steady source of equal-magnitude extensional stress over thousands of kilometres. Moreover, the existence of several tectonic episodes, as shown by, e.g. Plescia and Saunders (1982) and in numerous papers discussed throughout this study, makes restoration of the relationships between stress and strain difficult. Furthermore, some tectonics features which seem to have formed in one stage, may have formed in several stages which cannot be distinguished in light of stratigraphic and geomorphological observations. As an example, most of the Valles Marineris depth dates back to upper Hesperian (Tanaka, 1986), and a general implicit agreement has been found that Valles Marineris formed at the same time from one end of the trough system to the other end. However, no clear evidence, either stratigraphic or morphologic, indicates that all the troughs formed contemporaneously. Upper Hesperian lasted for 1.3 Ga on Neukum and Wise's scale (Neukum and Wise, 1976), and still 150 Ma on Hartmann *et al.*'s scale (Hartmann *et al.*, 1981). Morphostructural methods of fault scarp analysis has revealed the existence of three distinct tectonic events on the Valles Marineris' Candor Chasma northern wall (Mège, 1994; Mège and Masson, 1994a). This considerable uncertainty on the succession of tectonic events during graben opening should be kept in mind, and shows the need for careful interpretation of the Valles Marineris driving processes.

A key problem also is the significance of several Hesperian, and particularly Noachian intricate fault sets (e.g. Scott and Dohm, 1990) observed in several places beneath the recent Tharsis lava flows (Ulysses Fossae, Acheron Fossae, Uranius Fossae, and some fault sets in Tempe Terra), the meaning of which in a regional tectonic context has been poorly discussed. The fault sets around Syria Planum are a strong concern as well. Crosscutting relationships are often contradictory (Masson, 1980), so that Tanaka and Davis (1988) finally used surface dating by crater counts in addition to crosscutting relationships to establish chronological relationships. Another major difficulty in stress modelling is that all the lithosphere support models can account for the topography and gravity data (Sleep and Phillips, 1979)—it should be noted however that gravity data may not help constraining stress models very much since it may not be representative of that when the tectonic activity of Tharsis took place. In 1976, Mutch estimated that "the driving force for the Tharsis uplift is largely speculative" (Mutch *et al.*, 1976); this opinion appears to be strikingly modern.

Banerdt (Banerdt *et al.*, 1992) has provided another review of the current views on the Tharsis stress models. His study summarizes basic concepts on loading, and presents a synthesis of major stress models based on elastic theories. The present paper wishes to be closer to structural interpretations than to theoretical models, so that both contributions are complementary. In Section 2 the pioneer studies are discussed, which were only based on Mariner 9 data. The latter played a key role in the history of Martian geology because the images of Mars obtained earlier (Mariner 4) showed intensely cratered areas only. A second series of works, discussed in Section 3, were carried out in the late 1970s and in the first part of the 1980s, after the Viking data had been obtained. Several of these studies used elastic modelling of stress trajectories and stress magnitudes (Banerdt *et al.*, 1982, 1992; Sleep and Phillips, 1985). Of particular interest is the detached crustal cap model, by Banerdt and Golombek (1990) and Tanaka *et al.* (1991). The elastic models have suggested that lithosphere loading should have been responsible for the formation of most structural patterns in the Tharsis province. However, many problems were found in comparing the elastic models and the observed structures in the Tharsis region (Banerdt *et al.*, 1982; Schultz, 1985). This may be one of the reasons why some other studies focused on what could be used as tests, boundary conditions and other clues for better constraining the dynamics of Tharsis formation (Section 4). New computational possibilities allowed to study the influence of mantle convection on Tharsis formation through numerical modelling of mantle circulation (Kiefer and Hager, 1989; Schubert *et al.*, 1990) (Section 5). Intuitively, that dynamic lithosphere uplift above a convecting mantle took place and played a role in the early evolution of Tharsis is not unreasonable. Indeed, making a thick load in a lithosphere in the absence of large horizontal plate movements probably requires the previous occurrence of a strong mantle plume first, and subsequent formation of voluminous magmatic materials. Only once frozen, can magma behave as a load in the lithosphere. This point is of considerable importance because before loading, the evolution of a lithosphere in which magma was emplaced may already be able to produce stress exceeding the rock strength. In Section 6 some structural arguments in favour of such an early Tharsis uplift associated to intrusion by magmatic materials is discussed, and a brief summary of the results obtained up to now is given.

Each of Sections 2–5 covers a specific research topic. However in subsections we have chosen, when possible, to summarize and discuss each contribution separately to keep an historical perspective.

2. Early models (Table 1)

2.1. *Mantle plume/convection*

Hartmann (1973) suggested that plume-like convection provided the driving forces for Tharsis uplift. He also suggested that "if the Martian temperature gradient [increases in the future], Mars may evolve further toward a differentiated crust with continental blocks, and lose its primitive cratered surfaces". The latter hypothesis, though attractive, seems unlikely from more recent works. It is likely that Hartmann meant that a possible future widespread basaltic differentiation would start a new geologic period characterized by horizontal plate movements, as suggested in some models of continental crust formation and plate tectonics initiation on Earth. Such a differentiation would have been made easier by mega-impacts, but more difficult by rapid planetary cooling due to the small planetary size. It is accepted since the lunar exploration that the mega-impacts occurred during a heavy meteoritic bombardment that peaked 3.9 Ga ago (e.g. Mutch *et al.*, 1976). Mega-impacts have been suggested to have contributed to trigger plate tectonics on Earth (Frey, 1980). Heavy bombardment certainly did occur on Mars at the same time, as suggested by the intensely cratered uplands. The dichotomy formation may be a more or less direct consequence of mega-impacts, early in the planet history (discussions in McGill and

Table 1. Review of stress models of Tharsis: early models (1973–1979). Numbers refer to chronological order of events according to authors

Authors	Origin of stress and support mechanism	Crust and elastic lithosphere thickness
Hartmann (1973)	mantle convection; regional extension NW–SE	
Carr (1974)	asymmetric updoming related to mantle convection	
Wise (1976)	Alba Patera: volcano load; regional extension E–W	
Wise *et al.* (1979b)	1. underplating beneath lowlands by a first-order convection cell 2. isostasy 3. volcanism related to hot underplating and energy of core formation	
Phillips *et al.* (1973)	crustal thickening	Tharis crust 70–120 km if the mean crust is 50 km thick
Phillips and Saunders (1975)	isostasy Airy compensation (appearing to be an insufficient support)	crust $\leq$ 100 km lithosphere 200–500 km
Sleep and Phillips (1979)	isostasy Airy + Pratt compensation	

Squyres (1991), and Breuer *et al.* (1993)), but no evidence of movements at the dichotomy boundary has been reported.

The thermal evolution of planets is also unfavourable to plate tectonics in the future. Classic studies on Mars' thermal evolution (e.g. Toksöz and Hsui, 1978) predict an early core formation and crustal differentiation, then planetary thermal expansion and further cooling. According to Toksöz and Hsui, cooling would now last from about 1.5 Ga ago—but some parameters influencing on the absolute timescale are poorly constrained. Nevertheless, this scenario of evolution appears to fit observations: planetary expansion would have led to partial resurfacing of the crust during lower Hesperian (more than 3 Ga ago), when most of the currently observed volcanic material was extruded (Scott and Tanaka, 1986; Tanaka *et al.*, 1988; Greeley and Schneid, 1991). Lower Hesperian might also correspond with the dichotomy formation according to McGill and Dimitriou (1990), and Hesperian is the age of the northern hemisphere resurfacing (Scott and Tanaka, 1986).

The thermal evolution may however be more complex in details, since the residue left from partial melting under a planetary crust protects this area from a strong coupling with mantle convection, producing mantle heating, and then widespread magmatism (Stevenson and Bittker, 1990). The concept behind this idea is not far from that of the continental tectosphere on Earth (the suggestion that a boundary layer exists in the upper part of the upper mantle and isolates the convective layer below from the lithosphere above) by Jordan (1975). Watters (1993) attributes the lower Hesperian widespread volcanism to such a phenomenon as an alternative view to the mere heating/cooling history of Mars because it takes into account the "catastrophic" character of Hesperian magmatism and subsequent rapid wrinkle ridge formation. Whatever the reason for the Hesperian volcanic peak, the comparison between the Earth, the Moon, Venus, and Mars histories indicates that all these bodies probably underwent a first differentiation early in their history, forming the crust, and then basaltic overplating (complete or partial); all seem now to cool down slowly (Lowman, 1989). Intense cooling has a tragic issue for plate tectonics: it thickens the crust and lithosphere, strongly decreases the probability of block individualization in the future, and then plate movements. Nevertheless, according to the theory of Stevenson and Bittker (1990), Mars may not be secure from a new widespread volcanic event in the future. Halls (1987) compiled the age of many terrestrial tholeiitic dyke swarms formed above thermal anomalies in the mantle and found that dyke swarm emplacement is an episodic phenomenon with peaks at 2.9, 2.5, 2.1 and 0.1 Ga. On Venus, the lava plains have been frequently interpreted to have formed during a catastrophic resurfacing event *c.* 500 Ma ago. Thus, although Stevenson and Bittker's theory may not be fully valid for the Earth because of crustal recycling, and, in fact, whatever the reason for these peaks, magmatic peaks in the evolution of terrestrial planets seem not to be linearly dependent on time. It is for sure however that, due to the small diameter of Mars, it cools more rapidly than the Earth and that the possibility of future partial or complete resurfacing should be weak. Carr (1974) considered that the Tharsis radial fault system may be linked to an asymmetric updoming resulting from mantle convection. Asymmetry was made necessary in order to account for preliminary stretching calculations, the first ones ever calculated on Mars. South of Tharsis, upwarping of an area 4000 km in diameter around the 7 km uplifted Syria Planum requires 0.1% of crustal stretching. In contrast, northward, in the Alba Patera region, stretching, deduced from fault counting and from the assumption of 60° dipping faults, was estimated to 2%. A careful stretching analysis for the Alba Patera region by Plescia (1991) from Viking data gave about 2% stretching in the Ceraunius Fossae area around 110°W, but only a bulk 0.5% in the whole Alba Patera region, so that the updoming asymmetry would be less developed than expected by Carr. Carr suggested that the Tharsis shape could be due to a mechanism responsible for the symmetric component of bulging, and another one responsible for the asymmetric component. The symmetric component would be produced by doming, producing radial patterns, like in uplifted magmatic regions on Earth. The asymmetric component should be due to another phenomenon: Carr sug-

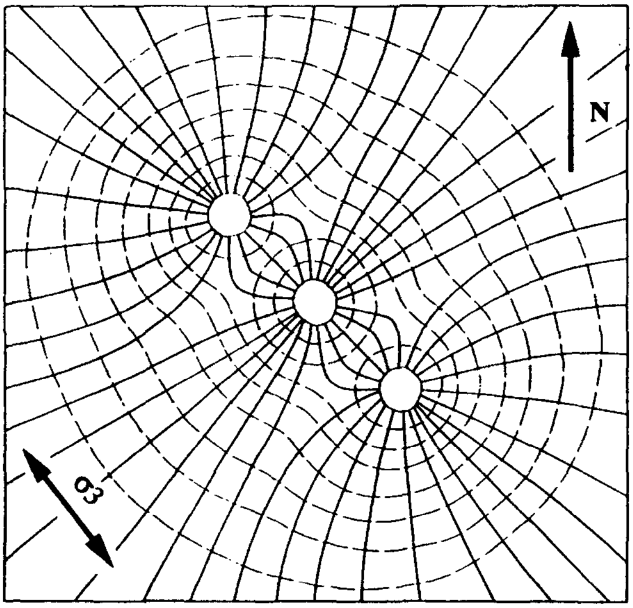

Fig. 2. Horizontal principal stress trajectories generated by simultaneous inflation of three magmatic chambers located beneath the three main Tharsis volcanoes endowed with the same overpressure. Note that the stress patterns are significantly different from the observed radial patterns on Fig. 1 (Carr, 1974)

gested it to be due to the crosscutting of Tharsis by the dichotomy (the flanks of the three main Tharsis volcanoes appears to cross a rift line, e.g. Crumpler and Aubele (1978), which could be linked with the dichotomy boundary), or to a planet-wide stress system whose minimum principal stress trajectory is NW–SE oriented, an idea that was taken from Hartmann (1973) for explaining the alignment of the Tharsis volcanoes. More recent data (Viking) and works (e.g. Scott and Tanaka, 1986) show that several stages of extension occurred in the Alba Patera region. Plescia and Saunders (1982) noted that at least two stages of crustal extension occurred there, one during Noachian, and one later, but the Noachian episode does not require uplift of the Alba Patera area. This means that some faults measured by Carr (1974) may not have contributed to topographic doming.

