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Thermo-Mechanical Models of Lithosphere and Asthenosphere: Can a Change in Plate Velocity Induce Magmatic Activity? [and Discussion]

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Thermo-mechanical models of lithosphere and asthenosphere: can a change in plate velocity induce magmatic activity?

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The thermal and mechanical structure of the upper mantle is investigated on the basis of the law of deformation of olivine combined with appropriate equations of motion and of heat balance. The effect of changes in plate velocities is shown to generate temperature anomalies underneath the lithospheric plate. The time delay is not the usual thermal time constant but represents the time necessary to generate enough heat to approach a new steady state situation. It depends strongly upon the magnitude of the velocity jump and upon the amount of radiogenic heat sources present in the mantle. The correlations described by Briden & Gass (*Nature, Lond.* (1974) **248**, 650) for Africa between polar wander and magmatism can in principle be explained by our models.

Introduction

The motion of tectonic plates at the Earth's surface occurs at constant velocities for long periods of time. This justifies attempts to construct steady state models of mass transport in the mantle. The global surface velocity pattern can, however, suffer rather abrupt changes, for instance after the disappearance of one plate or the creation of new plate boundaries. This raises the question of what happens to the thermal state underneath a lithospheric plate after its velocity has changed. Our assumption in this discussion shall be that the velocity change has been imposed by forces acting at the boundaries of the plate so that the transient thermochemical effects in the asthenosphere underneath the central regions of the plate can be regarded as a consequence of this change in velocity with respect to a less mobile lower mantle.

The important coupling between mechanical and thermal state of the mantle arises partly from the strongly temperature dependent rheology of minerals like olivine. On the other hand, larger velocity gradients lead to greater heat production by viscous dissipation. Can this last mechanism lead to magma generation? What time constant is involved? These are the questions we attempt to answer quantitatively before comparing our time dependent models with major magmatic events in Africa.

Physical models of the upper mantle

On the basis of the creep properties of olivine (Post 1973; Kohlstedt, Goetze & Durham 1976), of the temperature equation and of the equations of motion, one can model the thermo-mechanical structure of the upper mantle. Schubert and co-workers (Schubert, Froidevaux & Yuen 1976; Schubert *et al.* 1977) have done this for the oceanic case. They impose simple boundary conditions like temperature and velocity at the surface, a temperature gradient, and a vanishing horizontal velocity component at great depth. The temperature, velocity and shear stress can then be computed for any depth and age. One solution is shown in figure 1 picturing isotherms and streamlines. A top layer thickening with age, i.e. with distance from the ridge axis, moves at the imposed surface velocity; it defines the lithosphere. Underneath it, a zone of shear, the asthenosphere, ensures the decoupling of the lithospheric plate from the lower mantle. In the particular solution shown here a certain amount of return flow occurs in the asthenosphere, the rest of the material being circulated through the lower mantle. The computed dragforce under the lithosphere varies with the assumed deep mantle adiabat and with the values taken for the rheological parameters. The latter depend strongly upon the amount of water present in the mantle. This dragforce increases with age; it can amount to a few bars† or several tens of bars. Such detailed two dimensional models can be used to calculate the topography of the ocean floor. They predict the flattening of the old ocean basins. The asthenosphere, as defined above, is a zone where the mantle material is warm enough to allow solid state creep to occur. It requires no partial melting and does not necessarily coincide with a region of low seismic velocities and high seismic attenuation. Indeed the computed seismic low velocity zone is at shallower depth.

