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THERMAL EFFECTS OF CONTINENTAL COLLISIONS: THICKENING A VARIABLE VISCOSITY LITHOSPHERE

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ABSTRACT

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Based on data that indicate that the lithosphere and the crust are anomalously hot in some convergence zones, simple calculations of the thermal effect of crustal thickening are presented. The results indicate that post-tectonic melting of the lower crust requires high levels of mantle heat flux. It has been suggested that this high heat flux can be explained by the convective removal of thermal boundary layers thickened by convergence. To test this hypothesis a series of numerical experiments were carried out on the removal of thickened thermal boundary layers whose viscosity depends on temperature and pressure. It is found that the boundary layers can be removed quickly (10 m.y. in some cases). However, these layers are too thin to greatly alter the thickness of the lithosphere or the heat flux at the top of the mantle. A set of calculations was done with internal heat sources which give equilibrium lithospheric thicknesses half that of the thickened lithospheres. These indicate that the time scale for convective thinning by one half is extremely long (at least 100 m.y.) for initial lithospheric thicknesses of 100 km. Therefore, the lithosphere must be anomalously hot and thin before deformation for extensive crustal melting to occur after thickening in observed times (40 m.y. for Tibet). Finally, it is suggested that subduction-induced convection may raise the general mantle temperatures over a broad region behind a convergence zone. This effect can inhibit the cooling of the lithosphere for more than 50 m.y. after the subduction induced convection has ceased.

INTRODUCTION

The growth of continents occurs primarily by the collision and suturing of continental fragments. Subduction of the lithosphere under a continent may eventually lead to the convergence of two continents. Unlike the oceanic crust and lithosphere, a continental plate cannot be subducted easily because of the lower density of the continental crust and the resulting buoyancy forces. If the convergence continues after the initial collision of continents, this leads to faulting and

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mountain ranges (Dewey, 1980; Dewey and Bird, 1970; Bird et al., 1975). A consequence of such a collision is a thickening of the crust, not only in the collision zone, but also under elevated plateaus (such as Tibet and the Iranian Plateau) that may develop behind the mountain ranges.

Theoretical modelling of the thermal state of the thick crust and the stability of the thickened lithosphere is the subject of this paper. Specifically, we will examine the suggestion that the lithosphere which has been thickened in such a region will be rapidly thinned by convective removal of the thermal boundary layer at the base of the lithosphere. The data on crustal thickness and present thermal state of Tibet are reviewed as are calculations which indicate the amount of mantle heat flux required for crustal melting (from Toksöz et al., 1981).

The question of the stability of thickened lithosphere has been treated for the case of a lithosphere made of several constant viscosity layers by Fleitout and Froidevaux (1982). They found that such a thickened lithosphere is unstable. Houseman et al. (1981) looked at the long-term evolution of an unstable thickened lithosphere. They modeled the lower lithosphere and asthenosphere as having the same constant value of viscosity. In the present study the viscosity was taken to be dependent on temperature and pressure. This caused these calculations to require much more computer time, but the need for arbitrary placement of flow boundaries was obviated. The position of the boundary between the asthenosphere, where heat is primarily transferred by convection, and the rigid lithosphere, where the heat transfer is all by conduction, is critical to this problem. In the Houseman et al. (1981) treatment this boundary was set at a given depth and does not depend on the temperature there. When the lithosphere was thickened this depth was not changed, although the temperature at that depth decreased by nearly a factor of 2. Figure 1 shows the effect of thickening the lithosphere on a temperature and a viscosity profile of the mantle. In the treatment used in this paper the boundary between the rigid and flowing regions was not specified, but came out of the calculation. This was a result of the temperature dependence of viscosity. In low-temperature regions the viscosity increases to values so high as to preclude significant flow. Heat had to be conducted into these areas to allow them to flow. As cooling or heating of a region occurred, the boundary moved. It should be emphasized that these results do not depend on particular values of the viscosity dependence. They merely result from some level of dependence of the viscosity of the mantle on temperature. That the viscosity of the mantle depends strongly on temperature has long been accepted (e.g., Stocker and Ashby, 1973; Weertman and Weertman, 1975). For lithospheric boundary layers to go unstable at all the viscosity in the region must be lower than the average mantle value of about 10^{22} poise determined from post-glacial rebound (Cathles, 1975). Neglecting the effect of pressure on viscosity, Yuen et al. (1981) find that boundary layers should not be unstable. As noted in Buck (1983) the dependence of viscosity on pressure allows the viscosity at the bottom of the boundary layer to be much lower than the value at greater depth.

