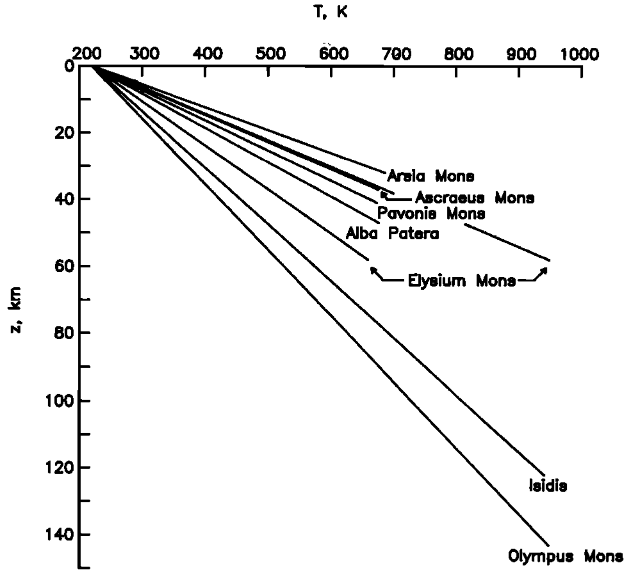

Fig. 7. Estimated thermal gradients in the mechanical lithosphere of Mars beneath major loads. These gradients correspond to the time of local loading and flexurally induced faulting (Table 1). Each temperature distribution is assumed to be linear with depth and is constrained to match a surface temperature of 220 K and the thermal gradient and mechanical lithosphere thickness T_m given by relations of the sort depicted in Figures 5 and 6. Two gradients are shown for Elysium Mons, corresponding to cases in which the ductile portion of the strength envelope is limited by the creep strength of either olivine (top line) or diabase. The gradients shown for the Isidis mascon and Olympus Mons are upper bounds.

faulting. Superimposed regional or global membrane stresses would act to move the bending stress curve (Figure 4) relative to the strength envelope, resulting in a somewhat larger mechanical lithosphere thickness for given values of T_e and curvature [*McNutt*, 1984]. While a number of models for regional and global membrane stress have been proposed for Mars [e.g., *Solomon and Chaiken*, 1976; *Banerdt et al.*, 1982; *Sleep and Phillips*, 1985], the magnitude and time dependence of such stress fields are too uncertain to be included in the calculations reported here.

Results

For the Isidis mascon and Olympus Mons the maximum values of flexurally induced curvature and strain are modest (10^{-8} m^{-1} and 0.5 to 1 × 10^{-3}, respectively), even for the lower bounds on effective elastic lithosphere thickness (Table 1). Under these conditions the depth T_m to the base of the mechanical lithosphere is approximately equal to T_e [*McNutt*, 1984] and, by the above assumption on crustal thickness, is determined by the ductile strength of the mantle (Figure 5). The minimum values of T_e for the Isidis mascon and Olympus Mons correspond, by this line of reasoning, to mean lithospheric thermal gradients of 5–6 K km^{-1} (Table 1 and Figure 7). These values are upper bounds to the mean vertical thermal gradients in the mechanical lithosphere beneath these features. Values for T_e in the range 200–300 km for the Isidis mascon and Olympus Mons regions correspond to mean thermal gradients of 2.5–4 K km^{-1}. Similarly, the range in elastic lithosphere thickness (100–400 km) obtained from the global response to the Tharsis rise [*Willemann and Turcotte*, 1982; *Banerdt et al.*, 1982] implies lithospheric thermal gradients of 2–7 K km^{-1} for characteristic strain rates of 10^{-18}–10^{-20} s^{-1}.

For the flexural loads of the Tharsis Montes volcanoes and Alba Patera the maximum flexural curvatures and strains are larger (2 to 5 × 10^{-7} m^{-1} and 4 to 5 × 10^{-3}, respectively, for the best fitting values of T_e) than for Olympus Mons or the Isidis mascon. Under these conditions the mechanical lithosphere thickness T_m exceeds T_e and, by assumption, is limited by the strength of crustal material (Figure 6). The mean thermal gradients consistent with the best fitting values of T_m for these loads under this assumption are in the range 10–14 K km^{-1} (Table 1 and Figure 7). Consideration of upper and lower bounds on T_e [*Comer et al.*, 1985] expands the range of possible thermal gradients to 7–27 K km^{-1} (Table 1).

The thermal gradient corresponding to the value T_e = 54 km determined for Elysium Mons [*Comer et al.*, 1985] depends strongly on the thickness of the Martian crust. In Table 1 and Figure 7 we show the gradient implied by the alternative assumptions that the thickness of the crust is significantly less than or significantly greater than 54 km. With either assumption the gradient falls between those for Olympus Mons and Isidis and those for the Tharsis Montes and Alba Patera. If T_e is similar to the crustal thickness, then the gradient will depend on the degree of coupling between the strong layers of the crust and the mantle. *McNutt et al.* [1988] have shown how the presence of a decoupling zone in the lower crust can substantially change the bending moment for given values of K and dT/dz. Given our limited understanding at present of the Martian interior, further consideration of such complicating factors does not seem warranted.

IMPLICATIONS

The thermal gradients in Table 1 and Figure 7 provide new constraints on Martian heat flux on both global and more limited regional scales. The magnitude and lateral scale of thermal heterogeneity in the Martian lithosphere also provide useful information on interior heat transport processes. We comment on these implications in turn.

Global Heat Flux

While the heat flow and thermal structure of Mars are not known, estimates are available from thermal history models, from studies of shergottites, nakhlites, and Chassigny (SNC meteorites), and from considerations of magma hydrostatics. Published thermal history models for Mars give present global mean heat flux values of 20–45 mW m^{-2}, with the large range a function primarily of different assumptions regarding internal heat production [*Lee*, 1968; *Johnston et al.*, 1974; *Toksöz et al.*, 1978; *Toksöz and Hsui*, 1978; *Turcotte et al.*, 1979; *Davies and Arvidson*, 1981; *Stevenson et al.*, 1983]. For comparison, if Mars loses heat at the same rate per mass as the Earth [*Sclater et al.*, 1980], then the mean heat flux would be 31 mW m^{-2}.

From interelement correlations in SNC meteorites, several workers have estimated the abundances and ratios of major heat-producing elements in the SNC parent body [*Dreibus and Wänke*, 1985; *Laul et al.*, 1986; *Treiman et al.*, 1986]. Under the premise that the SNC meteorites were derived from Mars [e.g., *McSween*, 1984; *Bogard et al.*, 1984; *Becker and Pepin*, 1984], these estimates provide useful constraints on global heat flux. *Dreibus and Wänke*