modest areas immediately surrounding localized lithospheric loads. The areas within major volcanic provinces for which the enhanced thermal gradients of Table 1 and Figure 7 are appropriate, unfortunately, are poorly constrained. Nonetheless, an approximate measure of the lateral extent of enhanced lithospheric heating is the area of regionally elevated topography arising either from remaining excess heat or, more likely, from a crust permanently thickened by the addition of volcanic and intrusive material. We assume that the areas of excess thermal gradient in the central Tharsis, central Elysium, and Alba Patera regions are 3×10^6 km^2, 0.9×10^6 km^2, and 0.8×10^6 km^2, respectively. These areas correspond to the regions encompassed by the 7.5-km, 1-km, and 5-km elevation contours in the respective provinces [*U.S. Geological Survey*, 1989]. The 7.5-km contour in the Tharsis region encloses the area of the Tharsis Montes, but not Olympus Mons, for which a lithosphere thickness and thermal gradient nearer to the global average values are indicated (Table 1). For excess thermal gradients of about 10 K km^{-1} in central Tharsis and 5 K km^{-1} in the other two volcanic provinces (Table 1) and a crustal thermal conductivity of 2.5 W m^{-1} K^{-1} the heat delivered by these excess gradients is at a rate of 1×10^{11} W, with 80% of the contribution to this figure coming from the Tharsis region. This figure is 3–5% of the estimated present global heat loss, $2\text{–}4 \times 10^{12}$ W for the probable range of average heat flux values given above. By way of comparison the fraction of mantle heat flux delivered by plumes at present on Earth has also been estimated to be about 5% [*Davies*, 1988; *Sleep*, 1990]. The heat delivered by plumes on Mars must be supplied by cooling of the Martian core. In the thermal history models of *Stevenson et al.* [1983] the Martian core delivers heat to the mantle at a rate of 2×10^{11} W during the Amazonian, sufficient to supply the estimated heat transported by plumes beneath major volcanic provinces.

We may compare these figures to the fractional heat flow delivered by volcanism and associated igneous intrusions. Estimates of the surface area of volcanic material at each major stratigraphic stage, including corrections for later burial, have been given by *Greeley* [1987] and *Tanaka et al.* [1988]. Both find 2×10^8 km^2 of volcanic material, though the two analyses differ in detail, particularly in the relative strength of a "peak" in the flux curve at early Hesperian times (corresponding to the formation of the Martian ridged plains) about 3–3.5 b.y. ago [*Tanaka*, 1986]. *Greeley* [1987] has suggested that the volume of volcanic material may be estimated by multiplying the area by an average thickness of about 1 km. A volume V of 2×10^8 km^3 may be converted to equivalent heat Q by means of the relation $Q = \rho(C_p\Delta T + \Delta H)V$, where ρ is the density of volcanic material, C_p is the specific heat, ΔT is the difference between eruption and ambient temperatures, and ΔH is the heat of fusion. We take $\rho = 3$ Mg m^{-3}, $C_p = 1.2$ kJ kg^{-1} K^{-1}, and $\Delta H = 0.4$ MJ kg^{-1}, and we adopt $\Delta T = 1450$ K [*Bertka and Holloway*, 1988]. Averaged over 3.8 b.y., 2×10^8 km^3 of volcanic material delivers only 10^{10} W, or less than 1%, of the estimated present global heat loss of $2\text{–}4 \times 10^{12}$ W. Accompanying any volcanic eruption, of course, is usually a significant intrusion of magma that cools and solidifies at depth. The ratio of intruded to extruded volumes is as great as 10:1 in terrestrial eruptions [*Crisp*, 1984]. The combined volumes of volcanic and intrusive material on Mars, for this ratio, would have delivered heat at an average rate of 3–6% of the total global heat loss. Note that the intrusive component of igneous activity may find expression in the inferred conductive gradient, depending on the depth of intrusion, but the volcanic component of activity will not.

The volcanic flux has decreased sharply with time over Martian history [*Greeley*, 1987; *Tanaka et al.*, 1988], so it is worth making a similar calculation both for the periods of high volcanic output as well as for the Amazonian epoch corresponding to most of the estimates of lithosphere thickness and thermal gradient given in Table 1 and Figure 7. For both the mid-Noachian and early Hesperian epochs, volcanism resurfaced large areas in widespread regions over the planet. With the areas of volcanic material given by *Tanaka et al.* [1988] for these periods, the 1-km thickness estimate of *Greeley* [1987], and the time intervals given by *Tanaka* [1986], the heat delivered by volcanism was at rates of 2×10^{11} and 3×10^{10} W, respectively. For a 10:1 ratio of intrusive to erupted material the heat flux delivered by magmatism during the mid-Noachian may have been as much as one third of the global heat flux, with the precise ratio depending on mantle heat production, global thermal history, and the uncertain duration of the epoch. In contrast, the heat delivered by volcanism and plutonism estimated by this same procedure for the Amazonian contributed on average no more than 4×10^{10} W, a figure equal to about 40% of the excess conducted heat at major volcanic centers, estimated above, and only 1–2% of the global heat loss.

CONCLUSIONS

Estimates of the thickness of the effective elastic lithosphere of Mars appropriate to a variety of locations and times have been converted to estimates of lithospheric thermal gradient and surface heat flow by means of strength envelope considerations. Local thermal gradients and heat flow values were 10–14 K km^{-1} and 25–35 mW m^{-2} at the time of formation of load-induced graben surrounding the Tharsis Montes and Alba Patera, while gradients and heat flow values of less than 5–6 K km^{-1} and 17–24 mW m^{-2} characterized the lithosphere beneath the Isidis mascon and Olympus Mons at the time of emplacement of these loads. On the basis of the thickness of the global elastic lithosphere required to support the Tharsis rise, inferred on thermal grounds to characterize a late, rather than an early, stage in the evolution of the Tharsis province [cf. *Sleep and Phillips*, 1985], as well as heat production estimates obtained from SNC meteorites, we suggest that the present global heat flux on Mars is in the range 15–25 mW m^{-2}. Approximately 3–5% of this heat flux during the Amazonian epoch has been contributed by excess conducted heat in the central regions of major volcanic provinces, a figure at least a factor of 3 greater than the heat transported solely by volcanism and shallow igneous intrusions. This excess heat flux is plausibly attributed to the action of mantle plumes on the base of the lithosphere beneath volcanic province centers. The fractional mantle heat transport contributed by plumes during the last 2 b.y. on Mars is therefore comparable to the present situation on Earth.

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