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CONVECTIVE THINNING OF THE LITHOSPHERE: A MECHANISM FOR RIFTING AND MID-PLATE VOLCANISM ON EARTH, VENUS, AND MARS

by

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ABSTRACT

Thinning of the Earth's lithosphere by heat advected to its base is a possible mechanism for continental rifting and continental and oceanic mid-plate volcanism. It might also account for continental rifting-like processes and volcanism on Venus and Mars. Earth's continental lithosphere can be thinned to the crust in a few tens of million years by heat advected at a rate of 5 to 10 times the normal basal heat flux. This much heat is easily carried to the lithosphere by mantle plumes. The continent is not required to rest over the mantle hot spot but may move at tens of millimeters per year. Because of the constant level of crustal radioactive heat production, the ratio of the final to the initial surface heat flow increases much less than the ratio of the final to initial basal heat flow. For large increases in asthenospheric heat flow, the lithosphere is almost thinned to the crust before any significant change in surface heat flow occurs. Uplift due to thermal expansion upon thinning is a few kilometers. The oceanic lithosphere can be thinned to the crust in less than 10 million years if the heat advection is at a rate around 5 or more times the basal heat flow into 100 Ma old lithosphere. Uplift upon thinning can compensate the subsidence of spreading and cooling lithosphere. The swells will rise to a common depth of about 4 km below sea-level if the uppermost mantle beneath the mid-oceanic ridge is partially melted to a slightly larger degree than the uppermost mantle beneath the swells. Uplift upon thinning on Mars is up to 10 km or more, thus explaining the height of volcanic regions like Tharsis. On Venus, uplift upon thinning would range from 0.5 - 4 km, in agreement with observed topographic variations.

INTRODUCTION

On Earth, almost all volcanic activity is confined to the plate margins. However, scattered around the globe on oceanic and continental plates are regions of isolated volcanic activity called hot spots. These surface hot spots are manifestations of mantle hot spots. Insofar as geologically recent plate motions are concerned the hot spots are quasi-stationary with respect to the lower mantle (Minster et al., 1974; Solomon and Sleep, 1974; Kaula, 1975). Consequently, when lithospheric plates pass over mantle hot spots, swells and sometimes trails of volcanoes are left behind. Examples on oceanic plates include the Hawaiian and Emperor swells and island and seamount chains, the Austral and Tuamotu swells and islands, and the Iceland transverse ridge. On continents, hot spot tracks are more difficult to identify because continental plates move slower and are thicker than oceanic plates. Morgan (1981) succeeded in extending hot spot tracks from the Atlantic into the Americas and Africa. Hot spots on continents are often accompanied by dome or plateau uplift and continental rifting (see the hot spot map recently provided by Vogt (1981)). Prominent examples include the East African dome and rift zone, the Tibesti Plateau, the German Rheinische Schild and the Rheingraben Rift, the Rio Grande Rift and the Colorado Plateau, the Siberian Platform and the Baikal Rift.

Tectonic structures that resemble continental domes, plateaus, and rifts on Earth are also found on Venus and Mars. On Venus, doming and rifting is thought to have caused the uplifted Beta and Phoebe Regiones which have been compared to the topographically similar Kenyan and Ethiopian domes (Malin and Saunders, 1977; Wood and Head, 1978; Masursky et al., 1980; Phillips et al., 1981; McGill et al., 1981). Two more rift-type linears have been recognized from Pioneer-Venus radar altimetry by Schaber (1982): the "Aphrodite-Beta linear" extending northeast from Aphrodite Terra to Beta Regio and the "Themis-Atla linear" trending northwest from east of Themis Regio to the northeastern tip of Aphrodite Terra (the location of Atla Regio). Schaber (1982) suggests that Beta and Atla Regiones are dynamically supported volcanic terranes located at the intersections of the currently active major Venusian rift systems. Vigorous volcanic activity has shaped the surface of Mars throughout most of its history (Greely and Spudis, 1981). There are large uplifted plateaus and huge shield volcanoes. The Tharsis plateau is the site of Olympus Mons, perhaps the largest shield volcano in the solar system. There is also a huge canyon complex, Valles Marineris, that resembles the East African Rift system.

It has been proposed that hot spots are the surface manifestations of mantle plumes, possibly of deep origin (Wilson, 1963a,b; Morgan, 1971). Other causes for hot spots may be thermal runaway in asthenospheric shear flow (Melosh and Ebel, 1979) or the subduction of active ridges under continents (Damon, 1979). Yuen and Peltier (1980) propose that mantle plumes could rise from the unstable core-mantle boundary layer of a whole mantle convection cell. This origin of the plumes may also explain the geochemical peculiarities of hot spot volcanism (Hofmann and White, 1982; Spohn and Schubert, 1982a). According to Anderson (1981), plumes could originate in the upper 220 km of the mantle. He suggests that thermal blanketing by the thick continental lithosphere could heat and partially melt the upper asthenosphere causing upwelling of the partially melted rock.

Steady-state models of mantle plumes have been developed by Parmentier et al. (1975) and Yuen and Schubert (1976). Plumes can have upward velocities as large as meters per year and advect heat at significant rates, up to 50 times the basal heat flux into the lithosphere. Where a plume impinges on the lithosphere the excess heat causes the lithosphere to thin. Heating and replacement of lithosphere by hot plume material cause uplift due to thermal expansion and isostatic equilibration. On continents, a dome may form and rifts may develop in the dome (Burke and Wilson, 1976; Neugebauer, 1978). Eventually, the continent may split along a line of connecting rift arms of several domes. It has been shown by Spohn and Schubert (1982b) that the continental lithosphere can be thinned to the crust on the time scale of tens of million years which is usually associated with continental rifting. Other authors have also attributed continental rifting to lithospheric thinning by enhanced mantle heat flow (e.g. Crough and Thompson, 1976a; Gass et al., 1978).

As for the oceanic lithosphere, Detrick and Crough (1978) and Crough (1978) have shown that the subsidence histories of oceanic mid-plate swells can be explained by a model in which the lithosphere is reheated and thinned as it passes over a mantle hot spot. However, using the thinning model of Crough and Thompson (1976a) and a heat flux anomaly of 2 times normal, they found that it would take around 100 Ma to conduct enough heat into the lithosphere to lift the Hawaiian swell to its maximum observed height while the swell originates in only a few million years. Our results are in accord with this conclusion (Fig. 5). Thinning rates from Crough and Thompson's (1976a) model are typically even higher than those computed from our model (Spohn and Schubert, 1982b). Thus, higher heat flux anomalies are required. We will show that a factor of 10 increase will suffice. Observations of conductive heat flow at the surface of the oceanic crust are not a good guide to the variations of heat flux at lithospheric depths. High heat flux anomalies at depth may never be observed as conductive heat flux at the surface. For one thing, it will take 20 Ma or more before the thermal anomaly reaches the surface because of conductive heat transfer across the thinning lithosphere. Furthermore, a considerable part of the advected energy from the plume will be consumed by magma formation upon pressure release melting once the plume reaches a shallow depth. A heat flux of 1 W m^{-2} can be consumed by a layer of magma growing at the rate of 20 mm a^{-1}. This heat will eventually reach the surface via volcanism. Another efficient mode of heat extraction is hydrothermal convection from magmatic intrusions.

Turning to Mars and Venus, we note that uplift of the Tharsis region on Mars can be isostatically compensated if the crust is underlain by a mantle of anomalous low density (Sleep and Phillips, 1979). Isostatic compensation of uplift on Venus is possible if asthenospheric density anomalies are kept from relaxing (Phillips et al., 1981).

In this paper we apply our convective thinning model (Spohn and Schubert, 1982b) to the oceanic and continental lithospheres of Earth and to the lithospheres of Venus and Mars. We begin by briefly reviewing the model and its general consequences. Some new results are presented for the continental lithosphere. We then proceed to discuss the convective thinning of the oceanic lithosphere and show that thinning to the crust can occur in only a few million years. We give an explanation of why swells rise to a depth of about 1800 m below the mid-oceanic ridge crest. Finally, we apply our model to Mars and Venus.

THE MODEL

The model consists of a horizontally infinite lithosphere of thickness ℓ_0, uniform density ρ, specific heat c, and thermal conductivity k. The lower boundary of the lithosphere, the lithosphere-asthenosphere boundary LAB, is a rheological boundary located at a depth where the viscosity is $O(10^{21}$ Pa s), the characteristic magnitude of the sublithospheric mantle viscosity (Cathles, 1975; Peltier, 1981). The viscosity of mantle rock is strongly temperature dependent and the LAB is conveniently marked by an isotherm or a given fraction of the mantle solidus. We assume that the LAB temperature T_ℓ varies linearly with depth z according to

$$T_\ell(z) = m_\ell z + T_{\ell s} \quad . \tag{1}$$

It is also possible that the solidus is reached at the LAB. There is evidence for incipient melting in the upper mantle (Anderson and Sammis, 1970).

The upper layer of the lithosphere is the crust of thickness z_0. Radiogenic heat production is assumed to be confined to this layer; it varies with depth according to

$$A(z) = A_0 \exp(-\frac{z}{D}) \quad , \tag{2}$$

where A_0 is the surface volumetric heat production rate and D is a characteristic depth for the near surface concentration of radioactive elements (Lachenbruch and Sass, 1977). Initially, the lithosphere is at thermal equilibrium with the temperature $T_0(z)$ given by

$$\begin{aligned} T_0(z) &= \frac{q_i}{k} z + \frac{D^2}{k} (A_0 - A(z)) \quad , \quad 0 \le z \le z_0 \quad , \\ &= \frac{q_i}{k} (z - z_0) + T_0(z_0) \quad , \quad z_0 \le z \le \ell_0 \quad , \end{aligned} \tag{3}$$

where q_i is the basal heat flux from the asthenosphere. The quantities q_i, A_0, and D are related to the surface heat flux q_{si} by

$$q_{si} = q_i + DA_0 \quad . \tag{4}$$

It can be easily demonstrated from (1) and (2) that

$$\ell_0 = k \left(\frac{T_{\ell s} - T_R}{q_i - km_\ell} \right) \quad , \tag{5}$$

where

$$T_R \equiv \frac{D^2}{k}\,(A_0 - A(z_0)) \tag{6}$$

is the temperature rise across the crust due to radiogenic heat production.