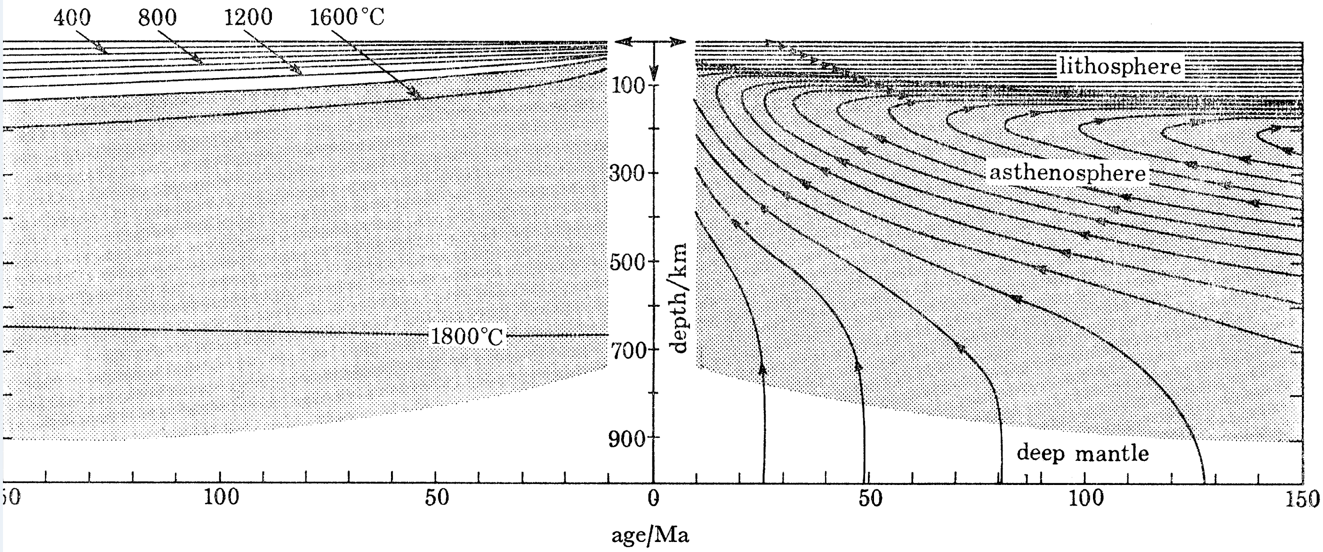

FIGURE 1. Model of oceanic upper mantle with partial shallow return flow. Isotherms (on the left) and streamlines (on the right) have been calculated by Schubert *et al.* (1977) for a dry olivine rheology and for the following boundary conditions: $T_0 = 0$ °C and $u_0 = 10$ cm/a at the surface, $T \rightarrow 1600$ °C + (0.3 °C/km) $|y|$ at great depth and a return flow in the asthenosphere amounting to 60 % of the forward flow in the rigid lithosphere. The latter is thickening with age and the asthenosphere is indicated by the shaded area.

TIME DEPENDENT MODEL FOR CONTINENTS

Under old continental shields the global flow of mass may be assumed to be horizontal, the asthenosphere acting as a zone of mechanical decoupling. In this case the equation of motion simply states that the shear stress is independent of depth z. The temperature equation reduces to:

$$\rho\, c_p \frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left(k\frac{\partial T}{\partial z}\right) + \tau\frac{\partial u}{\partial z} + Q, \tag{1}$$

where T is the temperature, τ the horizontal shear stress, t the time, ρ the density, c_p the specific heat at constant pressure, k the thermal conductivity of olivine which depends upon T

† 1 bar = 10^5 Pa.

(Schatz & Simmons 1972), and Q the radiogenic heat generation. The term $\tau\ \partial u/\partial z$ represents shear heating and couples this equation to the creep law. The latter can be written as:

$$\frac{\partial u}{\partial z} = \frac{B\tau^3}{T} \exp\left(-\frac{E^* + pV^*}{RT}\right), \tag{2}$$

where B is a measured constant, E^* the activation energy and V^* the activation volume. The confining pressure p can be expressed in terms of the depth z.