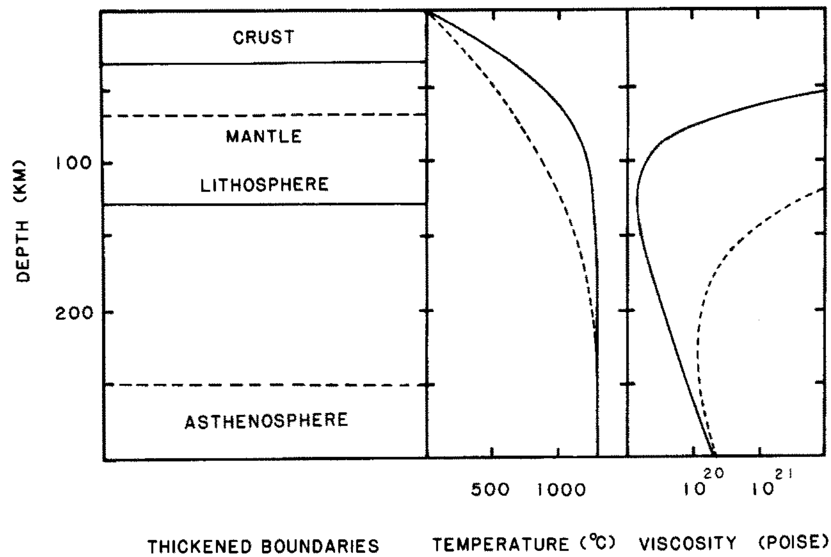

Fig. 1. The effect of doubling the thickness of the lithosphere on the temperature and viscosity profiles. Solid lines show the original thickness to the bottom of the crust and lithosphere as well as the temperature and viscosity profiles. Dashed lines show the thicknesses and profiles after thickening. The temperature profile is for conductive cooling for 40 m.y. of an isothermal crust and mantle using the diffusivity and temperature scale of Table I. The viscosity profile results from the viscosity parameters used for case 2 in Table II.

THERMAL EFFECTS OF CRUSTAL THICKENING

Among the mountain ranges which have clearly resulted from continental collisions are the Zagros, the Himalaya, the Alps, the Urals and the Southern Appalachians (Bird, 1978). Thrusting and folding in the crust caused thickening of the crust in all these ranges and may have been accompanied by thickening of the lithosphere through internal deformation. The Himalaya are associated with a large region, the Tibetan plateau, where the crustal and lithospheric deformation is apparently of this type. Geological and geophysical data indicate that the crust there is abnormally hot today. Since Tibet is the only mature example of such a plateau it is discussed in detail.

Tibet is an area where the crust has been thickened over 700,000 km^2. The crustal thickness over most of the plateau is about 70 km, nearly twice as great as normal continental crust. This has been determined by surface wave studies (Bird and Toksöz, 1977; Feng and Teng, 1983) and refraction lines done by Teng et al. (1980) in the plateau and by Jobert et al. (1982) in the northern Himalayas. This is consistent with the average elevation of five kilometers being isostatically compensated (Bird, 1978). The crust of Tibet was accreted onto Asia in several sections over about 150 million years as estimated from the ages of syntectonic granites in regions taken to be the suture zones between these sections (Gansser, 1980; Zhou et al., 1981; Allègre et al., 1982). The timing of the crustal uplift and thinning is disputed, but paleobotanical evidence indicates that the uplift post-dates the collision of India and Asia (Xu, 1981). The Indian collision occurred about 40 m.y. ago (Gansser, 1966; Powell and Conaghan, 1973; Molnar and Tapponnier, 1975). The northern part of the plateau has widespread calc-alkaline volcanics of Cenozoic age, presumably derived from melting of the lower crust (Dewey and Burke, 1973). There are also numerous hot springs indicating higher than normal crustal temperatures (Tong and Zhang, 1980). Furthermore, the attenuation of certain periods of surface waves may be an indication of partial melting of the lower crust (Bird and Toksöz, 1977).

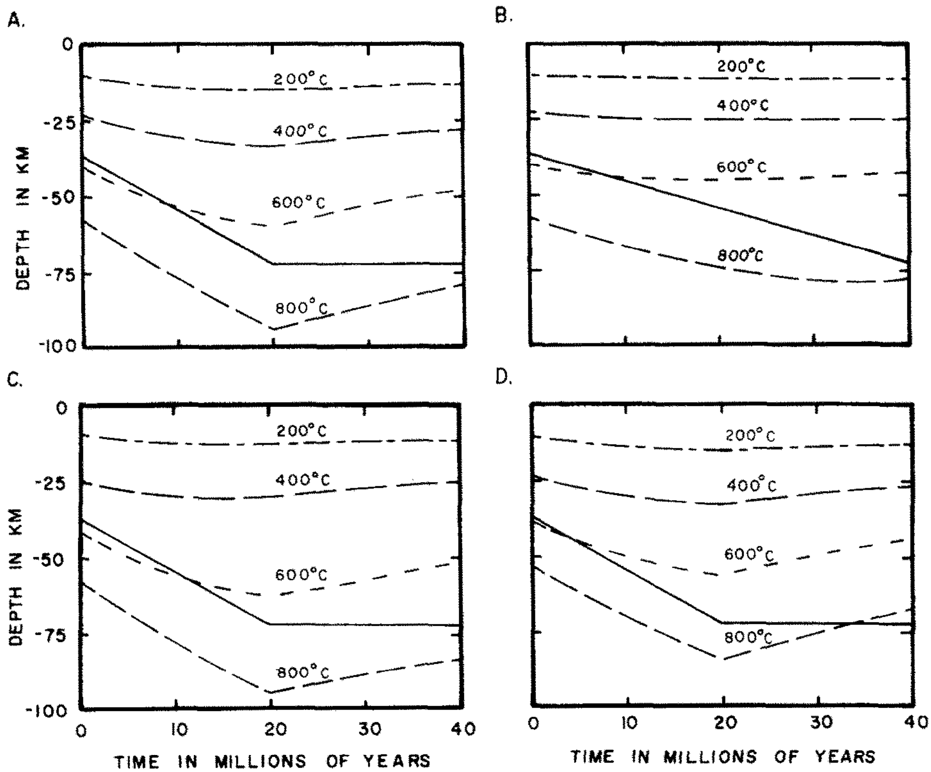

Fig. 2. Results of crustal thermal calculations in Toksöz et al. (1981). Dashed lines are isotherms against time since the beginning of crustal thickening. The solid line is the crust–mantle boundary position. All cases have the doubling of the crust taking 20 m.y. except case B which is for twice that time. Case A is for a uniform radioactive distribution in the crust and 1 kbar shear stress during deformation. Case B differs only in that the shear stress is 2 kbar. Case C shows that when the concentration of crustal radioactive sources is twice as large in the upper crust than the lower that temperatures remain lower in the crust. Case D has one third more mantle heat flux than the others.