At $t = 0$ we suddenly increase the heat flux from the asthenosphere from q_i to q by placing a plume beneath it, for example. Thermal equilibrium is perturbed and the LAB starts to rise to a new equilibrium position corresponding to a final lithospheric thickness ℓ_f given by

$$\ell_f = k\left(\frac{T_{\ell s} - T_R}{q - km_\ell}\right) \quad . \tag{7}$$

The space left by the boundary will be immediately filled by the plume. The rising lithosphere is underlain by a stagnation point boundary layer (Schlichting, 1960) across which temperature increases by $T_p - T_\ell$. It is also assumed that T_p is a linear function of depth

$$T_p = m_p z + T_{ps} \quad . \tag{8}$$

We study the time-dependent position of the LAB along the plume centerline $x = 0$, where x is horizontal distance. Here, $\frac{\partial T}{\partial x} = 0$ both within the plume and the stagnation point boundary layer. Also, $\frac{\partial^2 T}{\partial x^2} << \frac{\partial^2 T}{\partial z^2}$ within the lithosphere. Thus, the LAB at $x = 0$ will be horizontal and its upward velocity will not depend on derivatives with respect to x. Heat is transported across the thinning lithosphere purely by conduction. Temperature along $x = 0$ in the lithosphere therefore satisfies

$$\rho c\,\frac{\partial T}{\partial t} = k\,\frac{\partial^2 T}{\partial z^2} + A(z) \quad , \tag{9}$$

and is subject to the initial and boundary conditions

$$T = T_0(z) \text{ at } t = 0, \quad T = 0 \text{ at } z = 0, \quad T = T_\ell(z_\ell) \text{ at } z = z_\ell(t) \quad , \tag{10}$$

where $z_\ell(t)$ is the position of the LAB at time t.

Energy conservation at the LAB provides an equation for its motion. If the base of the lithosphere moves upward a distance δz_ℓ in time δt, then the quantity of heat $-\rho c(T_p(z_\ell) - T_\ell(z_\ell))\,\delta z_\ell$ per unit horizontal area must be provided to allow the plume with temperature T_p to follow the upward motion of the lithosphere base. If the LAB is at the solidus and there is some partial melt, an amount of heat $-\rho L\delta z_\ell$ per unit horizontal area must also be supplied, where L is the latent heat. This energy is provided by the difference between the upward heat flux into the base of the lithosphere q and the upward heat flux out of the base of the lithosphere $k(\frac{\partial T}{\partial z})_{z=z_\ell}$. Thus we can write

$$\rho\left\{L + c(T_p(z_\ell) - T_\ell(z_\circ))\right\}\frac{dz_\ell}{dt} = -q + (k\frac{\partial T}{\partial z})_{z=z_\ell} \quad , \quad z_\ell \geq z_0 \quad . \tag{11}$$

Equations (9) through (11) constitute a nonlinear boundary value problem which we solve numerically using a finite difference scheme. Convergence has been tested by decreasing the spacing of the grid points. The Courant criterion has been used to guarantee stability of the solutions.

Uplift due to thermal expansion of the thinning lithosphere and replacement of lithosphere by hot plume material can be calculated from our model by balancing the masses in lithospheric columns before and after thinning. After the LAB has reached its final equilibrium position isostatic uplift in continental regions is given by

$$u = \Delta\ell_f - \frac{\Delta\rho_p}{\rho_{\ell o} + \Delta\rho_p}(\ell_o - \ell_f) \quad , \tag{12}$$

where $\Delta\ell_f$ is the expansion of the thinned lithosphere due to heating

$$\Delta\ell_f = \frac{1}{2}\alpha\frac{(q - q_i)}{k}\ell_f^2 \quad , \tag{13}$$

and α is the volumetric coefficient of thermal expansion. The average density difference $\Delta\rho_p$ between lithosphere prior to heating and plume material in $\ell_o - \ell_f$ is given by

$$\Delta\rho_p = \rho_{\ell o}\,\alpha\,(T_o(\ell^*) - T_p(\ell^*)) - \Delta\rho_m \quad , \tag{14}$$

where $\rho_{\ell o}$ is the average density of the initial lithosphere, $\Delta\rho_m$ is the density change due to partial melting, and

$$\ell^* \equiv \frac{1}{2}(\ell_o + \ell_f) \quad . \tag{15}$$

The isostatic uplift in oceanic regions is

$$u = \frac{\rho_o + \Delta\rho_p}{\rho_o + \Delta\rho_p - \rho_w}\,\Delta\ell_f - \frac{\Delta\rho_p}{\rho_o + \Delta\rho_p - \rho_w}\,(\ell_o - \ell_f) \quad , \tag{16}$$

where ρ_w is the density of seawater.

SOLUTIONS TO THE MODEL EQUATIONS

The time required to thin the lithosphere to a given fraction of its initial thickness was determined by integrating (9) - (11) using a finite difference scheme. Figure 1 gives relative thinning $(\ell_c - z_\ell)/\ell_o$ as a function of the square-root of dimensionless time

$$\tau \equiv \frac{k}{\rho c \ell_o^2}\,t \quad , \tag{17}$$

for two examples. Lithosphere thickness decreases with the square-root of time for a large fraction of the total thinning time. Near the equilibrium states, the $\sqrt{t}$ - law is not valid. (In the related classical Stefan problem of a freezing half-space the thickness of the solid layer increases with the square-root of time.) We approximate $(\ell_o - z_\ell)/\ell_o$ as a function of time by

$$\frac{\ell_o - z_\ell}{\ell_o} = \begin{cases} \dfrac{0.95\,(\ell_o - \ell_f)}{\ell_o}\left(\dfrac{t}{t_{95}}\right)^{1/2} , & 0 \le t \le \left(\dfrac{1}{0.95}\right)^2 t_{95} \\ \dfrac{\ell_o - \ell_f}{\ell_o} , & t > \left(\dfrac{1}{0.95}\right)^2 t_{95} \end{cases} , \quad (18)$$

where t_{95} is the time to accomplish 95% of the total thinning. Its dimensionless equivalent τ_{95} is shown in Figure 2 as a function of the dimensionless parameters

$$\lambda \equiv \frac{\ell_o - \ell_f}{\ell_o} = \frac{q/k - q_i/k}{q/k - m_\ell} , \quad (19)$$

and

$$\Delta \equiv \frac{1}{(T_{\ell s} - T_R)}\left\{\frac{L}{c} + T_p(\ell^*) - T_\ell(\ell^*)\right\} \quad (20)$$

The parameter λ gives the total amount of thinning attainable and depends on three temperature gradients: the initial subcrustal temperature gradient q_i/k, the final subcrustal temperature gradient q/k, and the slope of the transition temperature profile. The parameter Δ is a measure of the total energy required for thinning from ℓ_o to ℓ_f. It depends on the average temperature rise $T_p(\ell^*) - T_\ell(\ell^*)$ across the stagnation point boundary layer, on the transition temperature at surface pressure, on the temperature rise due to radioactive heat production in the crust, and on the ratio of the latent heat to the heat capacity. Figure 2 shows that the thinning time is proportional to Δ and to $(1 - \lambda)^n$, with n slightly larger than 2.

Figure 1 also gives the ratio q_s^* of the increase in surface heat flow $q_s - q_{si}$ (q_s is surface heat flow and q_{si} is the initial value of q_s) to $q_{s\infty} - q_{si}$ the increase in surface heat flow after equilibrium has been reestablished. The ratio q_s^* increases from 0 to 1 as the lithosphere thins from its initial to its final equilibrium thickness. The change in surface heat flow lags behind thinning by an amount which increases with q/q_i. At the time the surface heat flow has increased by 10% of $q_{s\infty} - q_{si}$ the lithosphere has thinned by 50% for $q = 2q_i$ and by 90% for $q = 10q_i$.

DISCUSSION

We now apply our model to thinning of the continental and oceanic lithospheres on Earth and to thinning of the lithospheres of Venus and Mars. Geological observables will be calculated and inferences from the model calculations will be discussed.

It is assumed that the material parameters $\rho_{\ell o}$, α, c, k, and L take the same values for each planet and tectonic region. These values are $\rho_{\ell o}$ = 3300 kg m^{-3}, $\alpha = 3.0 \times 10^{-5}$ K^{-1}, c = 750 J kg^{-1} K^{-1}, k = 2.5 W m^{-1} K^{-1}, and L = 4.2×10^5 J kg^{-1} for complete melting. They have been adopted from Stacey (1977).

Two transition temperature profiles will be considered. In the first, the transition temperature is the 1000°C isotherm (models a). This is often thought to mark the transition from quasi-rigid to viscous behavior of mantle rocks over geological time (Oxburgh and Turcotte, 1978; Schubert, 1979) and it is also a reasonable approximation to the wet solidi of mantle rocks from laboratory experiments (Spohn and Schubert, 1982b). In the second, the transition temperature profile (models b) is 0.75 times a representative dry solidus for the mantle given by Chapman and Pollack (1977). Here $T_{\ell s}$ = 1100°C and m_ℓ = 3.1 K km^{-1}.

