GENERALIZED RHEOLOGICAL MODEL OF THE LITHOSPHERE AND PLATE TECTONICS

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ABSTRACT

The parabolic forms and temperature minima of the solidus curves for silicic, intermediate, and mafic rocks may be partly shaped by increase in CO_2 content of volatiles with depth. Suprasolidus states will occur where geothermal gradients pass through the local solidus and may lead to development of low-strength or "plastic" layers within otherwise elastic, subsolidus crust. The presence of such low-strength layers may greatly affect the rheological behavior of lithosphere.

INTRODUCTION

Over the last few years emphasis has been placed on tectonic layering of the lithosphere for solving the irreconcilability of regional geologic data with the main principles of the plate tectonic concept (Peive, 1980; Lobkovsky, 1988). A number of researchers have related observed geological phenomena to the heterogeneous structure of the lithospheric plates and, in particular, to the presence of intra-lithospheric plastic layers. These layers have been called "crustal asthenolayers" (Lobkovsky, 1988), "layers of paradoxical state of the crust" (Borisov, 1990), and "active layer" (Bakulin, 1990), among other names. In the opinion of these authors, the layers are ubiquitous and characterized by high plasticity, reduced seismic wave velocities, increased electrical conductivity, and increased permeability. Most researchers explain the occurrence of such a layer by the gradation of rocks at depth into a pseudo-plastic state at certain pressures and temperatures. It is also thought that actual plasticity arises only under dislocational creep, which in turn is produced only under subsolidus conditions (Turcotte and Schubert, 1982). The problem in explaining these layers illustrates the importance of examining the possible existence of plastic layers in the lithosphere.

MANTLE VOLATILES

Laboratory investigations have shown the great effect the presence of volatiles such as H_2O or CO_2 produce on the character of melting rock. However, there are different viewpoints, both on the possible presence of volatiles in the mantle (from complete absence to an H_2O content of 13% by weight in sepentinite), and on the form of their presence (free carbon in the form of graphite or diamond, carbonate minerals, or free gas phase). Since the preponderance of evidence seems to indicate that neither of the presently defined regional geotherms cross the anhydrous solidus of peridotite, the most logical viewpoint is that the mantle contains at least a small amount of volatiles such as H_2O and CO_2. Numerous data on the composition of gas inclusions in ultramafic basalts and nodules are also evidence of the presence of volatiles (Yoder, 1976). The preferred explanation is that these gases are the product of mantle melting and there is a continuous process of deg[illegible]ing of the Earth, which is the main source for the for[illegible] of oceanic water and atmospheric gases (Vinogr[illegible] 1964; Holland, 1984). It is not by coincidence that anomalously high $^3He/^4He$ ratios are characteristic not only of volcanic gases and their inclusions in basalts, but also of marine water (Yoder, 1976).

The influence of H_2O and CO_2 on the temperature of melting rock is ambiguous. Both reduce the temperature, but the effect of H_2O vapor is much greater than that of CO_2. Therefore, the ratio of these components in volatiles and the degree of their effect on the character of solidus curves becomes very important. A distinctive feature of the solidus curves for silicate rocks, in the presence of H_2O and CO_2, is the temperature minimum (Kadik and Frenkel, 1982). The temperature minimum is due to a sharp drop in the melting point of the rock with rising pressure and the consequent rise in temperature. With an increase in volatile water content, the minimum becomes more pronounced, but the curve flattens out and shifts to higher pressure (Fig.1). In general, the slope of the solidus curve lessens after the minimum. This implies that with predominantly CO_2 in the volatiles, the solidus minimum is not as significant, and with predominantly H_2O in the volatiles, the solidus curves display a greatly increased slope after the minimum and cross the continental geotherm even at low pressure. Thus, the melting of rocks, especially of silicic composition, should be ubiquitous at depths of 30-35 km, which does not agree with geophysical data.

After the deep minimum, the solidus curve seems to have a gradient close to that prior to the minima, but in the opposite direction. According to Kadik (1975), this may occur during the melting of granites when CO_2 content of the volatiles increases with depth. Kadik (1975) proposed that a change in the H_2O/CO_2 ratio of the volatiles with depth is the result of the variable degrees of solubility of these components in the melt as the total pressure drops. An additional potentially significant factor is the increasing effect of vadose waters on the composition of the volatiles released by the mantle melt as they migrate toward the Earth's surface.