In previous work (Toksöz et al., 1981) the thermal effects of the thickening of the crust have been considered. In that work the mantle heat flux was varied and only the temperatures in the crust were monitored. The same assumption was used in that work and the present one as to the way the crust deformed. The deformation was taken to be by internal strain rather than by crustal scale underthrusting suggested by Powell and Conaghan (1973). Figure 2 shows positions of isotherms in the crust as a function of time after the initiation of crustal thickening. It was found that neither shear heating nor redistribution of radioactive rich layers in the crust could lead to crustal melting without greater than normal heat flux from the mantle. An average of at least 0.8 HFU was required.

Three mechanisms to give a thin lithosphere and thus a large mantle heat flux have been proposed. (1) The entire lithosphere could be removed by "delamination" or peeling away from the crust (Bird and Baumgardner, 1982). This hypothesis is not advocated because seismic data for old continental cratons indicate that the lithosphere there is very thick (Leveque, 1980; Grand and Helmberger, 1982). The lithosphere could not be so thick if it were subjected to frequent removal. (2) An extremely large amount of the lithosphere could be convectively removed (McKenzie, 1978; Chen and Molnar, 1981; Houseman et al., 1981). This is critically examined using the numerical experiments which are described in this paper. (3) The lithosphere could have been thin at the time of the collision of India (Dewey and Burke, 1973; Toksöz and Bird, 1976; Toksöz et al., 1980; Buck and Toksöz, 1982). This third possibility is consistent with the observation that only tectonically young areas of lithosphere (which should be hot and thin) were deformed by the collision of India with Asia (Molnar and Tapponnier, 1981).

CALCULATION OF THE EFFECT OF INSTABILITIES IN A THICKENED LITHOSPHERE

The philosophy behind the approach used here is to consider the simplest possible set up for the calculations which retains the basic physics of the problem. As noted before, this requires the inclusion of temperature and pressure dependence of the viscosity. The non-dimensional equations of energy, mass and momentum conservation for the variable viscosity case are given in the appendix. They are solved explicitely on a finite difference grid with time stepping in the energy equation. The areas considered in the study correspond to a 700×700 km region of the mantle, when dimensionalized according to the parameters given in Table I. This size allows for three or more convective rolls to develop in all cases. The relation between temperature, pressure and viscosity is a standard one applicable to creep, taken from Weertman and Weertman (1975) and is given in the Appendix.

The variations in viscosity define the flow boundary for this problem, as was found in the linear stability analysis of variable viscosity boundary layers by Jaupart (1981), and no arbitrary boundary need be used. The boundary for the flow is the top of the lithosphere, but it is computationally more efficient to place a no slip

TABLE I

Parameters used for non-dimensionalizing equations

Symbol	Name	Value	Units
κ	diffusivity	0.01	cm^2/s
L	length scale	$7.0 \cdot 10^7$	cm
ΔT	temperature scale	1300	°C
α	thermal expansion coefficient	$3.0 \cdot 10^5$	1/°C
ν	viscosity	$1.0 \cdot 10^{22}$	poise
g	acceleration of gravity	980	cm/s^2
ρ	mantle density	3.5	g/cm^3

boundary at the depth in the lithosphere where the viscosity is 10^{22} poise. This is between two and three orders of magnitude above the minimum viscosity in the flow region. Runs with the boundary placed higher in the lithosphere, where the viscosity was 10^{23} poise gave the same results, but required at least twice the computer time. The side boundary conditions on the flow were stress free, except in the case of one model with periodic conditions (case 1). The bottom boundary is always taken to be stress free. The boundary conditions on the energy equation are fixed temperature (corresponding to 0°C) at the top and insulating on the sides and bottom. Case 1 had periodic conditions on the side temperatures.

The initial temperature profile for cases 1–7 was derived from purely conductive half-space cooling of an initially isothermal material, with diffusivity given in Table I, for a given length of time. This temperature profile was then stretched in depth by a factor of two to simulate lithospheric thickening. Figure 1 shows an initial and thickened temperature profile along with the corresponding viscosity profile. Next, a random temperature perturbation (between 0° and 1°C) was given to each grid point. Without perturbations the instabilities would not be initiated in the calculations. The initial flow velocities were zero. In the periodic test a periodic perturbation was given to excite a particular wavelength of flow. This thickening of a conductive temperature profile is not the same procedure used by Houseman et al. (1981), where the temperature structure resulting from a convection calculation was stretched vertically for the initial condition. This procedure of achieving a steady state is costly in terms of computer time for the variable viscosity calculations, where runs take 20–50 times the computer time for the comparable constant viscosity calculations. Thus, only in cases 8 and 9 were the initial conditions gotten by thickening a steady state temperature structure in the region of calculation. After steady state had been reached the top quarter of the box was doubled in thickness while the rest of the box was thinned so that the depth extent of the box was not changed.

In the first cases no heat sources were incorporated and only the falloff of the unstable boundary layer was followed. Later internal heat generation which would result in smaller equilibrium lithospheric thicknesses than existed at the start of the calculation were used. This was done to see how quickly equilibrium thicknesses would be approached. The runs were generally carried out to times that correspond to at least 40 m.y., since this is the time since the lithospheric thickening commenced in Tibet.