We assume that the plume temperature T_p increases with depth at a rate of 0.4 K km^{-1} which is approximately the adiabatic temperature gradient in the upper mantle. We further assume that $T_p(\ell_o)$ is 100 K above the transition temperature $T_\ell(\ell_o)$. The average over ℓ_o - ℓ_f of T_p - T_ℓ then takes values between 100 K and 600 K.

Finally, we must consider what are reasonable values for the increased basal heat flux q. Heat flow data for both continents and oceans on Earth vary by roughly a factor of 10 (Sclater et al., 1980). As is evident from (4) heat flow from the asthenosphere contributes only partly to the surface heat flow. A factor of ten increase in basal heat flow changes the surface heat flow by only a factor of five if q_i is approximately equal to DA_o. Mantle plumes can transfer heat at significant rates. A plume with flow velocity of 10 mm yr^{-1} and temperature 100 K hotter than the base of the lithosphere advects heat at the rate of about 1 W m^{-2}, about 50 times the basal heat flow in low heat flow provinces. As discussed in the introduction, this high heat flux might never be observed at the surface in the form of a conductive heat flow. The parameters q_i and T_R depend on the planet and the tectonic region under consideration. For the oceanic crust T_R may be 25 K; for the continental crust it may be approximately 100 K.

THINNING OF THE CONTINENTAL LITHOSPHERE

Data on continental and oceanic heat flow have recently been reviewed by Sclater et al. (1980, 1981). Representative values of the continental heat flow are close to 50 mW m^{-2} in heat flow provinces older than 250 Ma, while in younger provinces the mean is 77 mW m^{-2}. The scatter of data is considerable in young provinces; Sclater et al. (1981) give a standard deviation of 53 mW m^{-2}.

However, it is much smaller in provinces older than 250 Ma. Here the standard deviation is only 15 mW m^{-2}. Taking q_s = 50 mW m^{-2}, A_0 = 2.5 μW m^{-3}, and D = 10 km in (4), we get 25 mW m^{-2} as a representative value for the basal heat flux under continents. Equation (4) follows from (3), but it is also independently empirically valid (Birch et al., 1968). It is used to calculate a reduced surface heat flow from surface heat flow and radioactivity data. Sclater et al. (1981) give 25 mW m^{-2} to 29 mW m^{-2} for the reduced heat flow from continents. The reduced heat flow may be a somewhat crude estimate of the basal heat flow because there is radioactive heat production in the mantle. In addition, the continental lithosphere is still growing as a consequence of the cooling of the Earth and as a result the heat flux across the lithosphere is some percent larger than the heat flux from the underlying mantle (Schubert et al., 1979).

The temperature rise across the crust due to radioactivity is calculated from (6) to be 95 K, assuming a crustal thickness of 30 km. These values of q_i and T_R give initial lithosphere thicknesses ℓ_0 of 91 km (model a) and 146 km (model b) depending on the transition temperature profile assumed. The smaller value of ℓ_0 corresponds to the 1000°C isotherm. These values are on the low side of recent assessments of continental lithosphere thickness. It has been argued that the continental lithosphere might be more than 300 km thick under old shields (e.g. Jordan, 1975; Sipkin and Jordan, 1975; Chapman and Pollack, 1977). However, Sclater et al. (1980) concluded that the continental lithosphere thickness should not significantly exceed 150 km, an estimate in accord with other studies (Sleep, 1971; Crough and Thompson, 1976b; Turcotte and McAdoo, 1979; Blackwell and Chockalingam, 1980).

In order to determine the circumstances under which convective thinning of the continental lithosphere can be effective, we have calculated the time required to thin the lithosphere by 95%, t_{95}, for lithospheres 50 to 300 km thick subject to heat flux increases from $2q_i$ to $30q_i$. These results are given in Fig. 3. A few cases are reported in more detail in Table 1. Additional examples can be found in Spohn and Schubert (1982b). We list t_{95} whenever the completely thinned lithosphere is thicker than the crust. In cases where the final thickness is less than the crustal thickness we give the time to thin to the crust as obtained from (18). The time to thin to 50 km thickness is also listed for those cases in which the final equilibrium thickness is smaller. The values of Δ vary from 0.11 to 0.50. We have assumed that not more than incipient melting is involved.

Very thick lithospheres (ℓ_0 = 300 km) can be thinned to the crust in tens of million years, the time scale usually associated with rifting (Seidler and Jacoby, 1981), if basal heat flow increases by more than a factor of 10. Thinner lithospheres require less of an increase in basal heat flux, a factor of five can suffice. Because the radioactive heat production in the crust does not change substantially with time on this time scale, the increase in surface heat flow is less than the increase in basal heat flow. Surface heat flow data from the rift valleys of contemporary continental rifts have been compiled by P. Morgan (1981). The values range from 70 to 125 mW m^{-2}. This is somewhat less than the surface heat flow of 139 mW m^{-2} which obtains upon thinning the 100 and 150 km lithospheres of Table 1 with a factor of 5 increase in basal heat flow. In these cases the lithospheres are thinned to 50 km in only 19 and 37 Ma. The smaller observed heat flows can be understood by the time lag between the increase in surface heat flow and thinning (Fig. 1) and by the consumption of part of the advected heat by partial melting once the lithosphere is substantially thinned. This heat will eventually reach the surface via volcanism.

The uplifts given in Table 1 correspond to lithospheres thinned to their crusts or to their final equilibrium thicknesses, whichever are larger. Uplift increases with ℓ_0; it also increases with q/q_i provided that ℓ_f is larger than z_0. It ranges from 1 or 2 km for low q/q_i and small ℓ_0 to several km for the larger values of q/q_i and ℓ_0. The uplifts given here differ from those given in Spohn and Schubert (1982b) because here we have considered isostatic equilibration with ℓ_0 as the depth of compensation while in the earlier paper we considered uplift solely due to thermal expansion. Seidler and Jacoby (1981) give topographic heights of rift shoulders for many contemporary rift structures. With the exceptions of the Godawari rift, the Hon Graben, and the Rhine Graben, where uplifts are less than 1 km, uplifts range from 1 to 4 km. These values agree with the ones given in our earlier paper and those given here for $\ell_0 \leq 150$ km. We have not accounted for erosion, which can occur at rates comparable to the calculated average uplift rates. For example, with $\ell_0 = 150$ km and $q = 5q_i$ we get an average uplift rate of 0.06 km Ma^{-1}. Allowance for erosion would reduce the calculated uplifts and improve the fit to observation. However, our results argue against a 300 km thick continental lithosphere on Earth. It would not be easy to explain why plumes would avoid such thick parts of the lithosphere if they existed. However, Mars' lithosphere is likely to be around 300 km thick (see below) and uplifts around 10 km are observed at Tharsis and other volcanic regions.

VULNERABILITY OF THE CONTINENTAL LITHOSPHERE

Our model does not account for the lateral movement of the lithospheric plate with respect to the hot spot below. It has been argued by Gass et al.

(1978) and by Pollack et al. (1981) that the heat carried to the plate might be swept away by plate motion thus precluding significant thinning. Pollack et al. (1981) define a 'vulnerability' parameter V

$$V = \left(\frac{\pi u \rho c}{dk}\right)^{\frac{1}{2}} \ell_o \quad , \quad (2$$

where u is the plate velocity and d is the characteristic spacing of hot spots. Essentially, they compare the propagation length $(dk/\pi u \rho c)^{\frac{1}{2}}$ in time d/u of a thermal wave to the initial thickness of the lithosphere. This vulnerability parameter is derived from the conductive model of Gass et al. (1978) which follows the upward propagation of a temperature disturbance at the base of the lithosphere. While the vulnerability parameter varies from less than 1 to 14 for the Earth's lithosphere, Pollack et al. (1981) show that 80% of the Earth's hot spot population is located where V is less than 5, i.e. where the propagation length of the thermal wave in time d/u is more than 20% of the initial thickness.

We argue that the lithosphere should be more vulnerable to convective thinning than to conductive thinning. The relevant convective time scale t_{95} can be two orders of magnitude smaller than the conductive time scale for $\lambda \geq 0.85$ (Fig. 2). Thus the amount of thinning attainable in d/u should be considerably larger for convective models as compared with conductive ones. Taking 10^6 m as an order of magnitude value for d and 10^{-2} m a^{-1} as one for u we find that d/u is 10^8 years. This is a factor of 10 larger than the time scale we obtain for our successful models. Moreover, the lithosphere is thinned by more than 50% in this time The finding of Pollack et al. (1981) that hot spots are preferentially located in regions where V is relatively small together with the ten million year time scale associated with mid-plate volcanism and rifting suggest that the heat flux from asthenospheric hot spots is not highly anomalous. This is because we would expect the hot spots to be more evenly distributed over the Earth's surface if plumes advected much more than ten times the basal heat flux. The variations of heat flux into the lithosphere need not be much larger than the variations in the surface heat flux which are generally less than a factor of 10 (Sclater et al., 1981).

THINNING OF THE OCEANIC LITHOSPHERE

Heat flow data from the continents suggest that the mean basal heat flow is approximately constant for provinces older than 250 Ma. Mean oceanic surface heat flow values from well sedimented areas, on the contrary, decrease monotonically with crustal age. The data compiled by Sclater et al. (1980, 1981) fit the empirical relation

$$q_s \left[\mathrm{mW\ m^{-2}}\right] = 509.3\ t^{-\frac{1}{2}} \left[\mathrm{Ma}\right] \tag{22}$$

for crustal ages less than 200 Ma. If we assume that the upper 5 km of the oceanic lithosphere is a basaltic crust with a surface radioactive heat production rate per unit volume A_0 of 0.5 μW m^{-3} and a characteristic depth of heat production D equal to 5 km, we get a contribution to q_s due to radioactivity of 5 mW m^{-2}. The basal heat flow q_i is obtained by subtracting 5 mW m^{-2} from q_s. The temperature increase T_R due to crustal radioactivity is 25 K.