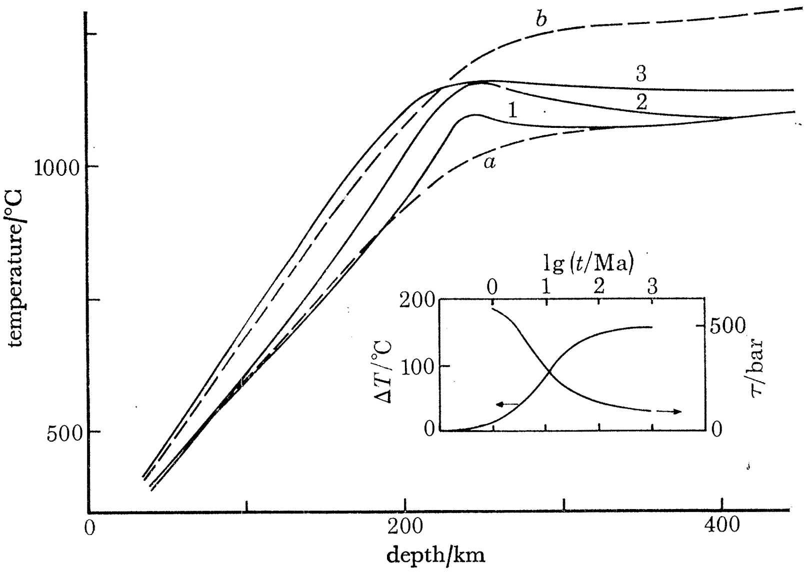

FIGURE 2. Steady state continental geotherms (dashed lines) for a mantle without radiogenic sources, and transient geotherms (solid lines) after a plate velocity jump from 1 to 6 cm/a. Solution a is for u_0 = 1 cm/a and b for 6 cm/a. Rheological parameters in equation (2) correspond to a wet olivine rheology: $B = 5.4 \times 10^{-15}$ cm^3 s^5 K/g^3, E* = 93 kcal/mol, V* = 20 cm^3/mol after Post (1973). Solutions 1, 2 and 3 are for 10, 100 and 1000 Ma after the velocity jump Δu_0. In the inset the temperature increase at 240 km and the shear stress τ are plotted against time. The time constant is about 10–30 Ma.

For steady state conditions, $\partial T/\partial t = 0$, Froidevaux & Schubert (1975) have given solutions of the coupled system (1) and (2) for simple boundary conditions specifying the temperature and heat flow at the surface and the heat flow at great depth. These models picture the Earth's mantle as an infinite olivine half-space overlain by a 40 km crust. The crust is divided into two layers with different radiogenic heat production. This yields $T(z)$, $u(z)$ and τ. The solutions show strong feedback effects between the temperature and velocity profiles. Two steady state temperature profiles a and b are pictured in figure 2. They both correspond to the same mantle rheology and to a mantle without radiogenic sources, $Q = 0$. The lower geotherm is for a plate velocity u_0 of 1 cm/a, the upper one for u_0 = 6 cm/a. The temperature adjusts itself so that a faster plate implies a warmer asthenosphere.

Let us now assume that the continental mantle is in a steady state corresponding to geotherm a, the plate moving at 1 cm/a, for $t < 0$. At time $t = 0$, the surface velocity jumps abruptly to a higher value, u_0 = 6 cm/a. Figure 2 also pictures the perturbed geotherms for $t > 0$, calculated on the basis of (1) and (2). Shear heating increases beneath the lithospheric plate where a temperature peak develops. The transient geotherms are labelled by lg t (Ma); thus curves 1, 2, 3, correspond to 10, 100 and 1000 Ma respectively. In the first hundreds of kilometers a new thermal régime is rapidly established. At greater depth the purely conductive model is somewhat artificial so that it takes a much longer time to achieve a new steady state. The change of temperature ΔT occurring in the central part of the shear zone, and the value of the shear stress τ are plotted as a function of time t in the inset of figure 2. It shows that the thermal perturbation in the asthenosphere occurs within a characteristic time of 10–30 Ma. During this time the shear stress has first risen above 500 bar and then has fallen toward the value corresponding to the steady state solution for $u_0 = 6$ cm/a, about 70 bar.

