effective Young's modulus is not well known for Mars, to first order the resulting relative uncertainty in T_e is only one third that in E. In general, of course, curvature is not a constant but varies along a flexural profile; it may be argued [*McNutt*, 1984], however, that the estimates of D and T_e in Table 1 are dominated by the portions of the profiles at which bending stress, and thus curvature, are near their maximum values, so we assume such maximum values in using Figures 5 and 6.

The effective surface temperature T_s has an uncertainty of perhaps 20–30 K. The mean microwave brightness temperature of Mars is about 200 K [*Morrison et al.*, 1969]. Climate models for Mars predict that the present seasonally averaged surface temperature should vary with latitude and should be 200–230 K at the latitudes (approximately 40°N to 10°S) of the features listed in Table 1 [*Kieffer et al.*, 1977; *Fanale et al.*, 1982]. Such models do not generally include variations in topography, which likely would add a superposed variation in surface temperature of perhaps 10–20 K [*Seiff and Kirk*, 1977] for the range in regional elevations of prominent lithospheric loads. An additional uncertainty is contributed by poorly constrained differences between the present climate of Mars and that at the time of load-induced graben formation. A final factor is the possible presence of a low-conductivity regolith layer, which can be accommodated in the simple model adopted here by use of a small increase in T_s [*McNutt and Menard*, 1982]. A 20–30 K uncertainty in T_s contributes only a 3–7% uncertainty to estimates of lithospheric thermal gradient and heat flow.

The assumption of a linear thermal gradient can also be questioned. In effect, for a given strain rate and flow law the quantity T_m essentially gives the depth to a particular temperature, and any temperature distribution that passes through that temperature-depth point and the surface temperature T_s (and that does not exceed the limiting temperature at depths less than T_m) should be regarded as possible. Effects such as upward concentration of crustal heat-producing elements and a thermal conductivity that increases with depth will tend to give a temperature distribution that is concave downward, so that near-surface gradients will be higher than those indicated in Figures 5 and 6. Further, there is also at least an order or magnitude uncertainty in the growth time of the lithospheric loads utilized to estimate T_e and thus in the effective flexural strain rate. A factor of 10 uncertainty in growth time, however, contributes only a 4% uncertainty to the derived values of T_m or dT/dz for either of the adopted flow laws.

A source of uncertainty is the abundance of water in the Martian interior. In the upper crust a significant pore pressure can lower the frictional strength [*Brace and Kohlstedt*, 1980], which would result in modest increases in T_m (or decreases in dT/dz) for given values of elastic thickness and curvature. Whether water is present in the lower crust and upper mantle of Mars is an issue for the estimation of ductile strength, since the presence of water can significantly affect creep rates in rock [e.g., *Ashby and Verrall*, 1977]. The adopted olivine flow law, although derived from laboratory measurements, has been reasonably well validated for the terrestrial oceanic mantle and geological strains rates, by means of both flexural estimates of mechanical lithosphere thickness [*McNutt and Menard*, 1982; *McNutt*, 1984] and centroid depths of intraplate earthquakes [*Bergman and Solomon*, 1984], although *McNutt and Menard* [1982] have suggested that the activation energy for creep in the mantle may be slightly less than that obtained from laboratory measurements on dry olivine. For Mars we note that significant water in the mantle and an activation energy less than that assumed here [*Goetze*, 1978] would both result in lower estimates of thermal gradient than shown in Figure 5; we comment on this point further below.

The flow law for crustal material is probably less well constrained in general than that for the mantle. To explore at least partially the effect of this uncertainty, we also considered the flow law of anorthosite reported by *Shelton* [1981]. Values of T_m and dT/dz differ by about 5–15% from those obtained with the diabase flow law of *Caristan* [1982] for strain rates in the range 10^{-16} to 10^{-19} s^{-1}. The straight-line approximation to the ductile portion of the strength envelope, while a reasonable simplification for the olivine flow law [*McNutt and Menard*, 1982], with flow laws for crustal material leads to an underestimate of ductile strength in the lowermost mechanical lithosphere and thus a slight increase in estimated T_m, for an assumed constant strain rate.

A considerable uncertainty in the application of relations such as those depicted in Figures 5 and 6 arises from our poor knowledge of the thickness of the Martian crust. Long-wavelength topography and gravity, if intepreted in terms of a crust of uniform density and variable thickness, indicate a mean crustal thickness of at least 30 km [*Bills and Ferrari*, 1978], which corresponds to zero crustal thickness beneath the Hellas basin. Models of the Viking line of sight (LOS) Doppler radio tracking residuals over the Hellas basin and the 370-km-diameter crater Antoniadi are consistent with essentially local Airy compensation if the crust beneath these impact features is 120–130 km thick [*Sjogren and Wimberly*, 1981; *Sjogren and Ritke*, 1982]. From the models of *Bills and Ferrari* [1978] a 130-km-thick crust beneath the Hellas basin [*Sjogren and Wimberly*, 1981] corresponds to a globally averaged crustal thickness of about 150 km. LOS tracking data over Elysium Planitia and Olympus Mons can be fit with varying degrees of Airy isostatic compensation and crustal thicknesses of 30–150 km [*Janle and Ropers*, 1983; *Janle and Jannsen*, 1986]. Available gravity data are thus permissive of a mean crustal thickness anywhere in the range 30–150 km, an interval consistent with, but not substantially narrowed by, the predictions of differentiation models of an early Martian magma ocean [*Warren*, 1988]. Whatever the mean crustal thickness, local variations of ±30 km or more are also likely [*Phillips et al.*, 1973; *Bills and Ferrari*, 1978].

Given these uncertainties, our approach is to assume that the large values of elastic lithosphere thickness ($T_e > 100$ km) determined from the local response to the Isidis mascon and Olympus Mons and from the global response to the Tharsis rise exceed the thickness of the Martian crust, while the values of 20–30 km obtained for T_e beneath the Tharsis Montes and Alba Patera (Table 1) are less than or comparable to the thickness of the crust. This assumption is consistent with known constraints on the crustal thickness and, because of the lesser creep resistance of crustal material, results in minimizing the spread in inferred thermal gradients.

Finally, it should be noted that the estimation of mechanical lithosphere thickness and thermal gradient is made under the assumption that flexural stresses dominated the local lithospheric stress field at the time of load-induced