Carr (1974) put forward the similarity between the Tharsis radial fault system and stress patterns in magmatic regions on Earth where dyke swarms intrude the continental crust (May 1971). At a first look, comparison between the Tharsis radial patterns and stress fields around magmatic centres suggests indeed that the Tharsis radial fault system could be a surficial expression of shallow dyke intrusions. Carr (1974) pointed out the analogy with the Spanish Peaks, Colorado, a double magma reservoir studied by Knopf (1936), Odé (1957), Johnson (1961), and Muller and Pollard (1977). The radial dykes of Spanish Peaks are different from many dykes around other volcanic centres (e.g. Weed and Pirsson, 1895) because at some distance they tend to follow a systematic trend. This preferential direction is due to the influence of regional stresses, as shown by Odé (1957). These points are raised in more details in Mège and Masson (1996a). Carr used a simplified version of Odé's (1957) numerical method for the Spanish Peaks to model the stress trajectories produced by overpressure in three magma chambers located beneath Arsia, Pavonis, and Ascraeus Montes (Fig. 2). He then considered that this analysis is "*questionable*" for two main reasons. First, the Spanish Peaks and Tharsis fault networks have lengths two orders of magnitude different. However dyke networks more than 2000 km long exist on Earth (reviews in Halls (1982) and Fahrig (1987)) and are expected on Venus (McKenzie *et al.*, 1992), and their mechanism of emplacement over such distances appears now to be well understood (Delaney, 1987; Lister, 1990; see also Bruce and Huppert, 1989). The second reason is that the Tharsis radial patterns numerically reproduced by Carr (1974) do not fit the observed patterns. Carr noticed that diminishing the effect of the central Tharsis volcano (Pavonis) would improve the correlation between the model and reality, because it would take the absence of radial faults toward Valles Marineris and Olympus Mons into account. That two of the three Tharsis volcanoes may have a common magma source cannot be dismissed indeed: assuming that the relation "1 volcano = 1 magma chamber" is necessarily correct would be an oversimplistic view of magma behaviour from analogy with the Earth. Deleting the influence of Pavonis Mons in dyke and graben formation is in apparent contradiction with results from structural mapping by Plescia and Saunders (1982), which revealed that most of the radial faults are radial about Pavonis Mons. This paradox can be understood if the influence of a Tharsis-independent stress source in graben geometry is accounted for. Based on Odé's work (Odé, 1957), Carr addressed the question of the possible existence of such a regional source of stress. However, he did not model what would be its influence on stress patterns in the Tharsis case. Results presented in Mège and Masson (1996a) tend to confirm its existence.

Wise, Golombek and McGill (Wise *et al.*, 1979b, modified from Wise *et al.*, 1979a) proposed an ingenious two-stage model for the Tharsis rise. Subcrustal erosion of the northern plains could have provided light material that would have been underplated beneath the Tharsis region by mantle movements, inducing rapid Tharsis uplift, thermal anomaly in the mantle, and subsequent volcanism. Subcrustal accretion cannot be ruled out since it is a means to thicken the Tharsis lithosphere which is expected to produce membrane stresses the same way as isostasy does in the current elastic models discussed below. All the faults of Tharsis one can attribute to isostatic stresses might also be interpreted by subcrustal accretion below Tharsis. Similarly, underplating would also be confronted with the difficulties of isostasy for explaining many structural features, as discussed by Schultz (1985) and in Section 3.3. The model by Wise *et al.* (1979b) presents a similarity with the model of dichotomy formation by Breuer *et al.* (1993), also discussed below. Breuer *et al.*'s model does not imply subcrustal erosion of the northern lowlands, but it does not exclude this mechanism.

2.2. *Loading models*

Like Carr (1974), Wise (1976) modelled the stress patterns in and around Alba Patera using a method derived from

Odé's approach (Odé, 1957) of stress modelling. Assuming that both the radiating and circumferential graben trends around Alba Patera should reflect intermediate principal stress trajectories, Wise noted that the grabens can be satisfactorily explained by the addition of stress generated by the Alba Patera crustal load to a remote extensional stress whose least compression trajectory is oriented north–south. Noting that the stress levels in his model cannot predict rock failure because they are too low, he suggested to replace the elastic behaviour of rock below the grabens by a long-term viscous flow behaviour, all boundary conditions kept equal. Stress would have concentrated in the elastic upper layer, anticipating the concept of stress amplification in the Earth's brittle layers developed by Kusznir and Bott (1977). He proposed that the graben system formed within this thin elastic surface layer, overlying a "massive homogeneous crust or mantle undergoing slow small viscous flow in response to gravitational loading". Although the stress patterns obtained by Carr (1974) and Wise (1976) both derive from Odé's work (Odé, 1957), the interpretation for the source of stress is different: Carr (1974) attempted to explain graben patterns through overpressure in magma chambers, like Odé (1957), whereas Wise (1976) put forward the role of volcanic loading in the crust. Another difference is that, contrary to Carr, Wise did not discuss the issue of hypothetical dykes beneath the graben patterns. The application of Odé's method to dykes is an important feature, since broad-scale dyke trends are generally good indicators of regional stress trajectories. Grabens are usually not indicators of stress trajectories, so that Wise's analysis can be believed only because the Alba Patera lava flows crosscut by the grabens appear not to have been horizontally sheared.

The three following loading models considered here deal with purely isostatic loads. Phillips *et al.* (1973) calculated an Airy compensation model along two complete equatorial profiles of Mars. They first calculated the Bouguer anomaly from the Martian gravity field deduced from Doppler tracking from the Mariner 9 probe. Topography, the effects of which have to be removed, was provided by occultation data and Earth-based radar profiles. Assuming a mean 50 km thick crust for the 0 level of gravity anomaly, Phillips *et al.* calculated crustal thickness variations along these two profiles, that they compared with Airy profiles. A quite good—but not perfect—fit between the lower crustal boundary provided by the Bouguer and Airy profiles was obtained, assuming typical terrestrial densities for the upper mantle. An important result was found, whatever the reasonable density values: the crust should be currently thicker in the Tharsis region than the mean crustal thickness. The authors also found that extrusives cannot account for this thickening, since the Noachian terrains (not defined as Noachian in 1973) beneath the Tharsis lava pile were uplifted. Phillips *et al.* (1973) proposed that the Tharsis surface uplift was accompanied by volcanism and radial faulting. Despite the partial correlation between Bouguer and Airy, the buoyant and flexural parts of the support were not discussed. However, many of the ideas developed later were suggested, including the buoyant component of the load, the likely importance of flexural loading, the role of intrusions compared with lava flows in crustal thickening (an alternative to underplating suggested by Wise *et al.* (1979b)), and the links between crustal thickening and radial faulting.

Phillips and Saunders (1975) discussed a pure isostatic model for Tharsis with a simple Airy compensation. The crust thickness would be equal to, or less than, 100 km. The calculations led to the conclusion that isostasy could not account for the Tharsis topography because of the excessive compensation depth required. Indeed, Tharsis primarily contributes to the second and third degree harmonics (Phillips and Lambeck, 1980; Schubert *et al.*, 1990), and these degrees would require 1000 and 600 km compensation depth, respectively.

Another purely isostatic model was given by Sleep and Phillips (1979). In this model, shallow mass differences are compensated by a crustal Airy model, while deep mass differences are Pratt compensated in the lower lithosphere. This lithospheric configuration was also used in the further Tharsis models. It implies a large quantity of crustal intrusions with respect to extrusive materials. Sleep and Phillips showed that the lithosphere should be 200–500 km thick, with 325 km as a preferred value, close to the values still considered. However, pure isostasy does not predict the whole tectonic structures observed, as discussed subsequently.

3. Flexural loading/isostasy combined elastic models (Table 2)

Three types of topographic supports have been studied through elastic modelling since 1982 (Fig. 3): flexural uplift (Fig. 4a), isostasy (Fig. 4b), and flexural loading (Fig. 4c) (Banerdt *et al.*, 1982, 1992; Sleep and Phillips, 1985). Different combinations of these models were extensively discussed. Almost all models proposed between 1982 and 1992 took into account some combination of flexural uplift, flexural loading and isostasy to explain the origin of the Tharsis rise. The flexural uplift model was rapidly dismissed, because clearly, the major tectonic features are in contradiction with the predicted stress ≡ patterns. A good illustration is given by Valles Marineris, whose direction of extension is perpendicular to what is expected from elastic modelling of flexural uplift. Banerdt *et al.* (1982) developed a thick shell approach, whereas Sleep and Phillips (1979, 1985), Willemann and Turcotte (1982), and Banerdt *et al.* (1992) used different approaches of the easier thin shell theory. The thin shell approach allows to consider isostatic stress as membrane stress, which is easily computed from elastic theories. Obviously, the problem is that the thin shell theory requires the shell to be thin with respect to the planetary radius, and how thin it must be has been discussed. Willemann and Turcotte (1982) keep within 10% of the radius, and Sleep and Phillips (1985) refer to 13%. Janes and Melosh (1990) think that 10% is the very maximum possible value relevant to the thin shell theory. However, the stress maps given by both thick and thin shell calculations are very similar, and the variations of crust thickness and lithosphere thickness are very close.

Table 2. Review of stress models of Tharsis: flexural load and isostasy combined models (1982–1990). Numbers refer to chronological order of events according to authors. *Flexure* always refers to a loading process

Authors	Origin of stress and support mechanism	Crust and elastic lithosphere thickness
Solomon and Head (1982)	1. isostasy (thin lithosphere) 2. flexure (thickened lithosphere) + local isostasy	lithosphere Alba Patera: 25–50 km Olympus Mons: > 50 km Ascraeus Mons: 40 ± 15 km Pavonis Mons: 30 ± 10 km Arsia Mons: 25–50 km
Willemann and Turcotte (1982)	partial isostasy plutonism more important than volcanism	crust: 40–70 km lithosphere: 110–260 km
Banerdt *et al.* (1992)	1. isostasy (Airy + Pratt) 2. flexure 3. flexure (recent deformations)	1. crust: 100 km lithosphere: 400 km 2. crust: 50–150 km lithosphere: 100–400 km
Sleep and Phillips (1985)	1. • initiation by flexure • isostasy (Airy + Pratt) • planetary expansion → most of the radial fractures 2. • isostasy (Airy + Pratt) • planetary retraction → some radial fractures, wrinkle ridges	crust average thickness: 50 km volcanoes ≅ 0 km lithosphere 450 km
Tanaka *et al.* (1991) and Banerdt and Golombek (1990)	1. magmatism 2. isostasy (inner regions—ductile lower crust) + flexure (outer regions—entirely brittle crust)	
Janes and Melosh (1990)	isostasy/flexure = 1.11 (support parameter) pure loading support considered only	

The importance of flexural loading compared with isostasy is not easy to estimate because the amount of stress reduced by flexural relaxation is unknown and there is no reliable mean for its estimation. From this point of view, an interesting approach has been provided by Janes and Melosh (1990). They pointed out that the tectonic response of terrestrial planets to loading is a function of the interaction between the shell thickness, the load width, and the buoyancy/flexural support ratio, called support parameter. In this respect, Mars is expected to be in a transitional state between two end members of planetary loading modes: loading of a thick lithosphere with a limited load, and loading of a thin lithosphere with a large load. The support parameter for Mars has been estimated to be 1.11.

Two kinds of models have been proposed. The first one suggests that the tectonic structures result from the succession of isostasy and flexural loading, and the second one suggests that both support modes existed contemporaneously at different places of the Tharsis province, allowing to form more structures at the same time.

3.1. *Time-dependent support*

3.1.1. *Models.* In the model by Solomon and Head (1982), 50 and 450 km crustal and lithospheric thicknesses are assumed. At the end of the heavy bombardment (3.8 Ga), the lithospheric strength had significant strength heterogeneities. The authors underlined that this could be an effect of a multi-ring buried impact basin under the Tharsis bulge, whose scarce clues were pointed out by Schultz and Glicken (1979). Schultz and Frey (1990) and Frey and Roark (1995) have suggested other locations for possible early mega-impact basins beneath the Tharsis lava flows. Mega-impacting in the Tharsis region would have thinned the lithosphere and enhanced magmatic activity. However, large impact basins on other planetary bodies (Venus, Mercury, the Moon) have never produced magmatic building like in the Tharsis area, so that such event likely did not play a key role in the evolution of Tharsis.

Before the end of the heavy bombardment, this magma would have produced a topographic rise, locally isostatically compensated by crustal subsidence into the thin lithosphere. Later, the lithosphere would have been strengthened because of its thickening with time, so that the larger load would have been supported by regional lithosphere flexure. Currently, Tharsis would be supported by a combination of flexure of the globally thickened lithosphere and local compensations due to local lithospheric thinning by volcanism. Table 2 indicates the predicted lithosphere thickness beneath Alba Patera, Olympus, Ascraeus, Pavonis, Arsia and Elysium Montes. This model is consistent with two others (Willemann and Turcotte, 1982; Banerdt *et al.*, 1982) that attempted to clear up the processes having led to the current Tharsis using structural observations, gravity and topography data.

Willemann and Turcotte (1982) postulated that, regarding the thin lava flows, the major part of the magmatic material should lie within the lithosphere as intrusions. This was also postulated by Phillips *et al.* (1973), Plescia and Saunders (1980), Sleep and Phillips (1979), and De Hon (1982). Willemann and Turcotte predicted partial compensation of the lithosphere, whose thickness range

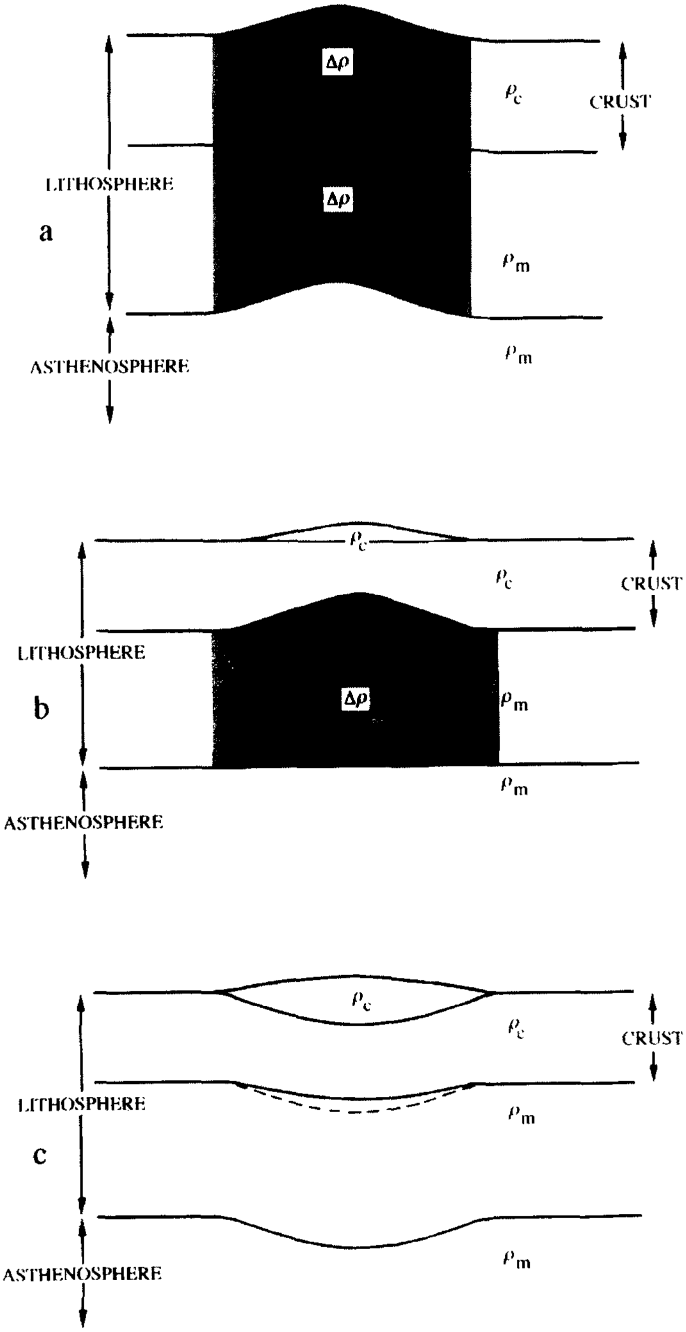

Fig. 3. Flexural uplift (a), Pratt+Airy isostasy (b), and flexural loading (c) are the three main ways to support the Tharsis topography in the whole Tharsis stress literature. In (a) and (b), Δr corresponds to a lateral density anomaly. In (c), the difference between the dashed line and the solid line at the crust–mantle boundary represents harmonically varying amount of crustal thinning. Figure from Banerdt *et al.* (1982)

(110–260 km, depending on the crust thickness and density) is lower than expected by Sleep and Phillips (1979), and a magmatic load thickness ranging within 40–70 km.