The solidus curves of other rocks besides granite can be assumed to similarly change, as shown in Fig.1. The solidus curves approximate a parabolic form with a temperature minimum for granites at a pressure of 5-10

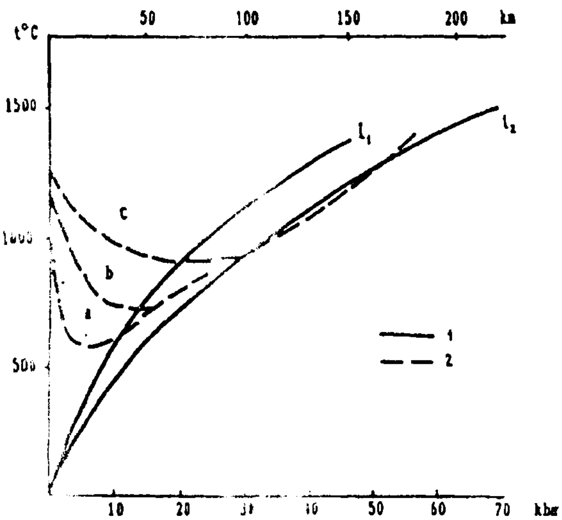

Fig.1. Conditions of rock melting in the continental crust and upper mantle. 1 - continental (I_2) and oceanic (I_1) geotherms according to Ringwood; 2 - solidus curves for (a) silicic and (b) mafic rocks and for (c) pyroxenite in the presence of water and carbon dioxide with the proportion of the latter increasing with depth. Constructed using the data of Kadik (1975) and Yoder (1976) data.

kbar, for mafic rocks at 10-15 kbar, and for ultramafic rocks at 20-25 kbar. Only the solidus curve for peridotite crosses the continental geotherm at a pressure of 30-35 kbar and a temperature of 1,000-1.250 °C. This implies that within the continents at depths of 100-150 km, there are conditions for the melting of upper mantle rocks that are responsible for the conductive layer there, generally characterized as "asthenospheric." The continental geotherm does not cross the solidus curves for silicic and intermediate rocks, though it approaches them. To initiate their melting, some additional factors are needed; in particular, a local rise in the geotherms (e.g., in the marginal parts of plates), and others.

RHEOLOGY AND STRENGTH

Established general properties should be used in rheological modeling. Resulting rheological profiles (e.g., Ranalli and Murphy, 1987; Molnar, 1988; Lobkovsky, 1988) for the generalized strength of the medium for various regions indicate the presence of one, two, or three rigid levels ("cores") within the lithosphere, but fundamentally they do not alter the idea of the presence of a thick quasi-elastic layer in its middle and lower part, e.g., the rigid (strong) base for the plates.

In calculations, such variables as effective strength, creep strength, etc., are used; and it is generally assumed that the temperature gradient does not change with depth.

Because the formulas used (Lobkovsky, 1988; Kirby, 1983) do not take into account the parabolic shape of the solidus curve, the calculated strengths of the middle and lower parts the lithosphere are artificially high. The plate tectonic interpretation of these calculations led to the conclusion that there is a lower crustal asthenosphere-like layer (Molnar, 1988) and, eventually, to the two-level model of plate tectonics (Lobkovsky, 1988).

In our opinion, it is expedient to use effective viscosity to characterize the generalized effective strength of the lithosphere, and to use the definition of Turcotte and Schubert (1982):

$$\mu = (RTh^2/24V_aD_o)\ \exp\ (aT_m/T)$$

where R is the gas constant; 8.314 J. mole^{-1} K^{-1}; T is temperature; h is depth; D_o is frequency factor; T_m is temperature of melting; V_a is volume of activation.

In this formula, the temperature dependence of the variables before the exponential is insignificant relative to the exponential function. Thus, this formula can be rewritten as:

$$\mu = C_1\ \exp\ aT_m/T$$

where C_1 is a constant.

Since

$$\exp\ (aT_m/T) = \exp\ (E_a+\rho V_a/RT)$$

where E_a is molar energy of activation, which according to laboratory investigations by Kirby (1983) for the "granitic" and "basaltic" layers and the mantle are 150, 250, and 540 kJ mole^{-1}, respectively, and T_{mo} is melting temperature at the Earth's surface,

$$a = (E_a + \rho V_a/RT_{mo}).$$

This can be simplified to

$$a = (E_a/RT_{mo}).$$

Conventionally, viscosity in a solidus state is identical for all rocks and is equal to 10^{19} Pa-s. Hence, the constant factor C_1 can be calculated. On the basis of data obtained, curves for the viscosity variations with depth have been constructed for the "granitic" and "basaltic" layers of the Earth's crust and upper mantle (Fig.2).