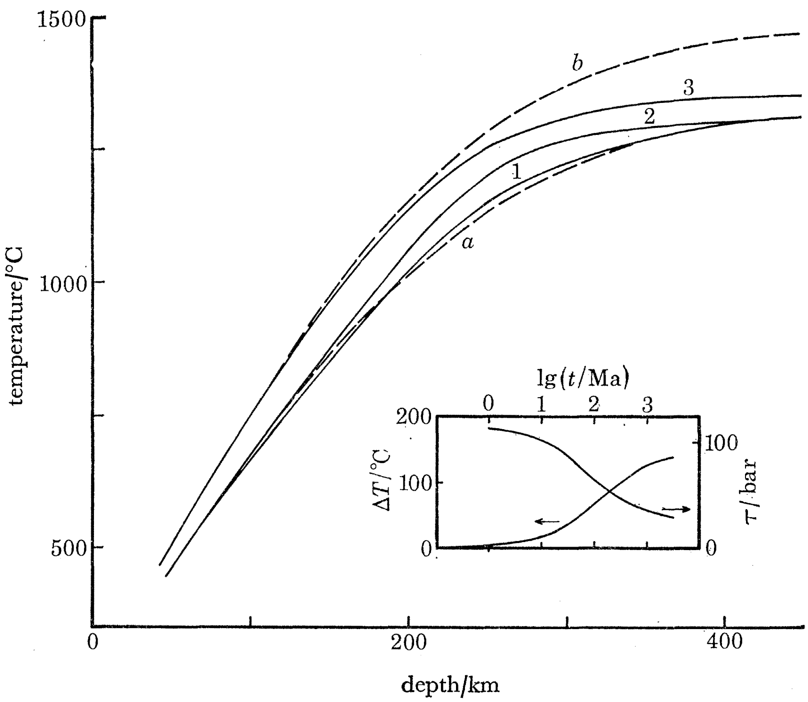

FIGURE 3. Steady state and transient geotherms for a mantle with radiogenic heat sources $Q = 0.06$ heat generation units (h.g.u.), 10^{-13} cal cm^{-3} s^{-1}. All other conditions are as in figure 2. The time constant is seen to have increased by one order of magnitude with the introduction of radiogenic heating.

A second case is illustrated in figure 3. The only difference from the previous case is that the mantle now contains radioactive heat sources. Their presence has an important consequence for steady state solutions (Froidevaux & Schubert 1975): it lowers the viscosity and hence the value of the shear stress τ. The time dependent solutions of figure 3 show that the temperature perturbation is more diffuse than in the previous case: the heat sources are not all concentrated in the shear zone. It again takes a long time to heat up the lower regions. The inset in figure 3 shows that τ rises just above 100 bar before dropping to 25 bar, the value for the final steady state. The small magnitude of τ explains why the characteristic time for the growth of the temperature perturbation, ΔT, at 250 km, is now 100–300 Ma, compared to 10 to 30 Ma for the previous case. The solutions just pictured represent two extreme cases of an upper mantle without or with a fair amount of radiogenic heating Q. The time delay to generate a thermal peak after an increase in plate velocity is quite sensitive to the values of Q and of the velocity jump Δu_0. We consider the amplitude of the peak to be large enough to yield significant geological effects. The absolute values of the temperature profiles discussed as illustration are, however, rheology dependent and could have been considerably hotter had we chosen a dry olivine rheology.

Finally, let us mention that numerical runs with a decrease of the absolute value of the plate velocity $\Delta u_0 < 0$, have produced a monotonic decrease of T for all depths, the geotherm shifting from the relatively hot solution corresponding to the high plate velocity, to the relatively colder solution corresponding to a lower plate velocity.

Discussion: thermal events in Africa

The model illustrated above predicts positive temperature changes in the asthenosphere after the plate velocity has increased by 5 cm a^{-1}. This temperature variation takes a considerable time, several tens to several hundreds of millions of years, to reach an amplitude of the order of 150 or 100 °C. The model assumes the upper mantle to be homogeneous. For the real Earth, lateral inhomogeneities may mean that such temperature variations can be more pronounced under certain regions.