The work by Banerdt *et al.* (1982) is the first having considered both loading and dynamic supports of Tharsis, which were used to quantify the resulting lithospheric stress (Fig. 4). Application of the thick shell theory to Tharsis led to a model consistent with the model proposed by Solomon and Head (1982). During a first stage, volcanic centres and the Tharsis rise would have formed. The magmatic load would have been completely compensated isostatically, assuming that the crustal mass before and after the bulge formation was preserved. Grabens would have begun to form in the topographically highest regions. The crust and lithosphere thicknesses at this stage would have been close to 100 and 400 km, respectively. During the second stage, the more voluminous magma would have behaved as a flexural load. This increasing load would have induced the formation of radial grabens. The most recent episodes of Tharsis volcanic activity would correspond to a further stage of loading which would have produced only minor additional deformations at the surface.

Sleep and Phillips (1985) did not consider dynamic uplift, dismissed by the works by Banerdt *et al.* (1982) on structural grounds. They used the wrinkle ridges in the Tharsis region to further constrain the support models. Indeed isostasy appears to be the only theoretical way among the investigated modes of support producing compressional stresses more or less consistent with the location of the wrinkle ridges. Structural studies showed that most of the ridges in the Tharsis region are older than, or contemporaneous with, the radial grabens in the same area (Watters and Maxwell, 1983). Therefore isostasy would be required to explain wrinkle ridge formation, but another major source of stress is expected to have existed more recently to produce the more recent extensional structures. Sleep and Phillips showed that this source of stress could have been flexural loading.

This extensional state of stress is considered by Sleep and Phillips (1985) to be necessary for explaining the occurrence of the Tharsis volcanism. However it must be underlined that volcanism appears not to be linked to the existence of an extensional state of stress on Earth. Volcanism sometimes occurs during crustal shortening, noticeably depending on the magma buoyancy (e.g. Glazner and Bartley, 1994), so that, e.g. alkali basalts are more likely to extrude in a compressive regime than tholeiitic basalts. Gomorphological observations suggest that many Tharsis lava flows should have a basaltic-type composition, but no one currently knows their exact composition, so that the requirement for contemporaneous extensional state of stress is still uncertain.

3.1.2. *Predicted stress versus observed structures.* Below are summarized some of the main inconsistencies between the observed structures and the stress predicted by elastic models of loading. Some of them were reported in Banerdt *et al.* (1982) and Schultz (1985).

1. The only Valles Marineris region summarizes a frequent problem for conciliating stress and strain: as long as the length of radial extensional structures does not exceed about 1000 km, and are located either close to its centre or much farther, they can frequently be readily predicted by isostatic stress or flexural stress, respectively, but forming them from one end to the other end in one stage (i.e. following a single type of support) is often not possible. This remark also applies to the radial grabens of Tempe Terra, and to Icaria Fossae, whose northern part, in the Amazonian lava flows, requires isostatic stress whereas the southern part requires flexural stress.

(a)

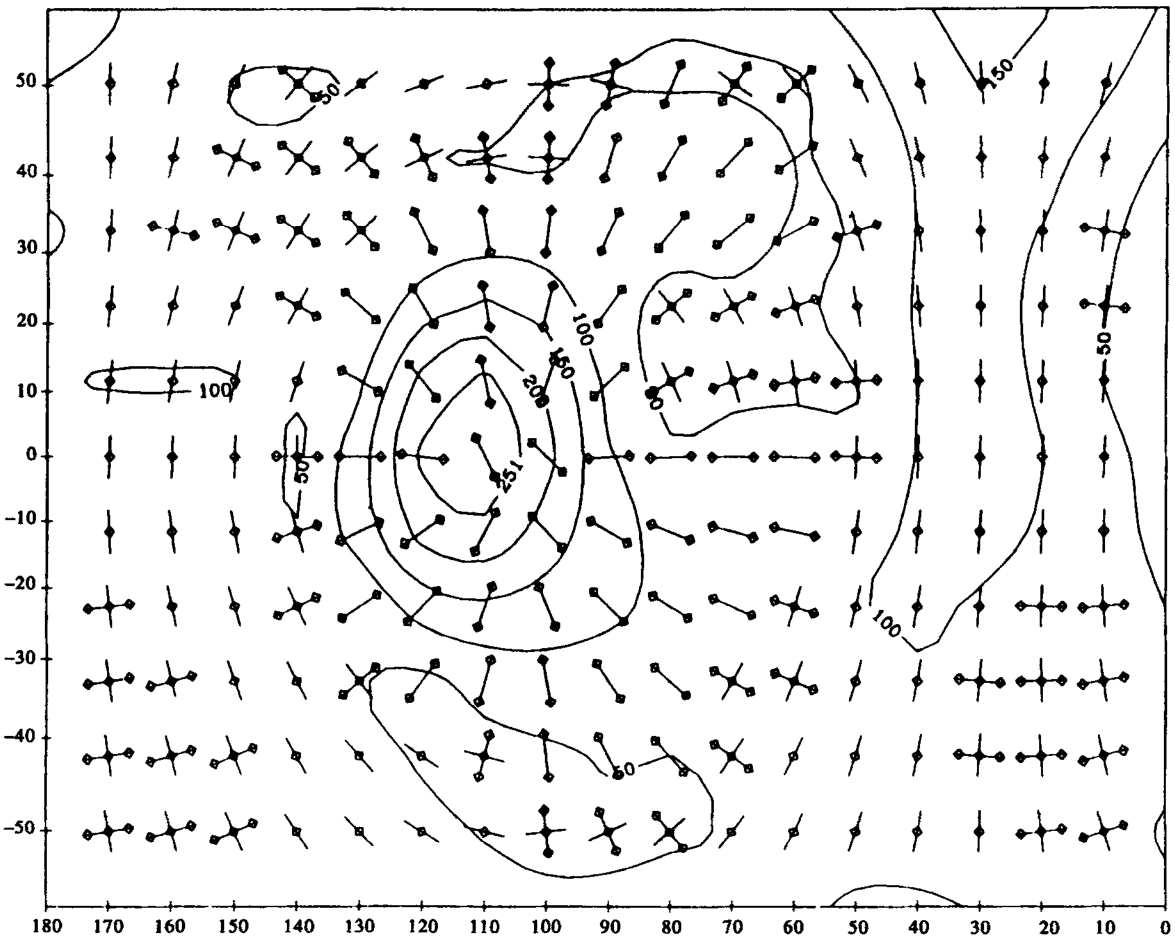

Fig. 4. Horizontal principal stress trajectories expected from the three modes of support presented on Fig. 3, plotted on a Mercator projection map (figure taken from Banerdt *et al.* (1992)). The stress orientation indicators with symbols at each end denote extension. Boundary conditions use topography and gravity harmonic coefficients up to degree and order 8. The same directions and roughly the same states of stress are also obtained from the calculations by Banerdt *et al.* (1982) and Sleep and Phillips (1985). Contours show the magnitude of the maximum stress difference at the surface in MPa. The elastic lithosphere thickness is 200 km, the mean crustal thickness is 100 km. (a) Flexural uplift stresses. Tharsis formed by buoyant uplift of the lithosphere below a locally thinned crust, due to density decrease in the upper 350 km of the mantle, most of which is below the lithosphere. (b) Isostatic stresses. The negative density anomaly in the mantle is taken to be 0.18 $Mg\,m^{-3}$, and the Tharsis crust is about 60 km thinned. (c) Flexural loading stresses. The topography of Tharsis is completed in adding material to the surface, causing a peak lithospheric deflection of 8 km. The local thickness of the crust is simultaneously adjusted in order to satisfy the observed gravity, reaching a maximum excess thickness of 28 km beneath Tharsis. The crust/mantle density contrast is assumed to be 0.5 $Mg\,m^{-3}$. The boundary of the possible detached crustal cap (Banerdt and Golombek, 1990) should be approximately located as shown on (b) and (c) (heavy dotted line)

2. Few parts of Claritas and Thaumasia Fossae only are predicted by whatever support mechanism. The faults north of Alba Patera are not explained by any of these models although the southern ones (Ceraunius Fossae) may be explained by isostasy. Formation of the South Tharsis Ridge Belt (Schultz and Tanaka, 1994) is not predicted either.
3. A very little discussed concern is the fault orientations in the westernmost part of Valles Marineris (Ius and Tithonium chasmata). Theoretical stress trajectories are consistent with the global N105°E Valles Marineris orientation. The change to the N090 trend eastward (Gangis Chasma and the eastern part of Coprates Chasma) is also predicted. But the stress trajectories do not predict the clear 300 km long N090 trend of the western part of Ius and Tithonium chasmata.
4. The existence of Noachian grabens in Tempe Fossae (Scott and Dohm, 1990), as well as the interpretation of Gordii Dorsum as a sinistral transcurrent fault zone (Forsythe and Zimbelman, 1988), are in agreement with a flexural support, however a flexural support during Noachian is in contradiction with the scenario of support evolution in the previously mentioned models.

(b)

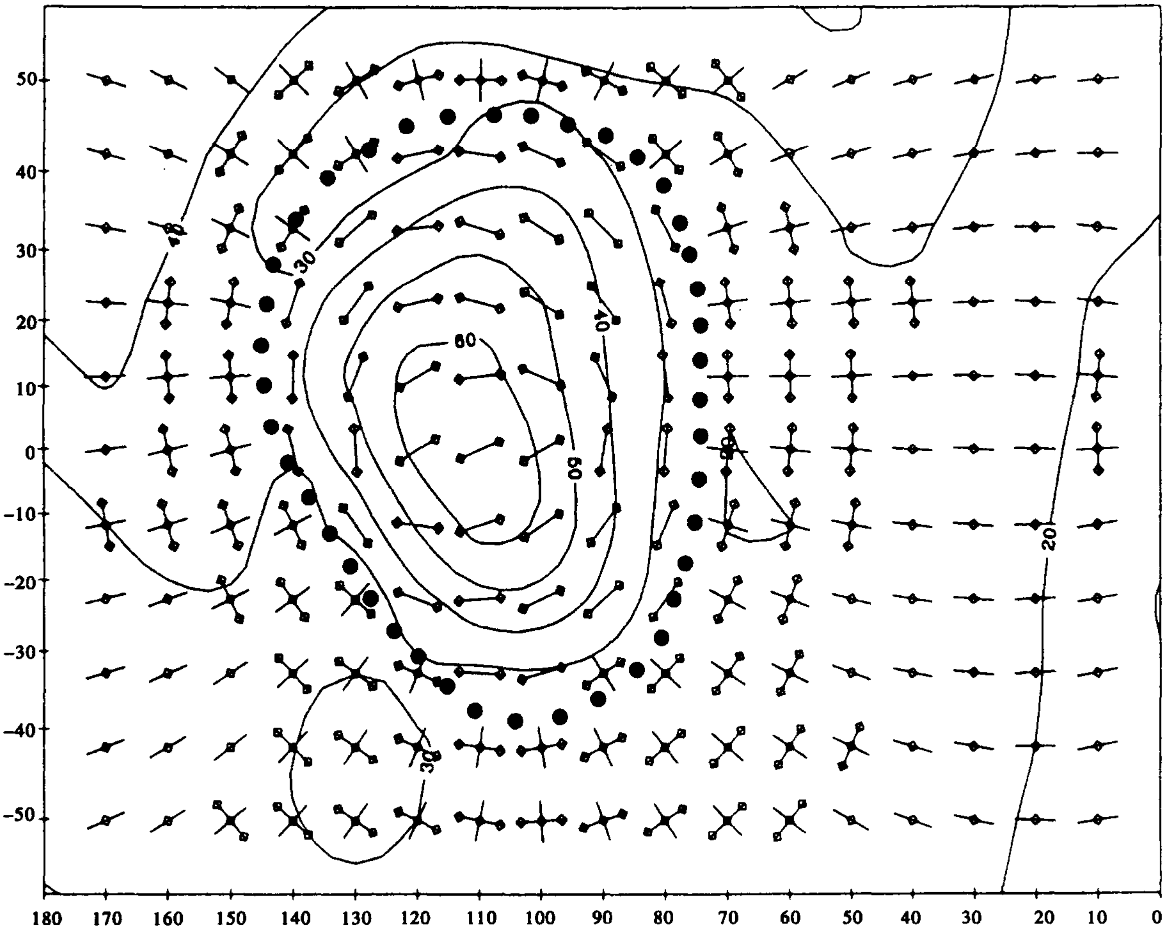

Fig. 4. (*Continued*)

3.2. *Geographically zoned support*

Banerdt, Golombek and Tanaka (Banerdt and Golombek, 1990; Tanaka *et al.*, 1991) proposed a mechanism of contemporaneous formation of all the radial grabens in order to take the series of structural inconsistencies with stress models raised in point 1 of the previous section into account. Their model is mainly inferred from the location of Valles Marineris in both the domain where its formation is predicted by isostatic models and the domain where it is predicted by flexural loading models. The Valles Marineris formation is assumed to have resulted from a single stage from its western end to the eastern end (Blasius *et al.*, 1977; Masson, 1980; Schultz, 1991). However, as pointed out in the introductory part, Valles Marineris opening in several stages cannot be ruled out because of the uncertainty on timescales. Anderson and Grimm (1994) suggested that the Valles Marineris opened by rift propagation; wallslope analysis and details of extension variations along the chasmata (Mège and Masson, 1994b, 1996b; Peulvast *et al.*, 1996; Schultz, 1995a) suggest a complex history that does not necessarily imply that all the Valles Marineris grabens formed simultaneously.