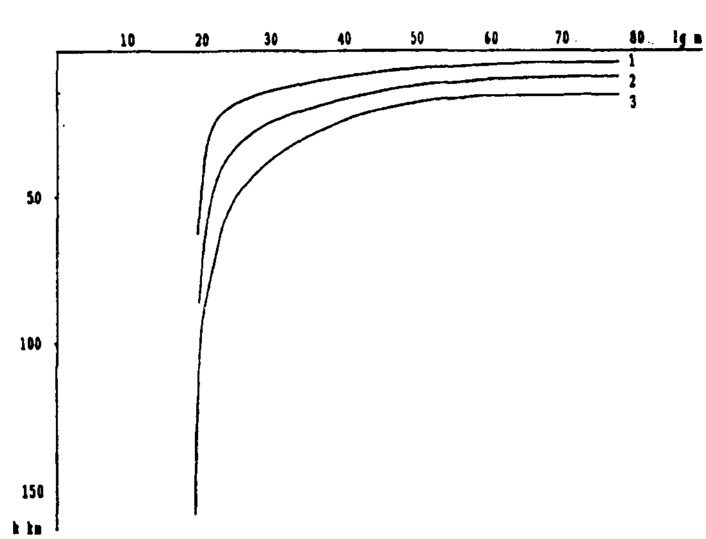

Fig.2. Viscosity dependence of the (1) "granitic" and (2) "basaltic" layers of the Earth's crust and (3) the upper mantle with depth.

The character of the curves will not change if we use creep stresses, for fixed deformation rates, as an index of the lithosphere's strength, e.g., the above Kirby (1983) formula written as

$$d\dot{e}/dt = A\sigma^n \exp(-Q/RT).$$

In this case, we should replace the exponential argument Q/RT by aT_m/T. Therefore,

$$d\dot{e}/dt = A\sigma^n \exp(-aT_m/T).$$

Thus,

$$\sigma^n = (d\dot{e}/dt)/A \exp(aT_m/T),$$

where $d\dot{e}/dt$ and A are set to constant values.

So, when $C_1 = (d\dot{e}/dt)/A$,

$$\sigma^n = C_1 \exp(aT_m/T).$$

Therefore, this formula is almost identical to the formula for determining viscosity.

Calculations on the strength variations for the "granitic" and "basaltic" layers of the Earth's crust and the upper mantle make it possible to construct generalized curves for the strengths of different types of lithospheric plates. Fig.3 shows the curves for a continental plate where the "granitic" and "basaltic" layers are each 25 km thick, and for an oceanic plate with a "basaltic" layer 8-12 km thick.

Examination of the curves in Fig.3 yields the following conclusions:

(1) Oceanic and continental plates are composed of upper, elastic, and lower, plastic (viscous), parts.

(2) The thickness of the elastic part of continental plates does not exceed 15 km. This conclusion is in agreement with seismic data that indicate that most continental earthquakes have focal depths of 5-15 km. The elastic part of the oceanic plates is not greater than 25 km in thickness. Therefore, the strength of the oceanic plate is higher than that of the continental due to the increased thickness of its elastic part.

(3) Within continental lithospheric plates, and especially at their margins, there may occur layers of rocks in a subsolidus state. These layers lie in the lower, plastic part of the lithosphere and differ slightly from the substratum enclosing them in their physical properties (elasticity, plasticity, etc.), i.e., they do not generally disturb the two-level rheological structure of the lithosphere (Fig.4).

In the oceanic plate, the highly plastic layer occurs at the boundary of the "basaltic" and "peridotite" layers within its rigid elastic part; i.e., it differs substantially from the enclosing medium in its physical properties and thus can be regarded as asthenosphere-like.

Under certain conditions, such as changes in the thickness of the individual lithospheric layers, local variations of the geothermal field, etc., the rheologic structure may change accordingly.