Again, we use the 1000^{o}C isotherm and 0.75 times the generalized dry solidus from Chapman and Pollack (1977) to denote the lithosphere base. Initial lithosphere thickness ℓ_0 as a function of age and q_s is given in Fig. 4. At age 50 Ma lithosphere thicknesses are 36 and 43 km for the respective transition temperature models. The thicknesses increase to 79 and 111 km at an age of 200 Ma. For transition temperature model a (1000^{o}C isotherm), the amount of lithospheric thinning depends only on q/q_i, i.e. $\lambda = (q - q_i)/q$. Thus, the lithosphere can be thinned by 95% if $q = 20q_i$, by 90% for $q = 10q_i$, by 80% for $q = 5q_i$, and by 50% for $q = 2q_i$. For transition temperature model b, $k\ m_\ell$ = 7.8 mW m^{-2} and λ depends on both q/q_i and q_i (see (19)). The dimensionless amount of thinning is 0.91 for $q = 10q_i$ and q_i = 67 mW m^{-2}, the basal heat flux into 50 Ma old lithosphere. This increases to 0.92 for q_i = 31 mW m^{-2}, the basal heat flux into 200 Ma old lithosphere. For $q/q_i = 5$, λ increases from 0.82 at 50 Ma to 0.84 at 200 Ma, while for $q/q_i = 2$, λ increases from 0.53 to 0.57. If $q = 20\ q_i$, λ is 0.96. The parameter Δ is roughly constant for model a at the value 0.10. For model b, Δ increases with both ℓ_o and q/q_i. Its maximum value here is 0.23 for ℓ_o = 111 km and q/q_i = 20 and its minimum value is 0.13 for ℓ_o = 43 km and q/q_i = 2. Again, we have assumed that not more than incipient melting is involved.

Figure 5 gives t_{95}, the time to accomplish 95% of the total thinning, as a function of the initial thickness ℓ_o and q/q_i for both transition temperature models. The thinning time increases with ℓ_o and decreases with q/q_i. It is essentially independent of the particular transition temperature profile. A comparison with Fig. 3 shows that t_{95} is also essentially independent of the thickness of the crust and the temperature increase due to radioactivity across it. A 100 km thick oceanic lithosphere can be thinned by 90% of its initial thickness in 16 Ma if the basal heat flux is increased by a factor of 10. This is the same as the time needed to thin an equally thick continental lithosphere subject to the same heat flux increase.

Taking a different approach, one may study lithospheric thinning over a plume delivering a constant heat flux q to the base of the oceanic lithosphere.

The ratio q/q_i is then a function of the age of the lithosphere rather than a constant parameter as in the previous discussion. Recalling that $\tau_{95} \sim (1 - \lambda)^n$ with $n \approx 2$ and $(1 - \lambda) = \ell_f/\ell_0$, we find that

$$\tau_{95} \sim q^{-2} = \text{constant} \tag{23}$$

for variations of $(1 - \lambda)$ that are not too large. As an example we choose $q = 250$ mW m^{-2}, which is the heat flow into 4 Ma old lithosphere. It also is about 5 times the heat flow into 100 Ma old oceanic lithosphere and ten times the heat flux into the continental lithosphere as discussed in the previous section. For ages of the oceanic lithosphere between 50 and 200 Ma, q/q_i then varies between 3.7 and 8.1. The oceanic lithosphere overriding such a plume will be thinned by 95% of its initial thickness in about 10 Ma.

Oceanic plates generally move faster than do plates that are largely made up of continental lithosphere. Thus the time scale d/u may be up to an order of magnitude smaller for oceanic plates than for continental plates (Pollack et al., 1981). Thinning of the oceanic lithosphere must occur at least as fast as d/u. In the case of the Hawaiian island chain the plate velocity is 100 mm a^{-1}, and d/u is about 10 Ma. Crough (1978) has shown that the Hawaiian swell did rise to its maximum height in 6 to 12 Ma. Slower moving plates like the Atlantic plate may not require such fast thinning times. It should be noted that Hawaii is anomalous because it is located in a region less vulnerable to thinning than the locations of 96% of all hot spots (Pollack et al., 1981). We conclude that the oceanic lithosphere, like the continental lithosphere, can be thinned in reasonable times by plumes delivering less than 10 times the undisturbed basal heat flux. In fact, summarizing the results for the continental and oceanic lithospheres we note that mantle plumes need not advect more than 250 mW m^{-2}. This is 25% of the heat carried by steady state models of mantle plumes (Parmentier et al., 1975; Yuen and Schubert, 1976).

UPLIFT OF THE OCEANIC LITHOSPHERE OVER HOT SPOTS

Uplift due to thermal expansion and isostatic equilibration of the oceanic lithosphere and replacement of lithosphere by hot asthenosphere is given in Fig. 6. It is assumed that the lithosphere-asthenosphere boundary has reached its final equilibrium position. The depth of isostatic compensation is the initial lithosphere thickness ℓ_0. Uplift increases with q/q_i and with ℓ_0. The uplifts range from about 500 m for q/q_i equal to 2 and young oceanic lithosphere to 3 km (model a) and 4 km (model b) for 200 Ma old lithosphere and $q/q_i = 20$. Uplifts for $q/q_i = 10$ and $q/q_i = 20$ differ by no more than 5%; they are approximately twice as large as uplifts for $q/q_i = 2$. The dotted line in Fig. 6a gives the uplift for the model that has a plume advecting heat at a constant rate of 250 mW m^{-2}. This uplift is close to that calculated for $q/q_i = 5$. It is smaller for ages less than 75 Ma and larger for older lithosphere.

Also given in Fig. 6 is subsidence as a function of age of the oceanic lithosphere relative to the height of the mid-oceanic ridge. This subsidence is a consequence of the cooling and thickening of the oceanic lithosphere as it moves away from the spreading center. On calculating the subsidence curves we have assumed that at the spreading center $\ell_0 = 0$ km and that the asthenosphere there is at T_p. From Fig. 6 it is seen that the reheating and thinning of the lithosphere over a hot spot fully compensates the subsidence if the increase in basal heat flux is at least as large as a factor of 10. If $q/q_i = 5$, less than 400 m would remain uncompensated. Uplift due to reheating and thinning over a plume advecting 250 mW m^{-2} would leave about 300 m of net subsidence.

Crough (1978) has compiled data on mid-plate swells. He points out that the crests of contemporary swells rise to a roughly uniform depth of 4250 m below sealevel, about 1750 m below the crests of the mid-oceanic ridges. This can be explained by postulating different degrees of partial melting beneath the mid-oceanic ridge and the swells. Assume that the density beneath the mid-ocean ridge and the swell is ρ_p except for a layer of thickness a beneath the ridge where the density is less than ρ_p by an amount $\Delta\rho_m$ due to partial melting. If Δs is the height of the ridge above the swell then

$$\Delta s = a \frac{\Delta\rho_m}{(\rho_p - \rho_w)} \quad . \tag{24}$$

In Fig. 7 we present the data compiled by Crough (1978) together with model determinations of the depths to mid-plate swell crests as a function of crustal age. The model assumes that the crust at the mid-oceanic ridge is underlain by a 35 km thick layer of asthenosphere in which the density is 3 to 6% lower than density at the same depth beneath the swell. This density difference can be due to a higher degree of partial melting in the source region of mid-oceanic ridge basalts (MORB) compared to the source region of oceanic island basalts (OIB). It is known that OIB are often richer in alkalis than MORB which can be due to a lesser degree of partial melting of the source rock.

THINNING OF THE LITHOSPHERES OF VENUS AND MARS

Because the thermal evolution of a planet is governed by its surface area to volume ratio, lithosphere thickness should be inversely correlated with planetary size (Kaula, 1981). This suggests that Mars' lithosphere is thicker than both Venus' and Earth's. Thermal history calculations of lithospheric growth have yielded a thickness of about 300 km for Mars (Schubert et al., 1979). The same thickness has been derived from isostatic models of the Tharsis region (Sleep and Phillips, 1979). If the Venusian mantle has the same rheology as the Earth's mantle and if the crusts of the two planets are similar in thickness and radiogenic heat production, then Venus' lithosphere would be thinner than Earth's because of Venus' high surface temperature. However, it is also possible that just the opposite is true. Comparative studies of Venus' gravity field and surface topography show that they are highly correlated (Phillips et al., 1981; Reasenberg et al., 1982) and that Venus' topography may be less compensated than Earth's. Thus, Venus' lithosphere might be thicker and possibly more rigid than Earth's despite the higher Venusian surface temperature. This could be the case if Venus' mantle is very dry compared with Earth's either because Venus had less water than Earth to begin with or it lost its initial water content early in its evolution as a consequence of a runaway greenhouse.

Crustal thicknesses on both Mars and Venus could be larger than on Earth. The crusts of these two planets may be as thick as 100 to 150 km (Wood and Head, 1978; Frey, 1979; Anderson, 1979; Sleep and Phillips, 1979; Phillips and Lambeck, 1980; Phillips et al., 1981; Schaber, 1982). Surkov (1982) has summarized the radiogenic element concentration of Martian and Venusian surface rocks inferred from in situ analysis and remote sensing. There are rocks in the surfaces of both planets with radiogenic element concentrations similar to those of basalts and granites on Earth although the Martian and Venusian concentrations are somewhat lower than typical Earth values. If the planets have the same abundance of heat sources per unit mass then radiogenic heat production scales with their mass. Consequently, if Venus and Mars are not significantly more differentiated than Earth, they should have lower abundances of crustal heat sources integrated over crustal depth than Earth, regardless of their crustal thickness. The temperature rise across the crust T_R of Mars and Venus due to radiogenic heat should then not exceed the value for the continental crust on Earth which is ~100 K. We can account for Venus surface temperature of 475°C by adding this temperature to T_R.