Resolution of the solutions on the grids used here were guaranteed in two ways. First, the numerical experiments were done on successively refined grids until the same results were achieved on two different grids. Second, the heat flux out of the grid and the internal heat generation were compared to the average temperature of the region to ensure conservation of energy. The grids used were at least 56×56 points with even spacing in the horizontal direction and variable mesh spacing in the vertical. The varible spacing of points allowed the needed resolution in the regions of the largest gradients of viscosity and flow, without an excessive number of points overall. In the region of highest resolution the grid spacing was uniform, so second-order accuracy in the finite difference approximations was preserved (Roache, 1982). The grid positions are shown in Fig. 3 as tick marks around the boxes.

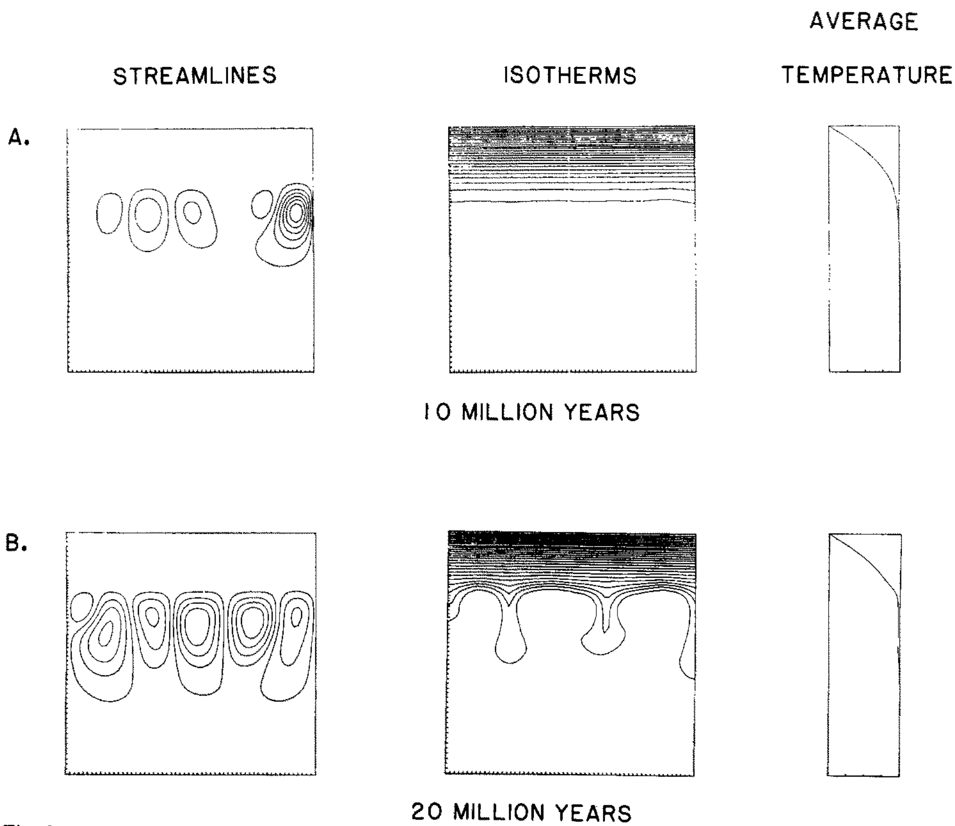

Fig. 3A, B. For legend see p. 60.

