sence of the oceans, ridges would rise 2 to 2.5 km above the surrounding abyssal plains [*Parsons and Sclater*, 1977]. Because this amount of thermally induced relief on earth requires that near-melting temperatures extend almost to the surface, we regard 2 km as a reasonable limit to the contribution of thermal effects to the relief of the Tharsis rise at any stage in the history of the province.

Compositional effects. Support of the Tharsis rise by density differences between the crust or mantle beneath Tharsis and those of adjacent areas could also be accomplished, in principle, by lateral variations in composition. *Wise et al.* [1979*a*, *b*] have proposed that lighter crustal material from the northern hemisphere of Mars was transported by mantle convective flow to the region over a zone of downwelling initiated by core infall. 'Underplating' of the sub-Tharsis crust, by this scenario, led to a permanent isostatic rise of the Tharsis region. Since isostatic crustal thickening involves simply the replacement of mantle material with a greater volume of crustal material, the subsurface excess mass required by gravity and topographic data [*Phillips and Saunders*, 1975] is not explained by this model.

The isostatic model of *Sleep and Phillips* [1979], as was noted above, involves a large anomaly in the density of the mantle beneath Tharsis, relative to adjacent regions, and a crust either thinner or denser than average for the planet. The long-term stability of such a lateral density contrast extending to depths of several hundred kilometers against the mixing effects of mantle convective flow has not been addressed. The only compositional mechanism explored quantitatively to date to account for the proposed density model is depletion of the sub-Tharsis mantle of its low-melting-point, presumably basaltic, component [*Finnerty and Phillips*, 1981]. Finnerty and Phillips have suggested, for instance, that removal by partial melting of 30% of the uppermost 140 km of mantle beneath Tharsis would decrease the mantle density sufficiently to produce a 10-km topographic rise. The volcanic extrusives or plutonic intrusives removed from the mantle after such an episode of partial melting would have a total thickness of 40–50 km. The isostatic model by this mechanism cannot therefore be attained until after a vast outpouring of volcanic material, reaching a total thickness far in excess of the present topographic relief. The Finnerty-Phillips model predicts that the crust beneath the Tharsis rise would be thicker than average, in contrast to the thinned crust suggested by *Sleep and Phillips* [1979], unless much of the basaltic component of the mantle beneath Tharsis was emplaced volcanically in regions adjacent to Tharsis [*Phillips et al.*, 1981]. We regard this last suggestion as unsupported by the geological history of the Tharsis province summarized above. Despite the emphasis of *Finnerty and Phillips* [1981] on the support of the Tharsis rise by chemical variations in the Martian mantle, their model has some strong similarities to the model of *Solomon and Head* [1980*b*], elaborated below, in which the Tharsis rise is dominantly a product of volcanism.

Summary. The lithospheric uplift model for Tharsis [*Carr et al.*, 1973; *Phillips et al.*, 1973; *Hartmann*, 1973; *Carr*, 1974] fails to predict lithospheric stresses in agreement with the observed pattern of fractures and ridges. The isostatic model of *Sleep and Phillips* [1979] can account for some of the observed tectonic features [*Banerdt et al.*, this issue], but only if at least a large fraction of the required mantle density anomaly is nonthermal in origin. The only nonthermal mechanism explored quantitatively [*Finnerty and Phillips*, 1981] requires as a first step the emplacement of tens of kilometers of volcanic material in the Tharsis region. A downward load on the Martian lithosphere is necessary to explain at least some of the observed tectonic features in the Tharsis province [*Willemann and Turcotte*, 1981; *Banerdt et al.*, this issue]. Thus areally extensive and voluminous igneous activity is a required element for both a loading model and the Finnerty-Phillips isostatic model. In the next section we show that volcanic construction, lithospheric loading, and lithospheric failure in response to load-induced stresses are widespread and interlinked processes on Mars. We then present a physical model for the evolution of Tharsis in which these same processes largely account for the geological history of the Tharsis province.

LOADING AND LITHOSPHERIC FLEXURE ON MARS

The eruption of volcanic deposits on a planet generates a load on the underlying lithosphere. The flexure of the lithosphere in response to loads of lateral extent small in comparison with the planetary radius is a strong function of the effective thickness of the upper elastic portion of the lithosphere. Stresses generated as a result of such flexure can exceed the strength of brittle lithospheric material, leading to failure. The processes of volcanic loading, flexure, and failure are well known on the earth [e.g., *Walcott*, 1976; *Watts*, 1978; *McNutt*, 1980], particularly in oceanic regions where the apparent thickness of the elastic lithosphere is inversely proportional to the mean thermal gradient in the uppermost tens of kilometers of the oceanic plate [*Caldwell and Turcotte*, 1979; *Watts et al.*, 1980]. These processes are also well documented on the moon, where volcanic deposits in the circular mascon maria have loaded the lunar lithosphere to levels sufficient to produce failure in response to induced bending stresses [*Solomon and Head*, 1980*a*].

Volcanic units similarly exert loads on the lithosphere of Mars. The loads produced by the youngest major volcanic constructs must be at least partially supported by the finite strength of the Martian lithosphere, since these constructs are sites of prominent positive free-air gravity anomalies [*Sjogren*, 1979]. The circumferential graben surrounding many of these constructs (e.g., Elysium Mons in Figure 6) indicate that brittle failure has occurred in response to load-induced stresses, acting to reduce the magnitude of stresses supported by finite strength. The Olympus Mons shield, in contrast, has a gravity anomaly sufficiently large that almost no compensation of the load has occurred in the time since the volcano formed [*Thurber and Toksöz*, 1978; *Comer and Solomon*, 1981].

By combining knowledge of the loads of individual volcanic constructs inferred from measured excess masses [*Sjogren*, 1979] with the positions of circumferential graben in response to those loads [*Scott and Carr*, 1978], we have estimated the thickness of the elastic lithosphere of Mars beneath a number of volcanic features at the time of graben formation [*Solomon et al.*, 1979; *Comer et al.*, 1980]. A map of the inferred thicknesses is shown in Figure 7. The figure presents strong evidence for the existence of lateral variations in the thickness of the elastic lithosphere of Mars. In particular, there appears to have been a dichotomy in lithospheric thickness that was insensitive to load age. Thin (25–50 km) elastic lithosphere is indicated for the regions immediately surrounding large shield volcanoes in the later