There is no reason to require lithospheric thinning on Mars to have occurred on the relatively fast 10 Ma timescale that is representative of rifting on Earth. There is evidence in the cratering record that the Tharsis region may be 3 Ga old and that volcanic activity has continued to the present or recent past (Wise et al., 1979). Other volcanic provinces like Elysium, Hellas, and Hesperia may also be 1 to 4 Ga old (Carr, 1976). The time scale associated with the growth of these regions could thus be as long as 1 Ga. On the other hand, a shorter time scale is certainly possible and Carr (1976) has suggested that Olympus Mons on Tharsis might have formed in only 200 Ma. Assuming ℓ_o = 300 km, z_o = 30 km, T_R = 100 K and all other parameters equal to their corresponding terrestrial values, we find that Mars' lithosphere can be thinned in hundreds of Ma by relatively modest increases in heat flux, i.e. $q/q_i < 10$ (Table 1, Fig. 3). If the Martian crust were 100 km thick, the lithosphere would be thinned to the crust in less time. Even smaller increases in heat flux would be required if thinning occurred early in the thermal history of Mars. Thermal history calculations suggest that the Martian lithosphere 3 Ga ago, for example, was only about 100 km thick (Schubert et al., 1979). It may have remained thinned under old volcanic provinces ever since their formation.

Turning to Venus, we note that for given q_i and q/q_i thinning would be faster than on Earth if Venus' model parameters are the same as Earth's. The reason is that Venus' lithosphere thickness ℓ_o would be smaller in this case because of its high surface temperature. According to Schaber (1982), the excellent preservation of Venusian rift vallcys and the height of their rims of about 2 km suggest that rift structures on Venus are young and form on a time scale comparable to that of continental rifting on Earth. If Venus has a thick lithosphere with $\ell_0 > 100$ km, large increases in heat flux ($q/q_i > 10$) would be required to thin the lithosphere in tens of Ma. On the other hand, Phillips et al. (1981) note that the Equatorial Highlands within which the rifts are located may be old features. Thus the arguments put forward for Mars may also apply for Venus. The lithosphere under the highlands may have been thinned by plumes of modest strength over times of hundreds Ma or even Ga and it may have remained thinned to the present day. (Note that the decay time of the lithosphere anomaly once the plume is switched off is much larger than the thinning time (Spohn and Schubert, 1982b)). Recent rift structures may then be caused by further thinning and disrupting the pre-thinned lithosphere.

Uplifts for models of lithosphere thinned to the crust on Mars and Venus are given in Tables 2 and 3. The crustal thicknesses are also the final equilibrium thicknesses of the lithosphere. The tables include uplifts for the crustal thicknesses of 30 and 100 km used above. Uplift increases with initial lithospheric thickness ℓ_0 and with q/q_i. It also increases with decreasing z_0. Uplifts are generally much larger for transition temperature models b, partly because q_i is larger for a given ℓ_0 and partly because $T_p(\ell^*) - T_0(\ell^*)$ is larger. (Remember that T_p is chosen so that $T_p(\ell_0) = T_\ell(\ell_0) + 100$ K). The larger value of q_i increases $\Delta\ell_f$ (13). The larger value of $T_p(\ell^*) - T_0(\ell^*)$ increases the absolute value of $\Delta\rho_p$ (14). In the models presented here $\Delta\rho_p$ does not exceed 3% of ρ_p.

Martian uplifts in Table 2 range from 1 to 11 km. They would be about 1.5 times larger if the thermal expansion coefficient of 4.5×10^{-5} K^{-1} were used as in some of the terrestrial models. A further increase of uplift is possible if the plume is partially melted. Density differences due to partial melting can easily increase the absolute value of $\Delta\rho_p$ by another 3% of ρ_p. Hence, the model may explain the height of Tharsis and other volcanic regions of Mars of 10 km above the planet's median level (Carr, 1976; Greeley and Spudis, 1981). The large Martian uplifts are mainly a consequence of Mars' thick lithosphere.

For Venus we calculate uplifts ranging from 500 m to 4 km. These match well the topographic heights of the ridges on Venus that rise 500 m to 2.5 km above the surrounding plateaus (Schaber, 1982). Uplifts on Venus are lower than uplifts on Earth and Mars for given ℓ_0 and q/q_i. The high surface temperature which adds to T_R reduces both $\Delta\ell_f$ and the absolute value of $\Delta\rho_p$ because q_i is smaller (5) and, more importantly for thin crusts, $T_0(z)$ lies closer to the plume temperature profile. Uplifts on Venus would be as large as uplifts on Earth and Mars if the lithosphere-asthenosphere transition temperature in the Venusian mantle were about 500 K larger than in both the terrestrial and Martian mantles.

We conclude from our calculations that the model of convective thinning of the lithosphere can also be successfully applied to structures on Mars and Venus that resemble hot spot regions on Earth.

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ℓ_o (km)	q_{si} (mW m^{-2})	$\frac{q}{q_i}$	$\frac{q_{s\infty}}{q_{si}}$	λ	Δ	t_{95} (Ma)	Thinned to z_o (Ma)	Thinned to 50km (Ma)	Uplift (km)	Transition temperature
300	41.1	33	14.0	0.98	0.50		34	29	7.8	b
		20	8.4	0.97	0.49		59	51	7.8	
		10	4.5	0.95	0.48		123	106	7.8	
		5	2.6	0.88	0.46	342		340	7.8	
		2	1.4	0.66	0.37	1256			5.2	
150	47.9	10	5.3	0.93	0.29		30	21	3.0	b
		5	2.9	0.86	0.27		53	37	3.0	
		2	1.5	0.60	0.22	310			2.1	
100	48.0	10	5.2	0.90	0.12		11	6	1.1	a
		5	2.9	0.80	0.12		38	19	1.1	
		2	1.5	0.50	0.11	158		158	0.8	

TABLE 1. Thinning times and uplift for various models of the Earth's continental lithosphere. The lithosphere-asthenosphere boundary is defined by 0.75 times the dry solidus for the models b and by the 1000°C isotherm for the model denoted by a. q_{si} is the initial surface heat flow and $q_{s\infty}$ is the surface heat flow after equilibrium has been obtained.

ℓ_o (km)	q/q_f	a		b	
		z_o (km)	u (km)	z_o (km)	u (km)
250	10	25	3.4	15	6.6
	8.3	30	3.3		
	5	40	3.1	30	6.1
	2.5	100	2.6		
	2	125	1.2	90	4.0
	1.8			100	3.6
300	10	30	4.1	16	8.6
	5	60	3.7	34	7.8
	3	100	3.1		
	2	150	3.0	102	5.2
350	10	35	4.7	17	11.
	6.1			30	10.
	5	70	4.2	37	10.
	3.5	100	3.8		
	2	175	2.7	113	6.6

TABLE 2. Uplift for models of the Martian lithosphere thinned to the crust. Models a use the 1000°C isotherm to denote the lithosphere base while models b use 0.75 times the dry solidus.

ℓ_0 (km)	q/q_i	a		b	
		z_0 (km)	u (km)	z_0 (km)	u (km)
100	10	10	0.64	6	1.5
	5	20	0.64	10	1.4
	3.3	30	0.63		
	2.3			30	1.1
	2	50	0.47	40	0.91
150	10	15	1.0	8	2.6
	5	30	1.0	17	2.4
	2.9			30	2.1
	2	75	0.67	48	1.7
	1.5	100	0.46		
	1.2			100	0.72
200	10	20	1.4	8	3.9
	6.7	30	1.4		
	5	40	1.4	18	3.6
	3.3			30	3.3
	2	100	0.90	58	2.6
	1.4			100	1.7

TABLE 3. Uplift for models of the Venusian lithosphere thinned to the crust. Models a use the 1000°C isotherm to denote the lithosphere base while models b use 0.75 times the dry solidus. T_R is 575°C, all other parameters are given terrestrial values.

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FIGURE CAPTIONS

Fig. 1 Relative thinning as a function of the square-root of dimensionless time $\tau \equiv (k/\rho c \ell_0^2) \cdot t$ for increased basal heat fluxes $q = 2q_i$ and $q = 10q_i$ (from Spohn and Schubert, 1982b). The relative thinning of the lithosphere as a function of time can be approximated by the indicated ramp functions (light lines). The dashed lines give $q_s^* \equiv (q_s - q_{si})/(q_{s\infty} - q_{si})$ where q_s is the surface heat flow, q_{si} is the initial surface heat flow, and $q_{s\infty}$ is the surface heat flow after equilibrium is re-established. Note that q_s^* lags considerably behind the thinning.

Fig. 2 Dimensionless time $\tau_{95} \equiv (k/\rho c \ell_0^2) \cdot t_{95}$ to achieve 95% of thinning as a function of relative thinning $\lambda \equiv (\ell_0 - \ell_f)/\ell_0$ (a) and Δ (b).

Fig. 3 Thinning of the continental lithosphere. Time t_{95} to achieve 95% of thinning versus initial thickness ℓ_0. The curve parameter is the basal heat flux ratio q/q_i. Circles refer to transition model temperature a wherein 1000°C denotes the base of the lithosphere. Crosses refer to transition temperature model b in which the lithosphere base is denoted by 0.75 times a representative dry mantle solidus.

Fig. 4 Thickness of the oceanic lithosphere versus age and surface heat flux. Model a uses the 1000°C isotherm to denote the lithosphere base while model b uses 0.75 times a representative dry mantle solidus.

Fig. 5 Thinning of the oceanic lithosphere. Time t_{95} to achieve 95% of thinning versus initial lithosphere thickness ℓ_0 with q/q_i as a parameter. Circles refer to transition model temperature a and crosses to model b. The triangles are from the continental lithosphere models of Fig. 3.