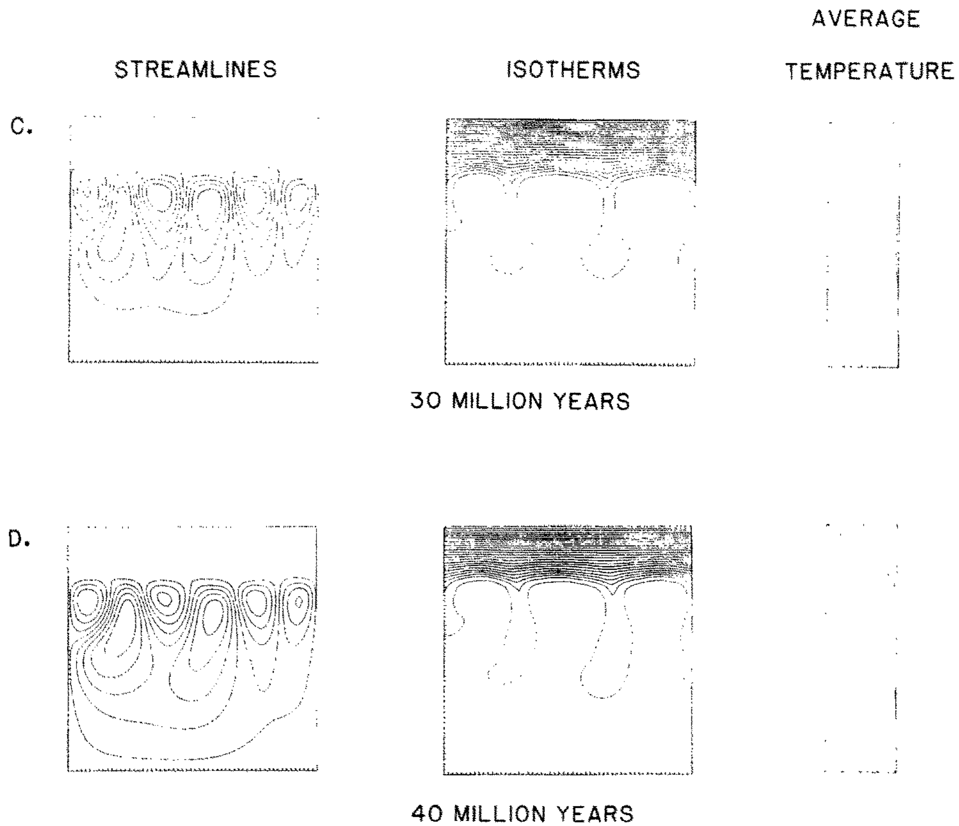

Fig. 3. Streamlines, isotherms and horizontally averaged temperature at 10 m.y. intervals for case 4. The boxes represent 700×700 km of mantle. The tick marks around the boxes show the grid point positions. The eight contours for the streamfunction have the following non-dimensional ranges: -4.3 to 11.3 for case A; -58.2 to 59.6 for case B; -33.4 to 25.9 for case C; and -40.2 to 24.4 for case D. The contour intervals for the temperature are 50°C. The range of the horizontally averaged temperature is 1300°C.

RESULTS

The cases considered here are summarized in Table II. The parameters varied were the boundary conditions, the reference viscosity, the heat sources, the activation volume (V^*), the activation energy (E^*) and the time of the initial conductive cooling (which effectively changes the initial lithospheric thickness). The heat sources are up to 3 times values estimated for the average mantle (Ringwood, 1975), but are meant to include the effect of both internal heat-production and heating from below. The activation volume, V^*, was varied within the range of experimentally determined values for olivine as was done for the activation energy E^*. The range of activation volume is 10–20 cm^3/mole (Kohlstedt et al., 1980; Sammis et al., 1981) and for activation energy is 70–125 kcal/mole (Goetze, 1978). The reference

TABLE II

Descriptions of the cases considered for the numerical calculations

Case	Reference viscosity (× 1019 poise)	Side boundary condition	Heat sources (erg/cm^3)	V^* (cm^2/mole)	E^* (kcal/mole)	Initial cooling time (m.y.)
1	2.5	periodic	0	10	100	40
2	1.0	free	0	10	100	40
3	1.0	free	$2.14 \cdot 10^{-7}$	10	100	40
4	1.0	free	$6.42 \cdot 10^{-7}$	10	100	40
5	1.0	free	$6.42 \cdot 10^{-7}$	20	100	40
6	1.0	free	$6.42 \cdot 10^{-7}$	10	100	80
7	1.0	free	$6.42 \cdot 10^{-7}$	10	120	40
8	1.0	free	$6.42 \cdot 10^{-7}$	10	100	steady state
9	10.0	free	$6.42 \cdot 10^{-7}$	10	50	steady state

viscosity was set so as to give viscosity minima in the asthenosphere under young lithosphere close to the estimates of the viscosity there (Passey, 1981; Richter and McKenzie, 1978). Each case required between 3 and 9 h of c.p.u. on a Vax 11/780 computer.

Case 4 resulted in the greatest convective thinning of the lithosphere for the cases with a conductive initial condition, so it will be discussed in detail and be used as a

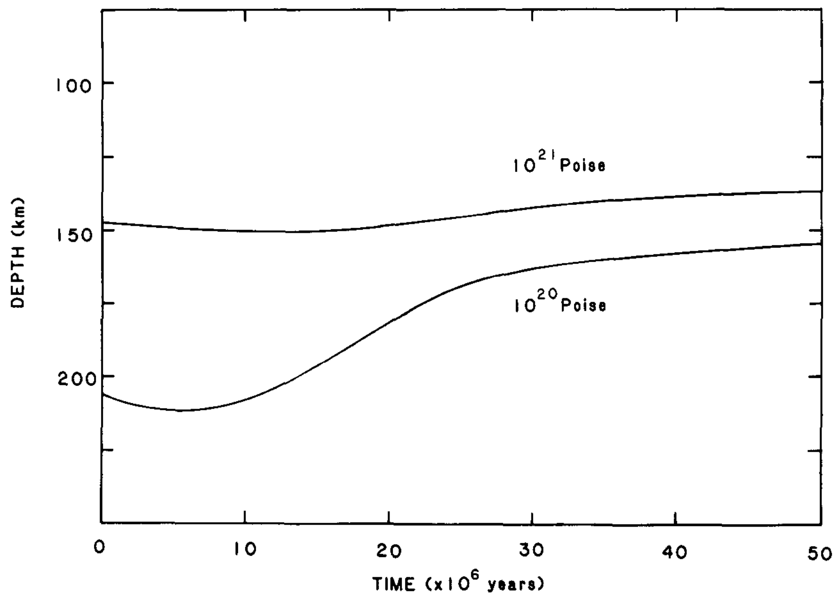

Fig. 4. The depth to the level where the horizontally averaged viscosity equals the indicated value for case 4.

reference when discussing the other models. Figure 3 shows snapshots at 10 m.y. intervals of the isotherms and streamlines of the flow for this case. The boundary layer can be seen flowing down in a droplet-like fashion. The horizontally averaged temperature profile shows that the thermal lithosphere is not greatly thinned over the 40 m.y. period. Figure 4 is a more quantitative picture of the variation in the thickness of the lithosphere through the duration of the run. It shows the variation to the depth where the horizontally averaged viscosity equals a constant value (either 10^{21} or 10^{20} poise). The base of the lithosphere could be defined as the place where the viscosity reaches such a value. Figure 4 shows that the depth to the 10^{21} poise level has changed negligibly during the calculation. The 10^{20} poise level has changed about 20% in this time.