The "detached crustal cap" hypothesis proposed by Banerdt and Golombek (1990) and Tanaka *et al.* (1991) attempts to reconcile the whole tectonic evolution of Tharsis, since late Noachian, and stress models (Tanaka *et al.*, 1991). The Tharsis lava plains would have resulted from early Tharsis volcanic activity associated with local extensional tectonics. Meanwhile, rapid planetary cooling would have produced a compressional stress field, and consequently wrinkle ridges formation. Differences in crustal structure are assumed to have existed beneath inner and outer parts of Tharsis in the following stages. The crust beneath the surrounding regions of Tharsis is supposed to be entirely brittle, whereas a weak crustal layer would decouple the brittle crust from the lower lithosphere in the Tharsis central areas. The stresses produced this way in the brittle cap would be akin to isostatic stresses (membrane stresses). Coupling between the brittle crust and the elastic upper mantle in the outer parts of the dome would lead to flexural stresses, similar to those computed in previous elastic models of flexural loading.

This suggestion of crust/mantle decoupling must probably be taken into account in further models. For simple rheological reasons, it appears to be highly probable that the crust beneath Tharsis is rheologically layered. In many regions of the Earth, especially in regions of crustal extension, the crust is divided into a brittle and a weaker layer. Since the paper by Phillips *et al.* (1973), the preferred mean Martian crust is taken to be about 50 km. Considering a mean 3000 kg m^{-3} crustal density, the lithostatic pressure

(c)

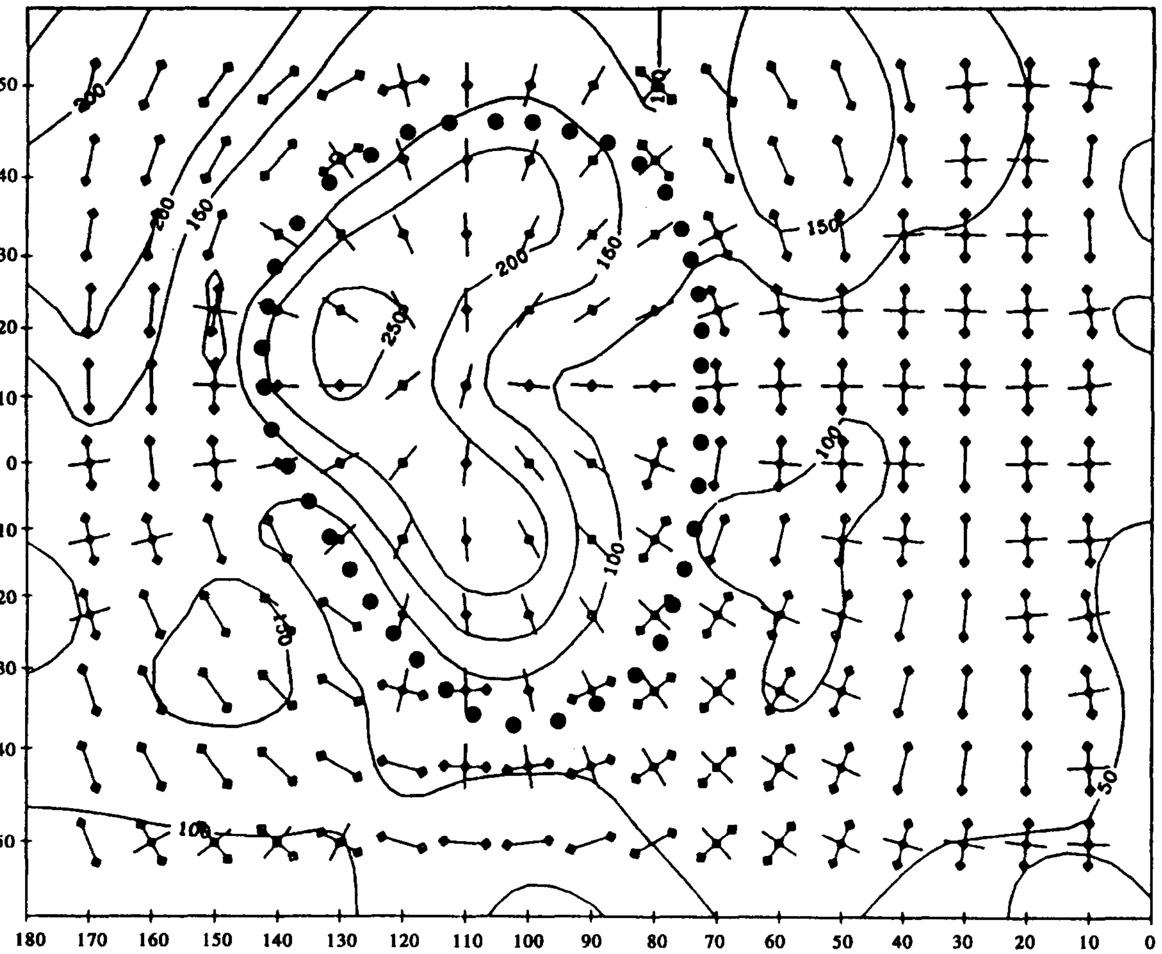

Fig. 4. (*Continued*)

at 50 km depth would be 550 MPa. The thickness of the crust flexurally supported is mostly between 50 and 70 km in the Tharsis region, and increases from the periphery to Syria Planum. The lithostatic pressure generated by a 70 km thick crust would be close to 770 MPa at the crust/mantle boundary. These values are well beyond creep (e.g. Lockner *et al.*, 1991), and even if the influence of heat flow and pore pressure have not been considered, and if the mean crustal depth is not really 50 km, the models using flexural loading should account for crustal creep at depth in order to be fully self-consistent. The detached crustal cap model accounts for this rheology in the Tharsis region, however, it might also be sensible to consider the effects of a weak crustal layer in most regions if its mean thickness is taken to be around 50 km. The "detached crustal cap" system would then work provided that the rheological characteristics of the lower crust are sufficiently different so isostatic-like stresses are still dominant within the cap, and that the flexural stresses still dominate outwards in the brittle crust. This might be possible considering that the heat flow in the Tharsis region should have been significantly higher than in the surroundings, making plastic creep in the Tharsis lower crust easier than around.

The Valles Marineris geometry may also be indirect evidence of lithospheric layering. According to Buck (1991), extension over a zone of limited width both in the crust and lithospheric mantle would produce a narrow rift made up of a single graben. The lithosphere below such rifts is expected to consist of four rheological layers: the upper crust (brittle), the lower crust (ductile), the part of the upper mantle just below the Moho (brittle) and the deepest lithospheric mantle (ductile). An example is the East African Rift System, which has been compared with Valles Marineris for a long time (Mutch *et al.*, 1976; Masson, 1977; Frey, 1979a; Schultz, 1991; Mège and Masson, 1996b). According to Buck (1991), wide rifts made up of several sets of parallel horsts and grabens, such as in the North Basin and Range, are produced by extension of the whole lithosphere over an area larger than the lithosphere thickness. Wide rifts have higher geothermal gradient and greater crustal thickness than narrow rifts. According to Buck (1991), the lithospheric structure is the same, the ductile lower crust being hotter and less viscous, and thicker than in the narrow rifting mode. The crustal thickness and the initial thermal state would be the two factors primarily determining the occurrence of the first or the second rift type, a less important factor being extension rate.

Following Buck (1991), the Valles Marineris case was investigated by Anderson and Grimm (1994), who found that the narrow rift mode should be the most likely because a wide rift would require the extension rate to be less than 1 mm yr^{-1}. We have calculated that, under various hypotheses of fault dips and sediment thicknesses in Valles Marineris, and including block tilting, the horizontal extension across the central part of Valles Marineris could be in the range 20–80 km (Mège and Masson, 1996b). In Mège and Masson (1996a) we note that, from comparison with analogue processes on Earth, a few million years only are required to open Valles Marineris. A value of 5 Ma would imply a mean 4–16 mm yr^{-1}, consistent with narrow terrestrial rifts.

Conversely, if Valles Marineris formed more slowly, e.g. 0.5 mm yr^{-1}, 40 Ma would be required to produce 20 km of horizontal extension, and 160 Ma would be required to produce 80 km of extension. Most of the extension occurred during upper Hesperian (Scott and Tanaka, 1986; Peulvast *et al.*, 1996). A value of 0.5 mm yr^{-1} is consistent with the 1.3 Ga long-lasting upper Hesperian according to Neukum and Wise's absolute stratigraphic scale (Neukum and Wise, 1976), and even with the 150 Ma proposed by Hartmann *et al.* (1981). Therefore there is no inconsistency, from this point of view, the central part of Valles Marineris to be compared with terrestrial wide rifts. Furthermore, the central part of Valles Marineris is 600 km wide, and so is unanimously larger than the current lithosphere thickness (Banerdt *et al.*, 1982; Willemann and Turcotte, 1982; Sleep and Phillips, 1985; Finnerty *et al.*, 1988), which is a condition required by Buck (1991) for the development of terrestrial wide rifts.

According to scaled analogue modelling of rifting (Allemand *et al.*, 1989; Allemand and Brun, 1991), producing several parallel grabens requires both the existence of two brittle and two ductile layers in the lithosphere and strong crust/mantle decoupling. In the experiments, these conditions favour the formation of four symmetric, or slightly asymmetric grabens (Allemand *et al.*, 1989). The exact number of grabens is probably of poor importance, nevertheless, it can be noted that Valles Marineris is also made up of four parallel and approximately symmetric chasmata (Melas, Candor, Ophir, and Hebes). Whether Ophir and Hebes chasmata are perfectly symmetric or not is unknown; however part of Coprates Chasma (Schultz, 1991), together with most of Ius Chasma (Mège, 1991) and Candor chasmata (Mège and Masson, 1994a,b), present a slight structural asymmetry (the southward-dipping boundary faults having vertical throws a few per cent higher than the northward-dipping boundary faults). Thus, the Valles Marineris structures fit the results obtained by Allemand *et al.* (1989) when a four-layered lithosphere is stretched.

Geophysical models of Tharsis consider an upper asthenospheric layer without rigidity (Banerdt *et al.*, 1982), and thus account for the lower ductile layer. Banerdt *et al.* (1992) proposed a strength envelope for the Martian lithosphere which accounts for a ductile crustal layer, however, Allemand *et al.*'s results (Allemand *et al.*, 1989) suggest that a stronger crust/mantle rheological contrast in tension should be probably considered in the Valles Marineris region when the grabens formed.

Although no quantification of the detached crustal cap model has been published yet, it does probably not solve some inconsistencies between stress and structures pointed out above (Section 3.1.2, points 2–4). Moreover, Tanaka *et al.* (1991) suggested that wrinkle ridge formation was produced by planetary cooling, their roughly concentric geometry about Tharsis in the Tharsis hemisphere being related to Tharsis-related stresses. However, planetary cooling should produce planetwide compressional stresses, which is in contradiction with the location of wrinkle ridges in lava plains only. Furthermore, Watters (1993) demonstrated that explaining the wrinkle ridge geometry needs taking both a Tharsis-related stress field and another deviatoric stress field into account. Existence of the latter can also be shown by stratigraphic arguments. Indeed, Tanaka *et al.* (1991) pointed out the agreement between the compressive state of stress predicted in the outer parts of the isostatically compensated regions, and the occurrence of wrinkle ridges there; however, the inner limit of the wrinkle ridges Tharsis is not known because it was buried by later upper Hesperian and Amazonian lava flows (Scott and Tanaka, 1986; Watters and Maxwell, 1986), and their formation there is inconsistent with the isostatic state of stress, which is extensional (Watters, 1993). Flexural loading stresses would provide a suitable state of stress, but if all the wrinkle ridges formed contemporaneously in a given lava field (Watters and Maxwell, 1983), these stresses should be contemporaneous to isostatic stresses outwards, in disagreement with all the published Tharsis models.

Several reasons may explain this persistent discrepancy between theoretical stress and the observed structures. Frey (1979b) and Plescia and Saunders (1982) clearly showed that two or three fracturing centres are requested at least to explain the formation of the tectonic structures, but the detached crustal cap model assumes that the tectonic activity during late Noachian to Amazonian can be approximated to one event, or to several events always related to the same fracturing centre, in order for the axisymmetric condition assumed by elastic models to be filled. Other reasons are given below.