Briden & Gass (1974) have shown that in the last 800 Ma the African paleomagnetic pole has gone through three periods of rapid wandering, and that each such period was followed by extensive magmatic and metamorphic activity in various parts of the continent. The periods of rapid polar wander are approximately the following: 680–600, 400–220 and 110–40 Ma B.P. These were followed by the pan-African tectono-thermal event, 650–400 Ma B.P., Mesozoic magmatism, 200–100 Ma B.P. and finally late Tertiary to Recent magmatism, 40 Ma B.P. to now.

Assuming that the periods of rapid polar wandering imply rapid plate motion, and that the intermediate periods correspond to slow plate motion, our model predicts that thermal activity under the plate was to be expected after characteristic times of the order of magnitude observed in Africa. Qualitatively, the fact that the thermal activity occurred during the periods of slow plate velocity is not puzzling: it just reflects the long response time after the onset of a high velocity episode. In the real Earth the thermal perturbation can reach the surface by more efficient mechanisms than conduction, as magmatic material migrates upward.

In conclusion we want to emphasize the fact that shear heating effects under tectonic plates are certainly very small as τ is found to be as low as a few tens of bars under continents and a few bars under oceans. However, on a geological time scale this yields enough heat to adjust the thermal régime in the asthenophere. Such adjustments, after a large increase of the plate velocity, could be at the origin of important geological events like metamorphism, magmatism and epirogeny. The effects we have studied under continental plates having changed their velocity relative to deeper parts of the mantle have turned out to occur after a long time delay, typically 100 Ma. The proposed triggering mechanism of thermal activity should be less efficient under oceanic plates as the shear heating effects are smaller there than under continents.

Finally, we call attention to the smoothness of the perturbed geotherms of figure 3, which correspond to a situation with radiogenic heat sources, as in the real Earth. This contradicts the 'kinked' pyroxene geotherm proposed by Boyd (1973) and attributed to transient effects of shear heating by this author. More recent analysis of phase equilibria in kimberlite nodules tend to question the reality of the kink in palaeogeotherms (Mercier & Carter 1975).

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Discussion

F. C. Frank, F.R.S. (*H. H. Wills Physics Laboratory, The University, Bristol*). I should like to sound a warning against placing total reliance on the magnitude of the Rayleigh number as a criterion for the occurrence of thermo-convection. One tends to give it absolute trust, perhaps just because it is due to Rayleigh, but it only applies to materials with isotropic properties, and anisotropy can make an enormous difference, as thermoconvection studies on liquid crystals (by Dubois-Violette and others) have shown. In the nematic (i.e. uniaxial liquid) substance called MBBA, for which the ratio of thermal conductivities parallel and transverse to the director (microstructural symmetry axis) is about 5:3, the critical Rayleigh number for onset thermoconvection, Ra_c, comes down from its normal value for isotropic fluids, of order 1000, to about 1, when the director is horizontal. When the director is vertical this modest anisotropy is sufficient not merely to enhance it above 1000, but to make it transfinite, i.e. negative: it takes a downward heat current to drive it into thermo-convection, though it has a normal positive coefficient of expansion. For anisotropic systems such as this, Ra_c can take any value positive or negative, only excluding the neighbourhood of zero: alternatively stated, $1/Ra_c$ may be anywhere between about 1 and about -1. I should have no confidence that in the mantle of the Earth (particularly the upper mantle), material properties are equivalent horizontally and vertically, so I do not think the critical Rayleigh number is so firm an anchor for decisions about the occurrence of gross convection as is usually assumed.

In particular, in addition to anisotropies of thermal conductivity, wherever there is incipient melting, the predominant transport of heat relative to matter will be by melt/percolation, for which predominantly vertical channels are likely to develop. This will powerfully stabilize such regions against convective overturn.

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