To quantify the vigor of the flow at different times, the average dissipation for the flowing region was calculated. The dissipation (D) is proportional to the integral of the strain rate squared $(\dot{e})^2$ over the area (A):

$$D = \int\int \dot{e}^2 \, \mathrm{d}A \tag{1}$$

Figure 5 shows that the maximum rate of dissipation in case 4 occurs at a depth just below the lithosphere, at about 150 km. It also indicates that there is a depth range over which the flow is much more vigorous than above and below. This is due to the viscosity structure having a minimum in that region. The reason for the low between two highs is that the vertical flow occurs only in narrow zones. The horizontal averaging emphasizes the top and bottom of the convection cells where the flow is

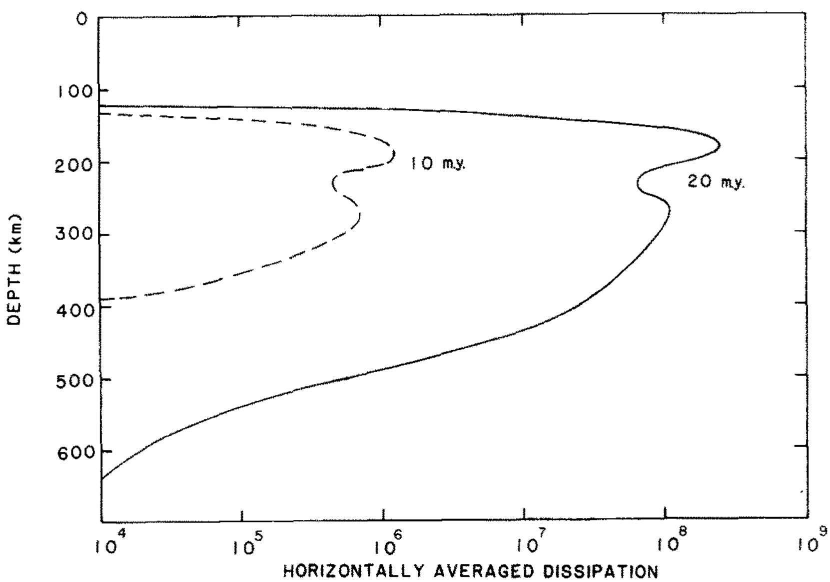

Fig. 5. The variation of the horizontally averaged dissipation with depth for case 4 at the indicated times into the runs.

horizontal. Figure 6 shows the average dissipation through time for all cases except number 5 which was not on scale and 8 and 9 where the average dissipation was nearly constant. In all cases the maximum dissipation was associated with the first convective removal of the conductive boundary layer. This maximum took over 20

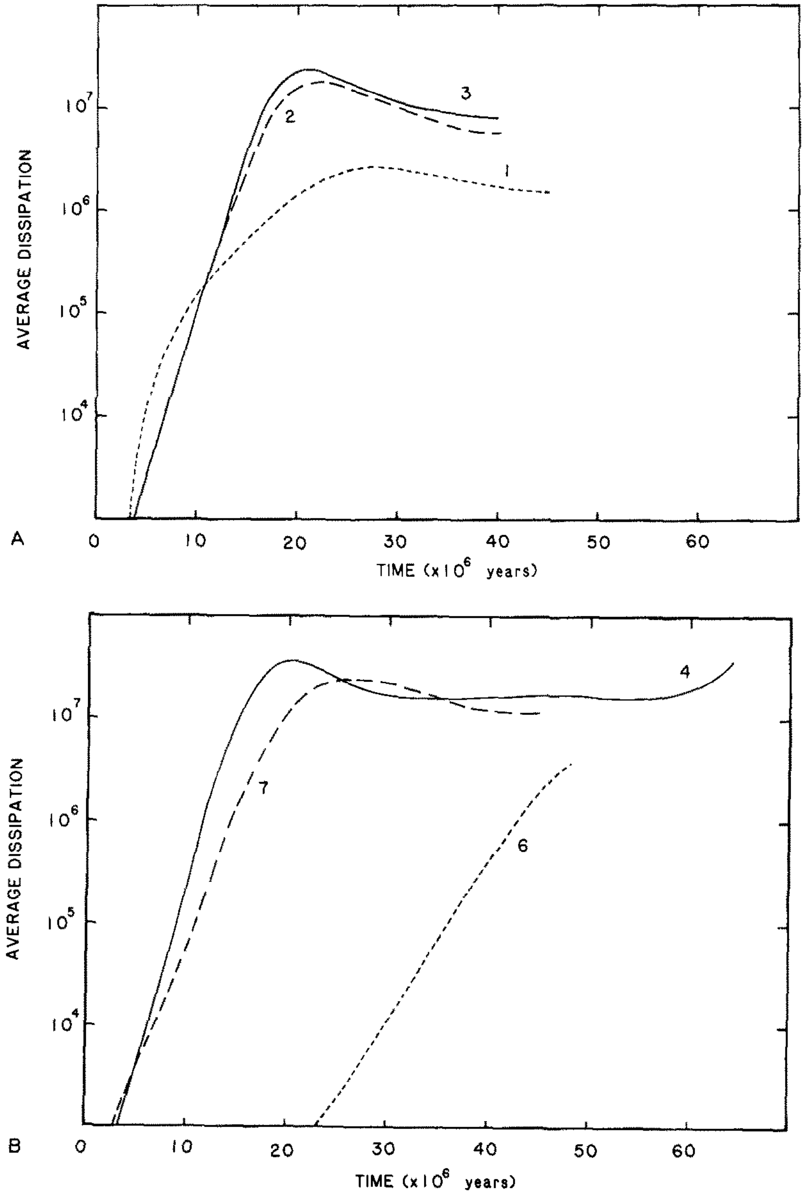

Fig. 6. The average dissipation versus time into the run for the cases indicated. Dissipation for case 5 was too low to be on scale.

m.y. to develop in runs 1–7. In the cases with internal heat sources the high rate of flow is maintained through the rest of the run. But, as shown in Fig. 4, the bulk of the convective thinning is associated with this boundary layer removal and not with ablation due to the heat sources and rapid flow rates.

