TABLE 1. Estimates of Effective Elastic Lithosphere Thickness and Lithospheric Thermal Gradient on Mars

Feature	Age of Deformation	D, 10^{23} N m	T_e, km	dT/dz, K km^{-1}
Arsia Mons	UA	0.5 (0.1–11)	18 (10–50)	14 (7–23)
Ascraeus Mons	UA	0.9 (0.05–11)	22 (8–50)	12 (7–27)
Pavonis Mons	UA	1.6 (0.1–11)	26 (10–50)	11 (7–21)
Alba Pateria	LA	3.2 (0.6–55)	33 (19–85)	10 (8–14)
Elysium Mons	LA	14 (10–120)	54 (48–110)	7–13 (6–14)
Olympus Mons	UA	50–240	140–230	<5
Isidis mascon	UN	>150	>120	<6
Tharsis rise	MN–UA	>100	>100	<7

Sources of data are as follows: D and T_e are from *Thurber and Toksöz* [1978], *Willemann and Turcotte* [1982], *Banerdt et al.* [1982], *Comer et al.* [1985], and *Janle and Jannsen* [1986]; ages are stratigraphic positions from *Tanaka et al.* [1988], defined as follows: N, Noachian; H, Hesperian; A, Amazonian; L, lower; M, middle; and U, upper. Parentheses denote bounds on parameter values. Thermal gradients dT/dz are derived under the assumption that strength at the base of the mechanical lithosphere is limited by the creep strength of diabase [*Caristan*, 1982] for $T_e \leq 50$ km and by that of olivine [*Goetze*, 1978] for $T_e > 50$ km; for Elysium Mons the ranges in dT/dz reflect the possibility that either flow law may limit the strength at the base of the mechanical lithosphere for the best fitting and lower bound values of T_e.

well as consideration of such important time-dependent effects as volcano growth during the formation interval of preserved graben (Table 1). The graben judged most likely to be of flexural origin formed in volcanic surfaces of Upper Amazonian age for Ascraeus, Pavonis, and Arsia Montes and Lower Amazonian age for Alba Patera and Elysium Mons [*Scott and Tanaka*, 1986; *Greeley and Guest*, 1987; *Tanaka et al.*, 1988]. While the assignment of actual age intervals to stratigraphic series from impact crater density carries considerable uncertainty, the Upper and Lower Amazonian stratigraphic levels correspond approximately to ages of less than 250 m.y. and 0.7–1.8 b.y., respectively, according to the *Hartmann et al.* [1981] cratering time scale [*Tanaka*, 1986].

For the Isidis basin region (Figure 1) the elastic lithosphere thickness exceeded 120 km ($D > 10^{25}$ N m) and may have been as large as 200–300 km ($D = 0.7$ to 2.4×10^{26} N m), at the time of mascon loading and graben formation [*Comer et al.*, 1985]. Basin concentric graben cut Upper Noachian surfaces and are buried by Lower Hesperian units [*Greeley and Guest*, 1987; *Maxwell and McGill*, 1988]. These relationships indicate a time of formation between 3 and 3.8 b.y. ago according to the *Hartmann et al.* [1981] time scale [*Tanaka*, 1986], similar to the time of mascon formation and associated graben development on the Moon [*Lucchitta and Watkins*, 1978; *Solomon and Head*, 1980].

The absence of circumferential graben around Olympus Mons (Figure 1) requires the elastic lithosphere to have been at least 150 km thick ($D > 3 \times 10^{25}$ N m) at the time of loading [*Thurber and Toksöz*, 1978; *Comer et al.*, 1985]. On the basis of an upward revision to the volume of Olympus Mons [*Wu et al.*, 1984], *Janle and Jannsen* [1986] were able to place both upper and lower bounds on D (5×10^{24}–2×10^{25} N m) and T_e (140–230 km); these ranges include an explicit uncertainty in E (20–80 GPa). These estimates for T_e apply to a late stage in the growth of the volcano, corre-

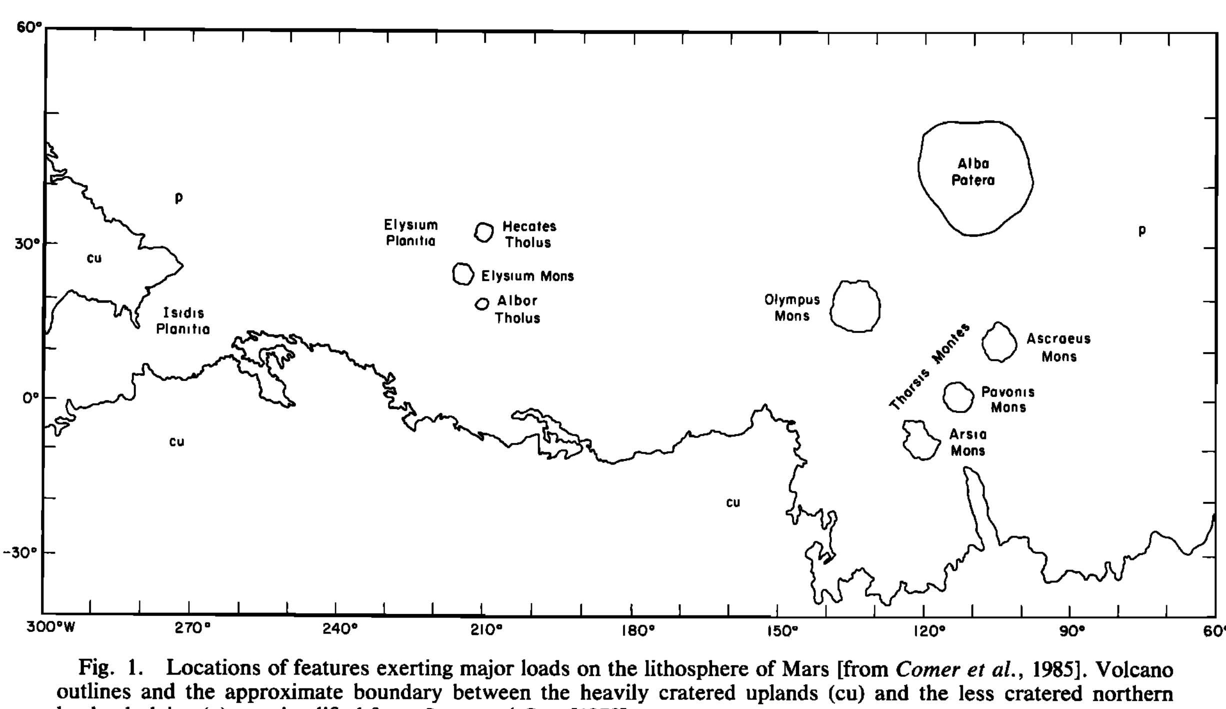

Fig. 1. Locations of features exerting major loads on the lithosphere of Mars [from *Comer et al.*, 1985]. Volcano outlines and the approximate boundary between the heavily cratered uplands (cu) and the less cratered northern lowland plains (p) are simplified from *Scott and Carr* [1978].