LITHOSPHERIC RHEOLOGY AND PLATE TECTONICS

The features of the rheological state of the lithosphere and differences in rheology between oceanic and continental plates determined above allow us to examine the plate tectonic mechanism as a whole, the character of the plate interactions, and of intraplate tectonics, from a new angle. We thus examine the model of convergent-divergent interactions of the Eurasian and Pacific plates using the above viewpoint (Fig.5).

In the convergent stage (Fig.5b), the thin, upper, rigid-elastic part of the Eurasian margin undergoes undulatory bending as a result of tangential stresses occurring in the boundary zone between the plates. In the initial stress stage, the lower, viscous-elastic part of the continental lithosphere is deformed with about the same amplitude.

As stress increases (Fig.5c), the upper, brittle part of the crust fractures. In the lower (viscous) part, the initiation of subduction of the oceanic plate beneath the continental plate results in continued folding. Subsequently, during the relaxation stage (Fig.5d), the decreased stresses result in a tendency for the system to occupy its original position; flattening of the plates and the "sheets" comprising them takes place. Eventually, the straightening out of the lower viscous-elastic part of the lithosphere results in the breaking of the upper brittle crust, fractured in the compression stage, into blocks and

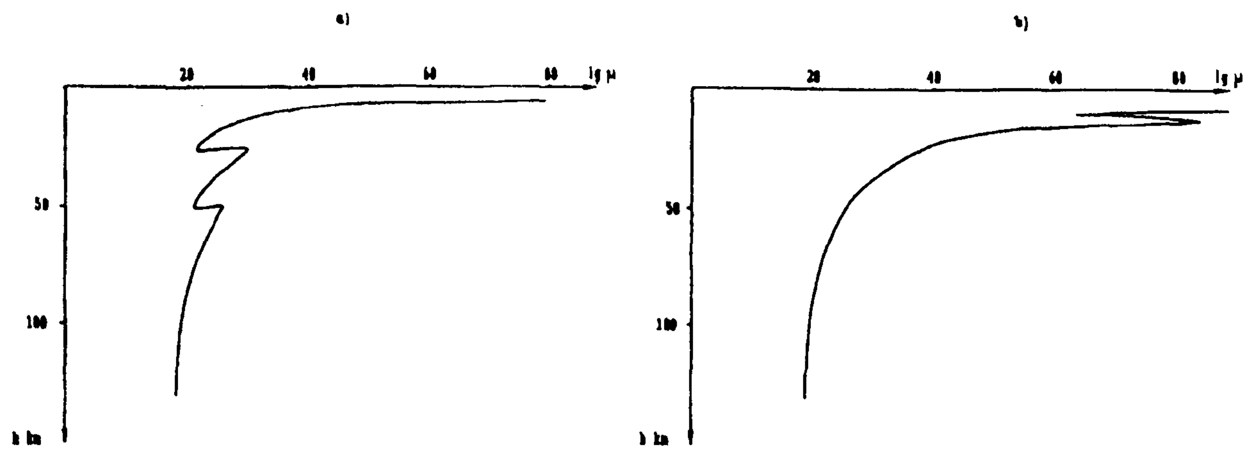

Fig.3. Curves of viscosity variations of the (a) continental and (b) oceanic lithosphere, assuming that continental crust consists of "granitic" and "basaltic" layers, each 25 km thick, and that the oceanic plate is composed of basalts 8-12 km thick.

