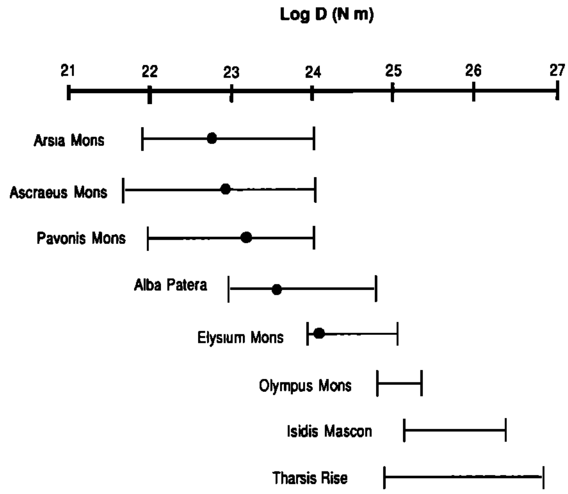

Fig. 2. Summary of estimates of the flexural rigidity of the Martian elastic lithosphere, from Table 1. The solid circles denote best fitting values [*Comer et al.*, 1985], while the bars delimit the ranges of possible values [*Willemann and Turcotte*, 1982; *Banerdt et al.*, 1982; *Comer et al.*, 1985; *Janle and Jannsen*, 1986].

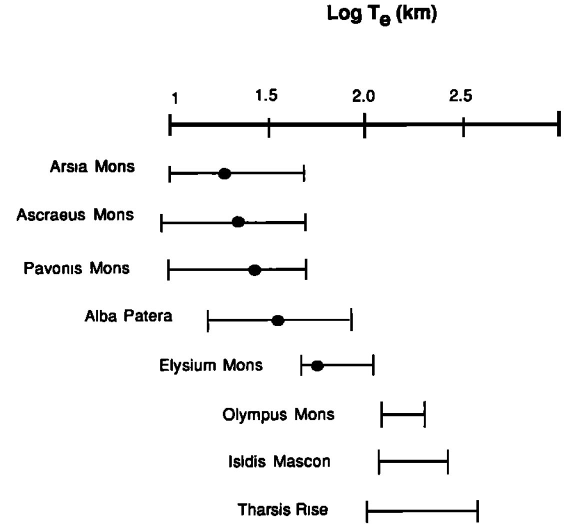

Fig. 3. Summary of estimates of the effective thickness of the elastic lithosphere of Mars, from Table 1. Symbol conventions are as in Figure 2.

sponding to at least the late Amazonian and perhaps to somewhat earlier epochs as well [*Scott and Tanaka*, 1986; *Tanaka et al.*, 1988].

A variety of models for the deep structure of the Tharsis rise have been proposed [e.g., *Sleep and Phillips*, 1979, 1985; *Banerdt et al.*, 1982; *Willemann and Turcotte*, 1982; *Finnerty et al.*, 1988]. The distribution and orientation of at least some of the tectonic features of the Tharsis province are best explained if the long-wavelength topography of the Tharsis rise was supported for a significant time by the finite strength of the global elastic lithosphere [*Willemann and Turcotte*, 1982; *Banerdt et al.*, 1982; *Sleep and Phillips*, 1985]. Quantitative models of this support that provide a reasonable fit to the geoid and to the distribution of tectonic features have an elastic lithosphere 100 to 400 km thick, corresponding to $D = 10^{25}$ to 7×10^{26} N m [*Willemann and Turcotte*, 1982; *Banerdt et al.*, 1982]. The time at which this estimate is an appropriate average for the planet depends on the evolution of the mechanism of support of long-wavelength topography [*Sleep and Phillips*, 1985] and can range from Middle Noachian (3.8 b.y. ago) to Upper Amazonian [*Tanaka*, 1986].

The best fitting values for D derived for various loads span at least 2 orders of magnitude (Figure 2), and the corresponding values for T_e span nearly a factor of 10 (Figure 3). As is apparent from Table 1 and from the discussion above, these values are not consistent with a simple progressive increase with time in the thickness of the elastic lithosphere of Mars. The largest estimates of T_e, for instance, are for the oldest (Isidis mascon) and youngest (Olympus Mons) local lithospheric loads considered. Spatial variations in elastic lithosphere thickness must therefore have been at least as important as temporal variations [*Comer et al.*, 1985]. In particular, there appears to have been a dichotomy in lithosphere thickness that spanned a significant interval of time, with comparatively thin elastic lithosphere (T_e = 20–50 km) beneath the central regions of major volcanic provinces and substantially thicker elastic lithosphere (T_e in excess of 100 km) beneath regions more distant from volcanic province centers and appropriate for the planet as a whole.

THERMAL GRADIENTS

Procedure

All of the values of T_e described above were obtained under the assumption that the elastic lithosphere of Mars behaves as a uniform elastic plate or shell. A better model for the lithosphere is an elastic-plastic plate, with the strength as a function of strain rate, plate curvature, and depth [*Goetze and Evans*, 1979; *Brace and Kohlstedt*, 1980]. To estimate the mean lithospheric thermal gradient from such an effective elastic lithosphere thickness, T_e must first be converted to the depth T_m to the rheological boundary marking the base of the mechanical lithosphere (Figure 4). This conversion is accomplished by adopting a representative strain rate and a flow law for ductile deformation of material in the lower lithosphere, constructing models of bending stress consistent with the adopted strength envelope, and finding for each model the equivalent elastic plate model having the same bending moment and curvature. This procedure has been described by *McNutt* [1984].

The temperature distribution in the mechanical lithosphere is assumed to be locally given by a surface temperature T_s of 220 K [*Kieffer et al.*, 1977; *Fanale et al.*, 1982] and a uniform vertical gradient dT/dz. We take the representative strain rate for the flexural response to each local load to be the quotient of the maximum horizontal strain given by the elastic model and the growth time of the load. On the basis of stratigraphic relations among the principal volcanic units associated with the major lithospheric loads [*Scott and Tanaka*, 1986; *Tanaka*, 1986] and the crater chronology of *Hartmann et al.* [1981], we take the load growth time to be 10^8 years. The brittle and ductile portions of the strength envelope are approximated by straight lines, so that the bending moment may be found analytically given the curvature and dT/dz [*McNutt and Menard*, 1982; M. K. McNutt, personal communication, 1988]. The brittle portion of the strength envelope is taken from the low-pressure friction law of *Byerlee* [1978], with a lithostatic pressure gradient appropriate to the Martian crust (11 MPa km^{-1}) under the assumption of negligible pore pressure. If the mechanical litho-