[1985] and *Laul et al.* [1986] have inferred that the mantle plus crust of the SNC parent body has a U abundance of 16 ppb, a Th/U ratio of 3.5, and a K/U ratio of 2×10^4. For an inferred mass fraction of 22% for an Fe-FeS core [*Dreibus and Wänke*, 1985; *Laul et al.*, 1986] the U abundance expressed as a fraction of the whole planet is 12.5 pbb. The present heat production for such bulk abundances is equivalent, under steady state, to a mean heat flow of 14 mW m^{-2}. *Treiman et al.* [1986] obtain element-element correlations similar to those of *Dreibus and Wänke* [1985] and *Laul et al.* [1986], with slightly smaller Th/U and K/U ratios (3.0 and 1×10^4, respectively), which for the same bulk U abundance would give a heat production about 20% lower.

These values of heat flow may be compared with the thermal gradients in Table 1 and corresponding estimates of thermal conductivity. For Tharsis Montes and Alba Patera, T_m is less than the crustal thickness, so we adopt a thermal conductivity appropriate for crustal material, 2.5 W m^{-1} K^{-1} [*Clark*, 1966]. Lithospheric thermal gradients of 10–14 K km^{-1} (Table 1) correspond to heat flow values of 25–35 mW m^{-2}. For the mechanical lithosphere supporting Olympus Mons, the Isidis mascon, and the Tharsis rise, most of the lithosphere probably consists of mantle material, so we adopt the thermal conductivity given by *Schatz and Simmons* [1972] for olivine ($Fo_{86}Fa_{14}$) at 560–930 K: 4 W m^{-1} K^{-1}. Mean lithospheric gradients less than 5–6 K km^{-1} (Table 1) then correspond to heat flow values less than 20–24 mW m^{-2}. These last estimates of heat flow, and the associated temperature distributions depicted in Figure 7, should be reduced to be consistent with the likely contrast in thermal conductivity and gradients across the crust-mantle boundary, but in order to do so a crustal thickness must be assumed. For a crust 30–50 km thick and the above values of crustal and mantle thermal conductivity the upper bound on heat flow consistent with mean lithospheric gradients less than 5–6 K km^{-1} is 17–21 mW m^{-2}. A global elastic lithosphere at least 100 km thick, indicated by flexural models for the support of the Tharsis rise [*Willemann and Turcotte*, 1982; *Banerdt et al.*, 1982], implies lithospheric gradients of 7 K km^{-1} or less. The corresponding upper bound on heat flow is 23–25 mW m^{-2} for a mean crustal thickness of 30–50 km.

Thermal gradients at or near the upper end of the range allowed for Olympus Mons, Isidis, and the globally averaged response to Tharsis thus give heat flow values similar to or slightly larger than that expected from U, Th, and K abundances in SNC meteorites under the steady state assumption. Such upper bounds on thermal gradients are also consistent with a modest contribution to heat flow from secular cooling of the mantle [e.g., *Schubert et al.*, 1979]. Most published thermal history models for Mars, in contrast, predict a heat flow larger than that permitted by these upper bounds, a result attributable to an overestimation of heat production in the interior. We suggest that the global heat flux on Mars during the Amazonian epoch (approximately the last 2 b.y. [*Tanaka*, 1986]) has been in the range 15–25 mW m^{-2}. This inference is in agreement both with the thermal gradients implied by the thick elastic lithosphere beneath Olympus Mons and globally supporting the Tharsis rise and with the bulk abundances of heat-producing elements in Mars inferred from SNC meteorites. Regional heat flow in the centers of major volcanic provinces of Mars during the Amazonian epoch reached highs of 35 mW m^{-2} or more, as much as twice the typical global value.

A final point of comparison with the thermal gradients in Table 1 and Figure 7 is provided by estimates of the thickness of the thermal lithosphere made from the heights of volcanic constructs and a hydrostatic model for magmatic overpressure [*Vogt*, 1974; *Carr*, 1976; *Blasius and Cutts*, 1976]. Such estimates are quite approximate, given uncertainties contributed by differential compressibility between magma and surrounding rock, viscous head loss in the magma conduit, and additional overpressure contributed by magmatic volatiles, as well as by the probably oversimplified notion of a continuous magma column extending from a basal magma chamber to the volcano summit. We consider such models here merely as tests of consistency. For instance, the thermal gradient shown in Figure 7 for Olympus Mons, if continued downward, reaches a temperature of 1670 K, the 2.3-GPa solidus temperature of a model Martian mantle [*Bertka and Holloway*, 1988], at a depth of 285 km. Inasmuch as the volcano stands about 24 km higher than the surrounding terrain [*Wu et al.*, 1984], this relief is consistent with the thermal gradient shown and with the hydrostatic head model if the average effective density contrast between magma and the surrounding rock column is about 8%, a quite reasonable value. The lesser relief (14–18 km) of the Tharsis Montes volcanoes [*Blasius and Cutts*, 1976] is consistent with the higher thermal gradients indicated for the lithosphere beneath these features (Table 1 and Figure 7), but quantitative agreement with the magma hydrostatic model would require consideration of distinct densities and thermal conductivities for the crust and mantle, lateral variations in crustal thickness [*Blasius and Cutts*, 1976], and the possible contribution of convective flow to heat transport in the lower lithosphere beneath the central Tharsis region.

Evolution of the Tharsis Province

As noted above, considerable attention has been devoted to the deep structure and evolution of the Tharsis volcanic and tectonic province. While the long-wavelength gravity and topography of the region are not consistent with complete local isostatic compensation by a single mechanism such as crustal thickness variations [*Phillips and Saunders*, 1975], complete local compensation is possible if a combination of Airy (crustal thickness variations) and Pratt (lateral density variations) mechanisms act in concert, but only if the crust is relatively thin (or is pervasively intruded by high-density plutonic material) beneath the Tharsis rise and substantial density anomalies persist to at least 300–400 km depth [*Sleep and Phillips*, 1979, 1985; *Finnerty et al.*, 1988]. Alternatively, the gravity and topography are consistent with the hypothesis that a portion of the high topography of Tharsis is supported by membrane stresses in the Martian elastic lithosphere [*Willemann and Turcotte*, 1982; *Banerdt et al.*, 1982; *Sleep and Phillips*, 1985].

These compensation models have been used to calculate lithospheric stresses for comparison with the observed distribution of tectonic features. The isostatic models for Tharsis predict stresses in approximate agreement with the distribution and orientation of extensional fractures in the central Tharsis region and of compressive wrinkle ridges oriented approximately circumferential to the center of tectonic activity, while the models involving lithospheric