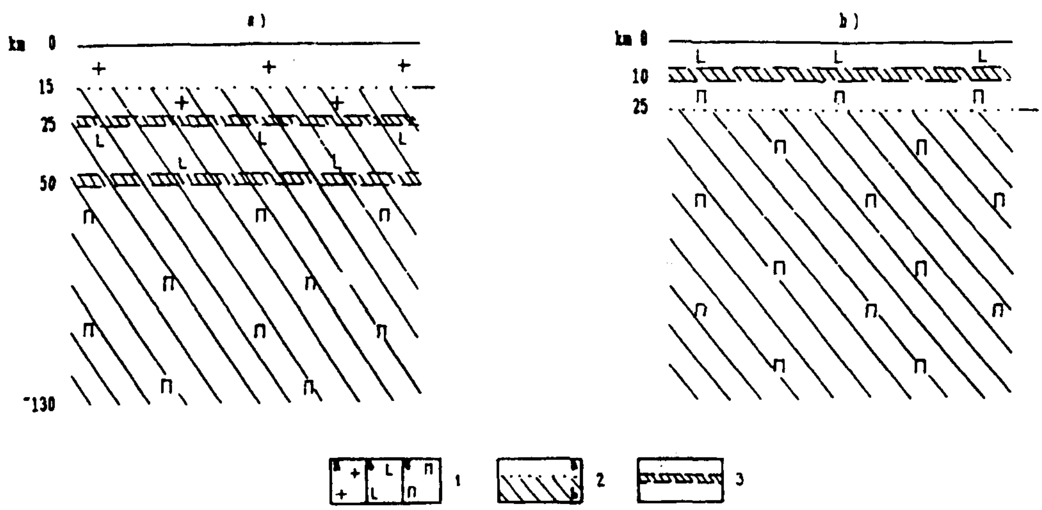

Fig.4. Models of rheological structure of the (a) continental and (b) oceanic lithosphere. 1 - (a) "granitic" and (b) "basaltic," (b) layers of the Earth's crust, and (c) mantle rocks; 2 - (a) rigid-elastic and (b) plastic-viscous parts of lithosphere; 3 - subsolidus layers of the crust.

pieces and the formation of extensional structures. These extensional structures are of rift origin and do not have a crystalline basement. They have a thinned basement, differential subsidence, and a variety of normal and strike-slip faults. The structures experience sedimentation and corresponding magmatism. In some of the open fractures, rising magma transports ore material to the upper horizons.

During the subsequent compressional stage (Fig. 5e), the detached blocks are pushed closer to one another, resulting in folding of the sedimentary cover and the formation of a fold system, similar to accretionary complexes.

The processes of folding, warping, and fracturing are poorly developed in the oceanic plate because it is stronger (thicker and more rigid). At some distance from the boundary between the plates, a small-amplitude arch ("outer rise") develops, and fractures formed above it are later filled with basalts and ophiolites from the asthenosphere-like intralithospheric layer. This active layer seems to be the source of most of the basalts of the Pacific floor, and feeds numerous active volcanoes.

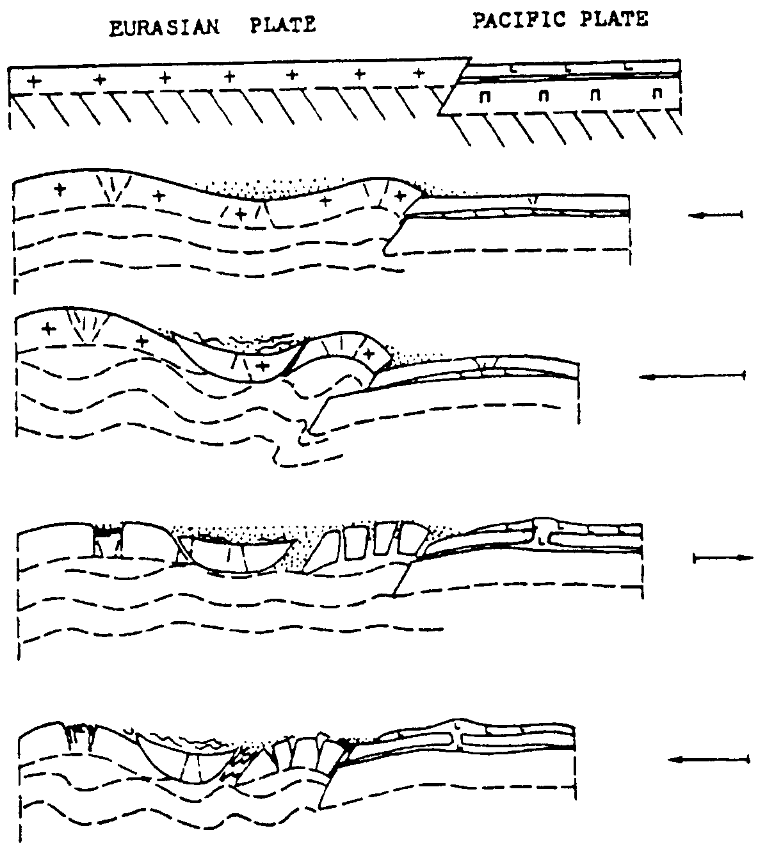

Fig.5. Model of the convergent-divergent interactions between the Eurasian and Pacific plates (same symbols as Fig. 4).

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