3.3. *Reasons for the difficulties encountered by elastic models of Tharsis loading*

3.3.1. *Discrepancy between stress and strain.*

3.3.1.1. *Stress magnitudes.* The models shown on Fig. 3 predict horizontal principal stress differences ranging between 20 and 60 MPa in the isostatic case, and between 50 and 250 MPa in the case of stress induced by a flexural load. Up to 30 MPa, and between 60 and 240 MPa are respectively predicted by Banerdt *et al.* (1982). Sleep and Phillips (1985) predict up to 20 MPa and between 50 and 75 MPa, respectively. However, tension failure occurs for a few MPa in traction, e.g. ~5 MPa for the volcanic substratum of Iceland (e.g. Gudmundsson and Bäckström, 1991). Even lower values have been found in basalts (Schultz, 1995b). Therefore, fracturing should have occurred from the earliest stage of loading. The stress magnitudes predicted by elastic models do as if Tharsis was instantaneously built, and implicitly considers the lithosphere had an infinite strength to resist them. Schultz and Zuber (1994) emphasized that the stress in excess cannot be redistributed to past events in order to study the initial failure conditions, because the evolution of stress states is not linear in fractured materials. Comparison with results of analogue modelling is a way to attempt to study how load growth affects fracturing history (Williams and Zuber, 1995). Numerically, relationships between stress computed in elastic models can be investigated using a failure criterion. A plastic or viscous rheology should be used as soon as the failure criterion is reached. McGovern and Solomon (1993) used the Coulomb failure criterion (Byerlee, 1978) for studying the style of fracturing around crustal loads, whereas Schultz (1995b) used the Hoek–Brown failure criterion (Hoek and Brown, 1980; Brown and Hoek, 1988), which is particularly adapted to jointed rocks such as basalt, a material which is likely to mantle wide areas in the Tharsis province.

3.3.1.2. *Stress states.* Many problems could be solved if concentric strike–slip states were not widely predicted by elastic models. Strike–slip movements are observed in very few locations. The clearest examples were given by Schultz (1989, 1990), and some other possible examples were found by Forsythe and Zimbelman (1988, 1989), and Mège (1994). All are anyway exceptional observations.

Why strike–slip is so rare has been explained by the influence of earlier fracturing. McGovern and Solomon (1993) studied the style of fracturing around loads, taking the Tharsis volcanoes as examples. Isostasy predicts that radial normal faulting occurs closer to the load centre than the theoretical strike–slip (Fig. 4b). But normal faulting is predicted prior to strike–slip, so that propagation of radial normal faulting from the inner extensional domain may require less energy than the formation of new strike–slip faults in the domain of the predicted strike–slip state. The paucity of observed strike–slip movements has been explained by fault initiation at a depth where lithostatic pressure favours the formation of normal faults in areas where strike–slip movements would be predicted at the surface (e.g. Golombek, 1985; Sébrier *et al.*, 1985) (Fig. 5), and then normal fault propagation toward the surface.

Schultz (1992a), and Schultz and Zuber (1992) showed that prediction of strike–slip faulting around loads mainly

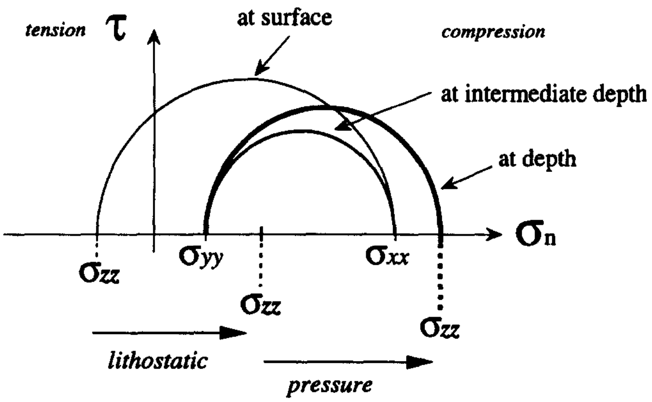

Fig. 5. Effect of increasing lithostatic pressure at depth on stress ellipsoid

results from the absence of failure criterion in computations. They showed that considering a failure criterion prevents prediction of strike–slip faulting at the surface. According to their results, zones of predicted strike–slip faulting occur at depth, contrary to the results obtained by Golombek (1985), and there is no obvious reason why the possible deep strike–slip faults should propagate up to the surface through zones of thrust faulting or jointing. Furthermore, if the mechanism of faulting is similar to that in Iceland, as suggested in the companion paper (Mège and Masson, 1996a), the Tharsis radial grabens would have initiated at the surface (Gudmundsson, 1992), and thus even if the state of stress at depth favours strike–slip faults, it would not affect the style of deformation observed.

Another possible reason for the difference between predicted stress states and deformation observed is that fluid pressure has never been accounted for in stress models. Subsurface water may have been widespread around Tharsis when most of the tectonic structures formed (Costard, 1993) and might have modified the stress ellipsoid predicted at depth in theoretical models in a number of places.

3.3.2. *Effect of active magma chambers before Tharsis loading.* Since most of the tectonics of Tharsis occurred during Precambrian time on Earth, it may be unnecessary that topography and gravity measured nowadays fit structural observations. In other words, the period of main magmatic and tectonic activity is old enough for the magmas to be now frozen and to behave as crustal loads, but this does not give information on the nature and trajectory of the stress that produced the tectonic features. The elastic models do not consider the previous liquid state of the chambers. In major tectonic provinces on Earth, magmatic overpressure events lead to emplacement of radial dykes around the magma chambers, like the well-known Spanish Peaks dyke swarm, which is theoretically able to produce tensile stress at the dyke tip (e.g. p. 24 of Anderson (1951)), and, in the field, suitable to produce extensional deformations above the dykes (e.g. Delaney *et al.*, 1986). The role of dyke emplacement in the Tharsis tectonics (Carr, 1974; Schultz, 1988; Tanaka and Golombek, 1989; Tanaka *et al.*, 1991; Mège, 1994; Davis *et al.*, 1995) and the role of magma overpressure in magma chambers (Carr, 1974; Mège, 1994) are assessed and discussed in Mège and Masson (1996a). Although dyke emplacement is shown to have been a major feature of Tharsis magmatism and tectonics, the graben depths (up to a few hundred metres) are not predicted by theoretical models of dyke emplacement, and further, such depths have never been observed to have formed in response to dyke magma pressure on Earth, suggesting that another source of stress is required.

3.3.3. *Regional extensional stress source.* A regional extensional source of stress, Tharsis-independent, has been shown in the companion paper (Mège and Masson, 1996a) to exist when the wrinkle ridge, the radial grabens, and Valles Marineris formed. The reader is referred to this paper for details.

3.3.4. *Buoyant subsidence of lava plains.* Many wrinkle ridges on the Moon can be explained by lava flooding, cooling, and buoyant subsidence in a thin brittle crust.

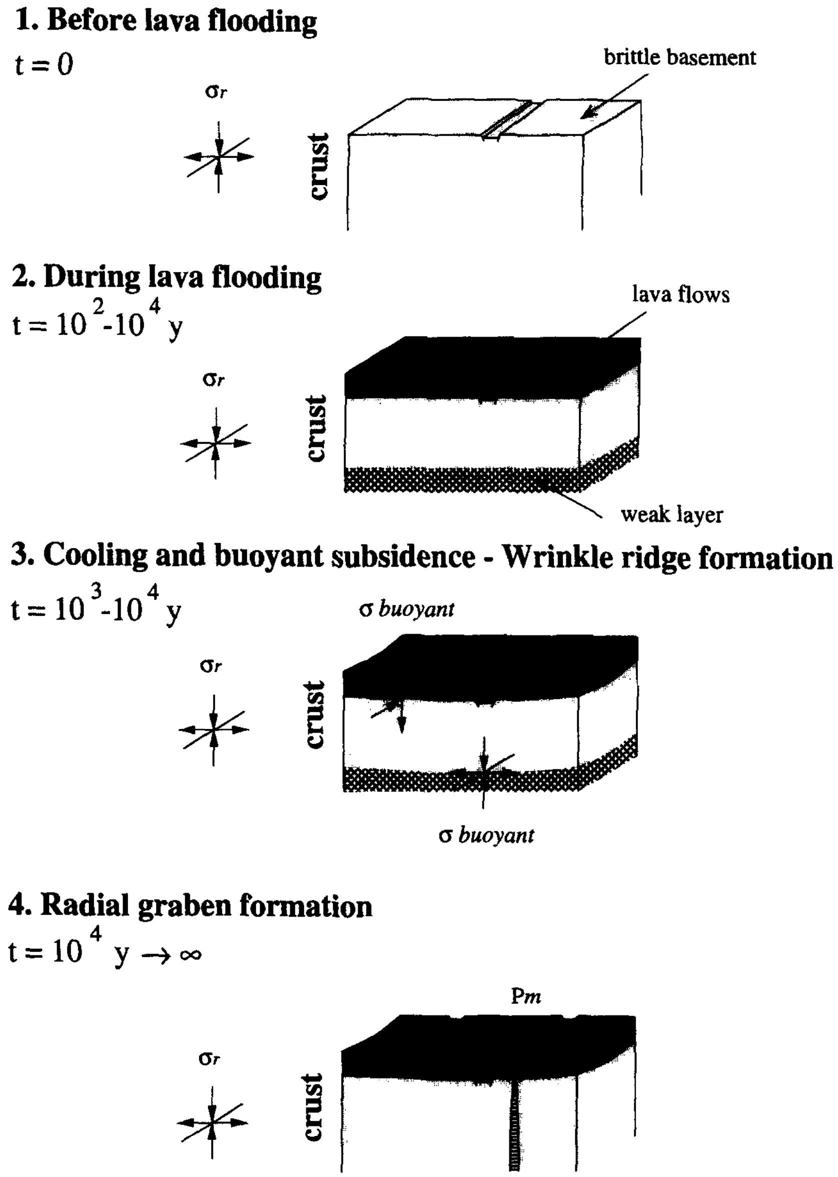

Fig. 6. Mechanism of contemporaneous wrinkle ridges and radial graben formation, according to the chronological relationships established by Watters and Maxwell (1983). The main sources of stress are indicated for each stage. σ_r, remote extensional stress; $\sigma_{\mathrm{buoyant}}$, stress generated by buoyant subsidence; P_m, magma pressure. Diverging arrows: σ_3 (tensional); converging arrows: σ_1; plain line: σ_2. Details in the text

Such a mechanism may also contribute to explain wrinkle ridge formation on Mars. It has been advocated for the eastern hemisphere (Watters, 1993). If this mechanism is also valid for the Tharsis hemisphere, a reason for thinning the Tharsis brittle crust must exist. It could result from thermal crustal weakening above a mantle plume (Hill *et al.*, 1992) centred on Tharsis. However, buoyant subsidence cannot explain wrinkle ridge distribution and geometry. It is horizontally hydrostatic, and therefore merely modifies the horizontal stress magnitudes, which become more compressional above a neutral plane in the crust in most of the subsiding area.

A mechanism for wrinkle ridge formation involving lava plain subsidence and other sources of stress previously mentioned is the following (Fig. 6, see also Fig. 22 in the companion paper (Mège and Masson, 1996a)). Lava emplacement (2) occurs in a remote extensional and deviatoric stress field (σ_r) partly Tharsis-dependent, and partly Tharsis-independent, which may produce extensional deformations (1). Gradual lava cooling produces buoyant subsidence of the upper crust. The resulting stress is akin to that produced by downward bending of a thin shell (e.g. p. 190 of Price and Cosgrove (1990)). It is extensional at the base of the upper crust, but compressional closer to the surface, where it is provisionally suitable to wrinkle ridge formation (3). During subsidence of the lava load, the driving stress is dominated by buoyant stress. In the absence of deviatoric stress, the wrinkle ridges would be randomly oriented. However, because of the existence of a remote deviatoric stress, wrinkle ridges are non-randomly oriented. After the lava load has reached the isostatic equilibrium, the state of stress in the brittle crust becomes progressively extensional again, and radial grabens can form, maybe in connection with dyke

Table 3. Review of stress models of Tharsis: petrologic approaches (1985–1990). Numbers refer to chronological order of events according to authors. See other values of lithospheric thicknesses beneath volcanoes compiled by Banerdt *et al.* (1992)

Authors	Origin of stress and support mechanism	Crust and elastic lithosphere thickness
Comer *et al.* (1985) Solomon and Head (1990)	flexure for volcanoes	lithosphere Alba Patera : 30 km Olympus Mons : > 150 km Ascraeus Mons : 20 km Pavonis Mons : 30 km Arsia Mons : 20 km
Finnerty *et al.* (1988)	magmatism partial melting : 26% crustal intrusions : 32% lithosphere/asthenosphere boundary deflection : 1 km + uplift on an undefined scale	crust 150 km DENSITIES crust : 3.03 ; mantle : 3.53 extrusives : 3.25 ; residuum : 3.41 crust + intrusives : 3.13
Phillips *et al.* (1990)	1. uplift 2. magmatism (plutonism dominant, min. = 95%) isostasy dominant, flexure secondary	

emplacement at shallow depth (4). Dykes may have assisted graben formation because of the extensional stress generated in the crust by magma overpressure in the dyke, or because of the focus of remote stress on dyke walls due to strength and density contrasts between dykes and the host rock (see Mège and Masson, 1996a).

4. Other approaches

4.1. *Petrologic approaches* (*Table* 3)

Finnerty *et al.* (1988) investigated the magmatic characteristics of the Tharsis load under the assumption that the current topography is entirely due to magmatic construction. The Tharsis region is considered as a laterally closed volume, in which three layers are distinguished, from bottom to top: the magma source region, the crust, and a magmatic pile (accounting for both volcanics and intrusions). Sleep and Phillips (1979) showed that isostasy can be completed only if the bottom of the lithosphere is compensated following Pratt's model, due to the lowering effect of partial melting and melt extraction on mantle density. Finnerty *et al.* first established a density model for Martian mantle and magmas derived from the composition of SNC meteorites. This approach has however to be taken with care because SNC meteorites have a controversial origin. The escape velocity required if they came from Mars must be very high. For this reason, Vickery and Melosh (1983) proposed an alternative hypothesis and suggested that they could originate from asteroids. This interpretation appears however to be marginal (McSween, 1994). The vertical density model was used to see how much the volume of the lithospheric column should increase (i.e. how much partial melting) to obtain the high Tharsis topography. The height to be obtained was estimated to correspond to 10 km of topographic uplift. It was completed in considering a constant Airy isostatic equilibrium during mass transfers between the three layers, then the effects on volume changes due to partial melting, and finally the effect on topographic rise. Of course, an infinite number of possibilities exists owing to the poor constraints on parameters (e.g. density of layers, crust thickness, elastic mantle thickness).