Another measure of the difference between the results is the dissipation weighted viscosity of the flow region. This average viscosity is defined following Parmentier (1978) to be:

$$\nu_{\mathrm{ave}} = \int\int \nu \dot{e}^2 \, \mathrm{d}A / \int\int \dot{e}^2 \, \mathrm{d}A \tag{2}$$

This quantity was calculated for all the cases and it is tabulated along with other output information in Table III. The average viscosity was fairly constant over the time of the calculations, but is tabulated a time 20 m.y. into the runs. This parameter is useful in calculating an average Rayleigh number for the variable viscosity flow region. The average Rayleigh number is defined as:

$$Ra_{\mathrm{ave}} = \alpha g \Delta T (L^3) / \kappa \nu_{\mathrm{ave}} \tag{3}$$

where the variables are defined in Table I, but here ΔT is the temperature difference over the length scale of the flow (L). Parmentier (1978) has shown that this parameter bears the same relationship to the heat flux across a region convecting in steady state as does the Rayleigh number in a constant viscosity case. There is ambiguity in the calculation of this parameter since the length scale of the flow is not clearly defined. A possible way to estimate this scale is to take the depth extent of the region where dissipation is within a factor of ten of the maximum horizontally averaged value. For case 4 it gives a value of 250 km for the length scale and $1. \times 10^5$ for the Rayleigh number. As it is difficult to find a consistent estimate of the length scale for the different runs, the Rayleigh number is not calculated for each case.

TABLE III

Results of numerical experiments described in Table I. See text for methods of calculating the parameters. The character * indicates the viscosity level was not present for that case.

Case	Effective viscosity (poise)	Max. average dissipation (dimensionless)	Change in depth (km) of viscosity level	
			(10^{21} poise)	(10^{20} poise)
1	$6.5 \cdot 10^{20}$	$2.8 \cdot 10^{6}$	−1.9	*
2	$3.0 \cdot 10^{20}$	$1.8 \cdot 10^{7}$	2.4	*
3	$2.8 \cdot 10^{20}$	$2.3 \cdot 10^{7}$	3.9	48.4
4	$2.5 \cdot 10^{20}$	$3.5 \cdot 10^{7}$	7.0	49.1
5	$1.4 \cdot 10^{21}$	$6.0 \cdot 10^{1}$	4.4	*
6	$8.6 \cdot 10^{20}$	$5.0 \cdot 10^{5}$	1.3	*
7	$2.3 \cdot 10^{20}$	$2.2 \cdot 10^{7}$	3.5	36.4
8	$6.0 \cdot 10^{20}$ *	$1.2 \cdot 10^{7}$	14.1 *	*
9	$3.6 \cdot 10^{21}$	$1.0 \cdot 10^{6}$	8.0	*

Case 1 shows that the higher the reference viscosity the lower the maximum dissipation rate for the flow. The high rate of dissipation early in the run was due to the periodic initial conditions which forced coherent flow quickly. This is the only case where the 10^{21} poise level moved down through the calculation. This was due to the high effective viscosity and lack of heat sources for the model. Case 2 indicates that the removal of the lower lithosphere can result in a thinner lithosphere, for a time, even without heat sources. Case 3 shows that the inclusion of small heat sources does not greatly affect the thinning of the lithosphere over 40 m.y. In case 5 the doubling of the pressure dependence of viscosity (V^*) raised the effective viscosity and narrowed the depth range of vigorous flow. The boundary layer was still unstable in this case but was not removed over the time of the calculation. Case 6 demonstrates that the thicker the lithosphere the slower the flow in response to boundary-layer instabilities. The effective viscosity and maximum dissipation were low in this case. Case 7 shows that increasing the temperature dependence of viscosity (E^*) by 20% had a small effect on the maximum dissipation, but that the lower lithosphere was made more resistent to convective thinning.

The two runs which started with thickened steady state profiles (8 and 9) showed that the rate of thinning of the lithosphere is not greatly increased relative to the cases with simpler initial conditions. This is due to the fact that the most rapid change in the lithospheric thickness is due to the removal of the thickened boundary layer. Case 8 was analogous to case 4 and showed a slightly greater decrease in the depth to the 10^{21} poise level (Table III) during the calculation. Case 9 shows that greatly decreasing the temperature dependence of the viscosity is offset by an increase in reference viscosity.

CONCLUSIONS

In all of the cases considered of the effect of the instability of thickened variable viscosity lithosphere on the rate of thinning of the lithosphere was not sufficient to bring it to equilibrium thickness in 40 m.y. when the lithosphere was initially about 100 km thick. This was true over a wide range of experimentally determined parameters controlling the viscosity of the lithosphere and asthenosphere. Inclusion of large internal heat sources in the model mantles maintained a higher level of vigor in the flow, but still did not result in rapid thinning of the lithosphere through ablation. For the cases considered which had heat sources, the lithosphere would eventually return to its equilibrium thickness, but on a time scale of greater than 100 m.y. The difference between these results and those of Houseman et al. (1981) are due to the inclusion of variable viscosity effects. In a study of the effect of mantle plumes, Spohn and Schubert (1982) find that 300 km thick lithosphere could be locally thinned to half this thickness in 50 m.y., but this required 5 times the normal mantle heat flux.