Fig. 6 Uplift upon thinning of the oceanic lithosphere for transition temperature models a and b as a function of age and initial lithosphere thickness ℓ_0 with heat flux ratio q/q_i as a parameter. Here models a and b differ also in the volumetric coefficient of thermal expansion which is 4.5 x 10^{-5} for model a. Also given (dashed lines) is the subsidence due to cooling of a column of undisturbed oceanic lithosphere as it moves away from the mid-oceanic ridge. With $q/q_i \geq 10$ uplift upon thinning can fully compensat this subsidence. The dotted line refers to a model in which the plume advects heat at the constant rate of 250 mW m^{-2}.

Fig. 7 Water-depth to the crests of mid-plate swells versus crustal age. The data are from Crough (1978). The crests rise to a common depth of about 4 km. The thin lines indicate the results of model calculations that explain this observation. The model assumes that the crust at the mid-oceanic ridge is underlain by a 25 km thick layer of asthenosphere in which the density is 3 to 6 % lower then the density at the same depth beneath the swell. SH: St. Helena, K: Kerguelen, M: Marquesas, CA: Cook-Austral, R: Reunion, S: Society, H: Hawaii, T: Trinidade, B: Bermuda, CV: Cape Verde.

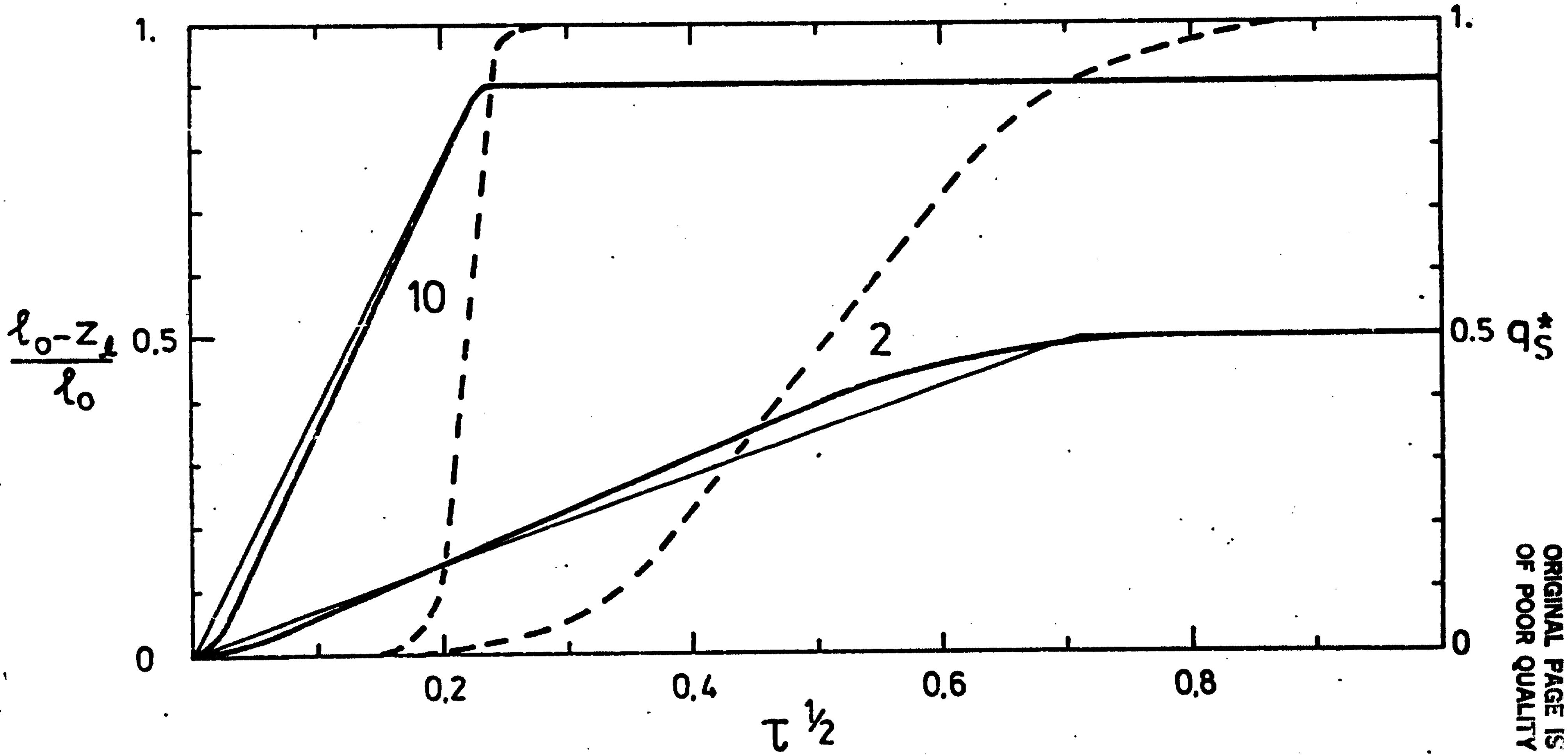
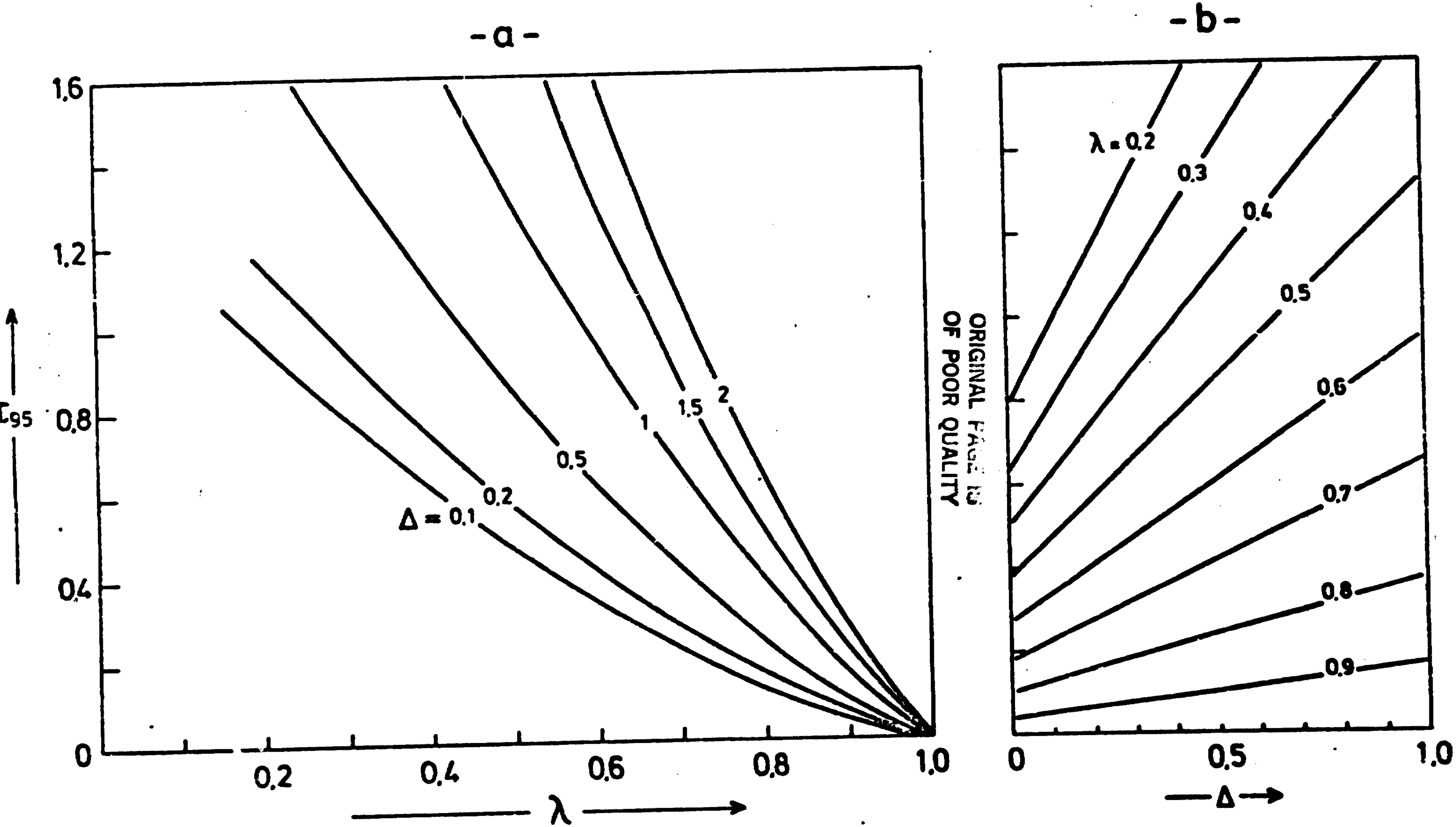


Fig. 1



-a-
1,6
1,2
0,8
0,4
0
τ_{95}
0,2
0,4
0,6
0,8
1,0
λ
Δ = 0,1
0,2
0,5
1
1,5
2
-b-
λ = 0,2
0,3
0,4
0,5
0,6
0,7
0,8
0,9
0
0,5
1,0
Δ