The thick shell theory used by Banerdt *et al.* (1982) was applied using a priori plausible parameters in order to calculate the vertical deflection of the bottom of the lithosphere and the residuum density. The mass of the lithospheric column was deduced, and the mass lost in the lower layer compared with the mass of intrusives and extrusives was estimated. The parameters were modified until a good fit was obtained. Many sets of parameters would have probably fit the percentages of partial melting considered to be reasonable; the set found by Finnerty *et al.* (1988) is shown in Table 3 and explains the Tharsis rise with 26% of partial melting. Of the fractionated material 32% would lie as intrusions in the crust. The melted layer could be 170 km thick. The value of 26% melting is close to that leading to mid ocean ridge basalts on Earth.

Much more melting could have occurred however: for instance, 50% melting would correspond to a melted layer 100 km thick and may not be unrealistic. Finnerty *et al.* underlined that the terrestrial komatiites resulted from much higher fractions of partial melting than 30%. In addition to higher melting due to the higher thermal gradient in the early history of planets than today, more melting can be expected in the Tharsis case than in the current Earth case because of the smaller size of Mars. McKenzie and O'Nions (1991) expect that melting is easier in silicated planets smaller than the Earth because rocks in the asthenosphere are less compressed than in larger bodies. For a thermal anomaly comparable to that of plumes on Earth, the magmas should be more voluminous, more mafic, and more fluid on Mars because of higher melt fractions produced by adiabatic mantle decompression. The possibility of such high amounts of partial melting is especially interesting to note when the possibilities for dyke swarms below the Tharsis radial grabens are investigated. Dyke swarms on Earth have a mafic composition (Halls, 1987), probably because only fluid magmas can travel over hundreds of kilometres. The length of the Tharsis grabens has been used as an argument against the interpretation of subsurface dykes (Carr, 1974; Cyr and Melosh, 1993). This argument does not stand from terrestrial evidence, since some of the dykes from the MacKenzie swarm are as long as many radial grabens of Tharsis (Fahrig, 1987; Ernst and Baragar, 1992; Ernst *et al.*, 1996), and all the less because higher partial melting should result in even more fluid magmas on Mars.

The model by Phillips *et al.* (1990) supports and develops the one by Finnerty *et al.* (1988). Finnerty *et al.* considered only vertical mass transfers, and Phillips *et al.* (1990) considered that Tharsis is an open system, which is subject to possible lateral mass loss. This point is considered in order to account for a possible early pyroclastic volcanism from a volatile-rich mantle, lasting until complete loss of volatiles (McGetchin and Smyth, 1978). Possible evidence of pyroclastic airfall materials have been suggested by Moore (1990), and Moore and Edgett (1993) on the opposite hemisphere of Mars. If correct, these materials might originate from major pyroclastic eruptions in the Tharsis or Elysium volcanic provinces and subsequent transport of small particles by wind. A circumstance favouring pyroclastic eruptions is the existence of a magma chamber at intermediate depth between the source region and the surface, in which part of the magma stays for a sufficiently long time to begin to fractionate. Mège and Masson (1996a) give evidence of such a common source region for the magmas of the three Tharsis Montes. It is frequent on Earth that the first eruption of a volcano chamber ejects the magma located in the upper part of the chamber, i.e. the part of the magma Na_2O-, K_2O-, and volatile-rich. The resulting materials are pyroclastic. That pyroclastic eruptions occurred in the Tharsis volcanoes is consequently highly probable from a mere analogy with the Earth, where pyroclastic eruptions are common in the large volcanic constructs as well as lava flows. The volume of materials cannot be estimated, and the amount of "lost magma" is therefore poorly constrained; fortunately, Phillips *et al.* (1990) showed this not to significantly affect their conclusions.

According to Phillips *et al.* (1990), the early Tharsis bulging would have been due to dynamic uplift. This assumption is based on the occurrence of several old normal faults in Claritas Fossae which are consistent with normal movements produced by the previously developed flexural uplift models. This argument is questionable: Claritas Fossae is one of the most complex regions of Tharsis, where faulting chronology from crosscutting relations are unclear (Masson, 1980), and where many directions of fractures were described (Tanaka and Davis (1988) distinguished a dozen fault sets). Thus, one would be quite unlucky not to find a fault trend consistent with one or other of the stress trajectory models.

Then, the Tharsis elevation would have been maintained by isostasy, like Finnerty *et al.* (1988) supposed it to be, the upper part of the lithosphere being supported by Airy compensation, and the lower part, by density variations. The parameters whose variations were particularly studied include the amount of molten material, of laterally lost material, the structural uplift, and the topographic uplift. All the calculations were carried out assuming that 10% of molten material was lost laterally by early pyroclastic eruptions (although attention has been turned to amounts reaching 50%). For amounts of partial melting ranging from 0 to almost 50%, 85% of intrusive material at least are necessary to induce structural Tharsis uplift. The remaining 15% must be extruded, partly as lava flows, and partly as pyroclastites. However, 90% of intrusive materials are required to produce a small topographic uplift. The latter is mostly due to the rising effect of the low density residuum: the intrusive/extrusive ratio is not high enough yet to cause topographic uplift. At 95% of intrusives, it is clear that uplift is produced both by the low density residuum and the higher intrusive/extrusive ratio. This result emphasizes the role of melting residuum in topographic uplift of magmatic regions, and is supported by results obtained by White and McKenzie (1989) in regions of extension on Earth.

It can be concluded from the results by Phillips *et al.* (1990) that whatever the amount of partial melting within the range 0–50%, and whatever the amount of magmatic material that escaped from the Tharsis system within the 0–50% range of extruded magma, at least 4/5 of intrusives are necessary to prevent crustal subsidence due to the heavy lava flows, and at least 9/10 are necessary to produce an uplift that can significantly affect the Tharsis rise.

4.2. *The lithosphere beneath the Tharsis volcanoes* (*Table* 3)

Comer *et al.* (1985) estimated the elastic lithosphere thickness beneath some large Martian volcanoes during their formation. Assuming that radial fractures around these volcanoes formed under stress related to the flexural load of the lithosphere, inversion of graben location led the authors to estimate the lithospheric thickness below Alba Patera, Olympus, Ascraeus, Pavonis and Arsia Montes, to be 30, > 150, 20, 30, and 20 km, respectively. For realistic rheological parameters, the possibility of formation of such grabens at a given distance from the magma chamber essentially increases as a function of two parameters: the magma chamber width and its proximity to the surface (Zuber and Mouginis-Mark, 1992).

The agreement between the Alba Patera surrounding grabens, including both circumferential and radial patterns, and theoretical modelling of flexural stress produced by the Alba Patera load within the framework of a remote extensional stress field is striking (Cyr and Melosh, 1993). But, surprisingly, topographic data (U.S.G.S., 1991a, 1992) indicate a rather poor correlation between the distribution of topography and what would be expected from volcano morphology and graben distribution. Graben formation corresponds to the latest stage of Alba Patera evolution (Mouginis-Mark *et al.*, 1988). If the available topographic data are reliable, despite the more than 1 km uncertainty in elevations, and if the grabens formed with the same volcano topography as today, they are not fully consistent with a flexural origin. Although flexural stress may have favoured graben formation, they could also be related to another mechanism. Raitala (1988) interpreted these gra-

Table 4. Review of stress models of Tharsis: comparison between Tharsis and Elysium related stress modelling and structural observations (1986). Numbers refer to chronological order of events according to authors

Authors	Origin of stress and support mechanism	Crust and elastic lithosphere thickness
	ELYSIUM REGION	
Hall *et al.* (1986)	1. flexural uplift at the Elysium scale • thermal → Elysium Fossae Hephaestus Fossae 2. magmatism, chronology undefined • Elysium flexure → concentric grabens • Tharsis isostasy → ridges East of Elysium • Tharsis flexure → Cerberus Rupes, ridges West of Elysium	lithosphere: 50 km
	THARSIS REGION	
	1. flexural uplift (limited extent) 2. not specified	

bens in a rift context. An important feature is that, whatever the origin for the curved grabens: flexural (Wise, 1976; Comer *et al.*, 1985; Cyr and Melosh, 1993; Watters and Janes, 1995), or due to rifting (Raitala, 1988), or maybe to both (Mège and Masson, 1996a), accounting for the absence of circumferential grabens on the northern and southern flanks of the volcano requires the existence of a remote extensional stress field whose trajectory of least compressional stress is difficult to relate to the stress produced by the load of the Tharsis dome southward as predicted by the theoretical models.

Thickness results exposed in Comer *et al.* (1985) represent the basic framework of the study by Solomon and Head (1990) dealing with thermal gradients and heat flow beneath the Tharsis volcanoes, Alba Patera, and Olympus Mons. When the radial grabens around Tharsis Montes and Alba Patera formed, the thermal gradient and heat flow would have been 10–14 K km^{-1} and 25–35 mW m^{-2}. They would have been 5–6 K km^{-1} for Olympus Mons. The current Mars global heat flow would be 15–25 mW m^{-2}. Roughly 3–5% of this value is assumed to represent conduction of excess heat under the major volcanic provinces during the Amazonian time. This value is three times higher than expected from heat transport in volcanic and shallow plutonic complexes, and would have resulted from plumes beneath the volcanic centres. Such a result leads to reconsider the role of magmatism as a stress source, for instance through mantle decompression, compared with "purely tectonic" stress sources, like those produced by magmatic loading.

4.3. *Correlation between the Tharsis and Elysium regions* (*Table* 4)

A stress study in the Elysium region, on the opposite side of Mars, by Hall *et al.* (1986) led these authors to suggest that the Tharsis formation could have influenced the Elysium development.

One of the first significant aspects of this result concerns the need for an early flexural uplift stage in the formation of Tharsis to account for the early evolution of Elysium. Hall *et al.* (1986) compared the structural features of the Elysium province with those expected from several elastic models, deriving from the following support modes: (1) loading by the Elysium and Hecates volcanoes, (2) regional loading of the Elysium province, (3) flexural uplift of the Elysium province, (4) Tharsis isostasy, and (5) Tharsis flexural loading. They showed that the concentric Elysium Mons fault system could be explained by a local flexural load of a 50 km elastic lithosphere. Surprisingly, loading of the whole Elysium region cannot predict any of the observed tectonic structures. According to Hall *et al.*, this disagreement could be explained by the following hypotheses: (1) volcanism may not have played an important role in the Elysium crustal evolution, (2) the flexural stresses at the surface were not strong enough to trigger graben formation, or (3) the whole load was isostatically compensated. Some ridges east of Elysium could be attributed to the isostatically compensated Tharsis load, and some others, together with Cerberus Rupes, to the Tharsis flexural load. Some remaining features were not predicted by any of these supports. Hall *et al.* suggested that they could originate from an early flexural uplift of the whole Elysium region. A number of stress sources, whose relative chronology is not clearly established, are thus required to explain the whole tectonic structures in the Elysium region. In their conclusion, Hall *et al.*, from analogy with Elysium, proposed that the first tectonic activity of Tharsis could be related to an early flexural uplift as well. As already noted, this hypothesis is not supported by the elastic models of Tharsis support, which cannot basically conciliate radial extensional structures and flexural uplift. In order to explain this inconsistency, Hall *et al.* argued that the extent of the Tharsis bulge could have been initially lesser than currently; the initial structural features, due to early uplift, would have been covered by later volcanic materials and obscured by later fracturing.

A second, and somewhat unexpected aspect of Hall *et al.*'s results, seems to be the unaptness of loading stresses to produce structural features in the Tharsis region. The Elysium load is extremely likely to be in equilibrium, since the magmatic activity ceased at the lower/middle Amazonian boundary (Greeley and Guest, 1987), e.g. 700 Ma ago at least (Neukum and Wise, 1976), maybe much more (2.3 Ga ago according to Hartmann *et al.* (1981)). Although tensile stresses higher than a very few MPa are not needed for surface failure (Section 3.3.1.1), the stress magnitudes resulting from the various loading models in the Elysium region are of the same order as to those predicted in the Tharsis province, i.e. up to 220 MPa. The lowest predicted magnitudes are of the order of 10–20 MPa. These stress levels should definitely produce surface deformations. In particular, loading of the Elysium and Hecates volcanoes and Elysium regional loading should produce widespread deformations which are not observed. Similar stress magnitudes are predicted around the current Tharsis load (Fig. 3) and radial grabens are observed. Thus, it would not be surprising that radial graben formation be due to another driving mechanism. Addition of the differential stress produced by Tharsis isostasy to the differential stress produced by the various Elysium loading models does not change this conclusion appreciably because of the negligible magnitude of isostatic stress in the stress magnitude balance. Results of addition of the differential stress produced by Tharsis flexural loading to the differential stress produced by the various Elysium loading models have not been published, however we expect the resulting differential stress to be still high and mainly influenced by Tharsis flexural loading, because in many areas of the Elysium region Tharsis flexural loading produces stress magnitudes that greatly exceed the Elysium-related loading stress magnitudes.

The influence of both Elysium-induced and Tharsis-induced loading stresses on tectonic patterns appears thus to have been negligible in the Elysium province. Consequently, it should have been weak in the larger Tharsis province as well. In addition, if the Tharsis-related stress were to produce significant deformations in the Elysium and Tharsis provinces, it should have been so in other, widespread areas of Mars as well, but no correlation with structural observations in other regions of Mars has been reported (see, e.g. Schultz, 1985).

