support of a topographic load predict stresses consistent with the more distal extensional features in regions adjacent to the Tharsis rise [*Banerdt et al.*, 1982; *Sleep and Phillips*, 1985]. An evolution in the nature of the support of Tharsis topography is the preferred explanation of the mixed success of individual stress models [*Banerdt et al.*, 1982; *Solomon and Head*, 1982], but the ordering depends on the relative ages of distal and proximal tectonic features. If the distal features are older, then a combination of faulting and viscoelastic relaxation of stresses associated with an early episode of lithospheric loading may have led to an essentially isostatic state at present [*Sleep and Phillips*, 1985]. If, instead, the distal features are younger, then a progression from local isostasy to lithospheric support as the Tharsis rise was constructed may have been the natural consequence of global interior cooling and lithospheric thickening [*Solomon and Head*, 1982; *Banerdt et al.*, 1982; *Sleep and Phillips*, 1985].

The timing of ridge formation on Mars provides a constraint on these two scenarios for the evolution of Tharsis. From a global analysis of the distribution of wrinkle ridges on Mars, *Chicarro et al.* [1985] find that ridges occur commonly throughout ancient terrain. In the volcanic plains, however, the distribution is highly uneven, with ridges strongly concentrated in the ridged plains units and in spotty occurrences in other regions. The lower Hesperian age (approximately 3–3.5 b.y. ago) for most major ridged plains units [*Tanaka*, 1986] and the contrast in ridge density between cratered uplands and young volcanic plains [*Chicarro et al.*, 1985] suggest that ridge formation may have been concentrated in a comparatively early stage in Martian evolution [*Watters and Maxwell*, 1986]. Examination of crosscutting relations between ridges and graben in the Tharsis province also supports the view that most ridge formation in that region was restricted to an early time period [*Watters and Maxwell*, 1983]. Given that the distribution and orientation of the ridges surrounding Tharsis are best explained by the isostatic model for support of topography, then any earlier episode of support by lithospheric strength [*Sleep and Phillips*, 1985] must have occurred prior to 3 b.y. ago.

The thermal gradient and heat flux implied by the large values of T_e in lithospheric support models for the Tharsis rise provide an independent constraint on this issue. As discussed above, values of T_e of 100–400 km [*Willemann and Turcotte*, 1982; *Banerdt et al.*, 1982; *Sleep and Phillips*, 1985] imply thermal gradients of 2–7 K km^{-1} or heat flow values of 7–25 mW m^{-2} (including a crust 30–50 km thick). The highest of these values for gradient and heat flow are more consistent with the values expected in the second half of Martian history than with the much higher values likely in the era before 3 b.y. ago, while the lowest of these values ($T_e > 200$ km or $dT/dz < 4$ K km^{-1}) are inconsistent with the heat budget expected from SNC meteorites and an olivine rheology for the Martian mantle whatever the age of loading. As noted above, a substantial water content in the Martian mantle would be expected to lower the creep resistance over that represented by the flow law of *Goetze* [1978] and *Evans and Goetze* [1979]. As a result, the thermal gradient implied by a given value of T_e would be even lower than that indicated here, strengthening these arguments further.

The average Martian heat flux prior to 3 b.y. ago can be estimated, but only in a model-dependent manner. With the bulk abundances of heat-producing radionuclides given by *Dreibus and Wänke* [1985] and *Laul et al.* [1986], radioactive heat production 3 b.y. ago was a factor of 2.7 higher than the present value. Thermal history models with parameterized mantle convection and secular cooling from a hot initial state give heat flow values at 3 b.y. ago higher than that at present by a similar factor [*Schubert et al.*, 1979; *Davies and Arvidson*, 1981]. If, as argued above, the present heat flow is 15–25 mW m^{-2}, then the heat flow prior to 3 b.y. for such thermal evolution models was in excess of 40–70 mW m^{-2}, too large to be consistent with an elastic lithosphere 100–400 km thick. Thermal history models with extended periods of global warming can also be devised [e.g., *Solomon and Chaiken*, 1976; *Toksöz and Hsui*, 1978; *Davies and Arvidson*, 1981], but such models are not consistent with early core-mantle differentiation as indicated by U-Pb systematics in SNC meteorites [*Chen and Wasserburg*, 1986] or the hot initial state implied by current models for planetesimal accretion [*Wetherill*, 1985].

These considerations suggest that the most likely evolutionary scenario for Tharsis was an early state of nearly isostatic equilibrium followed by additional loading of the lithosphere by extensive volcanic and igneous intrusive material and long-term support of a significant fraction of this load by the finite strength of the elastic lithosphere [*Solomon and Head*, 1982; *Banerdt et al.*, 1982]. Thermal gradient considerations support the view that the global elastic lithosphere presently supporting the Tharsis rise is no greater than 100–200 km in thickness.

Nature of Lateral Heterogeneity in Thermal Structure

As noted above, the differences in lithospheric thermal gradients implied by the different values of T_e (Table 1) must be partly, if not primarily, due to lateral variations in temperature within and beneath the lithosphere. Essentially contemporaneous temperature differences of as much as 300 K must have been present at 30–40 km depth at a late stage in the development of the Tharsis province (Figure 7). Two possible sources of these variations are lithospheric heating by mantle advection [e.g., *Carr*, 1974] and thermal differences remaining from large impacts [*Schultz and Glicken*, 1979]. Calculations of the cooling following formation of a large impact basin [*Bratt et al.*, 1985] indicate that the temperature differences in Figure 7 are far too large to be principally the consequence of basin-forming events several billion years earlier. In contrast, a lateral difference in temperature of 300 K or more at 30–40 km depth is similar to the temperature variation associated with lithospheric reheating beneath hot spot volcanic centers on Earth [*McNutt*, 1987]. We may conclude that the temperature anomalies beneath the central regions of major volcanic provinces on Mars are similarly the result of mantle dynamic processes, including convective upwelling and associated magmatism.

It is of interest to estimate the fraction of the global heat flux delivered by volcanic and magmatic processes and by enhanced conductive thermal gradients in the major volcanic provinces of Mars. Presumably, this fraction provides an approximate measure of the relative importance of comparatively narrow zones of mantle upwelling, or plumes, to mantle heat transport. Most of the regions for which estimates of elastic lithosphere thickness are available are