The problem of the thermal state of Tibet then is how it could be so hot today. The most reasonable possibility is that the deformed region was anomalously hot before the collision. As noted, this is consistent with the observations of the correlation of the tectonic age with deformation level in the region (Molnar and Tapponnier, 1981). Two related effects can have produced this hotter region. One is the rapid subduction which is known to precede the Indian collision (Molnar and Tapponnier, 1975) and the other is the small-scale convection associated with the cooling of the lithosphere. It has been suggested that the convection induced by subduction could bring a great deal of hotter than normal mantle into the asthenosphere behind a subduction zone (Toksöz and Hsui, 1978). This could even cause some thinning of the lithosphere, but as seen in the present calculations the lithosphere is resistent to ablation. Toksöz and Hsui (1978) found the major effect of the subduction induced convection is concentrated only a few hundred kilometers away from the subduction zone. Tibet, before the collision, was over 1500 km wide so some mechanism must keep the lithosphere thin in areas away from the active subduction. The small-scale convection associated with the cooling of the lithosphere can do this. The cooling of the lithosphere is slowed due to boundary layer instabilities compared to purely conductive cooling (Buck, 1983). If the subduction induced convection has brought hot mantle up beneath the lithosphere of Tibet then the cooling of the lithosphere could not proceed until the high temperatures of the mantle had been conducted through the lithosphere.

An inexact analogy which may clarify this mechanism is that of ice freezing on a lake. If the air temperature above the lake is below 0°C for a long enough time then the top of the lake will freeze. The lake a few meters below the ice is well stirred by convection and has a uniform temperature of 4°C (the temperature at which water is the densest). If a river feeding the lake brings in warm water (above 4°C) then this water must be cooled before freezing can continue. This heat has to be conducted through the lid of ice on the lake. In a similar manner, if subduction induced convection raises the temperature of the asthenosphere beneath Tibet, this excess heat must be conducted away before the cooling of the lithosphere can continue. The anology is inexact because ice freezing is a phase change which liberates latent heat and the raising of asthenospheric temperatures lowers the viscosity of the convecting region so more heat is carried to the lithosphere.

Although timings are uncertain, it is likely that subduction was not active under northern Tibet for 100 m.y. before the Indian collision, though it was active under southern Tibet. If previous subduction had left 300 km of the upper mantle there 50°C hotter than normal, then the lithosphere would not have cooled in 100 m.y., given a background mantle heat flux. This would not be the case if the cooling of the lithosphere proceeded only through conductive means. The small-scale convection below the lithosphere may not be able to thin the lithosphere quickly, as shown in this paper, but it can maintain the lithosphere at a given thickness as long as the mantle temperatures are high enough for vigorous flow.

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APPENDIX

For this study we solve the Navier-Stokes equations in two dimensions for mass, momentum and energy conservation (Batchelor, 1967). They are modified for the problem of flow in the earth's mantle by dropping inertial terms and terms that depend on material compressibility (Turcotte et al., 1973). They are written in terms of a stream function (ψ) and a vorticity (ω) as:

$$\frac{\partial T}{\partial t} = u \cdot \frac{\partial T}{\partial x} + v \cdot \frac{\partial T}{\partial y} + \nabla^2 T + H \tag{1}$$

$$\nabla^2(\mu\omega) = Ra\frac{\partial T}{\partial x} + 2\left[\frac{\partial^2 u}{\partial x^2}\frac{\partial^2 \psi}{\partial y^2} + \frac{\partial^2 \mu}{\partial y^2}\frac{\partial^2 \psi}{\partial x^2} - 2\left(\frac{\partial^2 \mu}{\partial x \partial y}\frac{\partial^2 \psi}{\partial x \partial y}\right)\right] \tag{2}$$

$$\nabla^2 \psi = \omega; \qquad \nabla^2 \equiv \frac{\partial^2}{\partial x^2} + \frac{\partial^2}{\partial y^2} \tag{3}$$

where the non-dimensional variables are: T = temperature; ν = viscosity; u, v = x and y velocities; $u = \delta\psi/\delta y$, $v = -(\partial\psi/\partial x)$ and $Ra = g\alpha\Delta TL^3/\nu\kappa$ (Rayleigh number).

The values of the variables in the expression for the Rayleigh number are given in Table I. To solve the temperature equation (1), a finite difference scheme with centered differences for the diffusion terms and upwind differences for the advection terms was used. Centered differences were used for the solutions of the coupled vorticity and streamfunction equations (1) and (2). The vorticity and streamfunction equations were iterated to convergence before advancement of the temperature equation. The convergence criterion was that the change in vorticity at all grid points between iterations divided by the maximum value on the grid be less than 0.001. The convergence criterion for the streamfunction was 0.0001. To calculate the viscosity at points in the grid we take the temperature (T) from eq. 1 and use the pressure (P) appropriate for that point using the following relation (Weertman and Weertman, 1975):

$$\nu(t,p) = A\,\exp((E^* + PV^*)/KT)$$

where: E^* = activation energy; V^* = activation volume; A = constant set according to reference viscosity at 1350°C and the pressure at 150 km, and K = Boltzman's constant.

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