5. The case for dynamic support

Processes related to dynamic lithosphere uplift, such as convection rolls or hot spots generating emplacement of large magma chambers in the crust, in producing stress and tectonic features has been neglected in many studies. Several recent works have however emphasized their role in the Tharsis volcano–tectonic history.

5.1. *Convection* (*Table* 5)

Although flexural uplift has been proposed to have played a role in the Tharsis tectonics by Wise *et al.* (1979b), Hall *et al.* (1986), and Phillips *et al.* (1990), the first convection models date back to a preliminary work by Kiefer and Hager (1989) based on finite element calculations. According to Kiefer and Hager, convection and subsequent uplift could have played a key role in the Tharsis support as well as loading processes. Some results were provided by numerical calculations of three-dimensional convection in Mars and Venus mantles by Schubert *et al.* (1990). The variation of several parameters were studied. The mantle lower boundary was expected to be isothermal shear stress free, whereas its upper boundary was considered rigid or shear stress free in order to assess the role of the lithosphere as boundary condition. Both heating from within and from the core, including transitional states, were considered. Small and bigger core hypotheses were studied to account for core formation.

Table 5. Review of stress models of Tharsis: recent dynamic support studies (1989–1990). Numbers refer to chronological order of events according to authors

Authors	Origin of stress and support mechanism
Kiefer and Hager (1989)	convection dominant past and currently isostasy and/or flexure secondary for the remaining current topography
Schubert *et al.* (1990)	1. convection 2. isostasy for the compensated topography + convection for the uncompensated topography

An interesting result is that assuming rigid upper mantle boundary conditions, downwelling currents are expected to form in the upper thermal boundary layer, and, according to Schubert *et al.* (1990), could be responsible for a surficial state of stress which could lead to formation of the Tharsis radial grabens. Unfortunately, no quantitative results of this effect have been given. According to these authors, a dynamic topographic support is currently sufficient to account for the non-compensated part of the topography, ruling out the requirement for flexural loading.

The convective approach may help constrain the chronological relationships between the formation of Tharsis and the dichotomy. The dichotomy could have formed by a convective system dominated by the spherical harmonics degree 1 because both highlands and lowlands represent about one third of the planetary surface. The topography and gravity field of the smaller Tharsis region are strongly correlated with the harmonics degree 2. Planetary cooling implies that the dominating harmonics degree of the convective system varies as a function of time, i.e. the number of convection cells gradually increases, suggesting that the dichotomy should be older than Tharsis.

Schubert *et al.*'s analysis (Schubert *et al.*, 1990) is consistent with a scenario of early planetary evolution which appears to successfully fit a number of previous observations, results, and proposals, especially as far as the dichotomy is concerned. Nevertheless, two series of difficulties need solving for more recent times. These points are developed below. A convective system of harmonics degree 1, which can be correlated with a structural dichotomy, requires that both the dichotomy and Tharsis are very old. The computations carried out in the small core case do not predict harmonics degree less than 2. This means that reaching degree 1 requires an especially strong heating from the core by a peculiar heating pulse, or accounting for an even earlier stage of core formation. A scenario of early Mars evolution consistent with Schubert *et al.*'s model has been proposed by Breuer *et al.* (1993), who considered that a plausible way to create the dichotomy consists in an early two-stage crustal differentiation, similar at the root to the model proposed by Lowman (1989) for terrestrial planets. The first stage includes the formation of the primordial, intensely cratered crust. It would have been followed by basaltic-type overplating, restricted to the northern Martian hemisphere. As shown by Schubert *et al.* (1990), this second differentiation would have been produced mainly by heating from the core since internal heating would have provided too many plumes, leading to a planet-wide volcanism. Thermal blanketing would explain preferential plume location beneath the northern hemisphere. This asymmetric blanketing could have originated from early giant impact in the lowlands, inducing crustal thinning, and then an asymmetric distribution of magma in an underlying "magmasphere" (Spohn and Schubert, 1991). This magmasphere asymmetry would have led to the focusing of internal heat preferably in the northern hemisphere. The role of impacting in this scenario reconciles the endogenic hypotheses (Wise *et al.*, 1979a,b) and exogenic hypotheses (Wilhelms and Squyres, 1984; Frey and Schultz, 1988; McGill, 1989; Schultz and Frey, 1990) of dichotomy formation, models somewhat reconciled by McGill and Squyres (1991). McGill and Dimitriou (1990) also proposed an endogenic hypothesis, but suggested the period of dichotomy formation to be upper Noachian/lower Hesperian, which seems to be less consistent with Schubert *et al.*'s proposal. The hypothesis suggested by Breuer *et al.* (1993) for the crustal dichotomy is thus attractive. It supports the model for the formation of Tharsis and the dichotomy by Schubert *et al.* (1990), and thus reinforces the hypothesis of an early Tharsis uplift.

A series of difficulties arises from that the reasoning of Schubert *et al.* (1990) is based on a gradual decrease in size and increase in number of cells with time, i.e. on linear planetary cooling, resulting in a cumbersome quantity of plumes late in the history of Mars. The long-lasting history of Tharsis requires a constant focus of heat below this region; this thermal anomaly should not decrease too much in magnitude because much of the magmatic materials formed in Hesperian and Amazonian. Following Schubert *et al.* (1990), many smaller volcanic areas should have randomly developed at the surface of Mars through the planet's history. Although small isolated volcanoes are observed in places, two suitable centres only exist: Tharsis and Elysium. In order to overcome this difficulty, the authors suggested that Tharsis and Elysium would have behaved as two weakness zones in a strong lithosphere, channelling most of the internal heat, and producing two magmato-tectonic centres only. However, there is no indication that the Elysium region is located above a particularly weakened lithosphere. Conversely, whatever the origin of the dichotomy boundary, the lithosphere beneath the uplands is certainly highly fractured and there appears to be no trace of suitable magmatic complexes there.

5.2. *Hot spots*

Contrary to hot plumes produced by convection in a mantle having normal temperature, hot spot plumes in the sense given by Morgan (1971) make it possible to focus heat in the same region for a long time without predictable implications for patterns of circulation in the mantle once the hot spot activity has ceased. Stratigraphic studies suggest that the Tharsis magmatic activity occurred during crises separated by periods of quiescence (Neukum and Hiller, 1981; Hiller *et al.*, 1982). The Tharsis magmatism would be explained by some recurring mechanism, which could be akin to the (still unclear) mechanisms for recurring hot spot activity on Earth (e.g. Heaman and Tarney, 1989). Breuer *et al.* (1996) have recently carried out numerical calculations of circulation in the Martian mantle accounting for mineral phase transitions. The endothermic spinel to perovskite phase change appears not to have decisive effects on mantle circulation, the exothermic olivine to β-spinel and β-spinel to γ-spinel transitions at depths of 1100 and 1400 km, respectively would accelerate the mantle flow and could generate a small number of large, long-lasting, and superheated plumes resulting in successive and distinct periods of magmatic activity.

The consequences of active recurring hot spot activity in the Tharsis region in terms of stress has been correlated with the observed tectonic and magmatic activity (Mège and Masson, 1996a). It has been found that hot spot activity appears to explain a number of tectonic features not predicted by other models. For instance, it can predict the Valles Marineris formation, location, and singleness.

6. Discussion

The discussion below, which attempts to summarize some of the main features of the Tharsis evolution as arising from the works mentioned above, emphasizes the role of dynamic processes early in the Tharsis structural evolution from new interpretations of some of the older tectonic structures observed in the Tharsis province.

6.1. *Succession of tectonic events*

From the works reviewed above a self-consistent scenario can be sketched. The results presented in Mège and Masson (1996a) complicate this story, but does not affect the succession of support modes. Tharsis would have primarily formed by dynamic lithosphere rising due to vigorous upwelling in the mantle producing high amounts of melt. As a consequence, loading processes would have taken over from dynamic processes, and Tharsis would have been supported by a combination of full or partial Airy and Pratt isostasy, primarily by injection of intrusive magmas, and secondarily by extrusion of lava flows and pyroclastic materials. Intrusions would have thickened and strengthened the lithosphere, and increased its flexural rigidity. Part of the load would then be flexurally supported. The crust thickness would have sufficiently increased with time the lower part to become ductile in the central part of Tharsis (Banerdt and Golombek, 1990). This scenario agrees with the current topographic and gravity data. According to Tanaka *et al.* (1991), numerous structures of Tharsis (radial grabens, wrinkle ridges, Valles Marineris) would have been produced during this stage. The above scenario addresses two major structural concerns.

- Did the early dynamic uplift stage produce tectonic structures? A number of structures were produced during Noachian and early Hesperian (Scott and Tanaka, 1986) that cannot be accounted for by loading models. In Section 6.2, we show that early dynamic uplift may have been responsible for stress and strain patterns more complex than expected in elastic stress models.
- Some structural observations which should stratigraphically fit the stress generated by the loading supports were shown to cause persistent problems. How to explain their formation? Important work has been done and has still to be deepened in order to better understand what could have been the influence of a possible ductile lower crust in terms of stress in the brittle upper crust, and how to relate this behaviour with surficial structures. Another approach is proposed in the companion paper (Mège and Masson, 1996a), which suggests that the loading stage during the evolution of the Tharsis support was reached (and the loading stresses would have been generated) too recently to have played a decisive role in the formation of structural patterns.

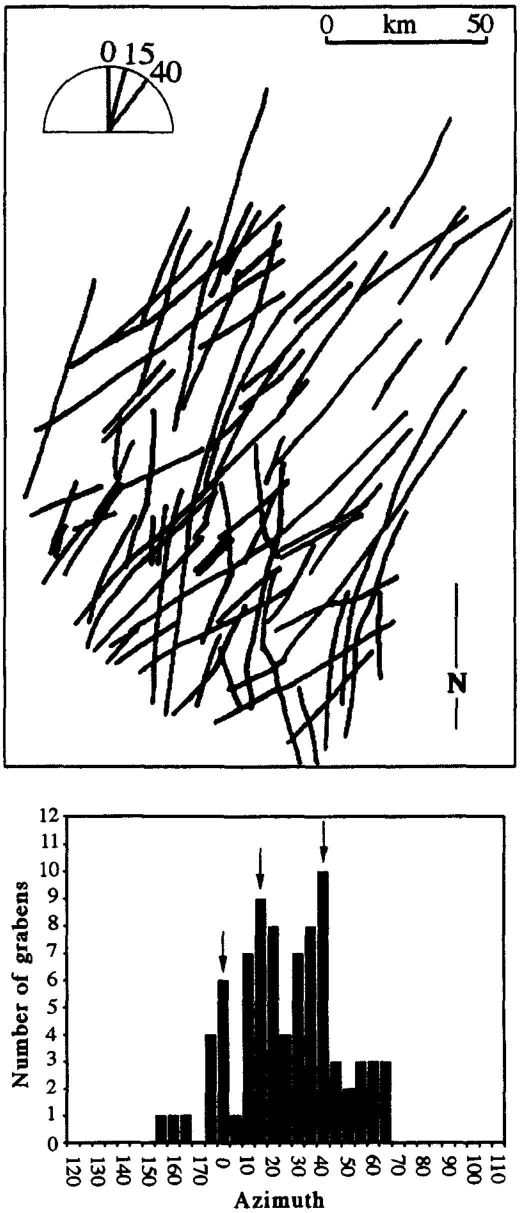

Fig. 7. Tectonic structures in Uranius Fossae and histogram of graben azimuth (class intervals: 5°; the class centres are numbered according to the azimuth of the lower boundary of each class). The arrows point to the classes representative of the three main graben trends (also displayed in the half-pie diagram). Significantly curved grabens are represented twice, each azimuth accounting for one graben tip

6.2. *Early flexural uplift: structural arguments*

6.2.1. *Structures due to bending forces and inferences for early flexural uplift.* Avoiding interpretation traps listed by Wise (1982), it is possible to describe the Noachian and some lower Hesperian deformations outcropping between more recent lava flows in terms of crosscutting sets of roughly parallel faults. Such deformed areas include Ulysses Fossae and Tempe Fossae (Scott and Dohm, 1990) as well as Uranius (Fig. 7), Acheron, and Ceraunius Fossae, and a small area south of Valles Marineris (80°W, 15°S) are other examples of Noachian or lower Hesperian grabens which can be analysed speaking of such coherent fault sets.

The reason for which the structures displayed in Uranius Fossae and the other Noachian terrains intensively deformed in the Tharsis area have been preserved from later lava flows seems to be related to their location on small topographic rises that have been wholly or partly conserved. The topographic maps of Tharsis (U.S.G.S., 1991a,b, 1992), despite uncertainties, indicate that Acheron Fossae and the Noachian part of Ceraunius Fossae lie on topographic heights, as well as Uranius Fossae to a small extent. Part of Ulysses Fossae appears to have been uplifted more than 1 km. This topographic information suggests that early Tharsis uplift should have partly occurred before lava flow emplacement, confirming the hypothesis by Hall *et al.* (1986) reported above.

Flexural uplift dominantly results from bending forces. Bending forces are here considered as the prominent source of stress when resulting from at least three possible processes. First, bending stresses may be due to "static" flexural uplift of the lithosphere, induced by buoyant forces due to Pratt overcompensation. This effect was studied by Banerdt *et al.* (1982, 1992) and corresponds to Fig. 3a. Secondly, bending stresses are also expected if the topography is supported by upwelling; according to Banerdt *et al.* (1982), the expected stress trajectories are the same as those predicted by buoyant uplift. Thirdly, bending stresses may be locally produced in the crust during pluton or batholith emplacement, through diapirism or other mechanisms such as emplacement and coalescence of multiple dyke-fed sills at a stress barrier in the crust (Gudmundsson, 1990). These magma bodies should not be too voluminous with respect to the lithosphere thickness, otherwise membrane stress dominates over bending stress (e.g. Banerdt *et al.*, 1992). The early Martian lithosphere was probably thinner than currently (e.g. Schubert *et al.*, 1992), so that limited intrusive volumes may have been enough to produce primarily bending stress in the crust in the early Martian history.