Fig. 2

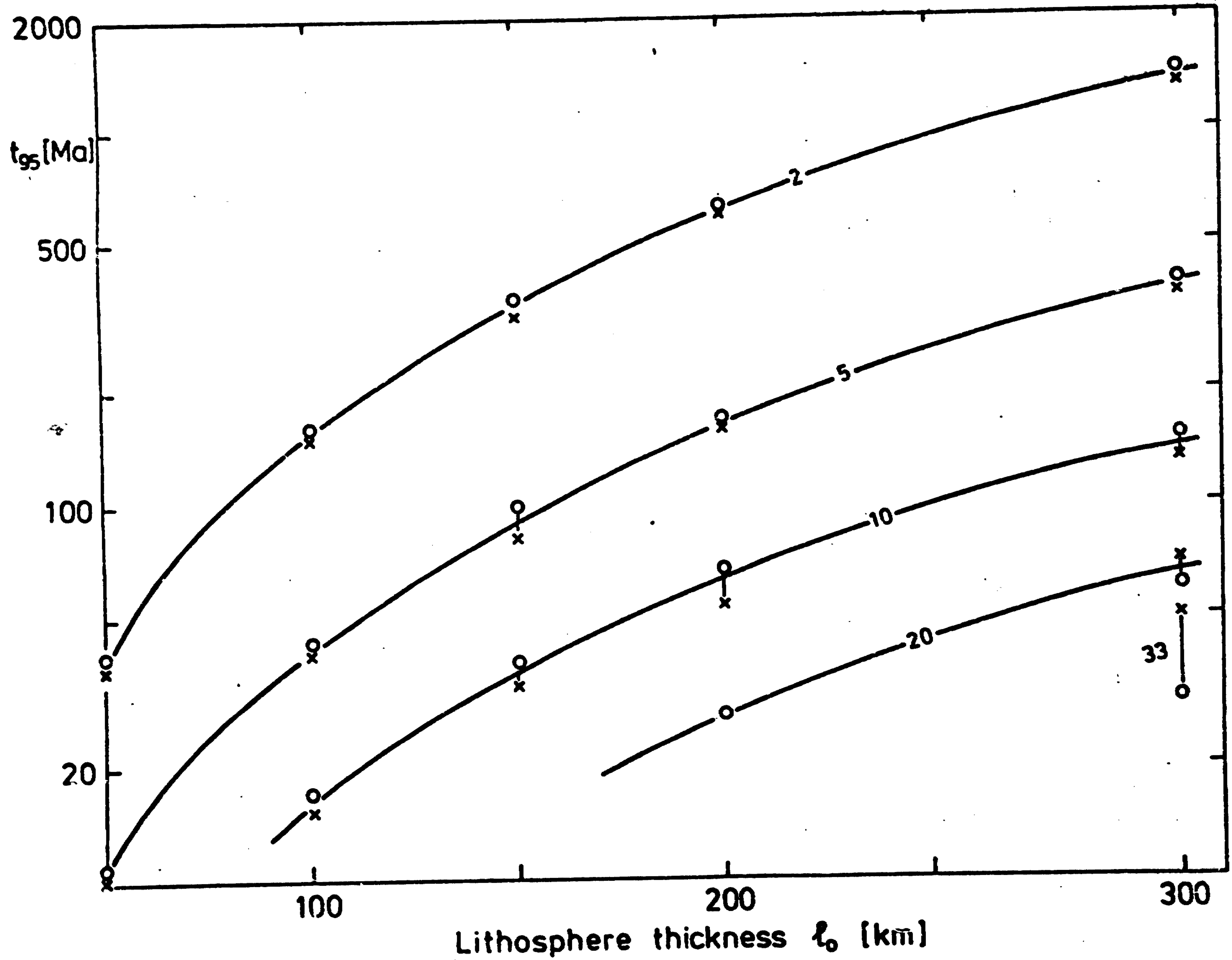

2000
500
100
20
t_{95} [Ma]
100
200
300
Lithosphere thickness ℓ_o [km]
2
5
10
20
33

Fig. 3

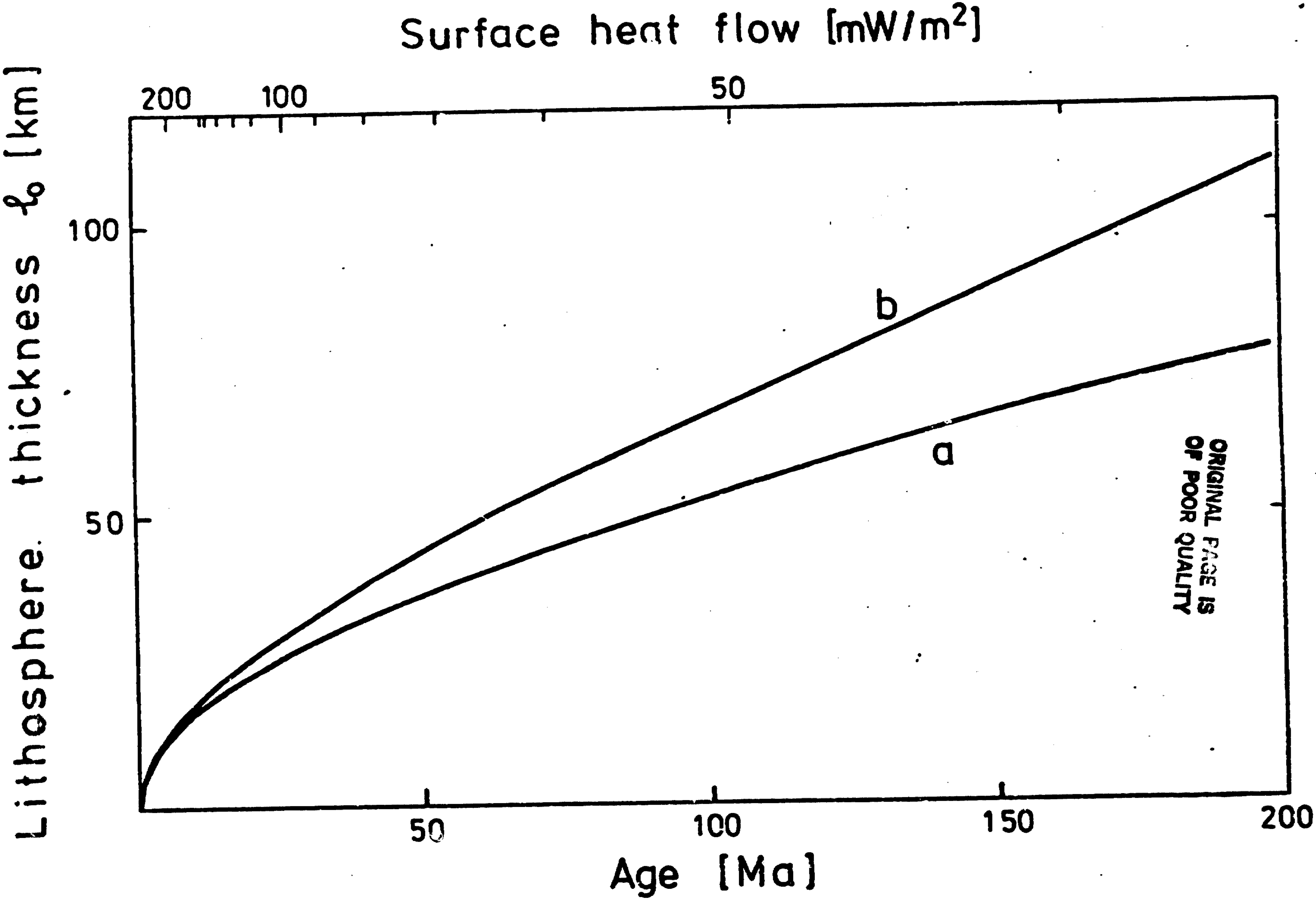
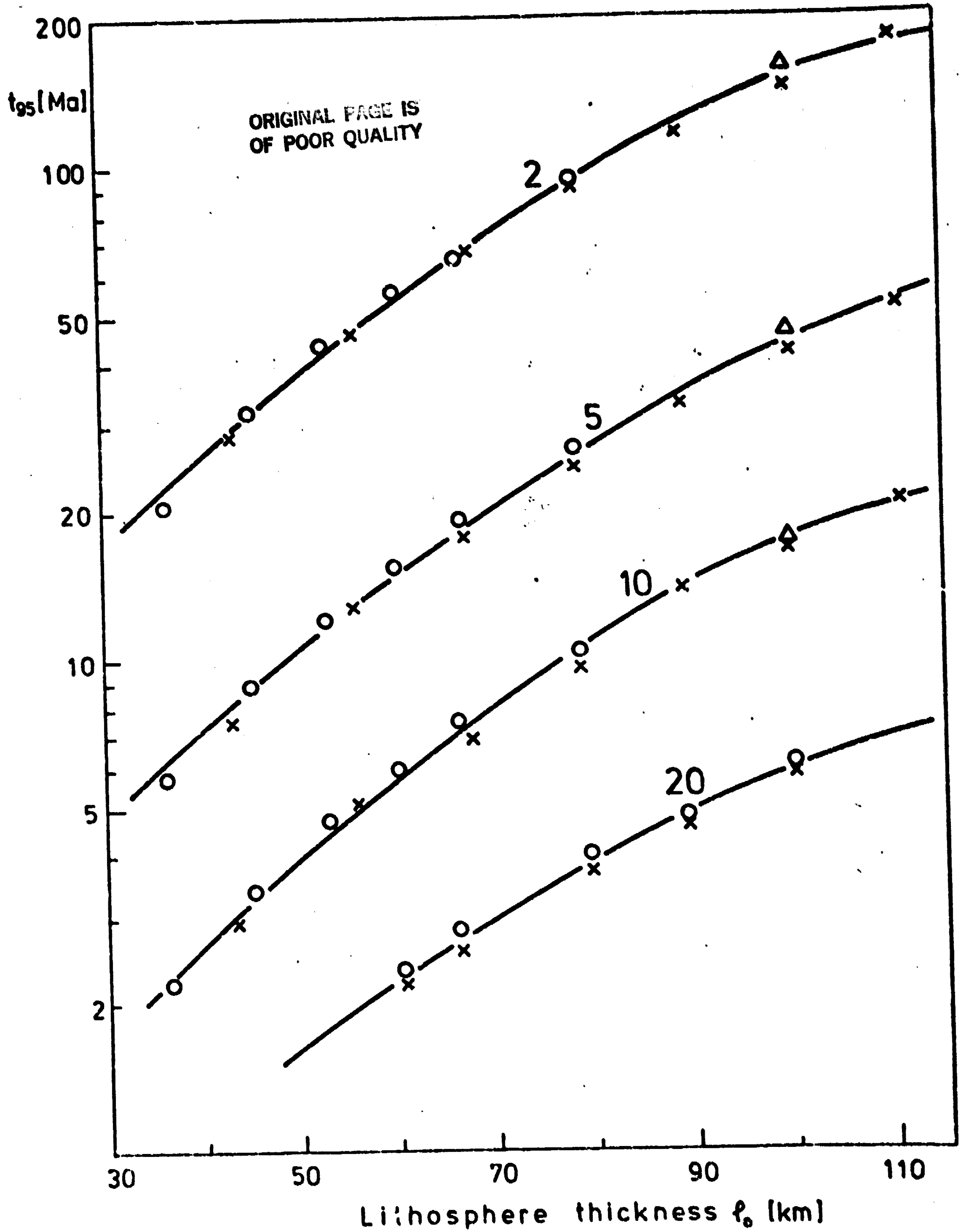
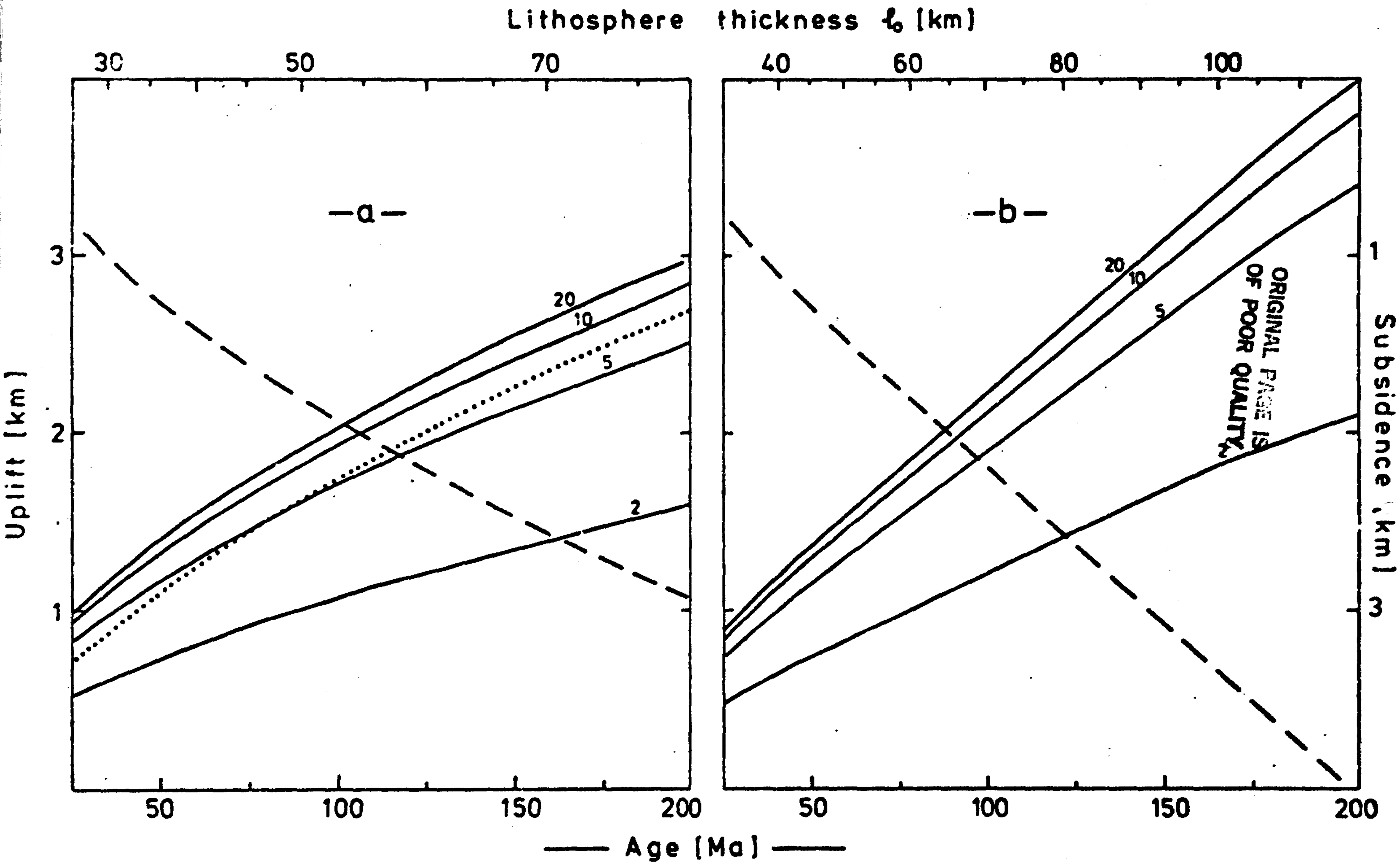

Surface heat flow [mW/m²]
200
100
50
Lithosphere thickness ℓ₀ [km]
100
50
b
a
50
100
150
200
Age [Ma]

Fig. 4.



200
t_{95} [Ma]
100
50
20
10
5
2
2
5
10
20
30
50
70
90
110
Lithosphere thickness ℓ_0 [km]



Lithosphere thickness ℓ_o [km]
30
50
70
40
60
80
100
-a-
-b-
20
10
5
2
20
10
5
2
3
2
1
1
3
Uplift [km]
Subsidence [km]
50
100
150
200
50
100
150
200
Age [Ma]
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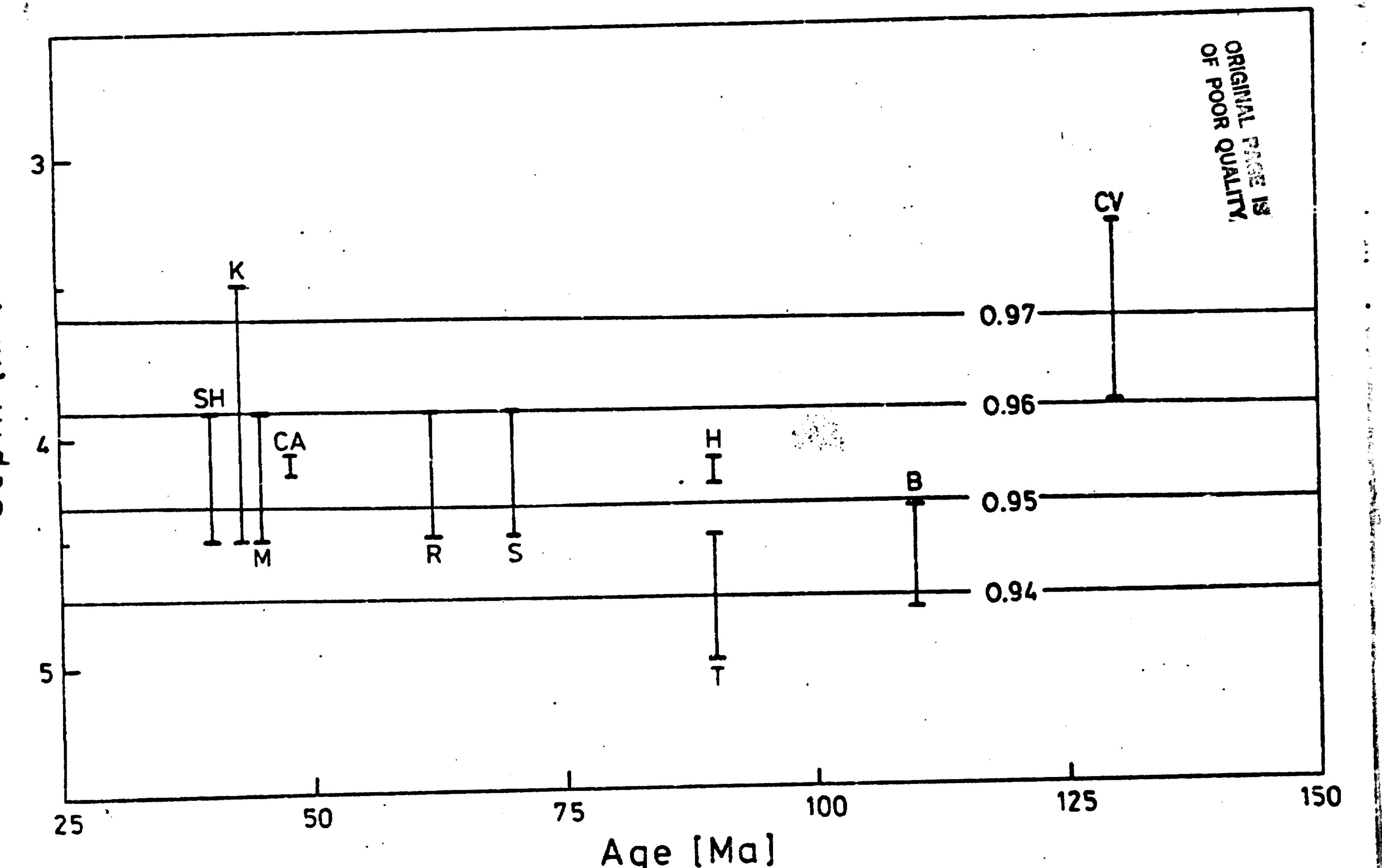
Fig. 6



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3
4
5
SH
K
M
CA
R
S
H
T
B
CV
0.97
0.96
0.95
0.94
25
50
75
100
125
150
Age [Ma]

Fig. 7