It could be argued against an early flexural uplift that the Noachian and early Hesperian tectonic patterns observed beneath the Tharsis lava flows are clearly inconsistent with the stress trajectories given by elastic modelling of flexural uplift. We show hereafter that this argument is not appropriate because the influence of the growing bending is not taken into account (Nur, 1982). Bending does not necessarily lead to structural deformations following two steady horizontal stress trajectories relating to a centre (one radial and one hoop). Numerous parameters govern emplacement of magma in the crust, and the size and geometry of plutons. Each stage of magma emplacement modifies the stress patterns in the host rock, depending on, e.g. variations of magma pressure and host rock properties. Permanent uplift above magma bodies may be favoured by the buoyant effect of a low density residuum beneath, following the mechanism exposed by Finnerty *et al.* (1988). The circulation in the mantle may have produced unsteady stress trajectories in the brittle crust through thermal crustal uplift above an upwelling region. Variations of heat flow in the mantle would have been responsible for variations in uplift amplitudes, and thus unsteady stress trajectories. If the models predicting gradual thickening of the Martian crust with time (e.g. Schubert *et al.*, 1992) are correct, the absence of ductile crust in early times would have enhanced the influence of asthenospheric movements on stress patterns in the crust. According to Blichert-Toft and Albarède (1994), the convection in the Earth's mantle may have been ten times faster in late Archean than today, and we expect a parallel evolution trend for Mars. Both a thin crust and more vigorous movements in the mantle would have enhanced the coupling between internal movements and near-surface stress.

6.2.2. *Mechanism of formation.* Nur (1982) noted that vertical tensile fractures forming two, three or four cross-cutting sets are observed east of the Gulf of Aqaba. These patterns as viewed on satellite images, show striking similarities to the old fractured terrains beneath the recent Tharsis lava flows such as Uranius Fossae. The spacing between the tensile fractures in the Gulf of Aqaba is typically several metres to several kilometres. In Uranius Fossae, it is typically ten kilometres, in good agreement with the Earth. Similar to our hypotheses for the old fractured terrains, Nur (1982) found that bending stresses are the most suitable for forming these fractures. The multiple set trends develop as follows. The first fractures form from the first bending stresses sufficiently large in magnitude to cause failure at the surface. They propagate downward. If the stress trajectories change, new fractures form perpendicular to the tensile direction only when the tensile stress is higher than the tensile strength S_0 of the unfractured rock. Conversely, the first fractures continue to open and propagate downward if their tensile strength S_1 is less than S_0. The critical angle ϕ_c beyond which a new fracture forms is given by

$$\phi_c = \cos^{-1}(S_1/S_0)^{1/2}. \quad (1)$$

For instance, when S_1 is close to S_0 and ϕ_c is null, so that no coherent sets can be distinguished. Three sets of faults may form at most if ϕ_c is systematically equal to 60° ($3 \times 60° = 180°$).

The fractured terrains display grabens instead of tensile fractures. Two mechanisms of graben formation are suggested below, depending on the host rock type close to the surface, either regolith or lava flows. The intense meteoritic bombardment intensely fractured part of the brittle crust during Noachian. Channels observed in the highlands show that water is likely to have existed at, or close to the surface (recent discussions in Squyres and Kasting (1994), and Carr (1995)), maybe due to a warmer climate (Pollack, 1979, 1991; Pollack *et al.*, 1987), or due to igneous processes (Squyres *et al.*, 1987; Wilhelms and Baldwin, 1989; Gulick *et al.*, 1991). Water circulation, and possibly magmatic fluids should have enhanced grain and fracture cementation in the intensely fractured upper part of the crust. Conversely, high impacting rate would have prevented efficient grain and fracture cementation. Both phenomena would have influenced the value of ϕ_c in equation (1). Tensile fractures would have initiated at the upper boundary of the basement. They would have propagated downward, following the mechanism advocated by Nur (1982). Closer to the surface, bending would have caused small-scale movements because of the poor regolith strength. Each fracture generates a strain field whose effect is similar to a "shadow zone", inside which the possibility for a new fracture to develop is inhibited. As a result, the number of fractures which propagate downward is inversely proportional to depth. Therefore, in the Tharsis case, a few wide fractures and many narrow fractures should have finally formed at the impacted/non-impacted transition in the crust in response to increasing bending. Conjugate shears (normal faults) would have developed within the regolith above the wider fractures, according to Anderson's model (p. 15 of Anderson (1951)) (Fig. 8). Such a mechanism for graben formation requires that the near-surface rocks possess a non-zero cohesion, and that no substantial anisotropy exists (Schultz, 1992b). Regolith should roughly satisfy both conditions.

However, it is not sure whether the grabens formed in the regolith; instead, they may have formed in early lava flows, suggesting another mechanism of graben development. Gudmundsson (1992) showed that normal faults in lava flows in Iceland may develop from tensile fractures and inclined joints. Normal faults nucleate once tension fractures originating at the surface have propagated downward to a maximum depth which is a function of the tensile strength of the host rock. The lava pile, which is horizontal at the surface, is vertically jointed. The lava pile becomes increasingly tilted at depth, so that the joints are oblique to the principal stress at depth, favouring shear fracturing. Joints start to propagate when the tensile stress around them reaches the tensile strength of the host rock; then coalescence of aligned joints results in a normal fault. Such a mechanism of faulting may correspond to fault sets like in Uranius Fossae if the graben formed in basaltic-type rocks. Basaltic joint tilting would result from crustal bending. It should be noted that this mechanism does not predict graben formation, as observed in Uranius Fossae. Conjugate secondary normal faulting could result from compressional stress concentration at the bottom of

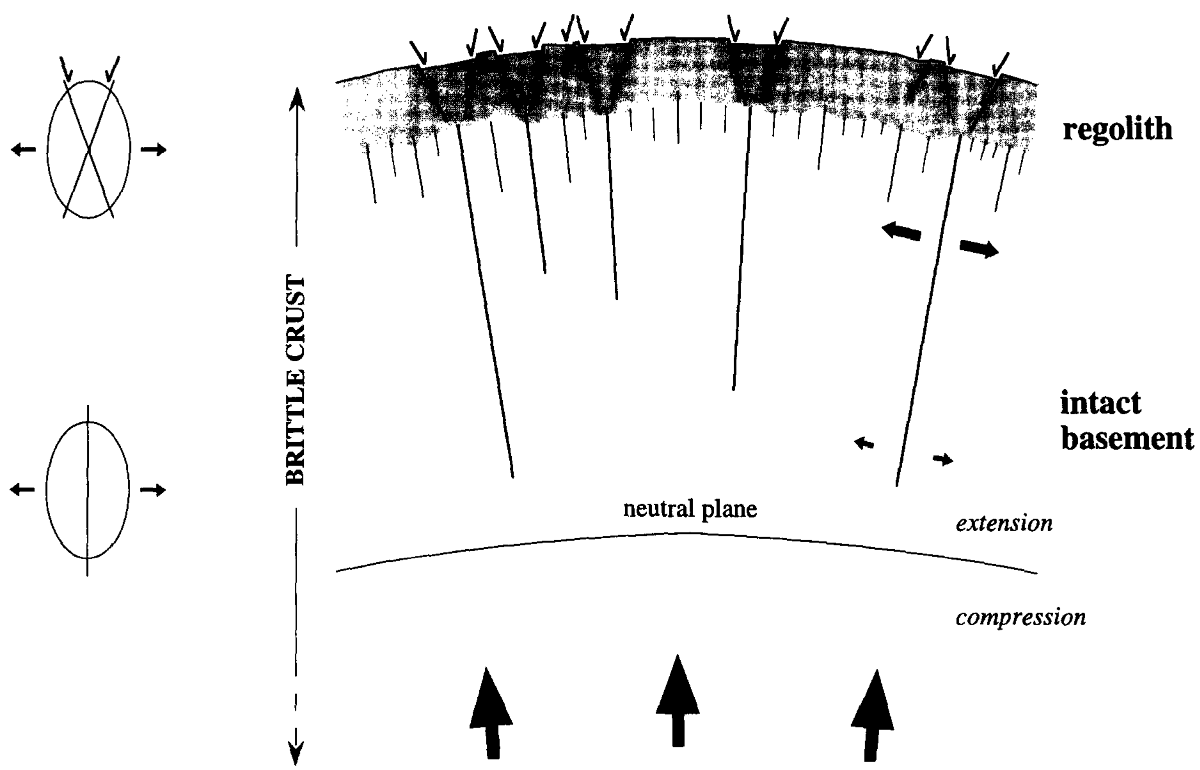

Fig. 8. Cross-section of the brittle crust in fractured areas such as Uranius Fossae (Fig. 7) illustrating one of the possible mechanisms of formation of one parallel graben set if crustal layering consists of regolith and intact basement. Crustal bending produces extension in the crust above a neutral plane. Tensile fracturing is expected in the intact basement whether shear fracturing is expected in the regolith. Drawing modified from Nur (1982)

the primary normal faults (p. 187 of Price and Cosgrove (1990)) and concentration of extensional stress at the surface at some constant distance from the latter, following a mechanism studied by Melosh and Williams (1989). Note that, in the absence of a lava pile, a similar mechanism of normal fault development and graben formation in regolith may be suggested, the main difference being that formation of tensile fractures leading to normal fault development would not use lava joints but suitably oriented pore discontinuities.

As on Earth, several fault sets may successively develop according to the variations of local stress trajectories, depending on the parameters of equation (1). In the Uranius Fossae case, at least three main graben sets are observed (Fig. 7) and should reflect these variations.

6.2.3. *Other clues to early bending.* Schultz and Tanaka (1994) found that a middle Noachian/lower Hesperian compressional South Tharsis Ridge Belt exists close to the bound between the cratered highland on the one hand, and the volcanic materials from the Tharsis lava flows and the lava plateaux south of Valles Marineris on the other hand. These structures appear to correlate well with a radially compressional stress state at or near the surface that would be expected to form around Tharsis submitted to flexural uplift (Price and Cosgrove (1990), see Mège and Masson (1996a) for details).

It seems thus that a possible early dynamic uplift in the Tharsis region can be correlated with some structural observations. The geometry of the uplifted region and its size remain poorly understood. The height of the uplift is unknown, and cannot be retrieved from the current topography owing to the crustal intrusions subsequently emplaced.

6.2.4. *Source of bending.* The source of early Tharsis bending which may have produced both the South Tharsis Ridge Belt and the intricate graben sets may be the same. Its nature is speculative; one of the possible scenarios is as follows. It can be suggested that a broad-scale uplift of the Tharsis region occurred at the onset of Tharsis formation, maybe similar to the thermal uplift preceding hot spot activity on Earth (Griffiths and Campbell, 1991). Whereas the uplift should have slowly decreased after the onset of volcanic activity (Hill *et al.*, 1992), variations of magma body distribution in the crust would have resulted in local topographic variations resulting from bending above magma bodies.

6.3. *Dynamic uplift and loading processes during Hesperian and Amazonian*

Although local structure sets exist, most of the tectonic activity of Tharsis during Hesperian and Amazonian has consisted in the formation of the radial grabens, the wrinkle ridges, and Valles Marineris. Two kinds of models have been proposed for explaining their formation. The first series of models attempt to correlate them with stresses generated by the Tharsis load. These models face several difficulties discussed in this paper, some of them solved by the detached crustal cap model of Banerdt and Golombek (1990), and Tanaka *et al.* (1991). In the companion paper (Mège and Masson, 1996a) we have suggested a plume tectonics model, which appears to us to be in better agreement with the structural record of Tharsis. For instance, it is in better accordance with the wrinkle ridge distribution (only in volcanic terrains (Watters, 1993)) and orientation (only roughly concentric to Tharsis), and with the length of the radial grabens. It predicts both the existence and the location of Valles Marineris, together with the uplifted surrounding plateaux. The Tharsis load should have now relaxed, generating loading stress whose cumulated magnitude may have reached 250 MPa (Banerdt *et al.*, 1982, 1992; Sleep and Phillips, 1985), but they are not required to explain the observed structures. Loading is a slow, approximately continuous process on geological scale, whose evolution follows the evolution of magma cooling. It could be suggested that loading stress has been gradually and slowly released in the crust during the load growth, by small-scale grain movements, e.g. in the regolith, and by working on discontinuities and faults previously formed, in the brittle domains. It might be correlated with the formation of recent tectonic features throughout Valles Marineris (Peulvast *et al.*, 1996). In domains of high confining pressure, creep should transmit stress in the brittle domains, enhancing the efficiency of the two processes (Kusznir and Bott, 1977; Kusznir, 1991).

The most realistic origin for the tectonic structures in the Tharsis province from the works analysed above, and from our own work, might be summarized as follows. The structures that can be explained by dynamic processes, according to the plume tectonics model presented in the companion paper (Mège and Masson, 1996a), include the Noachian/lower Hesperian South Tharsis Ridge Belt, the mostly Hesperian and Amazonian radiating patterns, and some Hesperian elliptic patterns on Syria Planum and Noctis Labyrinthus. According to the stratigraphy, the intricate grabens below the Hesperian and Amazonian Tharsis lava flows should also be related to an early stage of Tharsis evolution in this model, although their significance remains enigmatic in details. The ring grabens around the three Tharsis Montes, and the half-ring grabens around Alba Patera, fit the expected flexural response to the volcanic loads. The Noachian fault sets following the NE–SW Tharsis volcano trend, sometimes called "rift line", might be witnesses of an early major extension event of enigmatic origin, which requires an extensional state of stress over a large part of the Martian surface (Hartmann, 1973). Some of the intricate central Tharsis fault sets could also be linked to this event, together with the Noachian Ceraunius grabens, although no satisfactory explanation has been found.

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